Geology of the West-Central Part of the Southern Anthracite Field and Adjoining Areas, Pennsylvania

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A study of part of one of the classic areas of geology and of the surrounding region

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By GORDON H. WOOD, JR., J. PETER TREXLER, and THOMAS M. KEHN

ABSTRACT

This report describes an area of about 450 square miles in Berks, Dauphin, Lebanon, Northumberland, and Schuylkill Counties of the Anthracite region, east-central Pennsylvania. The stratigraphy and structure of about 100 square miles of the west-central part of the Southern Anthracite field and adjacent areas of the Klingerstown, Lykens, Minersville, Pine Grove, Swatara Hill, Tower City, Tremont, and Valley View 7½-minute quadrangles are described in detail.

The area, which is a part of the Appalachian Mountains, lies in the Delaware-Susquehanna Rivers segment of the middle section of the Valley and Ridge or folded Appalachian province. Tributaries of the Susquehanna River drain the central and western parts of the area; the eastern part is drained by the Schuylkill River. Resistant strata support a series of sinuous linear-strike mountain ridges separated by strike valleys which have been eroded in less resistant strata.

Middle Ordovician to Upper Pennsylvanian rocks that average about 24,500 feet thick crop out in the area. At many places, thin deposits of alluvium, soil mantle, and talus of Quaternary age overlie an erosion surface that has been cut into these rocks. Except for 700 to 1,800 feet of deltaic red beds of early Late Silurian age, the lower 8,000 to 9,600 feet of the Paleozoic sequence accumulated in the sea. Marine waters regressed about the middle of Late Devonian time, and the remainder of the Paleozoic sequence was deposited under continental conditions.

The upper 1,000± feet of the Martinsburg Shale of Middle and Late Ordovician age underlies the southeastern part of the area, but is everywhere covered by talus. Where the Martinsburg is exposed south and east of the area, rocks in the upper part of the formation consist of interbedded marine gray shale, siltstone, sandstone, and limestone. After deposition of the Martinsburg, the sea regressed westward, and the uplifting, folding, faulting, and erosion of the Taconic orogeny took place in Late Ordovician time.

At the end of the Taconic orogeny in earliest Silurian time the sea transgressed southeastward and deposited about 400 feet of the grayish detritus of the Tuscarora Sandstone. This detritus was eroded from a highland that lay to the southeast. The overlying Clinton Formation of Middle Silurian age grades downward into the Tuscarora and consists of 2,300± feet of red and gray detritus that was also derived from the highland. During early Late Silurian time the sea regressed northwestward, and 700 to 1,800 feet of the Bloomsburg Red Beds was laid down upon the Clinton in a deltaic environment. After deposition of the red beds the sea again transgressed towards the southeast in middle Late Silurian time, and the varicolored Wills Creek Shale, 0 to 400 feet thick, and the gray Tonoloway Limestone, 0 to 400 feet thick, accumulated successively. In latest Silurian and earliest Devonian time, the gray Keyser Limestone, 0 to 200 feet thick, was deposited with apparent conformity on the Tonoloway.

Definite Early Devonian sedimentation commenced with the Helderberg Formation, 0 to 100 feet thick, which consists of the gray Coeymans and New Scotland Limestone Members and the Mandata Member (shale). The Helderberg is overlain successively by upper Lower Devonian rocks of the dark-gray Shriver Chert, 0 to 80 feet thick, and the gray Ridgeley Sandstone, 0 to 60 feet thick. Stratigraphic relations between the Shriver and Ridgeley are conformable in the northwestern part of the area, but the Shriver and all subjacent Lower Devonian and Upper Silurian rocks down to and including the upper part of the Bloomsburg Red Beds are truncated in the central and southern parts by a pre-Ridgeley erosion surface. This surface developed because the Auburn promontory was uplifted south of the report area, following Shriver deposition. After erosion, the region subsided slightly, the sea transgressed southeastward toward the promontory, and sediments of the Ridgeley accumulated.

A minor rejuvenation of the Auburn promontory in the southern part of the area forced the sea to retreat northwestward in late Early or early Middle Devonian time. After the sea regressed, the upper beds of the Ridgeley were slightly eroded at most places, and locally the formation was completely removed. The sea then transgressed from the northwest upon this erosion surface, and the dark-gray Needmore Shale, 0 to 100 feet thick, was deposited. At the end of Needmore deposition the region adjacent to the promontory was again slightly elevated, and the Needmore was removed from the southern part of the area. The sea again transgressed southeastward, and the gray Selinsgrove Limestone, 20 to 100 feet thick, accumulated. Successively, the dark-gray to black Marcellus Shale, 60 to 700 feet thick, and the Mahantango Formation, 1,200 to 2,000 feet thick, of late Middle Devonian age were laid down on the Selinsgrove. The gray Montebello Sandstone Member, medial unit of the Mahantango, records a major rejuvenation of the Auburn promontory.

About the beginning of the Late Devonian or late in the Middle Devonian, mountain making that heralded the Acadian orogeny commenced southeast of the area and gradually uplifted a landmass of considerable size. Detritus derived from the landmass was transported northwestward to the sea and distributed by currents over a region much larger than the area of this report. It was consolidated into the Trimmers Rock Sandstone, 1,300 to 2,400 feet thick. As this influx of sediment gradually exceeded the capacity of the subsiding basin, the sea slowly regressed northwestward with many reversals. During regression, a sequence of intertonguing gray marine and red continental clastic rocks accumulated that ranges in thickness from about 200 to 2,250 feet. These marine and continental strata are the Irish Valley Member of the Catskill Formation. Near the end

of Irish Valley sedimentation, or early in the deposition of overlying sediments, the sea regressed, and almost all succeeding Devonian, Mississippian, and Pennsylvanian sediments are continental deposits.

From middle Late Devonian to Early Mississippian time the red and gray continental beds of the middle and upper parts of the Catskill Formation were laid down. This sequence in the northern part of the area is divided into the predominantly red Buddys Run Member, 4,200 to 4,800 feet thick, and the overlying gray and red Spechty Kopf Member, 0 to 1,000 feet thick. In the southern part it is divided, in ascending order, into the predominantly red Damascus Member, 2,600 to 4,650 feet thick; the gray and red Honesdale Sandstone Member, 0 to 700 feet thick; the predominantly red Cherry Ridge Member, 700 to 1,800 feet thick; and the gray and red Spechty Kopf Member, 0 to 2,400 feet thick.

At least three phases of the Acadian orogeny of the Late Devonian and Early Mississippian age are recorded by rocks of the Catskill Formation. The climaxes of these phases are marked by conglomerates in the Honesdale and the lower part of the Spechty Kopf and by an angular unconformity that bevels the Spechty Kopf. This unconformity shows that in Early Mississippian time, deformation associated with the orogeny advanced into the area from the southeast.

The Spechty Kopf is overlain unconformably by the Beckville Member of the Pocono Formation of Early Mississippian age which ranges in thickness from about 300 to 800 feet and consists of a gray basal conglomerate, gray finer clastic rocks, and a few thin beds of coal. The Mount Carbon Member of the Pocono Formation conformably overlies the Beckville and has a similar lithology. It ranges in thickness from about 400 to 1,000 feet. The basal conglomerates of these members seem to represent clastic detritus that was eroded from periodically rejuvenated parts of the Acadian Mountains.

Red and gray beds of the Upper Mississippian and Lower Pennsylvanian Mauch Chunk Formation, 3,000 to 7,500(?) feet thick, succeed the Pocono conformably. Three informal members are recognized in the Mauch Chunk. The upward decrease in grain size of the interbedded red and gray detritus of the lower member, 400 to 800 feet thick, records the gradual reduction of the Acadian Mountains. Most of the red beds of the Mauch Chunk are included in the middle member, which is 2,000 to 6,000 feet thick. These beds are composed largely of clay, silt, and fine sand. The fine grain size, plant fossils, and associated sedimentary features in these beds suggest that the middle member accumulated on flood plains or on a coastal plain at a considerable distance from a reduced source area. Clay- to cobble-sized detritus in interbedded red and gray beds of the upper member, 400 to 900 feet thick, indicate that the source area to the southeast was uplifted during latest Mississippian time. This uplifting marked the beginning of the Appalachian orogeny.

The deposition of red beds ceased at the beginning of the Pennsylvanian in the southern part of the area but continued for some time thereafter to the north. Red sediments of the Mauch Chunk were succeeded by intertonguing relations of 800 to 1,500 feet of gray fine- to coarse-grained sediments of the Pottsville Formation, which is divided, in ascending order, into the Tumbling Run, Schuylkill, and Sharp Mountain Members. The Tumbling Run consists of 275 to 600 feet of grayish pebble and cobble conglomerate, sandstone, and siltstone beds and several beds of anthracite. Although cobble conglomerates are virtually absent, the conformably overlying 300- to 700-foot thick Schuylkill Member of late Early Pennsylvania age is lithologically similar. The Sharp Mountain Member of Middle Pennsylvanian age rests conformably on the Schuylkill Member and consists of 100 to 315 feet of beds similar to those of the Tumbling Run. Pebbles and cobbles in the upper member of the Mauch Chunk Formation and in the Tumbling Run and Sharp Mountain Members of the Pottsville Formation contain the sedimentary record of the two earliest pulsations of the Appalachian orogeny in Pennsylvania.

The Llewellyn Formation of late Middle and Late Pennsylvanian age reaches its maximum known thickness of about 3,500 feet in the area. It is composed of grayish sandstone, siltstone, shale, and conglomerate and numerous thin to thick beds of anthracite. Sediments of the formation seem to have been laid down on a large poorly drained flood plain or coastal plain that lay northwest of a generally low but in places rejuvenated mountain belt. The plain was dotted by many swamps during deposition; some of these covered hundreds and perhaps thousands of square miles. Deposits of vegetation that accumulated in these swamps were partly decomposed to peat and eventually buried by succeeding layers of sediment. The peat gradually altered to lignite and then was metamorphosed through the bituminous-coal series to anthracite. Early metamorphic changes seem to have been caused by the pressure of increasing overburden, whereas later changes were primarily due to horizontal pressures related to folding and faulting, and secondarily, to pressure from increasing overburden.

The geologic structure is complex. Folding and faulting related to three periods of mountain building affected the area. During Late Ordovician time the southeastern part of the area was folded by forces of the Taconic orogeny. In Early Mississippian time, forces of the Acadian orogeny compressed the rocks of the area into a series of large-amplitude folds which trend northeast, and, in the latter part of the Paleozoic Era during the Appalachian orogeny, the area was intensively deformed. Relations of folds and faults formed during the orogeny indicate that mountain building commenced southeast of the area and that it gradually advanced towards the northwest with increasing vigor. By the end of the orogeny a multitude of complexly interrelated structural features had formed. Chief among these were a series of northeast-trending anticlinoria and synclinoria, numerous subsidiary doubly plunging en echelon folds, a multitude of low-angle folded and nonfolded thrust faults, numerous high-angle reverse faults, and a few large tear faults.

Most structural features in the report area formed during the Appalachian orogeny. The largest of these are the Minersville synclinorium, which splits westward into two subsynclinoria, and the Broad Mountain and New Bloomfield anticlinoria. The subsidiary anticlines and synclines of these composite fold systems are mainly on the limbs and in the troughs. They range in length from a few inches to many miles and in amplitude from several inches to thousands of feet. The principal subsidiary folds are the Branchdale, Eisenhuth Run, Frackville, Hooflander Mountain, Joliet, Mine Hill, Peaked Mountain, Powder Hill, Roedersville, and West West Falls anticlines and the Beury, Dauphin, Deep Creek, Donaldson, Heckscherville, New Boston, and Pine Grove synclines. Most long synclines are closed, asymmetric, and V-shaped in their central parts; in contrast, large anticlines are broad, open, and more nearly symmetrical. Near their ends the cross sections of many synclines are U-shaped, whereas those of most anticlines are inverted V-shaped.

The Minersville synclinorium and the Broad Mountain anticlinorium, which have amplitudes of 10,000 to 15,000 feet, are the largest structural features. Most subsidiary folds and faults are on the limbs of these large composite fold systems. The main trough and the south trough of the synclinorium separate generally upright strata to the north from generally overturned strata to the south. Rock sequences in the main trough and in the north and south troughs of the synclinorium were greatly thickened during folding. The development of the Mauchono and Pottchunk faults seems to be related to this thickening. Low-angle thrust faults formed on the north limb of the synclinorium and across the crest of the anticlinorium before extensive folding took place as the result of differential transmission of force through strata of differing compentencies. The thick, rather competent strata of the Pocono and Pottsville Formations were more capable of transmitting force than were the intervening thick incompetent strata of the Mauch Chunk Formation. During the early stages of folding, these competent strata tore loose from the less competent strata and moved differentially on the Pottchunk fault above and the Mauchono fault below. As faulting progressed, the upper plate of the Pottchunk fault was broken into imbricated blocks by the successive development of the Hans Yost, Dyer Run, Jugular, and Mine Hill low-angle trust faults. The area was then more tightly compressed; many subsidiary folds formed on the limbs of the anticlinoria and synclinorium; the planes of the Mauchono and Pottchunk faults were folded; and numerous high-angle reverse faults, such as the Newtown, South Newtown, Tremont, and Red Mountain, formed. As compression increased, the south limb of the Minersville synclinorium was overturned, perhaps into a nappe, and the low-angle Sweet Arrow and Blackwood faults formed.

The west-central part of the Southern Anthracite field has yielded about 318 million tons of anthracite valued at about $1.1 billion. Approximately 0.6 percent of the energy consumed in the United States before 1960 was from coal produced in the area.

Nearly 7.6 billion tons of anthracite is estimated to have underlain the area before mining began. Of this original content of anthracite, about 6.3 billion tons is estimated to have been in beds more than 42 inches thick, and about 5.7 billion tons of the original content lay within 3,000 feet of the surface. About 2.6 billion tons of anthracite was believed (1961) to be recoverable in the future within 3,000 feet of the surface; 1.9 billion tons of this recoverable anthracite is estimated to be in beds more than 42 inches thick.

INTRODUCTION

This report describes the geology of the Klingerstown, Lykens, Minersville, Pine Grove, Swatara Hill, Tower City, Tremont, and Valley View 7½-minute topographic quadrangles in Berks, Dauphin, Lebanon, Northumberland, and Schuylkill Counties (fig. 1). The area of these quadrangles, about 450 square miles, includes the west-central part of the Southern Anthracite field of the Anthracite region of eastern Pennsylvania. The report is a result of investigations by geologists of the U.S. Geological Survey from 1953 to 1963.

The structural geology of the report area is complex and is characterized by numerous folds whose limbs are broken by faults. Natural outcrops are rare, commonly small, and probably amount to less than 1 percent of the land surface; within the Southern Anthracite field, however, rocks are exposed at many places in strip pits and roadcuts. Because of these exposures, both natural and manmade, and the large amount of data provided by mine maps and cross sections, the report area is one of the better places in the central part of the Appalachian Mountains to study the geometry and relations of folds and faults and to observe and interpret the vagaries of "coal-measure" deposition.

Although anthracite had been mined in the area for 150 years, the area probably contains as much as 50 percent of the remaining reserves of coal in the Southern Anthracite field, and it probably contains greater remaining reserves than any other comparable area in the Anthracite region. The actual tonnage of coal produced is unknown because records that were kept during the early period of mining were chiefly for commercial district or county rather than for mines, because a considerable unreported tonnage was produced during the 1930's by independent or "bootleg" miners, and because official tonnage records are maintained by the amount of coal cleaned at preparation or breaker plants, by county, and by coal field, and only in some instances by mine. It is estimated that about 318 million tons of anthracite valued at about 1.1 billion dollars was produced in the area from 1790 to 1960.

Maps and structure sections of the eight described quadrangles do not appear in this report. They have been published by the U.S. Geological Survey as eight 1:24,000-scale Geologic Quadrangle maps with 16 structure sections (Trexler and Wood, 1968a–c; Wood, 1968; Wood and Kehn, 1968a, b; Wood and Trexler, 1968a; and Wood and others, 1968) and two Miscellaneous Geologic Investigations maps containing seven 1:12,000-scale coal maps and 44 structure sections (Wood and Trexler, 1968b, c). In this paper the geologic maps are hereafter referred to by their "GQ" number. The reader should consult these 15 geologic and coal maps and 60 structure sections in conjunction with this paper to derive maximum benefit from all.

The geology of two small areas in the Southern Anthracite field has been treated in preliminary reports published by the U.S. Geological Survey (Wood and others, 1958; Wood and others, 1962). Revised maps and cross sections of these two areas are included in the Miscellaneous Geologic Investigations maps mentioned above (Wood and Trexler, 1968b, c). The stratigraphy and nomenclature of some rock units and the structural development of the Anthracite region have been described in several short papers (Wood and others, 1956; Arndt and others, 1959; Arndt and Wood, 1960; Wood and Kehn, 1961; Trexler and others, 1962; Wood and others, 1962; Arndt and others, 1962). The conclusions reached in these short papers are incorporated in this report.

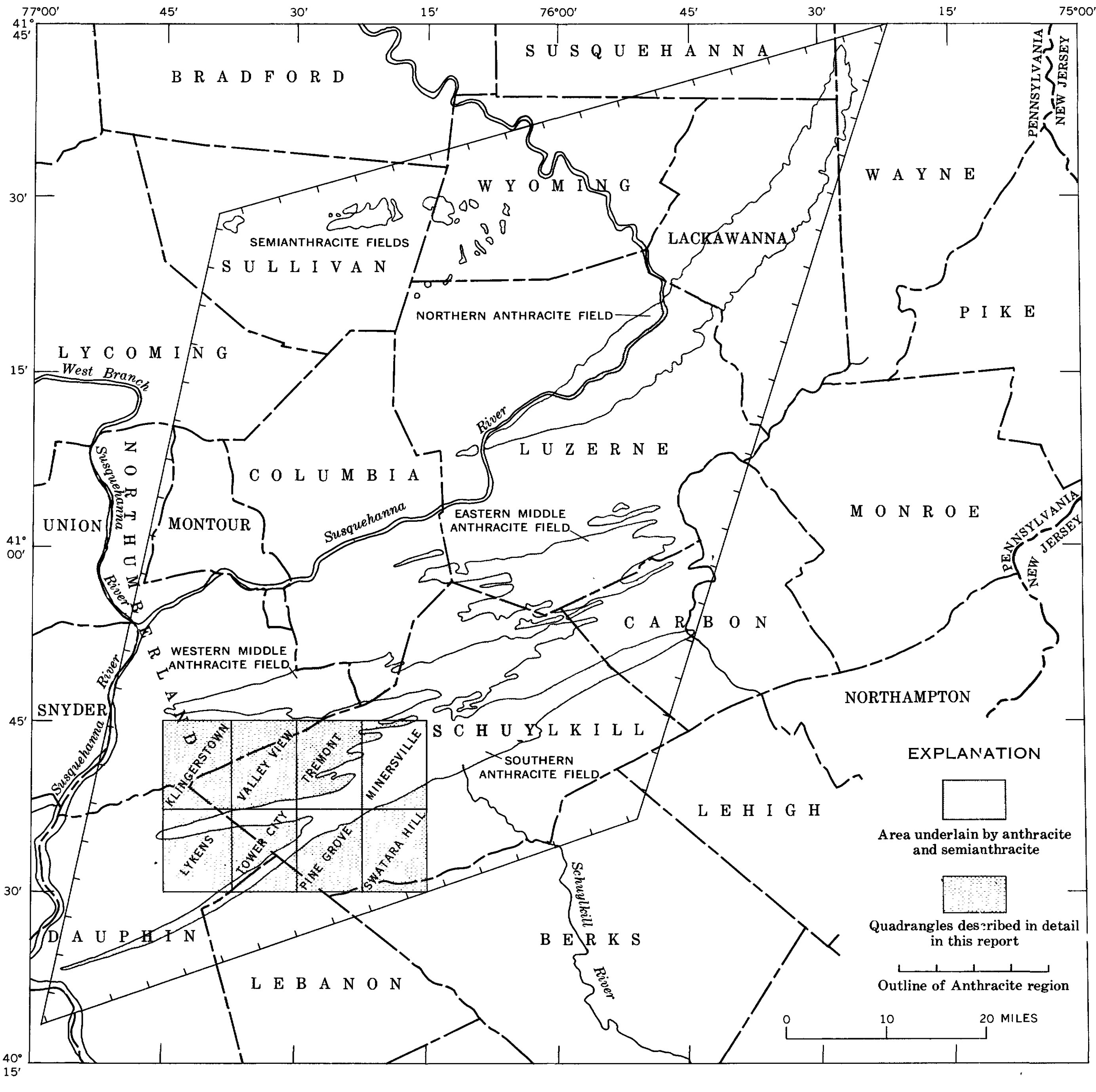

FIGURE 1.—Index map of the Anthracite region, east-central Pennsylvania, and location of quadrangles.

PHYSICAL FEATURES

The area is in the Delaware-Susquehanna Rivers segment of the middle section of the Valley and Ridge or folded Appalachian province of Fenneman (1938). Many northeastward-trending mountain ridges and valleys cross or extend into the area from the surrounding region. The Schuylkill River drains much of the Minersville quadrangle and a small part of the Swatara Hill quadrangle. Tributaries of the Susquehanna River drain the remainder of the area.

The mountain ridges are held up by gently to steeply dipping sandstone and conglomerate units, whereas the valleys are chiefly underlain by shale and siltstone units. Altitudes in the valley bottoms generally range from 500 to 900 feet above sea level and on the mountain tops from 1,200 to 1,850 feet above sea level.

The rocks are complexly folded and faulted. Regionally, the fold pattern plunges to the northeast; many of the smaller folds and some of the larger die out within the area. Many ridges and valleys double back and forth in a series of pointed zigzags, thereby outlining the structural configuration of the plunging fold pattern. The names of the ridges change at each sharp bend and in places at water gaps. The most persistent of these ridges is underlain by the Pocono Formation; others are underlain by the Tuscarora Sandstone, the Clinton Formation, the Montebello Sandstone Member of the Mahantango Formation, and the Pottsville Formation.

The only deep wind gap is in Fisher Ridge northwest of Klingerstown. However, shallow and incipient wind gaps are scattered throughout the area. Swatara Creek has cut a large water gap in Second Mountain north of Pine Grove and another in Sharp Mountain south of Tremont. The gap of Lorberry Creek in Sharp Mountain a few miles west of Tremont is similar in size and depth. Mill Creek has cut an impressive gap through Second Mountain north of Suedberg; by contrast, the gaps eroded in Sharp Mountain by its tributaries, Fishing, Black Spring, and Gold Mine Creeks, are considerably smaller. Rattling Creek flows through a small gap in Berry Mountain south of Lykens, and Bear Creek has eroded a similar gap through Big Lick Mountain north of the same town. From a distance, the most impressive gap is that cut by Pine Creek through Mahantango Mountain south of Klingerstown.

PRESENT INVESTIGATION

The study of the geology of the Southern Anthracite field began in the summer of 1953 and continued without interruption to 1963. The investigation has been under the direction of Gordon H. Wood, Jr., and has included at various times, J. Peter Trexler, Thomas M. Kehn, Julian Soren, and Andy Yelenosky. Areas of mapping responsibility are shown in figure 2.

Aerial photographs, at a scale of 1:20,000, were used for geologic mapping because they show more of the features related to mining than do topographic maps and because lithologic units and structural features too small to be identified on topographic maps can be traced through areas of heavy vegetation, mining activity, and farming. Within the anthracite field, geologic data were delineated on the aerial photographs and then transferred by vertical projector or multiplex to 1:10,000-scale topographic bases used for compilation of the coal maps. Mine data obtained from companies, engineers, and individuals were reduced photographically from scales of 1 inch = 100 feet, 1 inch = 300 feet and 1 inch = 400 feet to the 1:10,000 scale and then combined with the field data on the coal maps. Before the completion

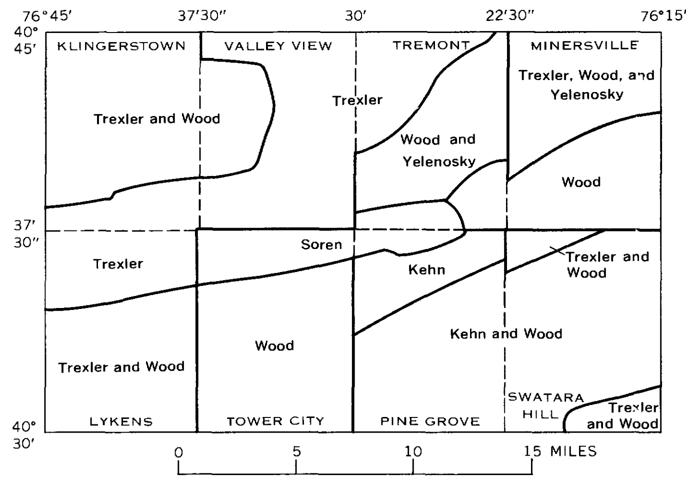

FIGURE 2.—Index showing quadrangle names and areas of responsibility. Geologic mapping by U.S. Geological Survey: Gordon H. Wood, Jr., 1953–60; J. Peter Trexler, 1953–58, 1960; Thomas M. Kehn, 1954–56, 1958–59; Andy Yelenosky, 1953–56; Julian Soren, 1954–57.

of each coal map (Wood and Trexler, 1968 b, c), many structure sections were plotted utilizing all available subsurface information. Some of these structure sections are included in GQ–689, 690, 691, 692, 698, 699, 700, and 701 and Wood and Trexler (1968b, c). Geologic data outside the coal field were delineated on the aerial photographs and subsequently transferred by similar methods to the 1:20,000-scale topographic bases used for compilation of the quadrangle maps. These latter maps were completed by transferring selected data from the coal maps with a vertical projector.

This report was prepared by Wood from 1961 to 1964. He compiled all maps and illustrations except the Klingerstown, Lykens, and Valley View 7½-minute quadrangles and the Lykens quadrangle coal map. Trexler hastened report completion by preparing the above-listed map exceptions and by writing a report (1964) on the western part of the area. Kehn was responsible for much of the fieldwork and preliminary office compilation of data and contributed materially to interpretation of stratigraphic and structural data.

Charles B. Read examined numerous plant collections and determined that floral sequences were missing locally. This knowledge aided in identification and interpretation of folded thrust faults at several places. Wallace de Witt, Jr., and George W. Colton studied the Devonian stratigraphy and thereby increased the certainty of regional correlations of rocks of this age.

Harold H. Arndt, geologist in charge of U.S. Geological Survey work in the Western Middle Anthracite field, cooperated fully with the authors by contributing pertinent data, ideas, criticisms, and conclusions. During many phases of the investigation he worked with

Southern Anthracite field personnel on problems of mutual interest.

PREVIOUS WORK

The Anthracite region has been considered by many geologists to be an area where the geology is known. This belief was based on the assumption that the geology was adequately explained in the reports of the First (1836–58) and Second (1875–95) Pennsylvania Geological Surveys. Since the completion of these surveys in 1895, much additional mining and geologic mapping has made it apparent that the reports of these surveys do not describe the presently known structural and stratigraphic complexities. In addition to the reports of these surveys, many articles have been published concerning parts of the region or limited phases of the geology, mining, and physiography; numerous others discuss the classification, origin, petrography, transportation, and utilization of anthracite, and hundreds contain references to the geology and mining industry.

The following paragraphs list in chronologic order some of the more important articles contributing to a geologic understanding of the area and the surrounding region. Many other reports are cited in the text and are listed in the "References."

The earliest known geologic reference to the area is a report on the Swatara mining district (Strong and others, 1839). Shortly thereafter, Taylor (1840, 1841) described the Dauphin syncline and published the earliest known cross section of folded rocks in the United States. One of these sections illustrated Taylor's concept of deformation in the western part of the report area.

Rogers in 1843 postulated that the coal of the eastern United States formed in a great, gradually sinking swamp adjacent to the sea. He explained the easterly decrease in volatile matter as the result of steam and gas being emitted through cracks as the Appalachian Mountains were folded. The report area was cited as the locality where the deformation of rocks and devolatilization of coal were greatest.

Bowen (1848) published the first map showing the extent of the coal fields of the Anthracite region. Within a decade, Rogers (1858a, b), released a geologic map of the region and published a comprehensive report on Pennsylvania which named many stratigraphic units, described most of the principal structural features, and described some relations between stratigraphic units and structural features in the area.

Nearly 20 years later, Lesley (1876) and Franklin Platt, and W. G. Platt (1877) renamed some of the stratigraphic units that were originally recognized by Rogers. Principal among these were the Pocono, Mauch Chunk, and Pottsville Formations of modern nomenclature.

From 1880 to 1895, numerous reports were published by the Second Pennsylvania Geological Survey on the geology of the report area and the surrounding region. Principal among these were: I. C. White's (1882, 1883) reports on the stratigraphy of Devonian and Mississippian rocks in eastern Pennsylvania; Ashburner's (1883-1885) reports on the Southern Anthracite field; the geologic maps of Dauphin (Sanders, 1881), Berks (Prime and others, 1884), and Schuylkill (Smith, 1891) Counties; and the summary descriptions of Silurian, Devonian, Mississippian, and Pennsylvanian rocks (Lesley and others, 1892–95, p. 721–1915; Smith, 1895, p. 2072–2146).

In 1893, Stevenson concluded that Pennsylvania anthracite differed from bituminous coal because of differences in original composition rather than because of metamorphism. Willis (1893) in the same year published an analysis of deformation in the Appalachians and concluded that the localization of folds in the Anthracite region was due to initial differences in thickness of sedimentary units.

A few years later C. D. White (1900, p. 749–930) described the stratigraphy, correlated the coal beds, and determined the floral succession of the Pottsville Formation in the report area and elsewhere in the Southern Anthracite field. In 1904, he correlated this formation from the type locality southwestward to Alabama and stated that the oldest rocks occurred in the area covered by this report and in the southern Appalachians.

Stevenson early in 1904 also correlated the Pottsville Formation from the area southwestward into Alabama, and in 1906 (p. 65–228), he correlated the overlying "coal-measures" (Llewellyn Formation) of the area with the "coal-measures" of western Pennsylvania and Ohio.

Shortly thereafter, Barrell, in reports published in 1907 and 1913–1914, concluded that the Catskill Formation in the vicinity of the report area was deposited on a great delta whose source lay to the east. He also postulated that the Mauch Chunk Formation accumulated on a subaerial plain and delta that was dry much of the year.

The carbon-ratio theory was advanced in 1915 by C. D. White. According to White, the eastward decrease in volatile matter and the increase in fixed carbon content of coal in the eastern United States was due to lateral compression by folding and thrusting. The relation between low volatile matter, high fixed carbon content, and structural complexity in the Southern Anthracite field was cited by White as proving his theory. In 1925, White published a comprehensive summary showing that the rank of coal is not due to original composition or depth of burial, but is due to progressive regional carbonization as the result of lateral compression. Turner, in 1934, supported White by postulating that carbonization in the southern part of the Anthracite region was brought about by horizontal thrust pressures applied prior to and from a different direction than the forces that formed the present fold system.

In 1933, Willard determined that in the region between the Lehigh and Susquehanna Rivers the basal beds of the Catskill Formation are younger towards the west. This westward decrease in age was attributed to intertonguing between continental red beds of the Catskill Formation and marine gray beds of laterally equivalent but locally subjacent rock units.

C. D. White (1934) and Willard (1936) independently theorized that an unconformity exists between the Catskill and Pocono Formations in the vicinity of the Southern Anthracite field. Neither, however, presented data supporting their theories.

During the years from 1931 to 1943, Willard, F. M. Swartz, C. K. Swartz, and A. B. Cleaves described in varying detail the Ordovician to Devonian rock sequence of the southern part of the Anthracite region. Willard (1934; 1935a–d; 1936; 1939a; 1941; 1943) concentrated his work largely upon Middle and Upper Devonian rocks and established much of the stratigraphic nomenclature used in this report. C. K. and F. M. Swartz studied the Silurian and Lower Devonian rocks east and west of the report area (Swartz, C. K. and F. M., 1931, 1941; F. M. Swartz, 1939). Cleaves (1939) described the Oriskany Group west of the area and with Willard (Willard and Cleaves, 1939) substantiated the unconformity at the top of the Ordovician sequence.

During 1944 and 1945, two estimates of reserves of anthracite were published. The first of these was by the Pennsylvania Geological Survey (Pennsylvania Bur. Statistics and Information, 1944) and the second, a correction of the first, was by Ashley (1945).

Commencing in 1950, a series of reports by members of the U.S. Geological Survey on the Western Middle and Southern Anthracite fields was published. This report is one of the series. Most of these reports describe the geology of parts of the Western Middle Anthracite field (Rothrock and others, 1950; 1951a, b; 1953; Haley and others, 1953, 1954; Kehn and Wagner, 1955; Danilchik and others, 1955; Maxwell and Rothrock, 1955; Wood and others, 1958; Danilchik and others, 1962; Arndt, Danilchik, and Wood, 1963; Arndt, Wood, and Danilchik, 1963). Others describe stratigraphic correlations and structural phenomena that are characteristic of the two coal fields (Wood and others, 1956; Arndt and others, 1959, 1962; Arndt and Wood, 1960; Trexler and others, 1961, 1962; Wood and Kehn, 1961; Wood and others, 1962). Two of these reports describe the geology of small parts of the report area (Wood and others, 1958, 1962).

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PRINCIPAL GEOLOGIC FEATURES

The rocks and unconsolidated sediments of the area range in age from Ordovician to Recent. Ordovician rocks consist of about 1,000 feet of shale and sandstone with a few beds of limestone. Rocks of Silurian age range in thickness from 2,700± to about 4,500± feet and are composed of a lower thin to thick sequence of gray conglomerate, sandstone, and shale overlain successively by a thick sequence of red sandstone, siltstone, and shale and a thin sequence of limestone. Lower Devonian rocks consist mostly of gray limestone, shale, and sandstone: Middle Devonian rocks are predominantly gray to dark-gray fine-grained clastics; Upper Devonian rocks are largely red sandstone, and shale. Devonian rocks range in thickness from 8,500± to 12,500± feet. Rocks of Mississippian age are composed of 1,000± to 3,800± feet of gray conglomerate and red sandstone, siltstone, and shale overlain by 3,000± to 7,500± feet of red sandstone, siltstone and shale. The Pennsylvanian sequence, which locally contains a few beds of red sandstone, shale, and siltstone at its base, consists largely of gray conglomerate, sandstone, siltstone, shale, and numerous beds of anthracite. It has a maximum thickness of about 5,000± feet. Deposits of unconsolidated Quaternary alluvium and talus mantle much of the area. These deposits locally are as thick as as a few tens of feet on some hillsides and along some valley bottoms.

The average thickness of the outcropping rocks is about 24,500 feet. This great thickness of rocks has been divided into 31 formations and members. In addition, within the thicker formations and members, many lithologic units have been mapped to aid in tracing structural features and in understanding and correlating the sedimentary facies of the thicker units.

The geologic structure is complex, the intensity of deformation increasing from northwest to southeast. Folding, faulting, and erosion related to the Taconic disturbance occurred southeast of the area in Late Ordovician time and are presumed to have occurred within the areas as well. Many east-northeast-trending overturned and recumbent folds and nappes have been identified southeast and east of the area (Drake and others, 1960; Pennsylvania Geol. Survey, 1954). A late phase of the Acadian disturbance deformed the rocks along a similar trend in Early Mississipian time. During the Appalachian orogeny in post-Early Pennsylvanian—pre-Late Triassic time, deformation again took place along the east-northeast trend. The structural features formed during the Taconic and Acadian disturbances were deformed, obscured, and in some places virtually obliterated by structural features developed in the Appalachian orogeny. The earliest and latest structural features of the Appalachian orogeny seem to have resulted from compression from the east-southeast and east, but those formed during the medial part of the orogeny seem to have resulted from compression from the southeast rather than the east.

The structural features of the Anthracite region that developed during the Appalachian orogeny formed in a progressive sequence. This sequence has been divided arbitrarily from northwest to southeast, on the basis of increasing complexity, into five structural stages. Initially, the horizontal strata of the region were flexed into broad, low folds (stage 1). The low folds were then successively broken by low-angle thrust faults, built into anticlinoria and synclinoria, and fractured by higher angle thrust faults (stage 2). The previously developed thrust faults and folds of stage 2 were further folded and broken by additional high-angle faults (stage 3). As deformation continued, the folds were overturned and further fractured by tear faults and additional high-angle faults (stage 4). As the orogeny drew to a close, recumbent folds and nappes formed south and east of the area (stage 5; Arndt and Wood, 1960). The interrelations of structural features in the Anthracite region show that any area containing folds and faults typical of the more complex stages and had previously undegone deformation characteristic of the preceding stages. The folds and faults characteristic of each stage are preserved in discrete geographic parts of the region. In a general way, structural features typical of stage 3 occupy the report area north of the trough of the Minersville synclinorium, and those typical of stage 4 occur between the trough and Blue Mountain. South of Blue Mountain, recumbent folds and nappes of stage 5 prevail.

STRATIGRAPHY

ORDOVICIAN SYSTEM

MARTINSBURG SHALE

The Martinsburg Shale of Middle and Late Ordovician age was named from Martinsburg, W. Va., by Geiger and Keith (1891), who stated that the formation overlay the Shenandoah Limestone and underlay the Massanutten Sandstone. Various authors (Rogers, 1858b, p. 105; Lesley and others, 1895, p. 525–624; Stose, 1909; Behre, 1927, p. 19–34, 1933, p. 138–148; Stose, 1930, p. 634; Willard, 1943, p. 1081; Pennsylvania Geol. Survey, 1954, p. 6) have used a wide variety of names and divisions for the Martinsburg. In this report the Martinsburg is considered to be those beds beneath the Tuscarora Sandstone.

The Martinsburg Shale underlies the southern slope of Blue Moutain in the Pine Grove and Swatara Hill quadrangles (GQ–689, 691) but is covered by talus of the overlying Tuscarora Sandstone which caps the mountain. It is poorly exposed in a few places a short distance southwest of Round Head in the Swatara Hill quadrangle.

The formation is the oldest stratigraphic unit exposed in the area; therefore, the nature of its basal contact is not known. The upper contact is covered by talus, but east of the area at Schuylkill Gap in Blue Mountain, the Tuscarora is believed to rest on the Martinsburg unconformably. West of the area, at Susquehanna Gap in the same mountain, Ordovician rocks younger than the Martinsburg underlie the Tuscarora and are presumably conformable with the Martinsburg (Willard and Cleaves 1938, p. 6–7).

The thickness of the Martinsburg Shale in the area is not known but probably greatly exceeds the 1,000 feet believed to underlie Blue Mountain. Many geologists have computed thicknesses of the formation in the belt of outcrop between the Delaware and Susquehanna Rivers to the south of the area. However, the structure of this belt is so complex that the thicknesses reported by different authors vary greatly even at the same locality. Reported thicknesses throughout the belt are highly variable and so erratic that no pattern of change in thickness can be established.

The authors were unable to divide the Martinsburg into members. However, it is our opinion that only the lower black shale unit originally recognized by Stose (1930, p. 634) is present, and that his upper sandstone member was removed by pre-Tuscarora erosion.

The Martinsburg consists mostly of gray to dark-gray shale interbedded with many beds of gray and yellowish-brown silstone and sandstone and a few beds of gray to dark-gray limestone. The shale is tabular bedded in units a few inches to many feet thick and is commonly silty or sandy. The siltstone is also tabular bedded in units of similar thickness, is argillaceous, and generally contains fragments of gray shale. The sandstone is in beds 4 inches to 6 feet thick and is of two types: a tabular-bedded very fine grained to medium-grained argillaceous sandstone commonly containing pellets and fragments of gray shale, and a cross-laminated to tabular-bedded medium- to coarse-grained sandstone. The limestone is dense, unfossiliferous, and tabular bedded in units 6 inches to 3 feet thick.

The sediments of the Martinsburg Shale are marine and are believed by the authors to have accumulated below the depth of intensive wave agitation. The even bedding of the finer clastics and the limestone suggest that the sediments composing these rocks were little affected by other types of strong currents during deposition, but the fragments of gray shale in the siltstone and sandstone units and the cross-laminations of the coarser clastics indicate the existence of currents of moderate strength during their deposition. The cross-laminations dip northwest and suggest that the source of the sediments lay to the southeast.

The Martinsburg Shale is generally considered to be of Middle and Late Ordovician age. E. O. Ulrich (in Stose, 1930, p. 643–644) and Willard and Cleaves (1938, p. 6) correlated the fossil fauna from the uppermost beds of the formation at Swatara Gap, 1 mile south of the report area, and Sterretts Gap, west of the Susquehanna River, with that of the Eden Group (of former usage) of Ohio, Indiana, and northern Kentucky. According to this correlation, the fauna is of Late, but not latest, Ordovician age (Twenhofel and others, 1954, chart 2); any younger beds, if deposited at the gap and in the area, presumably were removed by erosion that resulted from the Taconic orogeny and preceded deposition of the Tuscarora Sandstone.

ORDOVICIAN-SILURIAN UNCONFORMITY

The Martinsburg Shale and the overlying Tuscarora Sandstone of the report area generally are believed to be separated by an unconformity that resulted from the Taconic orogeny. Evidence of the unconformity is lacking in the area. Grabau (1921, p. 293), Miller (1926, p. 503–505), and Stose (1930, p. 646) concluded that an angular unconformity separates these formations at Schuylkill Gap in Blue Mountain a short distance east of the area. Chance (in White, I. C., 1882, pl. 5), Lesley (in Lesley and others, 1892–95, p. 674), Schuchert (1916, p. 549), Willard and Cleaves (1939, p. 1104), and Willard (1943, fig. 6) believed that the angular contact between these formations at the gap is a fault. The authors, after examination of the exposures in the gap, agree with Grabau, Miller, and Stose that the contact is an unconformity. We also believe that the unconformity extends at least as far southwest as Swatara Gap.

SILURIAN SYSTEM

The Silurian rocks that crop out in the report area are, in ascending order: the gray Lower Silurian Tuscarora Sandstone, the red and gray Middle Silurian Clinton Formation, the Upper Silurian Bloomsburg Red Beds, and the gray Upper Silurian and Devonian Keyser Limestone. The gray, olive, and red Wills Creek Shale and the gray Tonoloway Limestone of Late Silurian age lie between the Bloomsburg Red Beds and Keyser Limestone a short distance west of the Klingerstown quadrangle, but they are absent in the Pine Grove and Swatara Hill quadrangles.

Sediments that formed the Lower and Middle Silurian rocks were derived from a landmass east and southeast of the area which was uplifted during the Taconic orogeny of Late Ordovician time. The landmass at first contributed a great volume of coarse detritus to a restricted shallow inland sea. By early Late Silurian time the landmass had been eroded greatly, and the detritus, which had become much finer, accumulated on a delta that gradually forced the inland sea northwestward. The uppermost Silurian rocks are carbonates and shales that accumulated as the landmass was nearly base leveled and perhaps submerged as the sea transgressed southeastward onto the delta and open shallow-sea conditions prevailed for the first time since deposition of the Martinsburg Shale.

TUSCARORA SANDSTONE

The Tuscarora Sandstone of Early Silurian age was named by Darton (1896a, b) in his reports on the Piedmont and Franklin quadrangles of Maryland, West Virginia, and Virginia. The formation was originally named the Levant White Sandstone by Rogers (1858b, p. 105) and later the Medina White Sandstone by Lesley (in Lesley and others, 1892–95, p. 625). During succeeding decades the term "Medina" was used with so many stratigraphic and chronologic meanings that the geologic nomenclature and correlations of Upper Ordovician and Lower Silurian rocks in New York, Pennsylvania, West Virginia, and Virginia became confused. As a result, local names such as Tuscarora were adopted.

The Tuscarora Sandstone underlies principally the crest and southern slopes of Blue Mountain in the Pine Grove and Swatara Hill quadrangles (GQ–689, 691), but in a few localities, the upper part extends several hundred feet down the north slope of the mountain. The rocks of the formation generally break into large talus blocks which mantle both the Tuscarora outcrop and underlying and overlying stratigraphic units. The Tuscarora is poorly exposed. No complete or nearly complete sections have been found; therefore, the succession of strata and their stratigraphic relations are not known.

The Tuscarora Sandstone rests on the Martinsburg Shale unconformably. The upper contact is placed arbitrarily at the base of the lowest red bed typical of the Clinton Formation. It is rarely exposed in the area but is rather easy to trace in the field with the aid of aerial photographs because of the erosional-lithologic contrast between red beds of the Clinton and gray beds of the Tuscarora. The contact seems to be consistently at the same stratigraphic horizon and probably is conformable.

Computed thicknesses of the Tuscarora Sandstone along Blue Mountain in the Pine Grove and Swatara Hill quadrangles average about 400 feet. Available data indicate that the formation averages about the same thickness wherever recognized as a separate stratigraphic unit in eastern Pennsylvania. This consistency suggests that the topographic and structural irregularities of the Taconic disturbance were reduced to relative smoothness during the Ordovician and Silurian erosion cycle before the wide and even deposition of the Tuscarora.

Fresh exposures of the sandstone and conglomerate of the Tuscarora are very light gray, medium gray, yellowish gray, dark yellowish brown, light brownish gray, and brownish gray (Goddard, 1948). Joint surfaces in freshly exposed rocks commonly are stained by limonite, which seems to disappear gradually as weathering continues. Weathered rocks are predominantly very light gray; but grayish-orange, yellowish-gray, medium-gray, light-brown, and moderate-brown colors are common.

Beds of the Tuscarora are ½ inch to 6 feet thick and average about 1 foot in thickness. In the lower part of the formation the average bed is about 3 feet thick, but in the upper part, the average thickness is much less. Most beds are tabular or lenticular; however, wedge-shaped and irregular beds are rather common.

Cross-stratification is one of the principal characteristics of the bedding of the Tuscarora. Planar cross-stratification of the "torrential" type is very common and simple cross-stratification is slightly less common. Most planar cross-strata are small scale, being less than 1 foot long (McKee and Weri, 1953). Simple cross-strata range in length from 1 foot to about 25 feet. Most cross-strata originally dipped northwest, which suggests that the source of Tuscarora sediments lay southeast of the report area. Yeakel (1958) reached the same conclusion as to the direction of the source in his study of the formation in Pennsylvania and adjacent States.

The Tuscarora is composed largely of sandstone and a subordinate amount of conglomerate, mostly in the lower part of the formation. Medium- to coarse-grained sandstone is present throughout the formation and is the principal constituent of the upper part. A few thin beds of siltstone, silty shale, and fine-grained sandstone are intercalated in the conglomerate and coarser grained sandstone. Compositionally, the sandstone is chiefly subgraywacke, but protoquartzite and orthoquartzite are common.

Pebbles in the conglomerate average about half an inch in diameter but are locally as much as 3 inches in diameter. They are composed largely of light- to medium-gray, milky, and white vein quartz and light- to dark-gray, brownish-gray, and medium bluish-gray quartzite; small amounts of dark-gray and moderate-red chert; and a few irregularly shaped fragments of gray to light-gray shale and slate. The clasts of vein quartz, quartzite, and chert are mainly subrounded to rounded, those of shale and slate are principally subangular and were probably derived from the underlying Martinsburg Shale. In most conglomerate beds the clasts are scattered unevenly or are concentrated in trains. The matrix of the conglomerate is fine- to coarse-grained common quartz, vein quartz, quartzite, and chert and irregular fragments of mica and clay.

The orthoquartzite and protoquartzite are composed mainly of grains of vein and common quartz and quartzite and of lesser amounts of chert, tourmaline, and zircon. Interstitial material consists of biotite, muscovite, sericite, and unidentified clay minerals. The principal cementing material is silica. In a few thin sections, extensive quartz overgrowths are present on grains of vein and common quartz. The average diameter of grains is about 0.4 mm (medium), but the range in diameter is from 0.02 mm (silt) to 1.5 mm (very coarse). Some orthoquartzite and protoquartzite have a vague bimodal tendency; the lesser mode averages 0.25 mm in diameter and the greater averages about 0.5 mm.

The sorting of most orthoquartzite is excellent and that of protoquartzite is moderate to excellent. Grains in both types of sandstone are subangular to round but are chiefly subround.

Quartzite grains tend to be slightly more angular than grains of vein and common quartz, chert, tourmaline, and zircon. Some almost perfectly rounded grains of vein and common quartz were broken during transportation from the source area. These almost perfectly rounded grains probably were derived from sedimentary rocks in the source area, whereas the quartzite grains probably were derived from metamorphic rocks.

The mineral assemblage of the subgraywacke consists of grains of vein and common quartz; blebs and plates of sericite, biotite, chlorite, muscovite, and unidentified clay minerals; films of limonite and hematite; a few grains of leucoxene, rutile, tourmaline, and zircon; and varying amounts of fragments of quartizite, chert, phyllite, shale and slate. Silica is the principal cement, but some subgraywacke is cemented by hematite, limonite, and clay.

The diameters of the mineral grains in the subgraywacke range from 0.01 mm (silt) to 2.0 mm (very coarse). Many vein quartz grains and quartzite fragments reach diameters of 2.0 mm, but the average diameter is 0.5 to 0.75 mm (coarse). Common quartz grains have an average diameter of 0.25 to 0.35 mm, which is somewhat less than that of vein quartz and quartzite. Most grains of tourmaline and zircon are less than 0.25 mm in diameter. In much of the subgraywacke, quartzite, and quartz grains interpenetrate because of pressure solution. In addition, about 50 percent of the vein quartz grains have semicomposite extinction and comb structure.

Sorting in the subgraywacke is poor to very good. Grains are poorly to well rounded but average subangular. Most quartzite fragments are less rounded than are grains of vein and common quartz and chert, whereas zircon and tourmaline grains are commonly more rounded.

The siltstone and shale of the Tuscarora contain a high percentage of identifiable silt grains in addition to unidentified clay minerals. The silt grains are chiefly angular to subangular common quartz and smaller amounts of vein quartz, quartzite, chert, mica, and rock fragments.

The environment in which Tuscarora sediments were laid down has been discussed for years, and the conclusions range from entirely marine deposition to entirely nonmarine deposition (F. M. Swartz, 1948; Folk, 1960, p. 21; Yeakel, 1958, 1962, p. 1536). The fossils and sedimentary features of the rocks of the Tuscarora are suggestive of accumulation in an environment that extended from the edge of a coastal plain for an unknown distance into shallow marine waters. The following evidence supports such accumulation. *Arthrophycus* imprints, which are commonly believed to indicate deposition in a strandline environment (Pelletier, 1958, p. 1057), are numerous in the upper part of the formation. Stream-channel deposits are absent in the Tuscarora. Clay- and silt-sized particles are relatively sparse in rocks of the formation, which suggests that they were largely winnowed out and that sediment sorting was probably accomplished by currents other than streams. The predominance of northwest-dipping cross-strata suggests that deposition was from currents that moved consistently from the southeast, probably normal to the trend of a strandline.

Arthrophycus alleghaniensis, a worm burrow or animal track, is the only fossil found in the Tuscarora of the area. Clarke and Ruedemann (1912, p. 419) reported that Dr. Gilbert Van Ingen found *Eurypterus maria*, *Dolichopterus* cf. *D. ostius*, *Hughmilleria shawangunk*, *Pterygotus globiceps*, *Stylonurus* cf. *S. myops*, *Erettopterus* sp., and *Arthrophycus alleghaniesis* in float from the Tuscarora Sandstone at Swatera Gap. C. K. Swartz and F. M. Swartz (1930; 1931, p. 655) believe that this fauna firmly fixes the age as Early Silurian.

The Tuscarora Sandstone is generally considered to be the correlative of the Clinch Sandstone of the southern Appalachians, the Albion Group of western New York, and the lower part of the Shawangunk Conglomerate of eastern New York, easternmost Pennsylvania, and New Jersey.

CLINTON FORMATION

The Clinton Formation of Middle Silurian age was named for Clinton, N.Y., by Conrad (1842, p. 228–235). In terms of modern stratigraphic nomenclature, the formation as originally defined included the rocks lying between the Albion Sandstone, as it was then called, and the Rochester Shale. Shortly after the turn of the century, Ulrich (1911, pl. 28) and Kindle and Taylor (1913, p. 6–7) redefined the Clinton to include the Rochester Shale as the upper member. Many years before this redefinition, I. C. White (1883, p. 104–114) had correlated the Clinton southward into eastern Pennsylvania. Unfortunately, the stratigraphic units that bound the formation in New York by either definition are not recognized in eastern Pennsylvania. On the basis of regional correlations, therefore, the Clinton in the area is considered to include the rocks between the Tuscarora Sandstone and the overlying Bloomsburg Red Beds of Late Silurian age.

The Clinton Formation underlies the northern slope and, in a few places the crest of Blue Mountain in the Pine Grove and Swatara Hill quadrangles (GQ–689, 691). Outcrops are sparse and generally small. The rocks of the formation, as does the Tuscarora Sandstone, weather into large talus blocks that mantle the southern part of the outcrop belt of the overlying Bloomsburg Red Beds.

The contact between the Clinton and the underlying Tuscarora Sandstone is placed arbitrarily at the base of the lowest red bed of typical Clinton lithology. Although exposed at only a few places, it is easy to trace because of a slight swale that has been eroded in the basal beds of the Clinton and a small linear hill that has formed in the uppermost resistant sandstone beds of the Tuscarora. The contact between the red and gray beds of the Clinton and the much less resistant overlying Bloomsburg Red Beds is not exposed in the report area. The contact is believed to underlie a sharp topographic break in slope that resulted because of the marked difference in erosional resistance between beds of the Clinton and beds of the Bloomsburg. The break in slope is not well expressed on the topography of quadrangle maps GQ–689 and 691, but is easy to trace on the ground in the field with aerial photographs. The contact between these formations as shown on maps GQ–689 and 691 is also the boundary between the Clinton and overlying Quaternary talus deposits.

The thickness of the Clinton Formation in the report area and nearby localities cannot be determined because of thrust faulting. The maximum unduplicated thickness on Blue Mountain is about 2,300 feet; therefore, the formation is more than 2,300 feet thick at this locality, but how much thicker is unknown. The thickness of 2,300+ feet is the maximum reported for the Clinton in eastern Pennsylvania. It seems likely that the formation thins gradually to the northwest and west away from this maximum in the report area to the approximate 1,000-foot average thickness reported at other localities in eastern Pennsylvania (Swartz and Swartz, 1931, p. 630, 640; Burtner and others, 1958; Miller, 1961, pl. 2).

The Clinton is composed almost entirely of intercalated sandstone, silty shale, and shale. A few thin beds of conglomeratic sandstone are present in the lower part of the formation near the crest of Blue Mountain in the Swatara Hill quadrangle (GQ–689).

Rocks of the Clinton are predominantly grayish red, moderate red, and pale red; some are pale brown, light to medium dark gray, and light brownish gray. They commonly weather to the same colors and to grayish orange, light olive gray, and yellowish gray.

The sandstone beds in the Clinton Formation are commonly thicker than associated beds of silty shale and shale. Individual sandstone beds, which are generally about 1 foot 6 inches thick, range in thickness from ½ inch to 4 feet. Shale beds locally are as much as 2 feet thick, but are chiefly about 1 inch thick. Most strata are tabular; however, a few are lenticular and irregular. Small-scale planar and simple cross-stratification is present in some outcrops, but is uncommon.

The sandstone of the Clinton is subgraywacke, protoquartzite, and orthoquartzite in decreasing order of abundance. The subgraywacke consists of grains of common and vein quartz; fragments of quartzite, schist, slate, shale, and sandstone; grains and plates of biotite, chlorite, muscovite, sericite, and leucoxene; grains of magnetite, ilmenite, rutile, tourmaline, zircon; and films of hematite and limonite and unidentified clay minerals. Several of the subgraywacke sequences are highly ferruginous and contain as much as 15 to 20 percent by volume of hematite as cement and interstitial material. These sequences were named informally the Center and Swatara iron sandstones by C. K. Swartz and F. M. Swartz (1931, p. 638). Cementing media of the subgraywacke are chiefly silica and hematite, but in some beds are sericite, other micas, a binder matrix of quartz flour, and unidentified clay minerals. The grain size ranges from very fine to coarse, the average being fine. Sorting is poor to excellent but is mainly moderate to good. Grains are chiefly subround; however, subangular to round grains are rather common. Many quartz grains that lie in contact interpenetrate as the result of pressure solution; whereas others do not interpenetrate because they are surrounded by a matrix of mica fragments and clay which seems to have prevented pressure solution.

The principal mineral constituents of the protoquartzite are grains of common and vein quartz. Accessory constituents are fragments of quartzite, schist, slate, shale, and sandstone and grains of biotite, muscovite, sericite, chlorite, rutile, tourmaline, zircon, magnetite, ilmenite, and leucoxene. Some sandstone of this type contains as much as 10 percent by volume of hematite, which occurs chiefly as cement and interstitial material. Cementing media are silica and hematite and, in a few places, sericite, other micas, and unidentified clay minerals. Although the diameter of grains ranges from 0.1 to 0.8 mm, the average diameter is about 0.2 mm (fine). Sorting is moderate to excellent. The degree of original rounding of grains is difficult to determine in most protoquartzite because of interpenetration due to pressure solution. Despite interpenetration, the range of rounding seems to have been from subangular to well rounded and to have averaged subround.

The orthoquartzite of the Clinton is composed of a greater percentage of common and vein quartz grains than is the protoquartizite, but otherwise these two types of sandstone are similar.

The silty shales of the Clinton are predominantly quartzose and, to a much lesser extent, micaceous. They are composed largely of silt-sized grains of common and vein quartz embedded in a matrix of unidentified clay minerals. The clay shales are also quartz rich, but contain a much greater percentage of unidentified clay minerals than do the silty shales. Some beds of both types of shale contain a high percentage of hematite in the matrix.

Clinton sediments probably accumulated in a nearshore marine environment where oxidizing and reducing conditions predominated at different times. The marine environment is proven by fossils in the Clinton at Swatara Gap and in correlative beds in the Shawangunk Conglomerate of easternmost Pennsylvania (Swartz and Swartz, 1931, p. 636–637, 655). The red color of many of the shale, siltstone, and fine-grained sandstone beds of the formation is due to hematite cement and interstitial material. The hematite probably records oxidizing conditions that predominated during and following deposition of these fine clastics. The medium- and coarse-grained sandstone beds of the formation are commonly gray and contain considerably less hematite. The gray color of these beds probably records reducing conditions during and after deposition. A second, but less likely, possibility is that the marine currents that carried the finer clastics and iron either in solution or as hematitic material were localized by topographic features on the sea bottom or were forced to bypass localities where the coarser sediments were accumulating. A third, and much less likely, possibility is that the iron was deposited only from the marine currents either before or after deposition of the coarse sediments. Folk (1960, p. 42), in support of the latter theory, postulated that correlative red Clinton fine clastics in West Virginia were deposited a short distance offshore from the mouths of streams that carried large amounts of iron in solution. He postulated that the iron was precipitated as chlorite and glauconite along with the fine clastics and that at a later time oxygenated water converted these two minerals into hematite. The absence of glauconite and the small amount of chlorite in the fine clastic rocks of the Clinton suggests that Folk's postulate is not applicable in the report area.

Fossils have not been found in the Clinton Formation in the report area. A short distance to the south, however, at Swatara Gap, C. K. Swartz and F. M. Swartz (1931, p. 636–637) collected a large fauna. After studying this collection and additional collections from other localities in central and eastern Pennsylvania, they indicated on chart 3 of the "Correlation of the Silurian Formations of North America" (Swartz, C. K., and others, 1942) that the formation is correlative with the Niagara Series and is of Middle Silurian age. The U.S. Geological Survey considers the Clinton to be Middle Silurian.

BLOOMSBURG RED BEDS

The Bloomsburg Red Shale was named by I. C. White (1883, p. 106) for the red rocks exposed along the east bank of Fishing Creek at the northern boundary line of Bloomsburg, Pa. Subsequent geologists have abandoned the term "shale" and substituted "red beds" because the formation contains a considerable amount of red sandstone.

The Bloomsburg Red Beds are poorly exposed in the area. Outcrops are in two gently rolling, subdued linear belts in the Pine Grove and Swatara Hill quadrangles (GQ–689, 691). The southern belt underlies a series of valleys between Blue and Swope Mountains; the northern belt occupies the core of the Roedersville anticline in the central part of the Swatara Hill quadrangle.

The contact between the Bloomsburg Red Beds and the underlying Clinton Formation is not exposed in the area. It is placed arbitrarily at a sharp topographic break in slope at the lithologic change between resistant beds of the Clinton and less resistant beds of the Bloomsburg (see description of the Clinton Formation).

In the Pine Grove and Swatara Hill quadrangles, the contact between the Bloomsburg and the overlying Ridgeley Sandstone is unconformable. In its few exposures the contact is sharp; red beds of the Bloomsburg apparently parallel gray conglomerate beds of the Ridgeley. North and west of the report area, several hundred feet of uppermost Silurian and lowermost Devonian rocks intervene between these formations. The unconformity at the base of the Ridgeley must truncate these intervening rocks between the area to the north and west and the exposures in the Pine Grove and Swatara Hill quadrangles, but data are scanty to ascertain where the unconformity truncates these intervening rocks.

The Bloomsburg Red Beds range in thickness from about 700 feet in the northwestern part of the area to about 1,800 feet in the southwestern part. Figure 3 shows the thickness of the formation in the area and the surrounding region. It was prepared from thicknesses measured by the authors, C. K. Swartz and F. M. Swartz (1931), Miller (1961), Hoskins (1961), and I. C. White (1883). The formation thickens southeastward across the report area at a rate of about 35 feet per mile. This average rate of thickening was calculated without reference to the structural foreshortening between localities where measurements were made. Tectonic analysis based upon all available structural and stratigraphic data indicates that the Silurian and Devonian rocks of the area were shortened about 30 percent by folding and faulting; therefore, before the shortening, the Bloomsburg seems to have thickened southeastward at about 25 feet per mile.

Where freshly exposed, the shale, siltstone, and finer grained sandstone of the Bloomsburg are predominantly grayish red, some being pale to moderate red, moderate reddish brown, or yellowish gray. The coarser grained sandstone is generally greenish gray, but light olive gray, dark greenish brown, and dusky brown are not uncommon. Weathered rocks are mainly pale reddish brown to moderate reddish brown. Other weathered rocks are grayish red, dusky red, light brown, dark reddish brown, grayish brown, and moderate yellowish brown.

The beds of the Bloomsburg range in thickness from about ½ inch to about 2 feet 6 inches, the average being about 1 foot. Most are tabular but some are lenticular. Small-scale planar and simple cross-stratification is present in some medium- to coarse-grained sandstone.

The Bloomsburg is composed principally of siltstone and very fine grained sandstone intercalated with shale. Fine- to coarse-grained sandstone is rare, and granular to small pebble conglomerate is very rare. The sandstone consists, in decreasing order of abundance, of very fine-grained to fine-grained protoquartzite, very fine-grained to fine-grained orthoquartzite, and fine- to medium-grained subgraywacke.

Except for a difference in the amount of quartz, the mineral constituents of the protoquartzite and orthoquartzite are similar. Both are composed largely of common and vein quartz grains; scattered grains of quartzite, chert, shale, mica, tourmaline, rutile, zircon, ilmenite, magnetite, and leucoxene; sparse grains of andesine; and hematite and limonite. In either rock type many grains are embedded in a matrix of hematite and limonite, which in some protoquartzite constitutes as much as 10- to 15-percent of the rock. The andesine grains are the oldest observed occurrence of this mineral in rocks of the area.

The cement of the protoquartzite and orthoquartzite of the Bloomsburg is principally hematite and second-

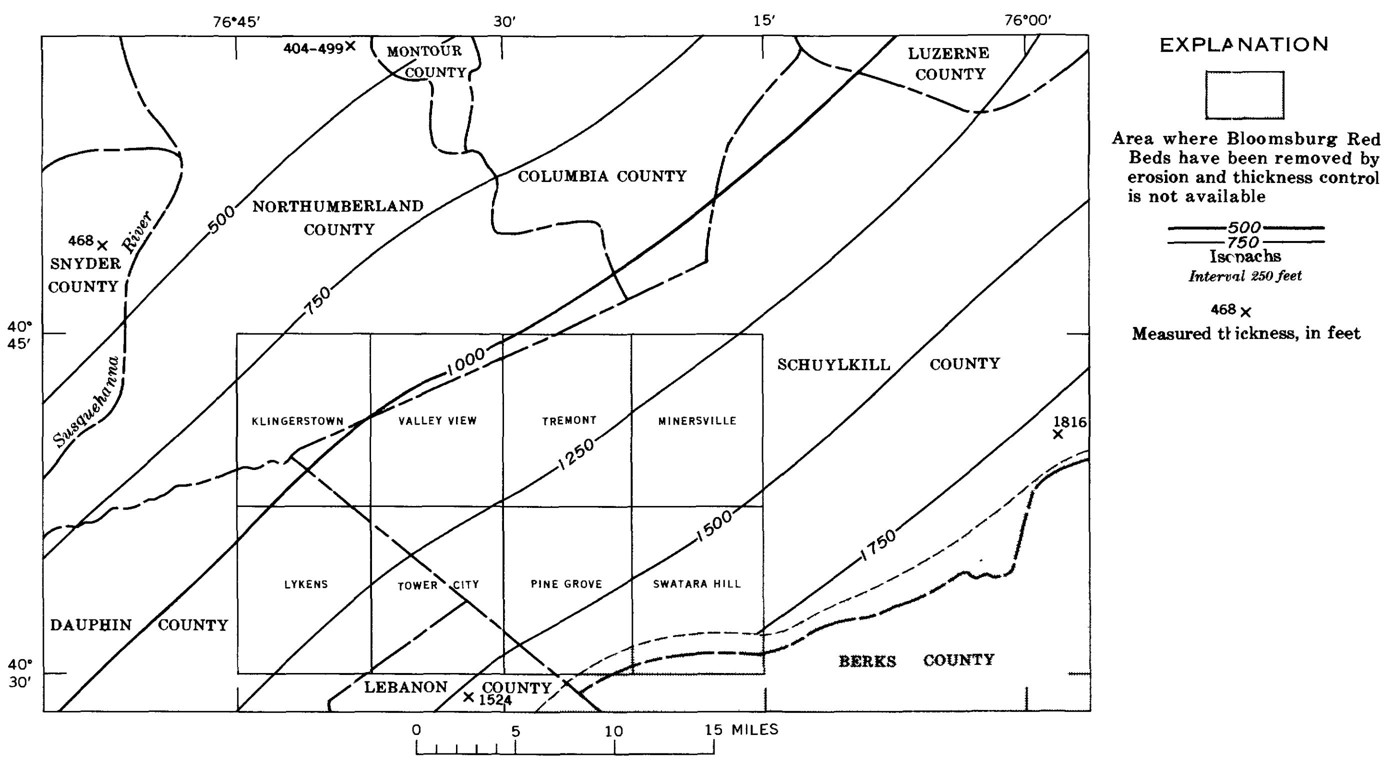


FIGURE 3.—Isopach map of the Bloomsburg Red Beds, east-central Pennsylvania.

arily silica. Sorting ranges from very good to excellent. Two distinct grain-size modes are common in both types of sandstone. The grains of the larger mode are sparse, constituting not more than 5 percent of the rock. They are usually well rounded, in contrast to those of the smaller mode which are subangular to subround. The diameters of grains in the larger mode range from 0.15 to 0.9 mm and those of the smaller from 0.07 to 0.125 mm.

The subgraywacke in the Bloomsburg consists of grains of vein and common quartz; small amounts of biotite, muscovite, chlorite, sericite, tourmaline, zircon, and unidentified clay minerals; also rock fragments of quartzite, chert, and shale. Some subgraywacke contains as much as 10 percent interstitial hematite. The cementing media are silica, hematite, and clay minerals. Sorting ranges from poor to good. The grain size ranges from 0.01 mm (silt) to about 1.0 mm (coarse sand) and averages about 0.25 mm (fine to medium sand). Many quartz grains have overgrowths, and others have deeply sutured margins because of pressure solution. The original rounding of grains is difficult to determine because of the sutured margins and overgrowths, but seems to have averaged about subround.

The siltstone in the Bloomsburg is similar in composition to the protoquartzite and orthoquartzite. Two distinct grain-size modes are present in much of the siltstone. The larger mode averages about 0.3 mm in diameter, and the smaller is silt size.

The shale in the formation is quartz rich, commonly silty, and locally micaceous. Clay-sized particles consist largely of quartz, mica, and illite.

The rocks of the Bloomsburg probably accumulated on a great delta whose source lay to the southeast. Some, however, may have been deposited in lagoons and shallow marine waters adjacent to the delta. No evidence points exclusively to any of these environments other than a general lithologic similarity between the rocks of the Bloomsburg and the rocks of the Catskill and Mauch Chunk Formations, both of which contain deltaic and subaerial deposits (Barrell, 1907, 1913–14). Alling and Briggs (1961, fig. 12) concluded that the Bloomsburg accumulated on a large delta whose main body lay in eastern New York and eastern Pennsylvania.

The marine depositional environment of parts of the Bloomsburg has been recognized for many years in central Pennsylvania where red beds assigned to it intertongue with marine gray shales and limestones of the McKenzie Formation and the Wills Creek Shale (Willard, 1939b, p. 14; Swartz, F. M., 1946, p. 20; Hoskins and Conlin, 1958, p. 156–161; Hoskins, 1961, fig. 6).

Fossils have not been found in the Bloomsburg Red Beds of the area, and the only age assignment that logically can be made is that the formation is younger than the Clinton Formation of Middle Silurian age and is older than the Ridgeley Sandstone of middle Early Devonian age. A short distance west of the area, Hoskins and Conlin (1958, p. 161) found in the Bloomsburg Late Silurian ostracode assemblages, which they correlated with similar forms in the McKenzie Formation and Wills Creek Shale of central Pennsylvania. C. K. Swartz and F. M. Swartz (1931, p. 659) also determined that the Bloomsburg in the region surrounding the area is of Late Silurian age and that it is laterally equivalent to the McKenzie and Wills Creek, and perhaps to the lower part of the Tonoloway Limestone. They additionally correlated the Bloomsburg with the Longwood and High Falls Formations of New Jersey and eastern New York and with the Vernon Shale of the Salina Group of central New York. Alling and Briggs (1961, fig. 12) correlated the Bloomsburg with the basal part of the Cayuga Series of Michigan, Ohio, and New York and also considered it to be an equivalent of the Vernon Shale.

WILLS CREEK SHALE

The Wills Creek Shale was named by Uhler (1905, p. 20–25) for Wills Creek at Cumberland, Md. In 1911, Ulrich (p. 522, 541, and pl. 28) correlated the formation into central Pennsylvania and stated that it underlay the Tonoloway Limestone and overlay the Bloomsburg Red Beds.

The Wills Creek is not exposed in the area but is believed to underlie much of it north of the Sweet Arrow fault zone. It is absent south of the fault zone. Where exposed in the surrounding region, the formation consists of medium- to olive-gray calcareous shale, argillaceous limestone, and siltstone; red to olive sandstone; and olive-gray to brownish-red shale, siltstone, and sandy limestone.

The thickness of the Wills Creek Shale is unknown in the subsurface of the area. The nearest outcrops are at Dalmatia, about 8 miles west of the area, where several hundred feet of the upper part of the formation are poorly exposed. Between Dalmatia and the Sweet Arrow fault zone, the Wills Creek thins in a southeastward direction to extinction. Synthesis of much regional geologic and stratigraphic data indicates that the formation thins in this direction because it has been truncated by pre-Ridgeley erosion, which also truncates all other pre-Ridgeley formations down to and including the upper part of the Bloomsburg Red Beds. Thus, the Wills Creek Shale in the subsurface of the report area probably is absent near the Sweet Arrow fault zone, but it may reach a thickness of 400± feet in the northwestern part on the Hooflander Mountain anticline.

According to Alling and Briggs (1961, p. 544 and fig. 21), the Wills Creek Shale accumulated during Cayuga time in an open sea that was bounded on the east by the Bloomsburg delta and on the west by Niagara reef banks. The fauna of the formation is of Late Silurian age. The formation is correlative with the middle part of the Bloomsburg Red Beds of eastern Pennsylvania (Swartz and Swartz, 1931, fig. 2) and with the upper part of the Salina Formation of New York, Ohio, and Michigan (Swartz, C. K., and others, 1942, chart 3).

TONOLOWAY LIMESTONE

The Tonoloway Limestone of Late Silurian age was named by Ulrich (1911, p. 590) from Tonoloway Ridge west of Rock Ford, W. Va. As presently defined, the formation rests conformably on the Wills Creek Shale and underlies conformably the Keyser Limestone of Late Silurian and Early Devonian age.

The contact between the Tonoloway and the underlying Wills Creek Shale is arbitrarily placed at the horizon where the amount of the limestone above exceeds that of the clastic material below. The contact between the Tonoloway and the overlying Keyser Limestone is placed where laminated limestone of the Tonoloway is overlain by nodular limestone of the Keyser.

The Tonoloway is not exposed in the area, but may underlie much of the area north of the Sweet Arrow fault zone. It is absent at outcrops south of the fault. The nearest exposure to the area is a few hundred feet west of the Klingerstown quadrangle on the axis of the Hooflander Mountain anticline. There and elsewhere in the Susquehanna Valley, the formation consists of very thin bedded to thick-bedded medium-gray laminated argillaceous limestone. Strata in the upper part of the Tonoloway generally are thick bedded and are characterized by vertical calcareous mud-filled cracks, which commonly unite to form prisms. At most localities, strata in the lower part are thin bedded and shaly.

The formation is absent south of the Sweet Arrow fault zone because of truncation by pre-Ridgeley Sandstone erosion or because of a depositional thinning and pre-Ridgeley erosion. F. M. Swartz (1939, fig. 10) showed that the area south of the fault was part of a highland mass which he named the Auburn promontory. In 1941, he (Swartz and Swartz, 1941, p. 1187) postulated that the Tonoloway thinned south and southeast toward the promontory because of depositional thinning and pre-Ridgeley erosion. The authors believe that the formation was truncated by erosion north of the Sweet Arrow fault zone before thinning out depositionally. However, they recognize that the depositional thinning indicates that the highland mass south of the fault (Auburn promontory) was being uplifted and eroded from Tonoloway time into early Ridgeley time.

Owing to the lack of exposures north of the Sweet Arrow fault zone, the thickness of the Tonoloway in the subsurface of the area is conjectural. Figure 4 depicts a series of measured sections which illustrate the lithology and thickness of the formation at localities near the area.

A map showing postulated thickness variations of the Tonoloway in the subsurface of the area is shown in figure 5. It was constructed by using the measurements shown on figure 4 and many other measurements in a region much larger than that of figure 5. The assumption was made in constructing the map that the formation thins evenly southward across the area to zero at the Sweet Arrow fault. Thus, north of the fault the Tonoloway is believed to thicken northward from 0 at the fault to about 250 feet in the subsurface of the northern part of the area. The average rate of thickening in that direction is about 20 feet per mile. The Silurian rocks of the area were foreshortened about 30 percent by folding and faulting. Before the folding and faulting, therefore, the rate of northward thickening was about 14 feet per mile.

The limestone of the Tonoloway accumulated in a sea that was bounded on the east by the Bloomsburg delta (Alling and Briggs, 1961, fig. 12) and on the south by a low landmass, the Auburn promontory. The sea in which the limestone was deposited seems to have been widespread, as shown by the large area underlain by the formation and its correlatives—Maryland to New York, but in places it became so shallow that mud cracks formed.

The Tonoloway Limestone contains a sparse marine fauna of middle Late to latest Late Silurian age (Reeside, 1917, p. 193; Swartz, C. K., and others, 1942, chart 3). It is correlative with the Bossardville Limestone of eastern Pennsylvania, New Jersey, and southeastern New York and with the Bass Island Dolomite of Michigan and Ohio (Alling and Briggs, 1961, p. 518).

SILURIAN AND DEVONIAN ROCKS

KEYSER LIMESTONE

The Keyser Limestone of Silurian and Devonian age was named by Ulrich (1911, p. 563, 590–591, and pl. 28) for Keyser, W. Va. Ulrich considered the Keyser to be the basal member of the Helderberg Formation. Many years later F. M. Swartz (1939, p. 49) proposed raising the Keyser to formation rank separate from the Helderberg and tentatively assigned it to the Silurian. In 1942, two committees that studied the correlations of Silurian and Devonian rocks in North America (Swartz, C. K., and others, 1942; Cooper and others, 1942) followed Swartz's recommendation and raised the Keyser Limestone to a formation. The committee on Silurian correlations assigned the Keyser to the Silurian but indicated that it could be Devonian. The Devonian committee considered it to be wholly Silurian. The U.S. Geological Survey recognizes the Keyser as a formation and designates its age as Late Silurian and Early Devonian.

The upper part of the Keyser Limestone crops out only in the western part of the Klingerstown quadrangle where it is poorly exposed (GQ–700). The lower part of the formation is well exposed in a series of quarries that lie in the Pillow quadrangle a few hundred yards west of this locality. There, it is a thin- to thick-bedded medium- to dark-gray fossiliferous nodular limestone with a few interbeds of laminated to medium-bedded light- to medium-gray yellowish-gray weathering shale. The limestone is finely to coarsely crystalline and is commonly transected by numerous thin calcite veins. The upper part is less well exposed in the quarries and consists of a lower sequence of even-bedded light-gray to medium-bluish-gray shaly to thick-bedded limestone with a few interbeds of medium-gray nodular limestone, and an upper sequence of laminated light-gray to medium-bluish-gray finely crystalline limestone.

West and northwest of the area, the Keyser seems to rest conformably upon the Tonoloway Limestone and to be overlain conformably by the Coeymans Limestone Member of the Helderberg Formation. However, Reeside (1917, p. 199) has suggested that both contacts of the Keyser may be unconformities marked by faunal and lithologic breaks. The lower contact lies between laminated limestone typical of the Tonoloway and nodular limestone characteristic of the Keyser. The upper contact, which is more difficult to determine, is placed above the laminated sparsely fossiliferous finely crystalline limestones of the upper part of the Keyser and below the thick-bedded finely to coarsely crystalline highly fossiliferous limestone typical of the Coeymans Limestone Member of the Helderberg.

The thickness of the Keyser Limestone in the subsurface of the area and in most of the surrounding region is uncertain. Figure 4 shows measured sections of the formation on the Hooflander Mountain anticline and at other localities adjacent to the area. It also shows that the formation is absent south of the Sweet Arrow fault zone (fig. 4, sections 3–5).

In the past, many geologists have not distinguished the Keyser Limestone from the overlying Coeymans and New Scotland Limestone Members of the Helderberg Formation. As a result, precise regional data concerning the thickness of the Keyser are sparse. In addition, the contact between the Keyser and these overlying rocks commonly is difficult to determine without faunal evidence. Therefore, the Keyser, Coeymans, and New Scotland are considered as a single stratigraphic unit between the Tonoloway Limestone and the Mandata Member of the Helderberg Formation. The map (fig. 6) was compiled from the measured sections illustrated in figure 4 and from additional thickness measured in the region surrounding figure 6. The Keyser-Coeymans-New Scotland sequence is believed to thicken northward in the subsurface of the area from a featheredge near the Sweet Arrow fault zone to about 250 feet in the northern part at a rate of about 22 feet per mile. The Silurian and Devonian rocks were shortened about 30 percent by folding and faulting during the Acadian and Appalachian orogenies. Thus, before deformation, the rate of northward thickening was about 15 feet per mile.

Much of the region south of the Sweet Arrow fault zone is believed to have been a part of a Silurian and Devonian landmass, the Auburn promontory. F. M. Swartz (1939, fig. 10 and p. 36, 39) postulated that the Keyser thinned depositionally southward towards the landmass. Regional evidence not only suggests southward thinning, but also strongly suggests that the formation was truncated north of the fault zone before deposition of the Ridgely Sandstone.

The Keyser Limestone was deposited in a shallow but widespread sea similar to that in which the underlying Tonoloway Limestone accumulated. It is correlative with the Decker, Rondout, and Manlius Formations of southeastern New York, New Jersey, and eastern Pennsylvania and with the Cobleskill Limestone and Decker, Rondout, and Manlius Formations of central New York (Swartz, C. K., and others, 1942, chart 3).

DEVONIAN SYSTEM

Devonian rocks form the thickest systemic sequence cropping out in the area. They range in thickness from about 8,500 feet in the central part of the area to about 12,500 feet in the Lykens quadrangle.

Lower Devonian rocks consist of marine limestone, sandstone, and shale. They are divided into the Helderberg Formation and the overlying Oriskany Group. The Helderberg is divided, from the base up, into the Coeymans Limestone Member, New Scotland Limestone Member and Mandata Member. The Oriskany Group is composed, in ascending order, of the Shriver Chert and the Ridgeley Sandstone.

Middle Devonian rocks are composed of shale, siltstone, sandstone, and conglomeratic sandstone and a thin limestone sequence near the base. In ascending order, they consist of the Needmore Shale, the Selinsgrove Limestone, and the Marcellus Shale and Mahantango Formation of the Hamilton Group. The Mahantango is divided into a lower shale and sandstone member, the

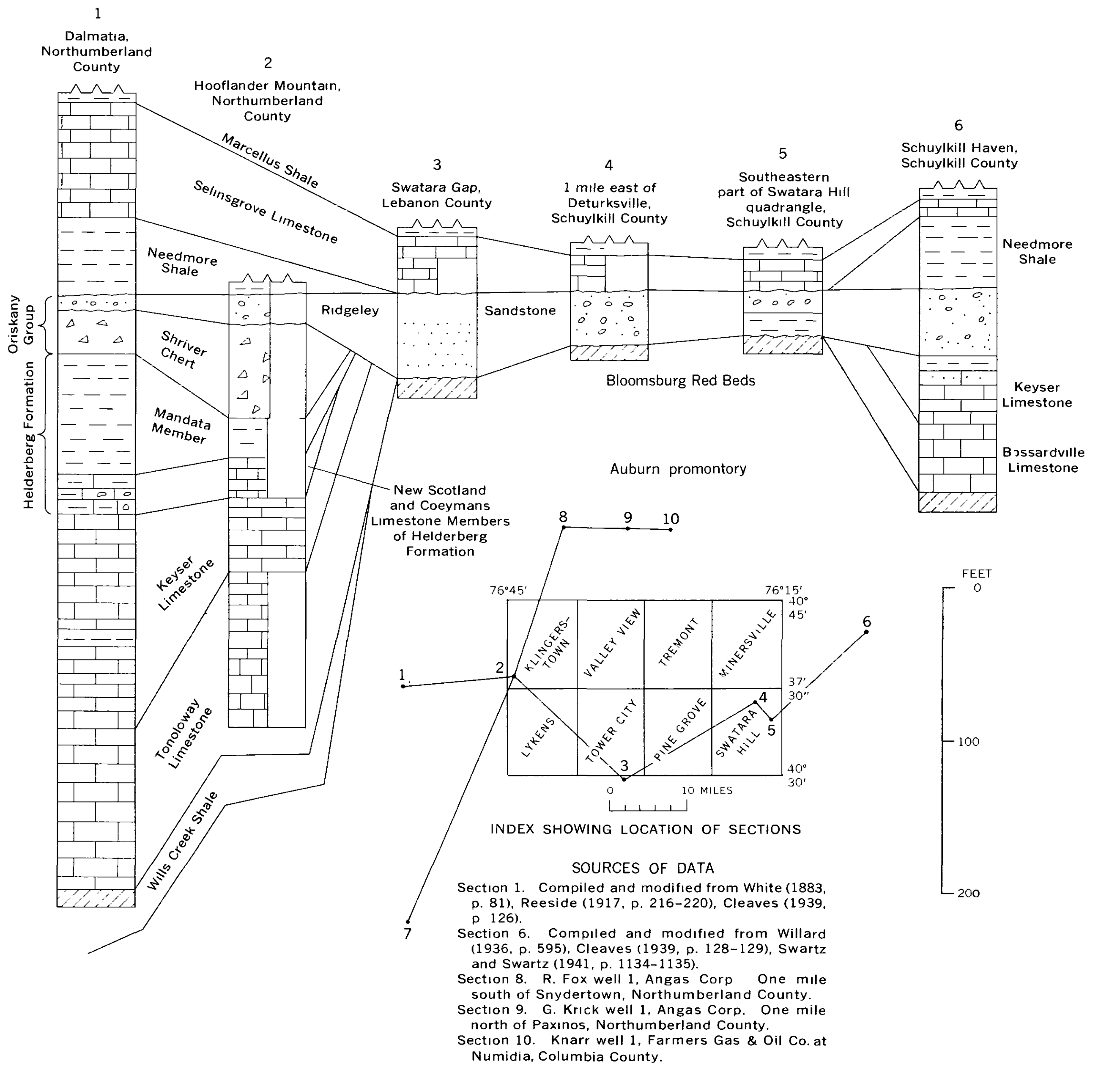


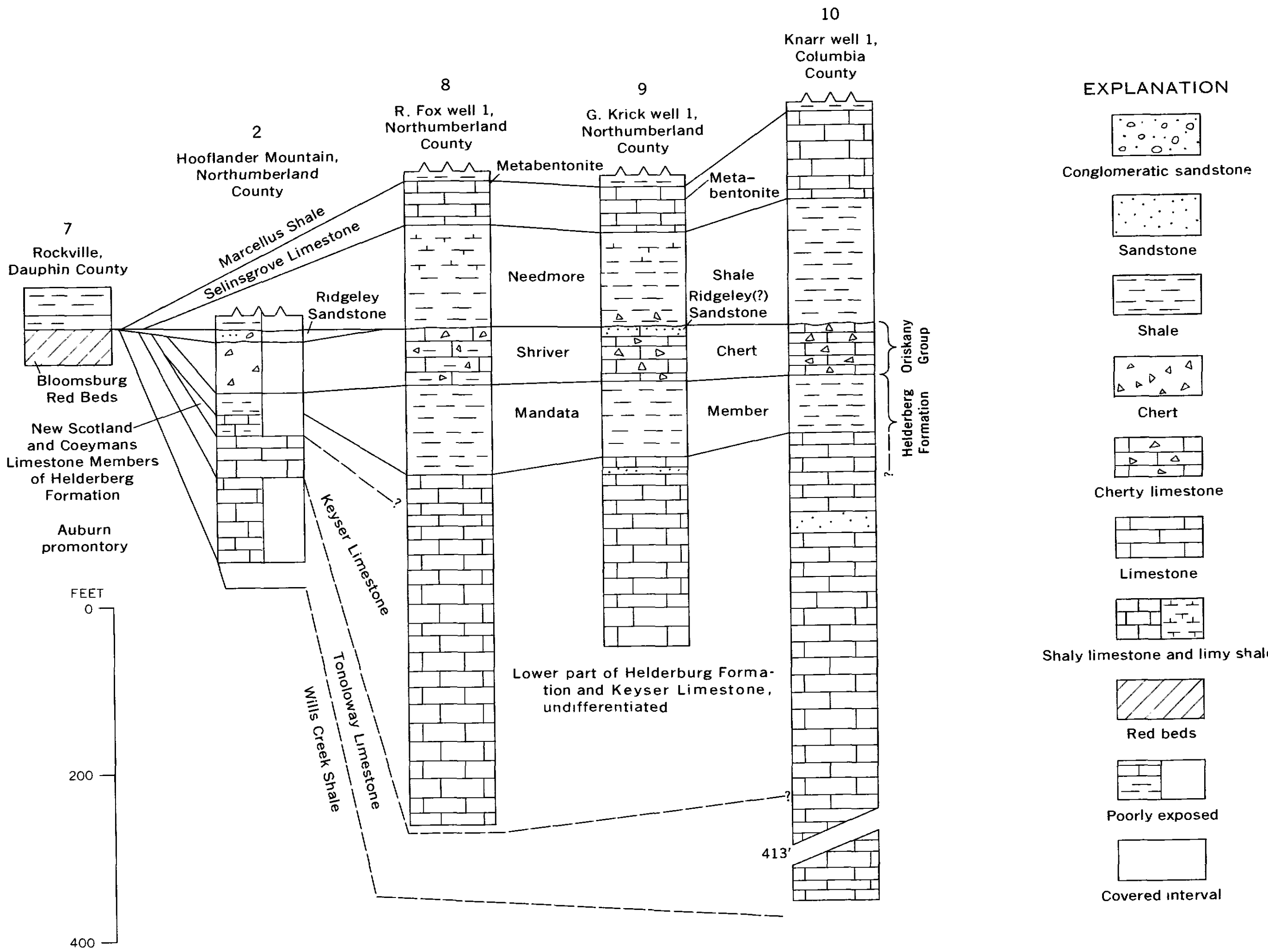
FIGURE 4.—Columnar sections of Silurian and Devonian rocks from the

Montebello Sandstone Member, and an upper shale member.

Upper Devonian rocks are included in the Upper Devonian and Lower Mississippian Susquehanna Group which is divided into the Trimmers Rock Sandstone and the overlying Catskill Formation. In the southern part of the area the Catskill is divided, in ascending order, into the Irish Valley, Damascus, Honesdale Sandstone, and Spechty Kopf Members. In the western and northern parts of the area, the Honesdale Sandstone Member thins out, and the indistinguishable Cherry Ridge and Damascus Members are mapped together as the Buddys Run Member.

HELDERBERG FORMATION

The name Helderberg was first used in 1842 by Conrad (p. 233) for the strata at the base of the Helderberg Mountains in Albany County, N.Y. Conrad applied the name to the rocks that, in terms of modern stratigraphic nomenclature, extend from the base of the Rochester Shale of the Clinton Group to the top of the Oriskany Group. Over the years, the rocks assigned to the Helderberg have been referred to by many names and have included a great variety of strata which, in their widest application, extended from the top of the Salina Formation of Late Silurian age to the base of the Marcellus



base of the Tonoloway Limestone to the top of the Selinsgrove Limestone.

Shale of Middle Devonian age (Wilmarth, 1938, p. 935–936; Rogers, 1858b, p. 107; White, I. C., 1883, p. 94–98; Swartz, C. K., 1913, p. 105–120; Reeside, 1917, p. 186–189; Swartz, F. M., 1939, p. 50–91).

In the area of this report the Helderberg is of Early Devonian age and includes the strata lying between the Keyser Limestone and the Shriver Chert of the Oriskany Group. It consists, from the base up, of the Coeymans and New Scotland Limestone Members and the recently recognized Mandata Member (deWitt and Colton, 1964, p. 34–35). It has been mapped as a formation because the Coeymans and New Scotland are not mappable units and so must be called members here, rather than formations as they have been called in other areas.

The Helderberg crops out at only one locality in the western part of the Klingerstown quadrangle (GQ–700). It is believed to underlie much of the area north of the Sweet Arrow fault zone but is absent south of this zone. Members of the formation are not shown separately on GQ–700 because they are too thin to depict adequately at the scale of the map.

In the stratigraphic sections illustrated in figure 4, the Coeymans and New Scotland Limestone Members

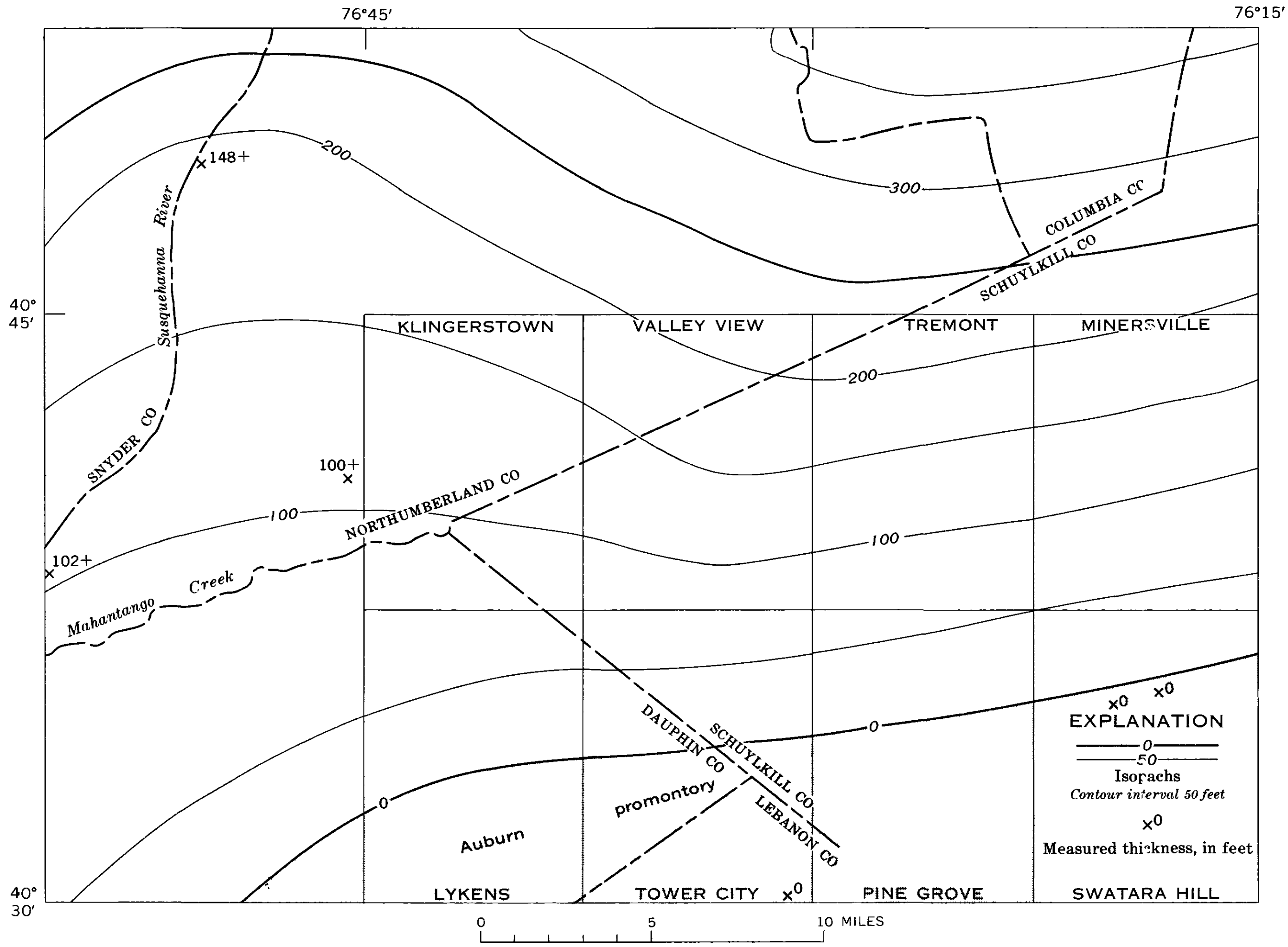


FIGURE 5.—Isopach map of the Tonoloway Limestone (shaded where absent).

are combined because the Coeymans New Scotland contact at Hooflander Mountain (fig. 4, section 2) is uncertain. In fact, the two members can be separated only at Dalmatia (fig. 4, section 1), 8 miles west of the area, where Reeside (1917, p. 217) recognized 2 feet of sandy limestone belonging to the Coeymans and 21.6 feet of limestone belonging to the New Scotland.

At Hooflander Mountain the Coeymans and New Scotland sequence consists, in descending order, of about 23 feet of dark-gray, medium bluish-gray to black, light-gray weathering laminated to medium-bedded finely to coarsely crystalline slightly cherty fossiliferous limestone and 1 to 3 feet of dark-gray coarsely crystalline sandy limestone.

The contact between the Helderberg and the underlying Keyser Limestone is covered in the area (GQ-700). It was mapped where the percentage of float from limestone typical of the Keyser equals the percentage of float from limestone typical of the Coeymans and New Scotland. The contact of the New Scotland with the overlying Mandata Member is also covered and was mapped where limestone and shale float are approximately equal in volume in the soil.

The Mandata at Hooflander Mountain is about 25 feet thick (fig. 4, section 2). Its outcrop belt has been eroded into a slight swale, which is almost completely covered by soil containing numerous chips of ash-gray or yellowish-gray calcareous chert and shale. The contact with the overlying Shriver Chert seems to be at about the place in the soil where the percentage of ash-gray or yellowish-gray chert and calcareous shale fragments from the Mandata equals the percentage of larger, very light gray and pinkish-gray fragments of vuggy chert from the Shriver. In wells north of the area (fig. 4, sections 8–10), the Mandata is a dark-gray calcareous shale, 70 to 107 feet thick, separating the limestone and shale of the Shriver from the limestone of the New Scotland and Coeymans.

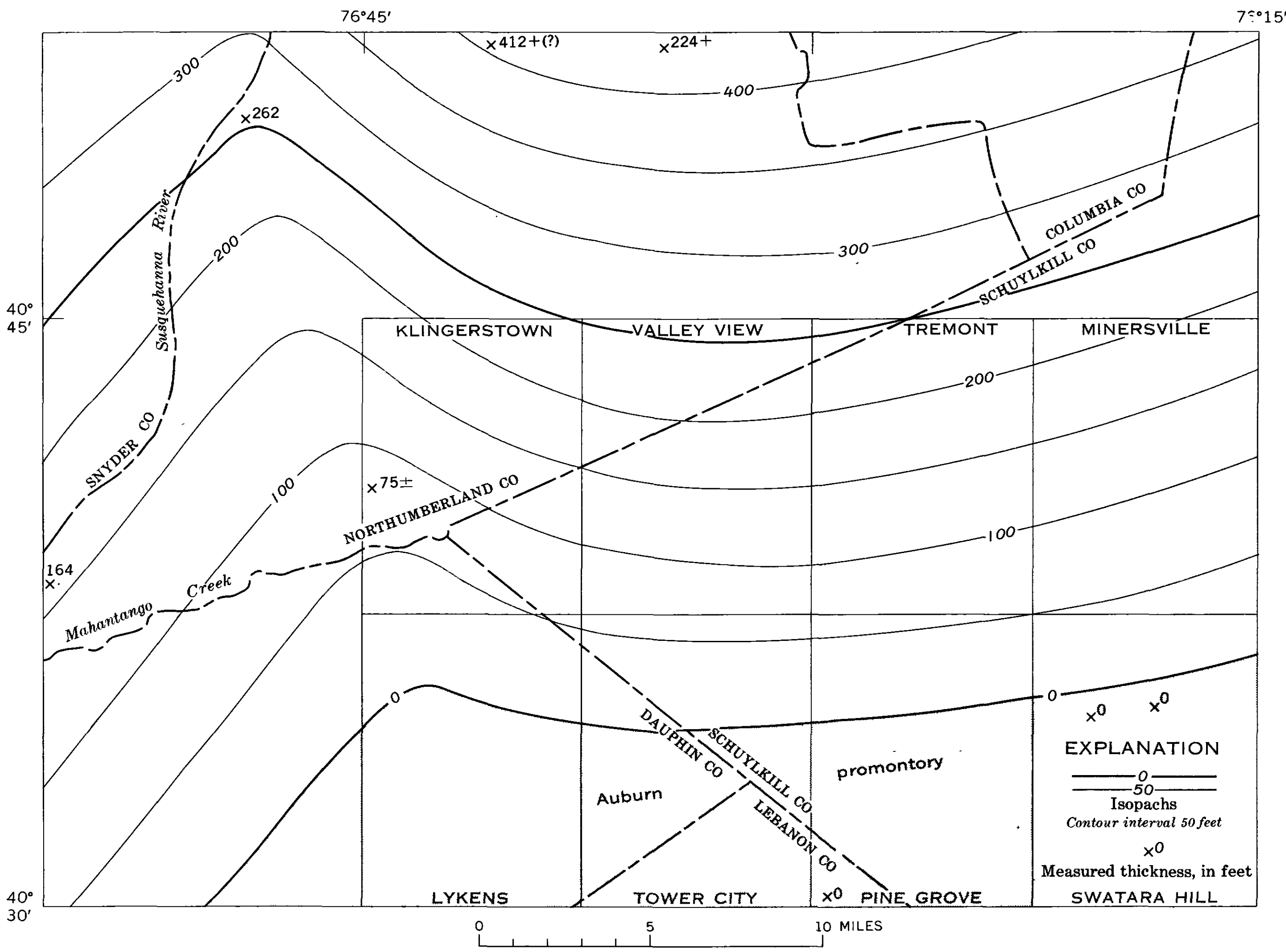


FIGURE 6.—Isopach map of the Keyser Limestone and Coeymans and New Scotland Limestone Members of the Helderberg Formation (shaded where absent).

Figure 7 is an isopach map of the Mandata Member. The zero isopach line is a short distance north of the Sweet Arrow fault zone. The map was compiled from thicknesses shown on figure 4 and from other thicknesses measured in the surrounding region. The Mandata may reach a maximum thickness of about 80 feet in the subsurface of the northern part of the area. It is believed to thicken northward at about 7 feet per mile. The authors determined by structural analysis that the Silurian and Devonian rocks of the area were foreshortened about 30 percent during the Acadian and Appalachian orogenies; thus, the pre-deformation rate of northward thickening was about 5 feet per mile.

The Helderberg Formation is absent south of the Sweet Arrow fault zone, presumably because of truncation by pre-Ridgeley erosion. The area south of the fault zone is a small part of a Silurian and Devonian positive mass, the Auburn promontory. F. M. Swartz (1939, p. 36, 39) noted a gradual increase in clastic detritus and an attendant depositional thinning of the Helderberg southward towards the promontory. Much regional stratigraphic evidence strongly suggests that the rocks of the formation were truncated a short distance north of the fault zone by pre-Ridgeley erosion. The thinning of the Helderberg and the increase in clastic detritus indicate, however, that the promontory was being uplifted and eroded at the time the formation was accumulating.

The Coeymans and New Scotland Limestone Members of the Helderberg were deposited in an open sea that had a large fauna. The Mandata apparently accumulated somewhat closer to shore where siliceous ooze formed and much silt and clay was deposited. The detritus and silica probably were derived from high parts of the Auburn promontory. The Mandata is less fossiliferous than the Coeymans and New Scotland.

The Helderberg Formation contains a large Early Devonian fauna. The Coeymans and New Scotland

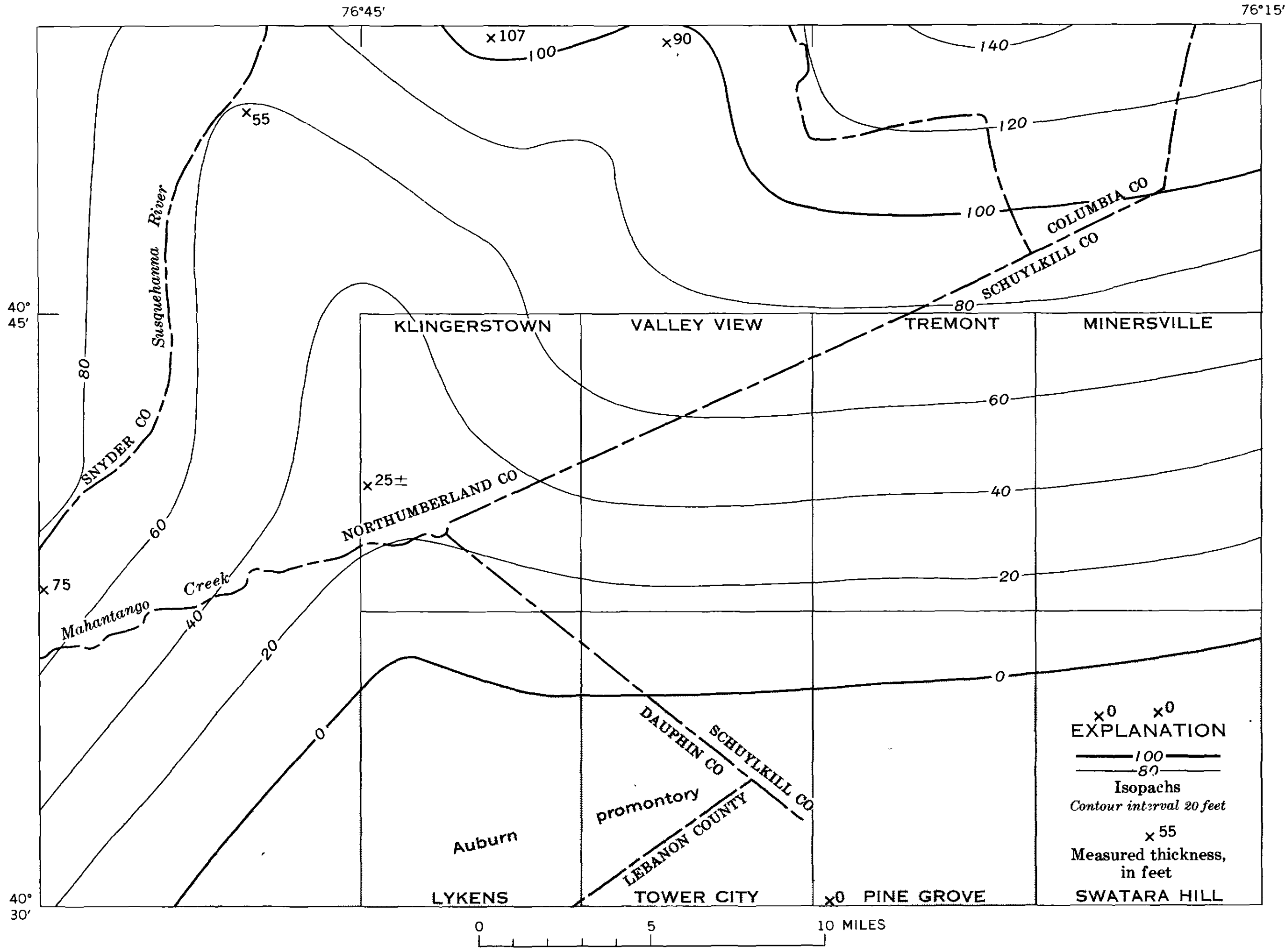


FIGURE 7.—Isopach map of the Mandata Member of the Helderberg Formation (shaded where absent).

Limestone Members are widely recognized from west-central Virginia to southeastern New York. The Mandata Member has been traced southwestward into northern Virginia and Maryland and is correlated to the north with the Port Ewen, Alsen, and Becraft Limestones of New York.

ORISKANY GROUP

The Oriskany Group of Early Devonian age was named by Vanuxem (1839, p. 273) for Oriskany Falls, N.Y. Throughout a long and involved nomenclatural history, beds as old as the Port Ewen Limestone of the Helderberg Group and as young as the Esopus Shale have been included by various authorities in the Oriskany of New York (Wilmarth, 1938, p. 1561–1562). In Pennsylvania the U.S. Geological Survey restricts the group to the Shriver Chert and Ridgeley Sandstone.

SHRIVER CHERT

The Lower Devonian Shriver Chert of the Oriskany Group was named by C. K. Swartz (1913, p. 91–92) for Shriver Ridge, Cumberland Mountain, Md. He (p. 121–122) correlated it from the type locality northeastward into central Pennsylvania and recognized equivalent strata in New Jersey and New York.

The formation crops out only on the flanks of the Hooflander Mountain anticline near the west border of the Klingerstown quadrangle and is poorly exposed (GQ–700). At this locality it is 30 to 60 feet thick and consists of thick-bedded dark-gray to black chert; thin-bedded medium- to dark-gray moderately crystalline fossiliferous cherty limestone; and dark-gray calcareous cherty shale. Relatively pure chert is largely concentrated in the upper part of the formation, whereas limestone and shale are more common in the lower part. Most of the chert in the upper part weathers very light gray, but pinkish-gray and grayish-orange-pink fragments are not uncommon. Most of the chert in the lower part weathers grayish yellow.

The lower contact of the Shriver at Hooflander Mountain was described in the section on the Helderberg Formation. The upper contact with the overlying Ridgeley Sandstone is gradational and is located where the rocks change from predominantly cherty limestone to predominantly coarse sandstone and conglomeratic sandstone.

The sequence of beds of the Shriver is similar to that at Hooflander Mountain throughout the region, except at Dalmatia (fig. 4, section 1), 8 miles west of the area, where tongues of sandstone similar to the Ridgeley Sandstone are intercalated in the upper part of the formation. At all places the upper part of the Shriver is highly fossiliferous, whereas the lower part is much less fossiliferous and locally barren. Cleaves (1939, p. 115) observed this difference in fossil abundance throughout central Pennsylvania.

An isopach map of the Shriver (fig. 8) was compiled from thicknesses at the localities illustrated on figure 4 and from thicknesses measured at other localities outside the area of figure 8. The formation is absent south of the Sweet Arrow fault zone, but may be as much as 60 feet thick in the subsurface of the northwestern part of the area. It seems to thicken northward at about 4½ feet per mile. Before the Acadian and Appalachian orogenies, which telescoped the rocks about 30 percent, the formation probably thickened northwestward at about 3½ feet per mile.

The Shriver is absent south of the Sweet Arrow fault because it was truncated by pre-Ridgeley erosion. Field evidence from many localities in eastern Pennsylvania indicates that the Auburn promontory was tilted northward about the end of the deposition of the Shriver, and the region was differentially eroded before the Ridgeley Sandstone was deposited. The region of thin Shriver shown on figure 8 west of the report area seems to have been caused by a slowly rising linear positive

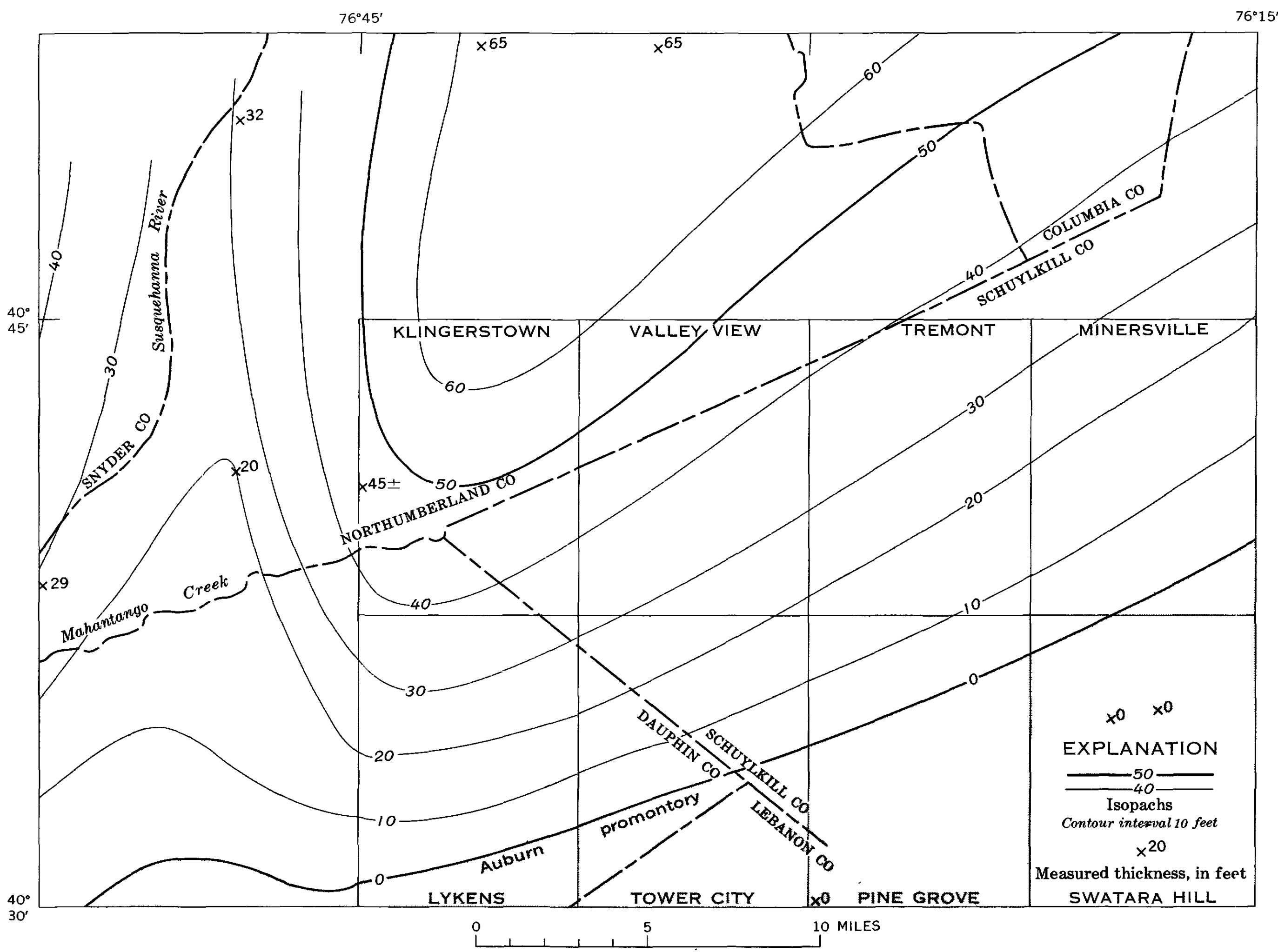


Figure 8.—Isopach map of the Shriver Chert (shaded where absent).

element or arch (Harrisburg arch of Willard, 1941), which extended northward from the promontory and which lay west of similar less pronounced arches during Helderberg and Keyser times (figs. 6, 7).

The upper part of the Shriver Chert seems to have been a near-shore deposit of siliceous ooze, silt, and clay like the underlying Mandata Member of the Helderberg Formation, except that the percentage of clay and silt was less and that of siliceous ooze, was significantly greater. The abundance of fossils, the general lack of fine clastic detritus, and the local presence of clay-free tongues of sand may indicate that this part of the formation was deposited in waters somewhat clearer than those of the lower part and the Mandata.

The fauna of the Shriver Chert is of Early Devonian age (Allan, 1935; F. M. Swartz, 1932; Cleaves, 1939). Cleaves (1939, p. 109) determined that 23 percent of the Shriver fauna is present in the underlying Helderberg Formation but that many common Shriver species are not present in the overlying Ridgeley and "vice versa" (p. 118). He also found that only 4 percent of the combined Shriver-Ridgeley fauna is recognizable in the overlying Needmore Shale and Selinsgrove Limestone and concluded that the Shriver-Ridgeley fauna was for the most part destroyed before the deposition of the Needmore and Selinsgrove.

The Shriver underlies much of central Pennsylvania, western Maryland, and eastern West Virginia. It is correlated with the lower part of the Oriskany Group, Formation, or Sandstone in Virginia, northeast Pennsylvania, and southeast New York.

PRE-RIDGELEY UNCONFORMITY

The Ridgeley Sandstone overlies the Shriver Chert conformably on the Hooflander Mountain anticline (GQ–700), but south of the Sweet Arrow fault zone (GQ–689, 691, 698) it rests on the Bloomsburg Red Beds. Between the anticline and the fault zone a pre-Ridgeley erosion surface cuts across intervening strata from the Shriver Chert to the Bloomsburg. Maps and reports of the Second Pennsylvania Geological Survey (Sanders, 1881; White, I. C. 1883; Prime and others, 1884; Smith, 1891) show that the Ridgeley rests on different formations in different areas, but the existence of an unconformity was not postulated until 1939 (Cleaves, 1939, p. 106; Swartz, F. M., 1939, p. 87). More than 400 feet of beds is truncated in a general southerly direction between the anticline and the fault zone. The average rate of truncation of about 28 feet per mile indicates a northward tilt of the report area of about a quarter of a degree.

RIDGELEY SANDSTONE

The upper unit of the Oriskany was named the Ridgeley Sandstone Member by C. K. Swartz (1913, p. 90) for a town in West Virginia. In 1939, Cleaves (p. 92–104) traced this unit from the Maryland-Pennsylvania border to Dalmatia in Northumberland County, 8 miles west of the area, and recommended classifying it as a formation in the Oriskany Group. The U.S. Geological Survey has accepted Cleaves' designation.

The Ridgeley Sandstone crops out on the flanks of the Hooflander Mountain anticline (GQ–700) in the northwestern part of the area and in two belts south of the Sweet Arrow fault zone (GQ–689, 691, 698); it is poorly exposed at all places. On the anticline the lower contact with the Shriver Chert is gradational, but in the area south of the fault zone it is an unconformity, the sandstone resting on the Bloomsburg Red Beds. In the subsurface between the outcrops on the anticline and the fault zone, the Ridgeley, from north to south, presumably overlies successively the Shriver Chert, Helderberg Formation, Keyser Limestone, Tonoloway Limestone, Wills Creek Shale, and a part of the Bloomsburg Red Beds. At Hooflander Mountain the upper contact of the Ridgeley with the overlying Needmore Shale is a poorly defined disconformity (Cleaves, 1939, p. 108). South of the fault zone, the Needmore is absent, and the Ridgeley is overlain unconformably by the Selinsgrove Limestone.

In the report area the Ridgeley ranges in thickness from 0 to slightly more than 50 feet (fig. 4). An isopach map (fig. 9) indicates that the formation probably is thickest in the subsurface of the report area. It also shows that the formation is absent locally north of the area. The thickness trends shown in figure 9 are oriented differently from the trends of the underlying Upper Silurian and Lower Devonian formations shown in figures 5 to 8. This difference in orientation of thickness trends may reflect the earliest deformation paralleling that of the later Acadian and Appalachian orogenies.

Most weathered sandstone of the Ridgeley is grayish orange. Fresh sandstone ranges from light gray to light olive gray, pinkish gray, medium gray, and yellowish gray. Most beds are thin bedded, tabular, and cross-bedded and have small-scale simple or planar cross-strata.

The Ridgeley is commonly a fine pebble protoquartzite. The pebbles are largely vein quartz and quartzite, generally about half an inch in diameter; in a few places they attain a maximum diameter of half an inch. The grain size, except for the pebbles, ranges from very fine to very coarse and averages medium. The grains range from angular to well rounded but average subangular. Cements are silica, limonite, hematite, and clay. At most localities the Ridgeley is a sparkling conglomeratic

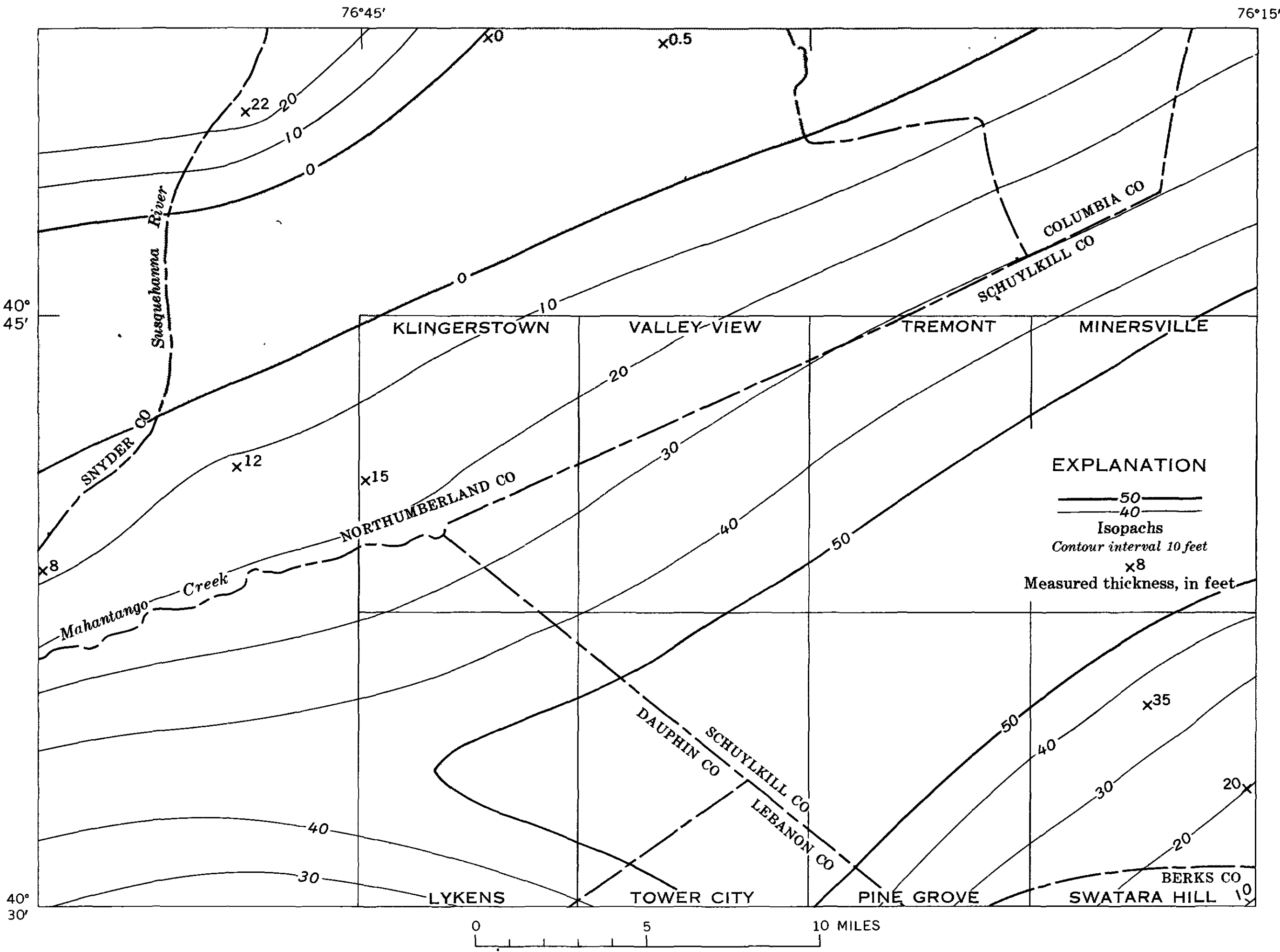


Figure 9.—Isopach map of the Ridgeley Sandstone (shaded where absent).

sandstone cemented by silica. South of the Sweet Arrow fault zone, however, it contains some subgraywacke, and in the Hooflander Mountain area it has a few thin layers of intercalated chert.

In addition to grains of vein quartz, the Ridgeley contains common quartz, biotite, chlorite, muscovite, sericite, leucoxene, tourmaline, topaz, zircon, unidentified clay minerals, and fragments of chert, schist, phyllite, quartzite, and slate. Feldspar is sparse, but about one-half percent andesine is present in some specimens. C. S. Ross (oral commun., 1960) identified small quantities of halloysite in thin sections from several localities.

The Ridgeley Sandstone accumulated in a sea that advanced southward upon the pre-Ridgeley erosion surface. The Auburn promontory, most of which lay south of the report area, probably was uplifted and eroded in late- or post-Shriver time. However, uplifting may have been so early that the Lower Devonian rock sequence may not have been deposited in the southern part of the area. This latter possibility is supported by the occurrence in the Ridgeley of the clay mineral hallosite, which is believed by some geologists to represent extremely weathered parts of shales (Rogers and Kerr, 1933, p. 287). The halloysite, however, may have been deposited after being eroded and transported from land areas to the south.

The formation is recognized from east-central Pennsylvania to West Virginia. The fauna has been intensively studied by many geologists, most of whom classify it as middle Early Devonian (Cleaves, 1939; Allan, 1935; Swartz and Swartz, 1941). The Ridgeley is correlative with the upper part of the Oriskany Sandstone of New York. It has been traced by the authors into the Silurian (?) Inwood Sandstone of C. K. Swartz and F. M. Swartz (1931, p. 635–638), whose type locality is at Swatara Gap, about 1 mile south of the central part of the area. The lack of pebbles in the Ridgeley at this gap led C. K. Swartz and F. M. Swartz to conclude that the Ridgeley was absent and that its stratigraphic position was occupied by an Upper Silurian sandstone, which they named the Inwood.

POST-RIDGELEY UNCONFORMITY

An unconformity has been postulated at the top of the Ridgeley Sandstone by Cleaves (1939, p. 108) and Willard (1939a, p. 150). They based their belief on (1) a thin limonite and hematite zone at the top of the Ridgeley, attributed to erosion or nondeposition; (2) regional variations in the thickness of the Ridgeley, presumably because of post-Ridgeley erosion; and (3) a faunal break between the Ridgeley and the Needmore Shale and Selinsgrove Limestone.

The existence of an unconformity is supported by the absence of the Needmore Shale, south of the Sweet Arrow fault zone. There, the Selinsgrove Limestone, rests upon the Ridgeley.

NEEDMORE SHALE

Willard (1939a, p. 149) defined the Needmore Shale as follows:

> Beneath the Selinsgrove limestone and supplanting it southward is a limy shale. * * * It is here dubbed the Needmore. * * * It is simply the downward continuation of the non-cherty limestone, the Selinsgrove member, and is distinguished therefrom solely upon the greater proportion of argillaceous material, and "pari parsu", the decrease in limestone content.

In the northwestern part of the area and farther north and west, the Needmore Shale rests unconformably on the Ridgeley Sandstone and is overlain conformably by the Selinsgrove Limestone. South of the Sweet Arrow fault zone, the shale is absent, presumably because it was overlapped to the south by the limestone which there rests unconformably on the Ridgeley.

The Needmore crops out only on the flanks of the Hooflander Mountain anticline in the western part of the Klingerstown quadrangle (GQ–700). There it is largely covered, and the continuity of the unit was determined by tracing shale float mixed with sandstone and limestone from the Ridgeley and Selinsgrove. The outcrop belt of the member generally underlies a gently rolling valley bottom.

The lithology of the Needmore on the Hooflander Mountain anticline is virtually unknown. West of the area at Dalmatia, however, the Needmore is well exposed and is about 50 feet thick (White, I. C. 1833, p. 81). There the upper 25 feet consists of alternating very thin and thin beds of medium- to dark-gray limestone and very thin beds of dark-gray calcareous shale. The lower 25 feet is largely dark-gray calcareous shale. The contact of the Needmore and the underlying Ridgeley is a disconformity between sandstone and shale. The upper contact of the member and the overlying Selinsgrove is gradational and is arbitrarily located where the amount of limestone exceeds that of shale.

The Needmore thickens from 0 feet near the Sweet Arrow fault zone to 140 feet several miles north of the report area (fig. 4). The thickness in the area and the surrounding region is shown in figure 10. The Needmore seems to thicken northward at average rates of 10 feet per mile in the eastern part of the area and 5 feet per mile in the western part. Deformation has shortened the Devonian rocks of the area about 30 percent. The rate of northward thickening was about 7 feet and 3.5 feet, respectively, before deformation.

The rocks of the Needmore apparently accumulated on an erosional surface of low relief (fig. 9) that was cut in the Ridgeley Sandstone as the result of a minor rejuvenation of the Auburn promontory. The sea advanced from the north; shale accumulated first; then limestone and calcareous shale alternately were laid down, and finally limestone predominated. The abundant fauna of the formation suggests that the sea was shallow and rather clear.

Willard (1939a, p. 156–160) and Cleaves (1939, p. 109) considered the Needmore to be of Middle Devonian age. The formation as recognized in the area probably correlates with the Schoharie Grit, the lower part of the Onondaga of New York (W. A. Oliver, Jr., oral commun., 1967), and with the lower part of the Needmore Shale of Maryland (deWitt and Colton, 1964, p. 40–42). Oliver considers the unit to be of Early and Middle Devonian age.

SELINSGROVE LIMESTONE

The Selinsgrove Limestone, herein adopted, was named for Selinsgrove Junction, Pa. (Willard, 1939a, p. 146). It rests conformably on the Needmore Shale on the Hooflander Mountain anticline and north and west of the area. The contact is gradational and is placed arbitrarily at the horizon where limestone predominates over shale. The Selinsgrove overlaps the Needmore southward and rests unconformably on the Ridgeley Sandstone with a sharp contact south of the Sweet Arrow fault zone.

The upper contact is too poorly exposed in the area to determine the relations, but farther north and west, the Selinsgrove grades upward into the Marcellus Shale. The contact is placed where shale predominates over limestone.

The Selinsgrove Limestone is present in three belts of outcrop in the area: the flanks of the Hooflander Mountain anticline (GQ–700); the southern part of the Swatara Hill, Pine Grove, and Tower City quadrangles (GQ–689, 691, 698); and the central part of the

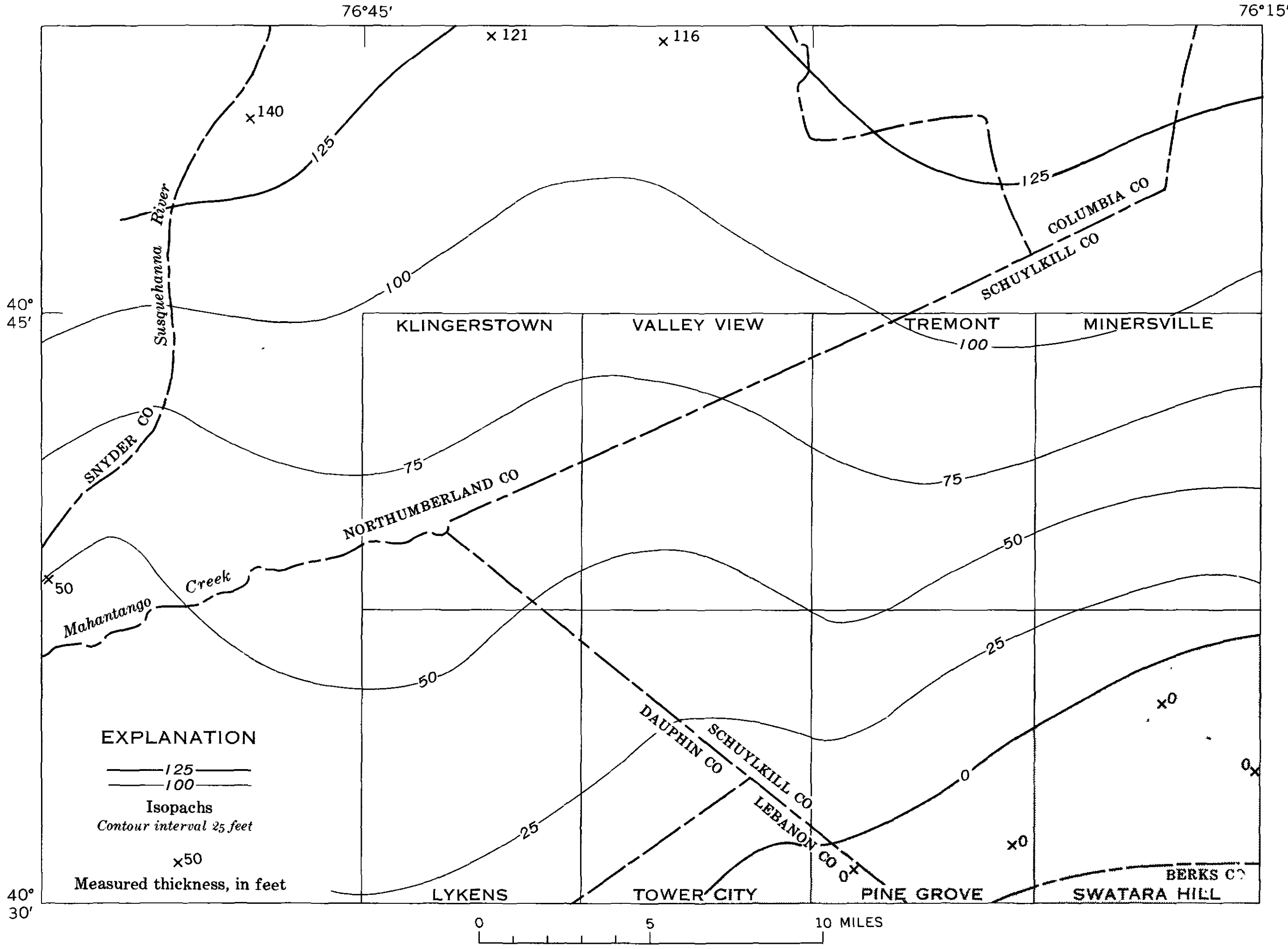


Figure 10.—Isopach map of the Needmore Shale (shaded where absent).

Swatara Hill quadrangle near Sweet Arrow Lake (GQ–689). It commonly underlies a subdued rolling or valley-bottom topography and is poorly exposed. The formation, however, is well exposed about 8 miles west of the area at Dalmatia. There, it consists of thin to thick beds of argillaceous fossiliferous finely crystalline limestone intercalated with very thin to thin beds of fossiliferous silty calcareous shale. The fresh limestone is dark gray, medium bluish gray, grayish black, and brownish black, but it weathers light gray, dusty yellow, and grayish yellow. Fresh and weathered shale, in contrast, is dark gray to black. Much of the weathered limestone splits along closely spaced bedding laminae. The lithology of the unit in the poor exposures of the area is consistent with that at Dalmatia.

The Selinsgrove is 0 to 105 feet thick in the area and surrounding region (fig. 4). Its thickness is shown by isopachs in figure 11. The unit thickens northward, but is thin in a belt that trends north through the central part of the area. This depositional pattern probably resulted from a slight rejuvenation of the Harrisburg arch part of the Auburn promontory.

The limestone of the Selinsgrove was deposited in a southward-advancing sea that overlapped previously deposited Needmore rocks and lapped onto Ridgeley rocks south of the Sweet Arrow fault zone. The prolific fauna of the limestone indicates that this sea was clear and shallow, and the lack of clastic detritus suggests that the Auburn promontory was low.

The Selinsgrove is early Middle Devonian in age (Willard, 1939a, p. 154–155; W. A. Oliver, Jr., oral commun., 1961). Willard (1939a, p. 154–155) considers the fauna older than, but similar to, that of the Marcellus Shale and Mahantango Formation. Cleaves (1939, p. 109) considers the fauna distinctly different from that of the underlying Oriskany Group. The Selinsgrove probably correlates with the upper part of the Onondaga Limestone of New York and the upper part of the Needmore Shale of western Maryland.

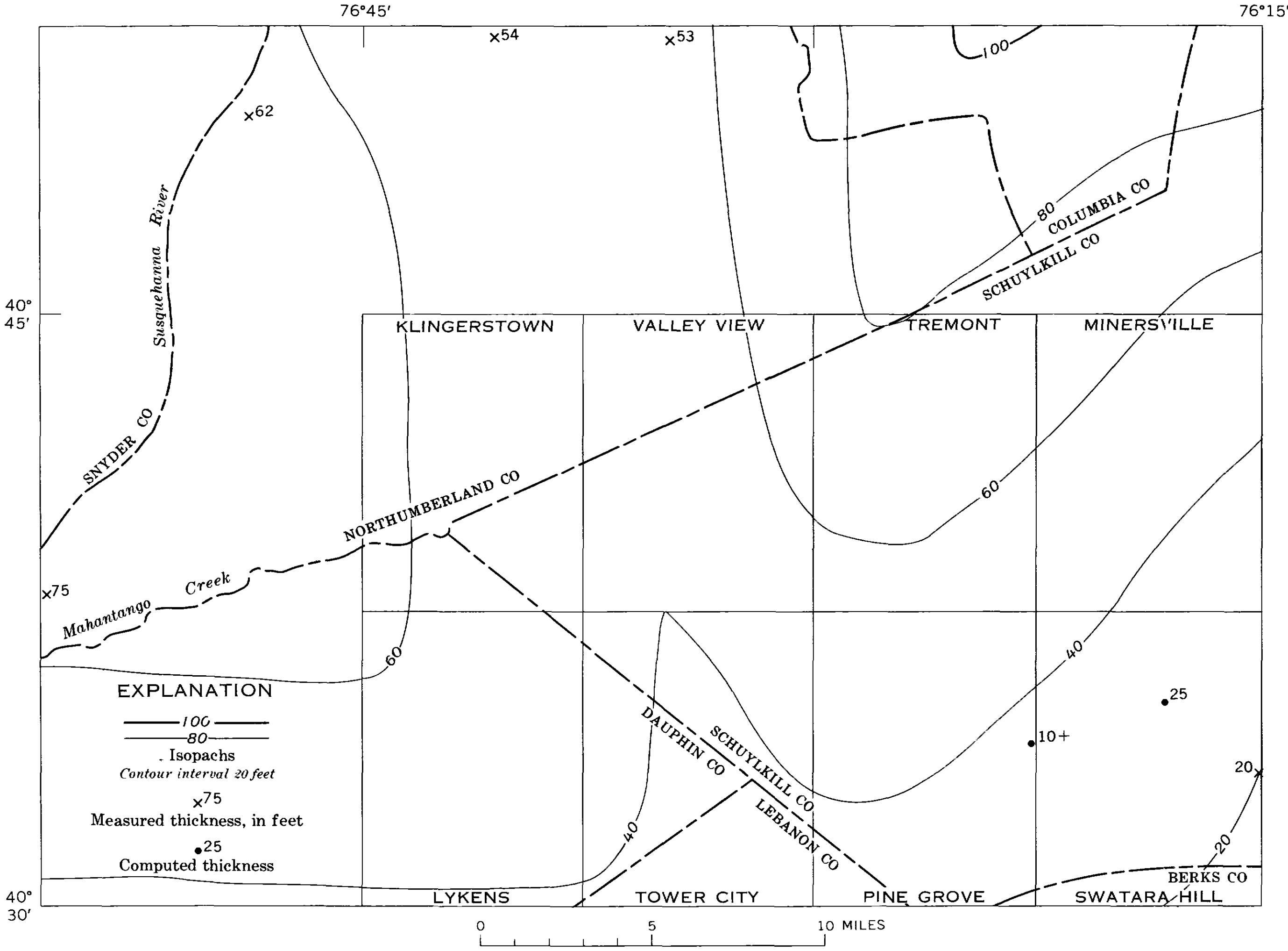


FIGURE 11.—Isopach map of the Selinsgrove Limestone.

HAMILTON GROUP

The Hamilton Group of Middle Devonian age includes, in ascending order in the report area, the Marcellus Shale and the Mahantango Formation. The latter is divided into a lower shale and sandstone member, the Montebello Sandstone Member, and an upper shale member. The group is underlain by the Selinsgrove Limestone and is overlain by the Trimmers Rock Sandstone of Late Devonian age.

Vanuxem (1840, p. 380) named the Hamilton for West Hamilton, N.Y. The group, as originally defined, included only the rocks later assigned to the Ludlowville Shale. In 1842, Vanuxem (p. 150–163) redefined the Hamilton to embrace the strata between the overlying Tully Limestone and the underlying Marcellus Shale. Clarke (1885) redefined the group in New York to include the Marcellus, and Willard (1939a, p. 165–200; 1939b, p. 9) proposed that in Pennsylvania the Hamilton Group include the Marcellus Shale and an overlying stratigraphic unit which he named the Mahantango Formation. The authors here accept Willard's definition of the Hamilton Group in Pennsylvania.

MARCELLUS SHALE

The Marcellus Shale of Middle Devonian age was named by Hall (1839, p. 295–296) for Marcellus, N.Y. As originally defined, the formation included the rocks between the overlying Ludlowville Shale and the underlying Seneca Member of the Onondaga Limestone. Cooper (1930) and Goldring (1931, p. 190–192, 369) redefined the Marcellus of New York to include the rocks between the Skaneateles Shale and the underlying Onondaga Limestone and assigned it to the Hamilton Group. In the area, the Marcellus rests on the Selinsgrove Limestone, correlative with the upper part of the Onondaga, and is overlain by the lower shale and sandstone member of the Mahantango Formation. The latter member contains beds that are probably correlative with the Skaneateles Shale (Willard and Cleaves, 1933, p. 766–774).

The Marcellus crops out in three belts, where it is poorly exposed. The northernmost belt underlies the flanks of Hooflander Mountain anticline in the Klingerstown quadrangle (GQ–700). The central belt crops out on the Roedersville anticline in the central part of the Swatara Hill quadrangle (GQ–689), and the southernmost belt underlies the slopes of Swope Mountain and adjoining hills in the southern part of the Swatara Hill and Pine Grove quadrangles (GQ–689, 691). The belts are generally eroded into a subdued rolling topography in valley bottoms or on adjacent hillsides.

The stratigraphy of the Marcellus Shale is not well known because of cover and structural complications; however, the upper part of the formation is well exposed a short distance downstream from the dam at Sweet Arrow Lake (GQ–689). Wherever exposed, the Marcellus is a grayish-black to black laminated to nonlaminated nonfossiliferous to slightly fossiliferous silty to nonsilty sooty fissile shale. The shale commonly weathers to light-gray and medium-gray paper-thin flakes. Septarian and calcareous concretions are abundant in some outcrops but are absent at most.

The contact of the Marcellus Shale with the underlying Selinsgrove Limestone is not exposed in the area, but in the surrounding region the change from limestone to shale is gradational over a few feet. It is arbitrarily placed at the horizon above which fissile nonlimy black sooty shale predominates over limestone and dark-gray limy shale.

The boundary between the Marcellus and the overlying lower shale and sandstone member of the Mahantango is generally covered, but where exposed, is located where fissile black sooty shale predominates over very fine grained sandstone or dark-gray, brown-gray, and olive-gray shale and siltstone. Locally, the lower shale and sandstone member of the Mahantango is absent in the area south of the Sweet Arrow fault zone, and typical Marcellus shale is overlain by the Montebello Sandstone Member.

The Marcellus Shale is 50 to 700 feet thick in the area and the surrounding region (fig. 12). The regional trends in thickness are shown by isopachs in figure 13. The formation is less than 100 feet thick over most of the southern part of the area and gradually thickens northward and north-northeastward to about 700 feet in the northeastern part at an average rate of 35 to 50 feet per mile. The Devonian rocks of the area were telescoped about 30 percent during the Acadian and Appalachian orogenies. The Marcellus Shale thickened northward in the northern part of the area before this deformation at an average rate of 25 to 35 feet per mile.

The rocks of the Marcellus Shale were deposited in a sea that overlapped upon the Auburn promontory farther south than any preceding Late Silurian and Devonian sea. The fine detritus in these rocks and the more extensive overlap of the sea probably indicate that the promontory was relatively low during Marcellus time. Fossils are rare, which suggests that the detritus from the promontory probably accumulated in rather muddy waters.

The sparse fauna of the Marcellus Shale is of Middle Devonian age (Willard, 1939a, p. 161, 172). The formation in the area is believed to be of the same age as the type Marcellus of New York because the fauna is similar and the formation is overlain and underlain by rocks that contain similar fossil faunas. It is correlated with the Delaware Limestone of Ohio and the Dundee Limestone of southern Michigan (Cooper and others, 1942, chart 4).

MAHANTANGO FORMATION

Willard (1935a, p. 202, 205–223) named the Mahantango Formation for the North Branch of Mahantango Creek in Perry and Juniata Counties. He included the rocks between the top of the Marcellus Shale and the base of his Portage Formation.

The Mahantango is the upper formation of the Hamilton Group in this area. It consists of a lower shale and sandstone member, the Montebello Sandstone Member, and an upper shale member. It includes the strata between the underlying Marcellus Shale and the Trimmers Rock Sandstone and all strata that locally are assigned to the upper part of the Middle Devonian Series. However, the Tully Limestone and the Burket Black Shale Member of the Harrell Shale, which overlie the Mahantango at the type locality but are absent in the report area, are of Middle Devonian age.

LOWER SHALE AND SANDSTONE MEMBER

The lower shale and sandstone member includes the strata from the top of the Marcellus Shale to the base of the Montebello Sandstone Member. These strata have not previously been accorded member status in the vicinity of the report area and have been assigned to the Marcellus by some authors and to the Montebello by others (Lesley and others, 1892–95, p. 1217; Willard, 1939a, p. 172–200; Miller, J. T., 1961, p. 27–30; Dyson, 1963, p. 17–19).

The lower member of the Mahantango Formation crops out on the Hooflander Mountain anticline (GQ–700), on the Roedersville anticline (GQ–689), and on the south slope of Swope Mountain and adjacent hills (GQ–689, 691). In all three areas this member underlies slopes capped by the Montebello Sandstone Member. It is largely covered by a heavy mantle of soil and talus blocks, and as a result, the lithology is not adequately known.

The lower contact of the lower shale and sandstone member is at the top of the black sooty fissile shale characteristic of the Marcellus. The basal bed of the member at some places is a very fine grained sandstone and at others is a dark-gray, brown-gray, or olive-gray shale or siltstone. The upper contact of the member is at the horizon where fine- to coarse-grained sandstone

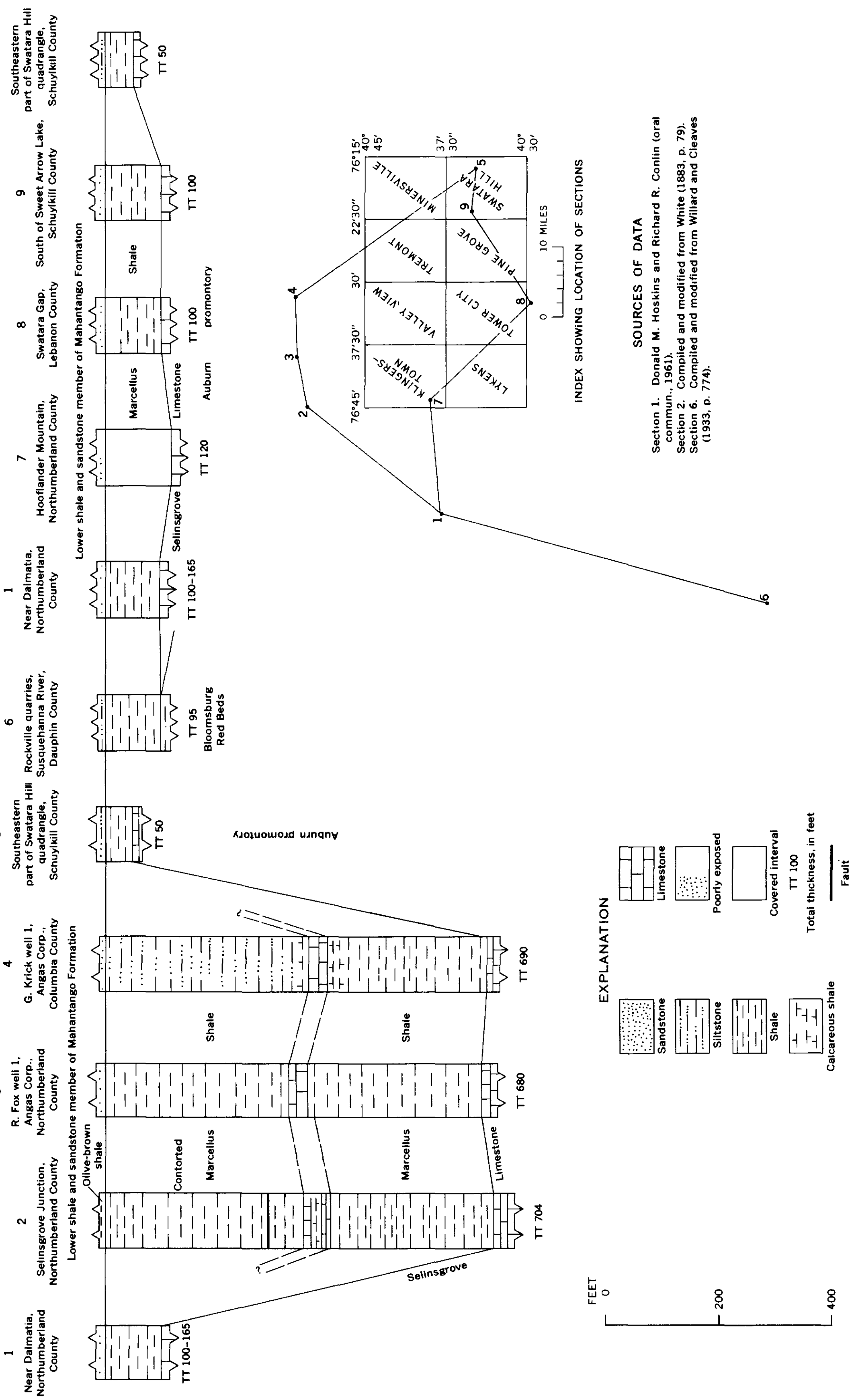


FIGURE 12.—Columnar sections of the Marcellus Shale.

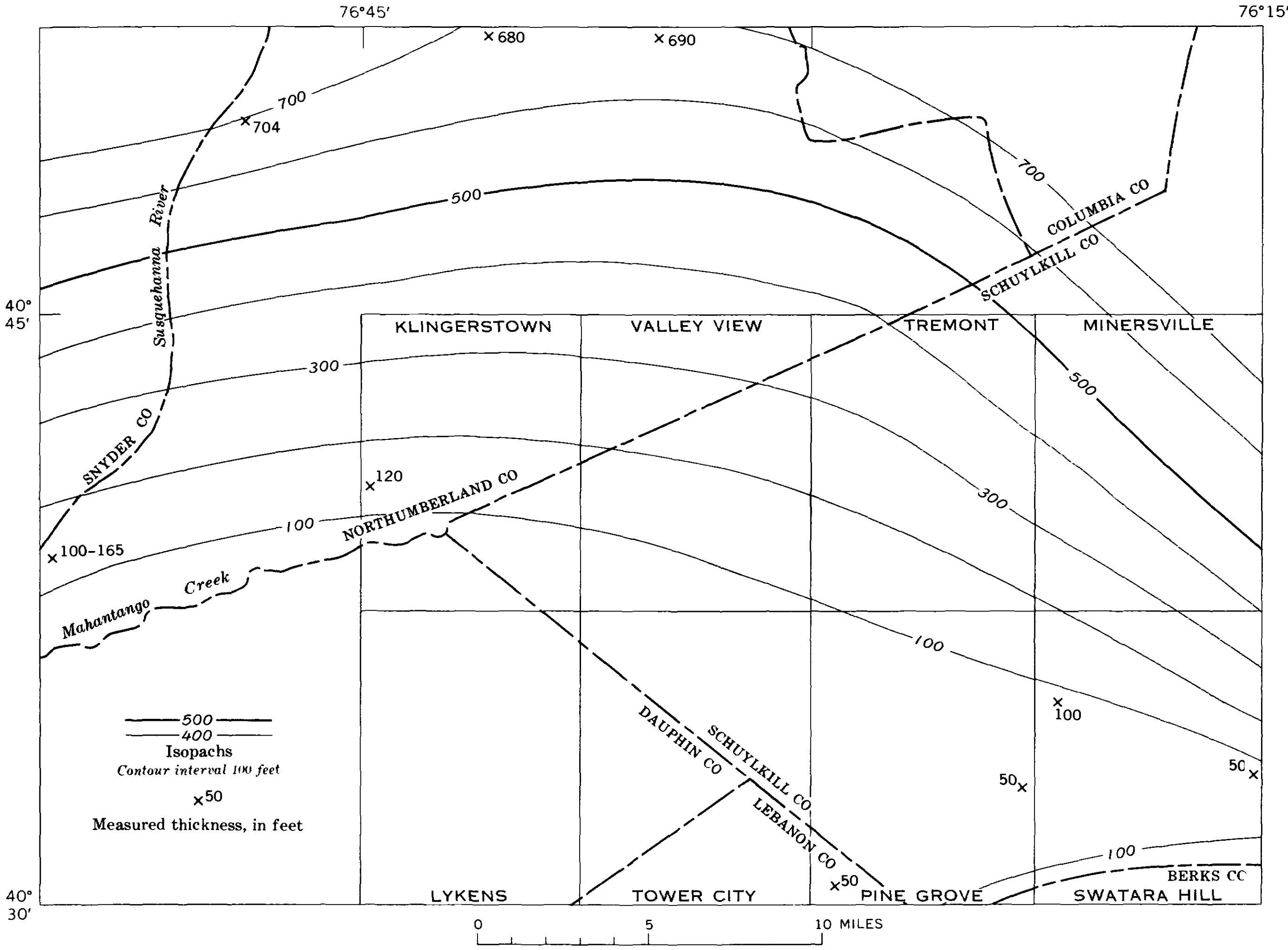


FIGURE 13.—Isopach map of the Marcellus Shale.

of the Montebello predominates over shale and siltstone.

The lower member is about 700 feet thick (fig. 14, section 1) on the Hooflander Mountain anticline (GQ–700). The upper 150± feet is poorly exposed, but seems to consist of olive-brown, dark-olive-gray, dark-greenish-gray, and medium-dark-gray siltstone and shale. A few thin beds of very fine grained medium-gray sandstone are present in the uppermost 50 feet of the member. The lower 550± feet of the member is covered, but a short distance west of the area, the rocks in this part of the member crop out and are composed of interbedded dark-greenish-gray to medium-dark-gray siltstone and shale and medium- to dark-gray very fine to fine-grained sandstone.

Part of the lower shale and sandstone member is exposed about 1.4 miles south of Pine Grove, where it consists, from the base upward of about 90 feet of intercalated laminated to thin-bedded yellowish-gray to light-olive-gray very fine grained sandstone and siltstone and about 20 feet of dark-gray shale (fig. 14, section 3).

The member is well exposed in the spillway below the dam at Sweet Arrow Lake (GQ–689). There, it consists, from the base upward of a 20± foot thin- to thick-bedded olive-gray to dark-gray very fine- to fine-grained sandstone overlain by a 95± foot unit of dark-gray to brownish-gray laminated silty shale and sandstone (fig. 14, section 4).

About 50 feet of dark-gray to brownish-gray laminated shale of the lower shale and sandstone member similar to the 95± foot unit south of Sweet Arrow Lake is poorly exposed in the southeastern part of the Swatara Hill quadrangle, a short distance south of Moyers (GQ–689) (fig. 14, section 6).

The lower shale and sandstone member in the report area ranges in thickness from 0 to about 700± feet (fig. 14). It is absent in much of central Schuylkill County (fig. 15) along a northeast-trending arch that originated before deposition of the Montebello. The member thickens northwestward and southeastward away from the arch at an average rate of about 50 feet per mile. Crustal shortening of Devonian rocks in

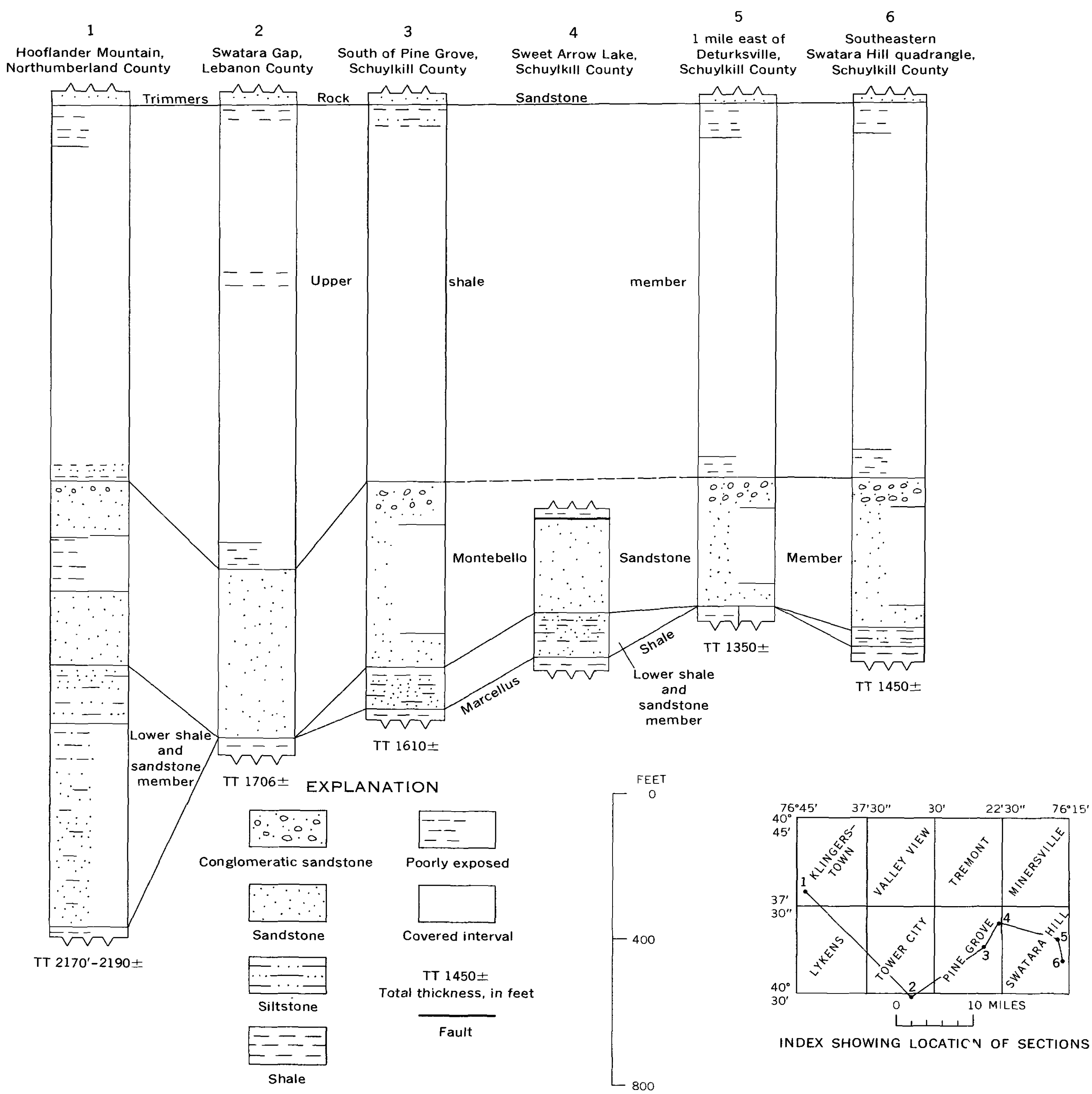


Figure 14.—Columnar sections of the Mahantango Formation.

the area averaged about 30 percent, and the average rate of thickening before shortening probably was about 35 feet per mile.

MONTEBELLO SANDSTONE MEMBER

Claypole (1885, p. 67–68) recognized a resistant sandstone in the midst of the generally soft beds of the Hamilton Group in Perry County and named it the Hamilton Montebello Sandstone for Montebello Narrows on the Little Juniata River. Subsequently, Willard (1939a, p. 180–200) defined the Montebello as a facies or sandstone member of the Mahantango Formation. The authors here adopt the Montebello Sandstone Member of the Mahantango Formation.

The existence of the arenaceous strata assigned to the Montebello was not recognized by geologists who worked in central Pennsylvania before I. C. White (1883, p. 78–82) and Claypole (1885, p. 67–68). This oversight is difficult to understand because some of the more prominent mountain ridges in that part of the

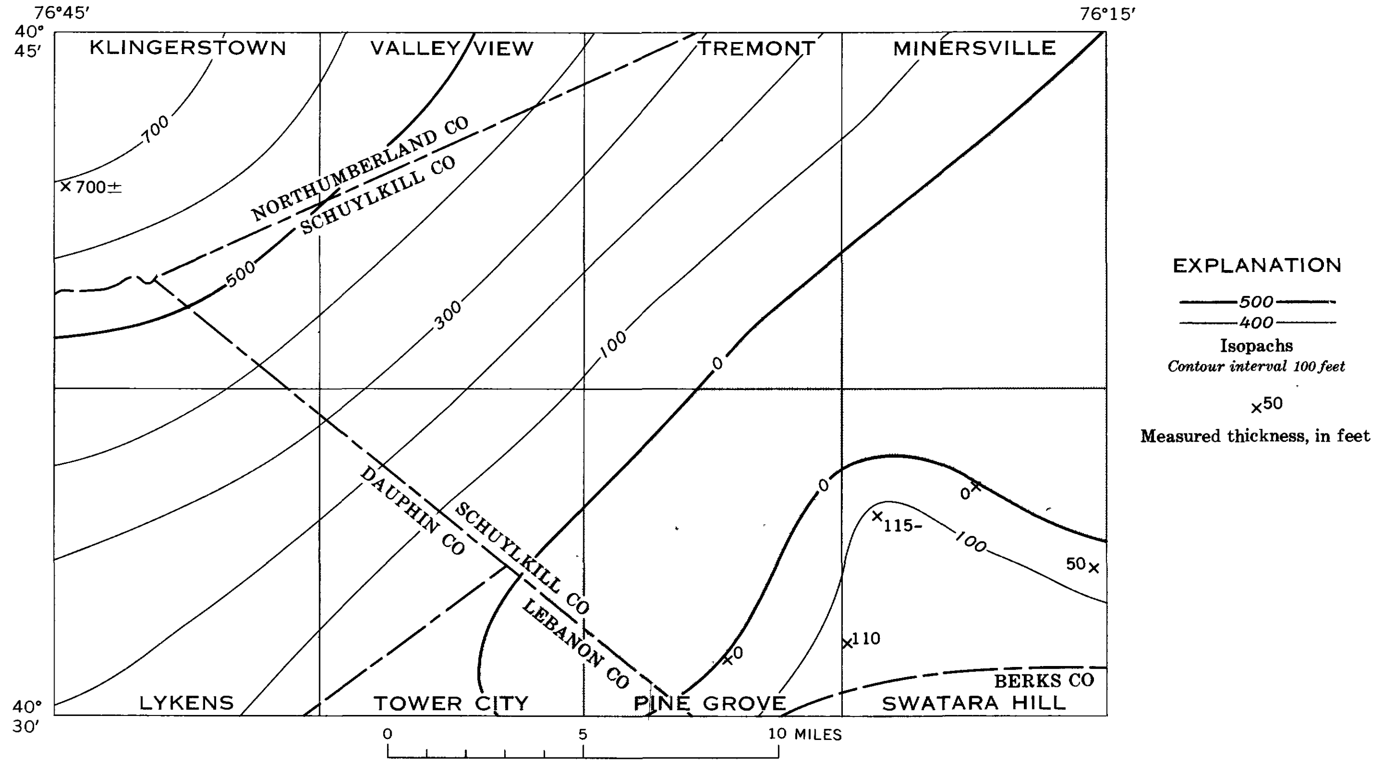


FIGURE 15.—Isopach map of the lower shale and sandstone member of the Mahantango Formation (shaded where absent).

State are underlain by the member.

The Montebello crops out on Hooflander Mountain and Fisher Ridge in the northwestern part of the area (GQ–700) and on Stone Ridge and Swope Mountain and adjacent ridges in the southeastern part (GQ–689, 691, 698).

The ridges underlain by this member generally are rough and steep, and their slopes are commonly covered by a heavy mantle of soil and talus.

The contact between the Montebello and the underlying lower shale and sandstone member is placed where fine- to coarse-grained sandstone predominates. In Hooflander Mountain and Fisher Ridge a 50± foot transition zone intervenes between distinct shale and sandstone lithologies, but is absent in the spillway of Sweet Arrow Lake and elsewhere in central Schuylkill County. There, the basal fine- to coarse-grained sandstone of the Montebello rests directly upon shale and siltstone of the lower member. The local absence of the lower shale and sandstone member of the Mahantango Formation in central Schuylkill County and the general absence of the transition zone (fig. 15) in this same general area suggest that this part of the county was subjected to pre-Montebello erosion that locally cut as deep as the upper part of the Marcellus Shale.

The Montebello Sandstone Member is 470 to 490 feet thick on Hooflander Mountain and Fisher Ridge (GQ–700). It consists, in ascending order, of 200± feet of fine- to medium-grained sandstone, 140± feet of covered rock, presumably shale, and 150± feet of sandstone that locally contains vein quartz and quartzite pebbles as much as 1 inch in diameter in the upper 20 to 40 feet (fig. 14, section 1).

About 1 mile south of Pine Grove (fig. 14, section 3; GQ–689, 691), the Montebello consists of about 500 feet of fine- to coarse-grained sandstone. The upper 50 to 60 feet of the unit contain pebbles of vein quartz and quartzite as much as 1 inch in diameter and abundant, but poorly preserved molds of brachiopods.

About 250 feet of the lower part of the Montebello crops out (fig. 14, section 4) near the dam at Sweet Arrow Lake (GQ–689), where it consists largely of fine- to coarse-grained sandstone and a few thin beds of siltstone. The upper part of the member is cut off by the Sweet Arrow fault.

About 1 mile east of Deturksville (fig. 14, section 5; GQ–689) the Montebello is about 350 feet thick, and near Moyers (fig. 14, section 6; GQ–689) it is about 400 feet thick. At these localities the lower part of the member consists of fine- to coarse-grained sandstone, the medial part is covered, and the upper part consists of conglomeratic sandstone containing pebbles as much as 1 inch in diameter and brachiopod molds.

The Montebello is well exposed a short distance south of the area at Swatara Gap (fig. 14, section 2), where it is 476 feet thick and consists of fine- to medium-grained sandstone with a few interbeds of siltstone.

The member is 290± to 710± feet thick in the area (fig. 16) and is thickest in an area extending from southern Schuylkill County northwestward into central western Schuylkill County (fig. 16). The thick area bifurcates in central western Schuylkill County, one subarea extending northwestward into Northumberland County and the other extending northeastward across northern Schuylkill County.

Willard (1939a, fig. 46), in his description of the Montebello, concluded that a landmass south of the area shed a large volume of coarse clastics during deposition of the member. He named this landmass "Cape Cumberland." It seems likely that the Cape Cumberland of Willard was a rejuvenated part of the Auburn promontory.

The discordancy of regional thickness trends between the Montebello and the lower shale and sandstone member (figs. 15, 16) suggests that the northeast-trending arch of late or post-lower shale and sandstone time supplied little sediment during Montebello time.

Fresh sandstone and siltstone in the Montebello are principally light olive gray and greenish gray, but light- to medium-dark-gray, olive-gray, and yellowish-gray hues are not uncommon. Most weathered rocks are medium- to medium-dark-gray; however, many are brownish gray, yellowish gray, and olive gray; a few are dark greenish gray, olive black, and grayish brown.

The beds in the Montebello are chiefly tabular and parallel laminated; large-scale cross-stratification is uncommon. The thickness of individual units ranges from 1/4 inch to 20 feet and averages about 1 foot.

Most of the sandstone in the Montebello is medium to coarse grained. Individual constituents range from very fine grains to 2-inch pebbles. The larger clasts consist almost entirely of vein quartz and quartzite. The grains range from angular to well rounded and average subangular. Many quartz grains interpenetrate. Others, mostly embedded in a clay and sericite matrix, do not interpenetrate and are rounded to well rounded. This difference in rounding and interpenetration suggests that the matrix protected the embedded grains from pressure solution.

The Montebello is composed mostly of subgraywacke and protoquartzite. The principal mineral constituents of both rock types are grains and fragments of vein and common quartz, sericite, quartzite, and unidentified clay minerals. A few grains of halloysite, leucoxene, muscovite, ilmenite, magnetite, tourmaline, and unidentifiable plagioclase and fragments of schist and siltstone are present in most thin sections. Much of the plagioclase seems to be altered to sericite and clay and may be the principal source of the unidentified clay minerals and halloysite.

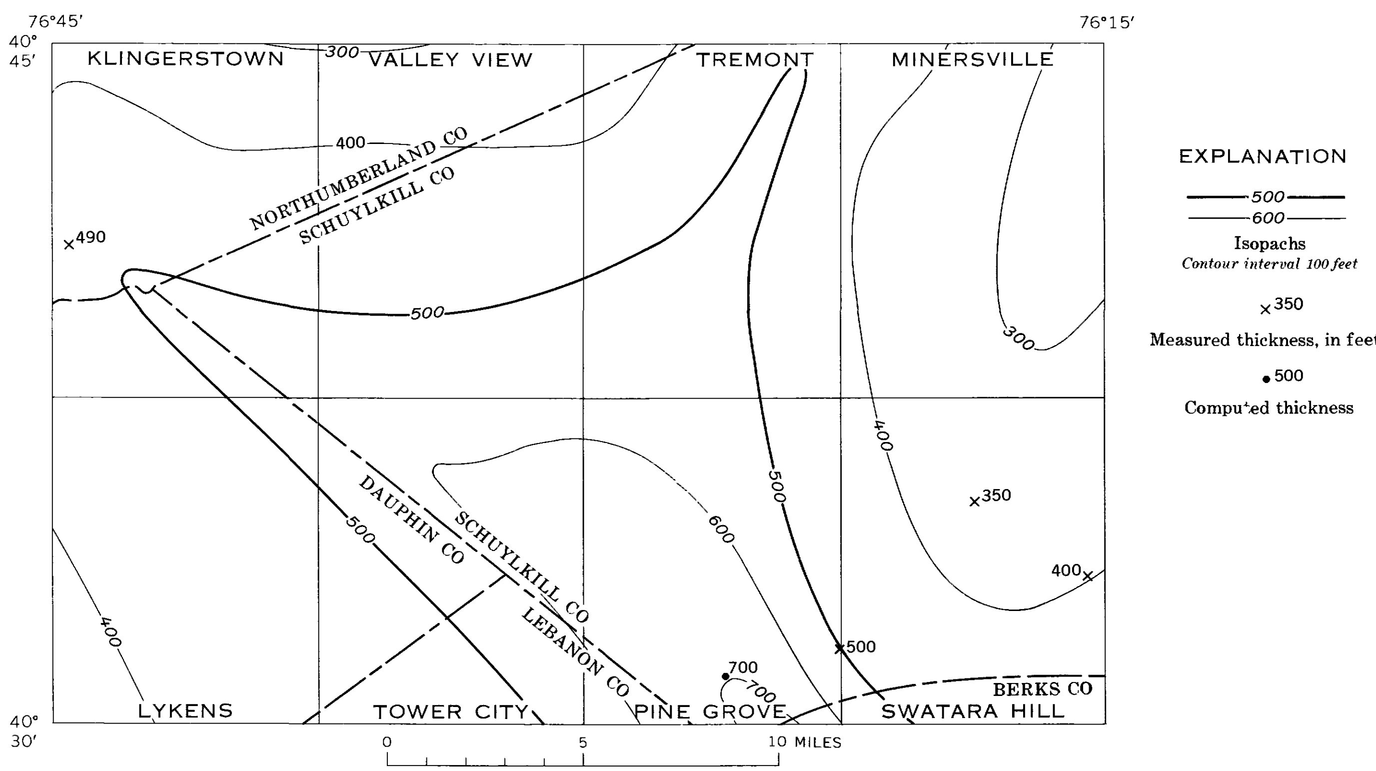


Figure 16.—Isopach map of the Montebello Sandstone Member of the Mahantango Formation.

UPPER SHALE MEMBER

The upper shale member of the Mahantango Formation includes the rocks from the top of the Montebello Sandstone Member to the base of the Trimmers Rock Sandstone. These are the strata that Willard (1935a, p. 202, 205–223) had in mind when he named the Mahantango. From a regional viewpoint, he considered the underlying Montebello Sandstone Member to be a local rock sequence of importance only in Perry, Juniata, and Dauphin Counties.

The member crops out in four belts in the area. The northernmost of these belts fringes Fisher Ridge and Hooflander Mountain (GQ–700). The member crops out in two belts on the north and south limbs of the Pine Grove syncline and in a fourth belt between the North Sweet Arrow and Sweet Arrow faults (GQ–689, 691). The member underlies valleys between ridges that are held up by the Montebello Sandstone Member and the Trimmers Rock Sandstone. These valleys in the southern part of the area are generally wide, flat bottomed, and alluviated. In contrast, the valleys adjacent to Hooflander Mountain and Fisher Ridge are steep sided and partly filled with talus from the Montebello and Trimmers Rock.

The contact between the upper shale member and the underlying Montebello generally is gradational and most commonly placed where shale predominates over sandstone. In the vicinity of Fisher Ridge and Hooflander Mountain, however, it is at the base of a 50-foot transition zone that consists of shale and thin sandstone beds. In the area south of the Sweet Arrow fault zone, an abrupt change from sandstone to shale marks the contact.

The contact of the upper member and the overlying Trimmers Rock Sandstone is conformable, and regional relations indicate that these units intertongue. It is well exposed at many places on the limbs of the Pine Grove syncline where it is at the change from shale, silty shale, or siltstone to sandstone.

Figure 14 shows five incomplete measured stratigraphic sections of the member. These sections and data from the maps (GQ–689, 691, 698, 699, 700) indicate that the member ranges in thickness from about 900 feet in eastern Schuylkill County to 1,250± feet in Dauphin, Lebanon, and central Schuylkill Counties (fig. 17) and that the member thins east, north, and west of Dauphin, Lebanon, and central Schuylkill Counties.

The rocks of the upper shale member are medium to dark gray, brownish gray, olive brown, and olive gray and weather to buff hues. Most beds are 1 to 2 inches thick, but some are as thin as one-eighth inch, and a few are as thick as 2 feet.

The member is composed principally of nonsilty shale, lesser amounts of silty shale, and small amounts

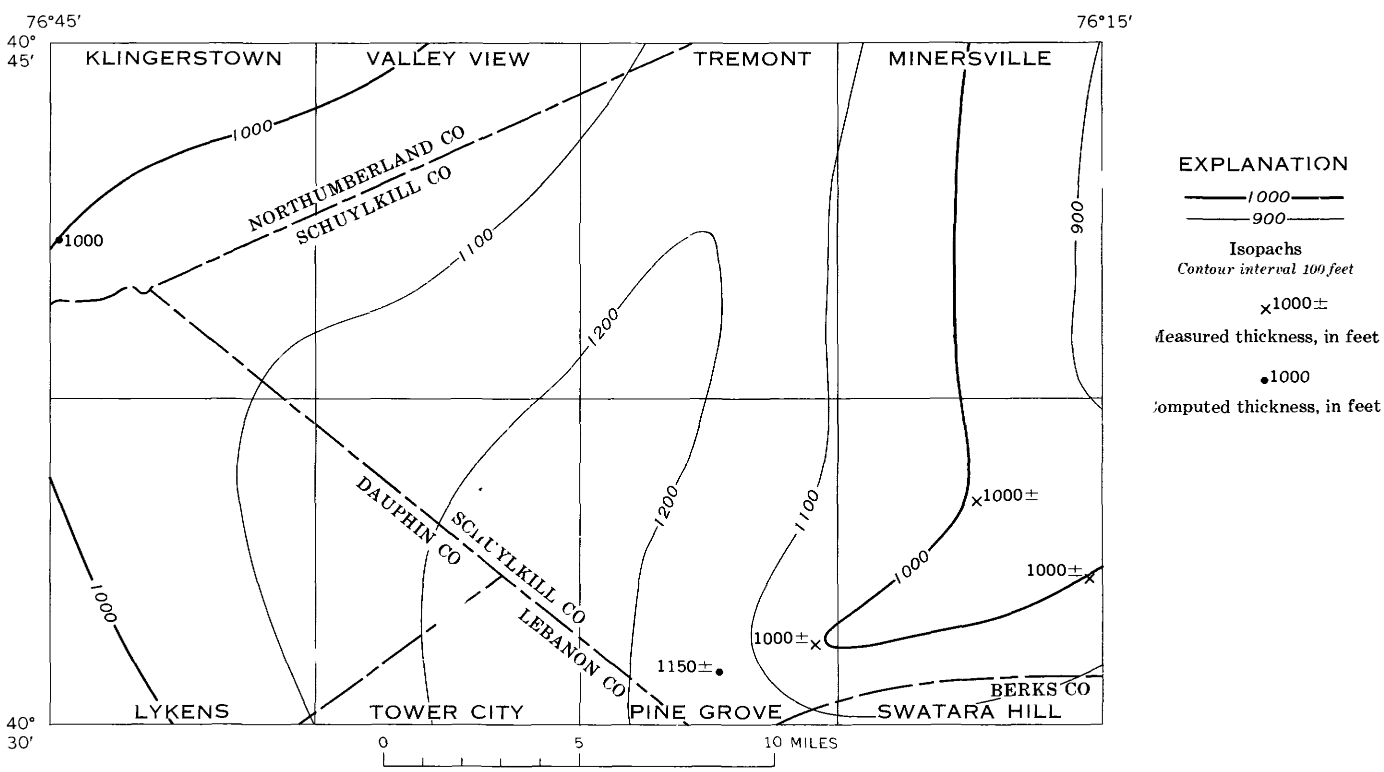


FIGURE 17.—Isopach map of the upper shale member of the Mahantango Formation.

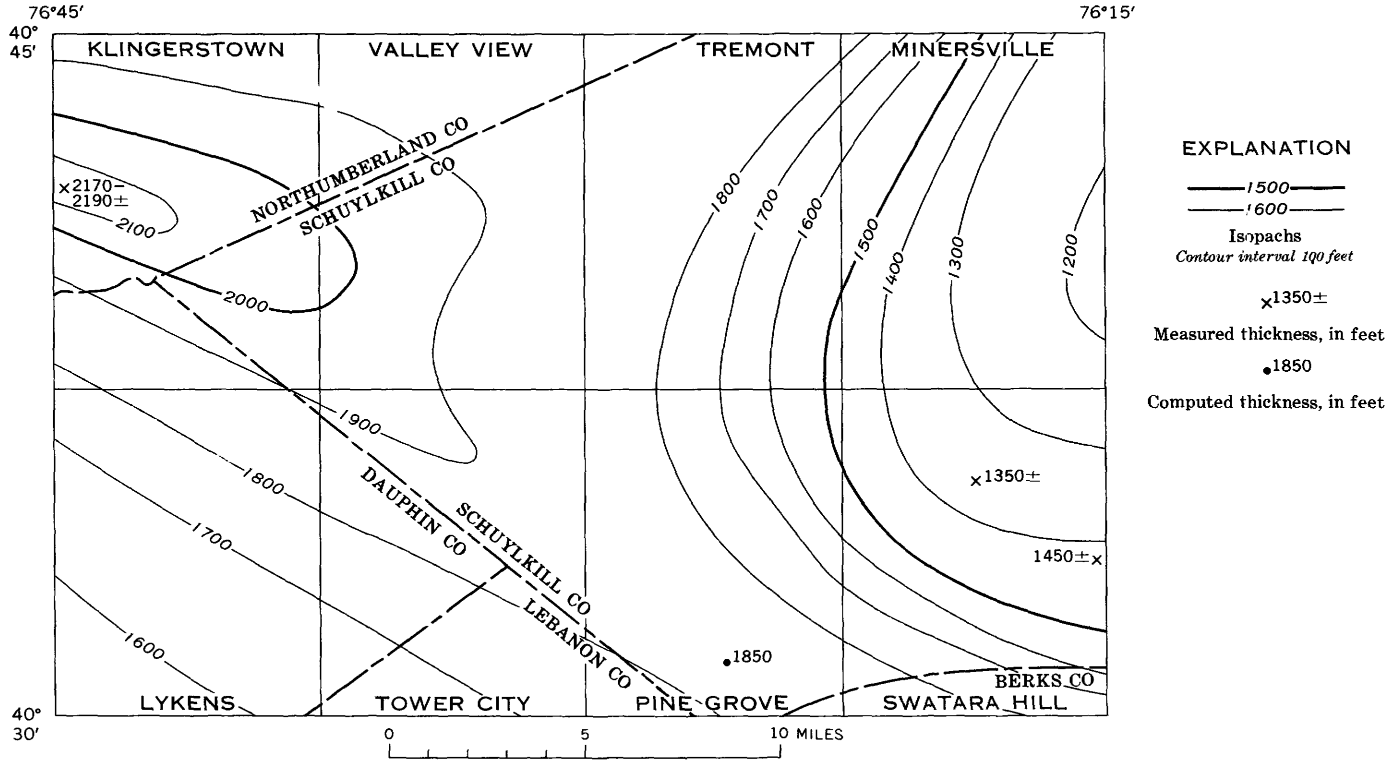


FIGURE 18.—Isopach map of the Mahantango Formation.

of intercalated very fine grained sandstone. The chief mineral constituents of these rocks are grains of vein and common quartz, unidentified clay minerals, and sericitic material.

THICKNESS OF THE MAHANTANGO FORMATION

The Mahantango Formation is 1,200 to 2,190 feet thick in the area (fig. 18) as shown by sections measured at the localities in figure 14 and from other measurements in the surrounding region. Relatively thick Mahantango flanked by thinner areas extends southeastward from the vicinity of Hooflander Mountain and Fisher Ridge into Berks County.

DEPOSITIONAL ENVIRONMENT, AGE, AND CORRELATION OF THE MAHANTANGO FORMATION

Rocks of the lower shale and sandstone member were deposited in the same transgressive sea as those of the Marcellus Shale. Relations between these stratigraphic units indicate that sedimentation continued without interruption during deposition. The average grain size in the lower shale and sandstone member, however, is slightly greater than that in the Marcellus. This characteristic suggests that the source lay somewhat closer to the report area, was more heavily eroded, or was slightly higher during lower member deposition than during Marcellus deposition.

The absence of the lower member of the Mahantango Formation in central Schuylkill County indicates that this part of the area was uplifted late in the deposition of the member and that erosion removed any previously deposited sediments of the member and a part of the Marcellus Shale. This uplift marked the beginning of a rejuvenation of the Auburn promontory, as recorded by the coarse clastics of the Montebello Sandstone Member.

The transitional sequence of shale and sandstone at Hooflander Mountain and Fisher Ridge indicates that the Montebello accumulated upon the lower member without sedimentary interruption. In central Schuylkill County, however, the Montebello was laid down on a erosion surface that had been cut through the lower member into the Marcellus Shale. Deposition of Montebello sediments at some localities took place first above, then below, sea level. For example, the lower beds of the member at the Rockville quarries, about 15 miles west of the area on the Susquehanna River, probably were deposited above sea level, as shown by remains of land plants (Willard, 1939a, p. 198); but the upper beds there and elsewhere were deposited below sea level, as indicated by brachiopod molds.

The upper shale member seems to have been deposited upon the Montebello Sandstone Member without an interruption in sedimentation. It accumulated in the sea as the influx of coarse clastics that had been eroded from the Auburn promontory during Montebello time lessened and finally ceased.

Willard (1939a, p. 176–194) described a large late Middle Devonian fauna from the Mahantango Formation. This fauna indicates that the formation is correlative with the Skaneateles, Ludlowville, and Moscow Shales of central and western New York, that is possibly is correlative with a part of the Skunnemunk Conglomerate of New Jersey and southeastern New York, and that it is correlative with the upper part of the Romney Shale of Maryland, Virginia, and West Virginia (Cooper, G. A., and others, 1942, chart 4).

SUSQUEHANNA GROUP

A thick sequence of marine and continental rocks, the Susquehanna Group, principally of Late Devonian age, rests with apparent conformity upon the Mahantango Formation in the area.

Inclusion of the Upper Devonian rocks of Pennsylvania in the Susquehanna Group was proposed by Ashley in 1923 (p. 1106–1108). His proposal did not gain widespread usage for many years, but in 1959 several geologists working in the Commonwealth accepted the group as a valid stratigraphic unit (Miller, J. T., and Conlin, R. R., 1959; Arndt and others, 1959, p. 3).

The Susquehanna Group, as recognized by the U.S. Geological Survey, includes the strata from the top of the Tully Limestone of Middle Devonian age to the base of the Pocono Formation of Early Mississippian. Where the Tully is absent, as in the report area, the base of the group is defined as the top of the Mahantango Formation. In the report area the group consists, in ascending order, of the Trimmers Rock Sandstone and the Catskill Formation.

The Trimmers Rock Sandstone and the Catskill Formation represent a sedimentary cycle that consists, in ascending order, of a marine rock sequence, a transitional marine and continental rock sequence, and a continental rock sequence. The gradual change from marine to continental sedimentation took place through several thousand feet of rock. The marine sequence is included largely in the Trimmers Rock. The transitional sequence is embraced largely by the Irish Valley Member of the Catskill, but it also includes some of the lower part of the Damascus Member (GQ–689, 691) and an equivalent part of the Buddys Run Member (GQ–699, 700). The continental sequence includes the middle and upper parts of the latter members and the overlying members of the Catskill.

TRIMMERS ROCK SANDSTONE

The Trimmers Rock Sandstone, basal unit of the Susquehanna Group in the area, was named by Willard (1935c) for Trimmers Rock, a prominent hill near Newport in Perry County. As originally defined, it included those beds underlying the Parkhead Sandstone member of the Jennings Formation and overlying Willard's Losh Run Shale. Subsequently, Willard (1935d) redefined the unit by reducing it to a member of his Fort Littleton Formation, which, in ascending order, consisted of the Harrell, Losh Run, Trimmers Rock-Brallier, and Parkhead Members. The Trimmers Rock is the dominant or only division of the Fort Littleton Formation of Willard that is present in the area and in the region to the east. Therefore, Klemic, Warman, and Taylor (1963, p. 15) reclassified the Trimmers Rock Sandstone as a formation in the Lehighton 7½-quadrangle, Carbon County. The authors concur with their decision and consider the Trimmers Rock of the area to be a formation that includes the strata between the Mahantango and Catskill Formations.

The Trimmers Rock crops out in three belts in the area. Each of these belts is hilly and has a topographic relief that averages about 200 feet, but which at places may be slightly more than 300 feet. Localities underlain by the formation commonly are mantled with a blocky sandstone rubble.

The largest belt of outcrop of the Trimmers Rock is about 15 miles long and underlies the limbs and trough of the Pine Grove syncline in the Pine Grove, Swatara Hill, and Tower City quadrangles (GQ–689, 691, 698). Outcrops are poor, except near the base of the formation where there are many good exposures in borrow pits.

North of the Sweet Arrow fault zone a second belt of outcrop extends from the eastern boundary of the Swatara Hill quadrangle to a point about halfway between Beuchler and Outwood in the Pine Grove quadrangle, where it is truncated at the fault zone (GQ–689, 691). This belt is about 0.4 to 0.5 of a mile wide. Outcrops are sparse and poor, except locally in steep-sided valleys.

The third belt of Trimmers Rock underlies both limbs and the crest of the Broad Mountain anticlinorium in the Klingerstown and Valley View quadrangles (GQ–699, 700). Exposures generally are poor, but along some valley walls several hundred feet of the formation crop out locally.

The Trimmers Rock Sandstone conformably overlies the upper shale member of the Mahantango Formation in the area. However, about 4 miles west of the area, near Mandata, the Burket Black Shale Member of the Harrell Shale intervenes, and about 8 miles west of the area, near Dalmatia, the upper part of the Harrell and the Brallier Shale also intervene (D. M. Hoskins and R. R. Conlin, oral commun., 1961). Detailed stratigraphic work between these exposures and the report area indicates that the Harrell and Brallier probably tongue out eastward into the lower part of the Trimmers Rock. Eastward intertonguing of the Harrel and Brallier into the Trimmers Rock was first suggested by Willard (1935b, p. 39), who noted that the basal beds of the Trimmers Rock east of the area, in Monroe County, contain *Hypothyridina venustula*, the guide fossil of the Middle Devonian Tully Limestone. On the basis of this fossil, he postulated that successively older Late Devonian Brallier Shale, Late Devonian Harrell Shale, and Middle Devonian Tully Limestone tongued eastward into the Trimmers Rock. Thus, in the area, the contact between the Trimmers Rock and the upper shale member probably is older than the Burket Black Shale Member of the Harrell and may be as old as the Tully Limestone. The contact is at the sharp change from shale, silty shale, and siltstone of the upper shale member to very fine grained to fine-grained sandstone and siltstone of the Trimmers Rock.

The contact of the Trimmers Rock with the overlying Irish Valley Member of the Catskill Formation normally is placed at the base of the lowest red bed. Within the area, these units intertongue at a scale so large that individual tongues can be shown on 1:24,000 maps (GQ–689, 691, 699, 700). In the core of the Pine Grove syncline (GQ–689) a discontinuous tongue of Irish Valley red beds locally splits the Trimmers Rock in half. At this one place, therefore, a sequence of olive and gray strata of the Trimmers Rock is shown to lie above the lowest red bed. Regionally, rocks at the base of the Irish Valley tongue to the east and southeast with successively older rocks at the top of the Trimmers Rock. The scale of intertonguing is so great that much, if not all, of the Irish Valley in the southeastern part of the area (GQ–689, 691) may be the time equivalent of the Trimmers Rock in the northwestern part (GQ–699, 700).

A few feet to several hundred feet of the Trimmers Rock Sandstone crop out at numerous places in the area. Only one of these outcrops, north of Pine Grove on the west side of Swatara Creek, is sufficiently well exposed so that a section can be measured and the lithology studied (fig. 19, section 1). There, the basal part of the formation is truncated by the Sweet Arrow fault zone, and only 2,011 feet is measurable. Regional mapping suggests that the original thickness of the formation at this locality was about 2,100 feet.

Another outcrop of the Trimmers Rock, 2,259 feet thick, which appears to be unfaulted, lies a few miles east of the area. The exposure is about 2.4 to 2.85 miles south of Pottsville on the east side of Schuylkill River in roadcuts of the Pennsylvania Railroad and a county road (fig. 19, section 2). The formation is not so completely exposed in this section as it is north of Pine Grove, but the lithology is similar.

The thickness of the Trimmers Rock Sandstone was calculated at many places in the area. These computed thicknesses, the thicknesses measured at the sections shown in figure 19, and many other thicknesses determined in the surrounding region were used to construct an isopach map of the formation (fig. 20). The Sweet Arrow fault zone separates the Trimmers Rock into two areas of contrasting thickness. North of the zone the formation thickens northeastward from about 1,800 feet in the southwestern part of the area to more than 2,400 feet in the northeast. South of the zone it thickens northwestward from about 1,300 feet to more than 1,500 feet. The contrasting thicknesses on opposite sides of the fault zone probably did not develop during deposition. Therefore, it is likely that faulting brought into juxtaposition parts of a depositional basin which previously were far removed one from another. This probability is supported by a similar difference in the thickness of the overlying Irish Valley Member of the Catskill Formation across the zone.

North of the Sweet Arrow fault zone the Trimmers Rock thickens northeastward at a rate of 27 to 33 feet per mile. Before the 30 percent structural foreshortening of the Devonian rocks of the area, it probably thickened northeastward at an average rate of 19 to 23 feet per mile. South of the fault zone, the Trimmers Rock thickens northwestward at about 80 feet per mile. Structural shortening in this part of the area seems to have been negligible, and the original rate of depositional thickening has not changed.

Freshly exposed rocks in the formation are chiefly light olive gray, light brownish gray, olive brown, and medium gray, but grays, browns, and greens are not uncommon. Weathered rocks are largely light olive gray, medium to dark gray, and brownish gray; a few are greenish-gray.

Most beds in the formation are tabular, but some are cross-stratified and wedge shaped. Beds range in thickness from 1/8 inch to 10 feet and average about 1 foot. Sole markings, current laminations, crossbeds, and other current-oriented sedimentary features indicate that the source of the clastics of the Trimmers Rock lay southeast and east of the area.

The rocks of the Trimmers Rock consist of slightly to moderately fossiliferous well-cemented sandstone and micaceous siltstone and some micaceous shale. The repetition of sandstone and siltstone beds of approximately equal thickness is the outstanding characteristic of the formation. The sandstones are protoquartzite and subgraywacke with a small amount of orthoquartzite. They are predominantly very fine grained; however, fine- to coarse-grained sandstone is not uncommon, and pebble conglomerate is present about two-thirds of a mile southwest of Twin Grove Park (GQ–698) in a roadcut of the Reading Co. The sandstones are composed of grains of common and vein quartz and subordinate amounts of quartzite, other rock fragments, halloysite, unidentified clay minerals, biotite, ilmenite, muscovite, sericite, and small percentages of andesine, orthoclase, tourmaline, and zircon. Although the average grain is subangular, the shape ranges from angular to well rounded. Cementing media are silica and clay with minor amounts of limonite and calcite.

The siltstone of the Trimmers Rock contains less quartz and more sericite and unidentified clay minerals than does the sandstone. The shale in the formation is similar to the siltstone, except that the percentage of sericite and clay minerals is greater.

The Trimmers Rock Sandstone seems to have accumulated without interruption upon the upper shale member of the Mahantango Formation. The basal beds of the sandstone in the area probably are laterally equivalent through intertonguing with the Tully Limestone, Harrell Shale, and Brallier Shale.

The clastic sediments of the formation were derived from a landmass that lay east and southeast of the area, possibly in New Jersey or eastern Pennsylvania. They

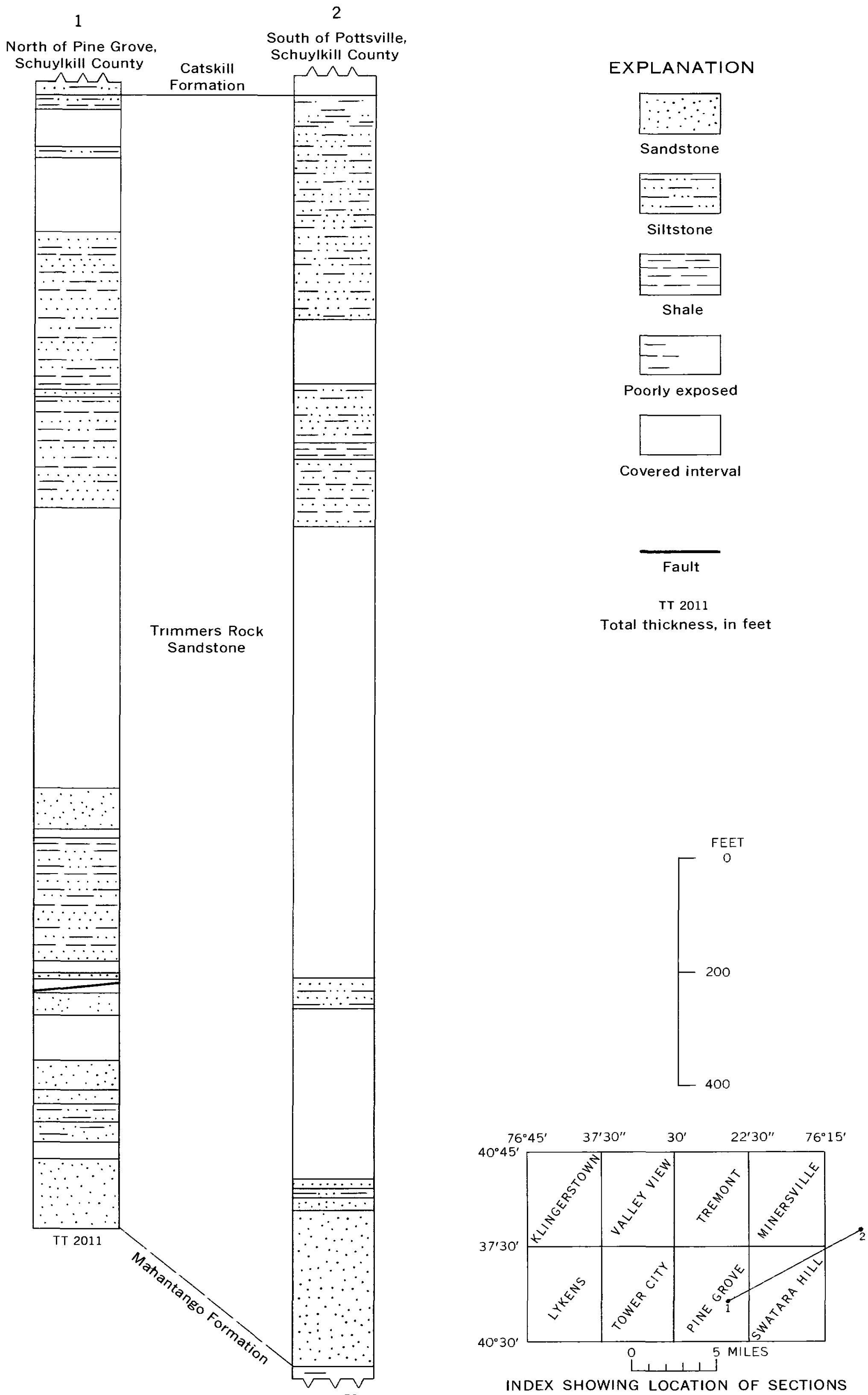


FIGURE 19.—Columnar sections of the Trimmers Rock Sandstone.

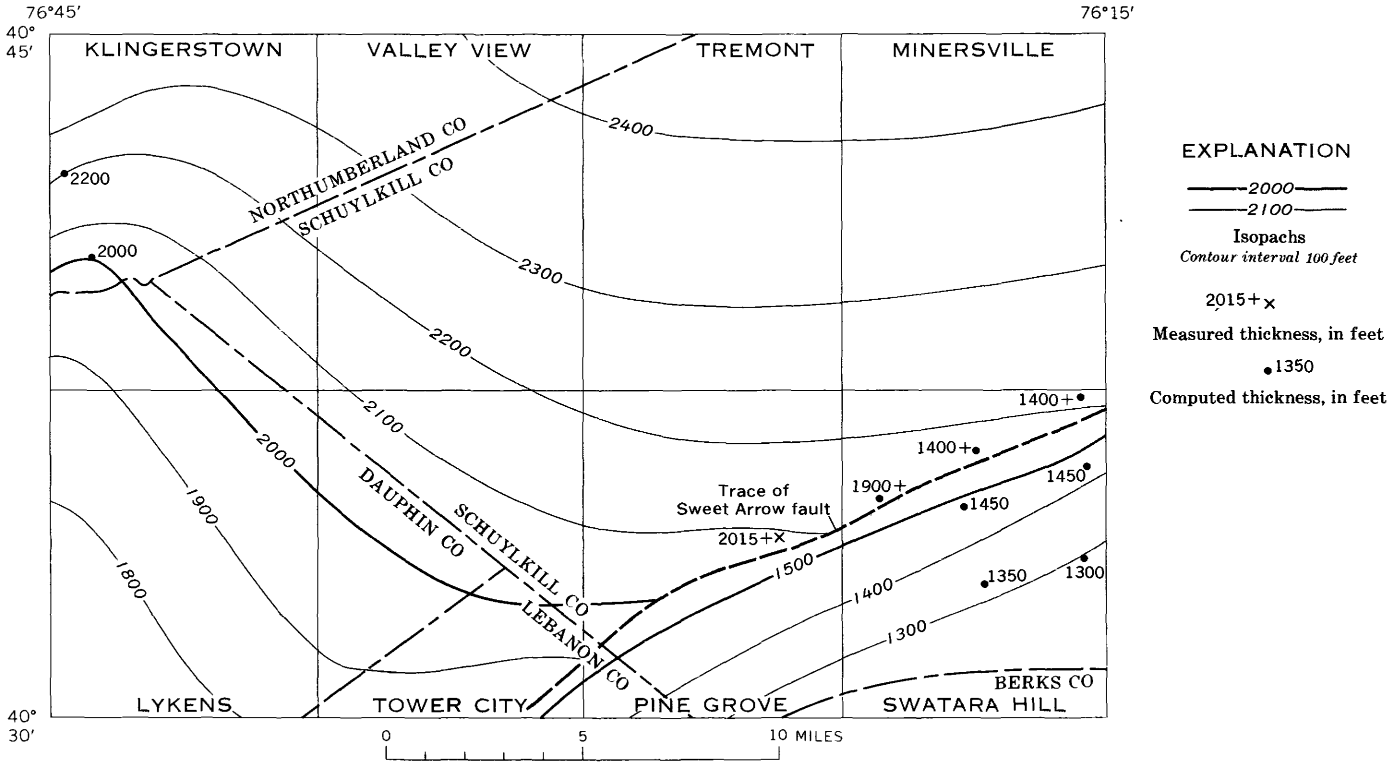


FIGURE 20.—Isopach map of the Trimmers Rock Sandstone.

seem to have been deposited in a westward-regressing sea that was being filled with debris from the landmass. The repetitious interbedding of siltstone and sandstone was probably caused by slight changes in the competency of the streams carrying sediment to the sea and by slight changes in the competency of marine and turbidity currents distributing the sediment.

Frakes (1961) postulated that the Trimmers Rock was deposited exclusively by turbidity currents. He cites as evidence of turbidity-current deposition: graded bedding, convoluted bedding, load casts, current laminations, oriented plant fragments, sole markings and alternating lithologies. Similar sedimentary features in other parts of the United States commonly are interpreted to indicate deposition by turbidity currents, but in many places they also can be interpreted to indicate variations in the fluvial supply of sediment and the marine currents depositing the sediments. The authors believe that turbidity currents were active during deposition of the formation, but they do not believe that the widespread alternation of sandstone and siltstone beds over thousands of square miles was accomplished solely by turbidity currents.

An extensive fauna has been reported from the Trimmers Rock (Willard 1939a, p. 212–215). The abundance of this fauna indicates that the sediments of the formation accumulated in relatively shallow waters where food, oxygen, and light were plentiful.

The uppermost gray beds of the Trimmers Rock Sandstone at any locality in the area are laterally equivalent through intertonguing east of that locality with the basal red beds of the Irish Valley Member of the Catskill Formation. The reason for the change in color at the Trimmers Rock and Irish Valley contact is not definitely known, but the following is postulated Sedimentary features and intertonguing relations indicate that the sediment in these stratigraphic units was derived from a source east and southeast of the area (McIver, 1960; Frakes, 1961, fig. 1). Much of the detritus from the source arrived at the strand line as a red sediment. Whenever deposition was rapid, the sediment was quickly buried and was not exposed for long periods to a marine reducing environment; it thus retained its red color. When deposition was slow, however, reduction of ferric iron to ferrous iron took place, and the red color of the sediment was changed to gray and olive. The basal red beds of the Irish Valley probably represent the oldest Upper Devonian rocks to accumulate so rapidly that they were buried before being subjected to reduction and a consequent change in color.

Although the contact between the Trimmers Rock and Irish Valley is placed at the lowest red bed in the Upper Devonian sequence, marine sediments similar to those of the Trimmers Rock were deposited as tongues in the lower parts of the Damascus and Buddys Run Members of the Catskill Formation. Most red beds between these tongues, and some red beds in the Irish Valley Member, contain macerated plant fossils, fragments of coalified vegetation, raindrop impressions, mud cracks, and animal tracks that are considered to indicate terrestrial deposition (Barrell, 1913–14, p. 455, 466). The interbedded marine and continental rocks in the Damascus, Buddys Run, and Irish Valley Members indicate that the lower part of the Catskill Formation accumulated in a generally regressive sea in which minor transgressive phases were not uncommon. The minor transgressions probably were due to slight readjustments in the altitude of the sea bottom and to variations in the amount of sediment being deposited in the sea.

At any one locality it is uncertain where continental accumulation began, but the intertonguing indicates that such accumulation began at different times in different localities. It is possible that at some places the oldest continental sediments coincide with, or slightly underlie, the basal red bed of the Irish Valley, but it is more reasonable to assume that terrestrial sedimentation began later in Irish Valley time. Despite this uncertainty, most geologists who have worked the Upper Devonian rocks of eastern Pennsylvania consider the basal red bed of the Irish Valley to mark the upper limit of wholly marine deposition which began in Silurian time.

The Trimmers Rock Sandstone contains an early Late Devonian fauna (Willard, 1939a, p. 211–215; Cooper, G. A., and others, 1942, chart 4). The Harrell and Brallier Shales of central Pennsylvania, the Woodmont Shale of western Maryland, and the Sonyea and Genesee Formations (de Witt and Colton, 1959) of central New York contain a similar fauna. The regional intertonguing of the Trimmers Rock with underlying rocks and the local occurrence of the fossil *Hypothyridina venustula* east of the area indicate that the basal beds of the formation are equivalent to the Middle Devonian Tully Limestone (Willard, 1935b, p. 39). The intertonguing between the rocks of the formation and overlying rocks is suggestive that the upper beds of the formation are equivalent with part, and perhaps with all, of the Irish Valley Member of the Catskill.

DEVONIAN AND MISSISSIPPIAN ROCKS

CATSKILL FORMATION

The Catskill Formation of Late Devonian and Early Mississippian age is the upper division of the Susquehanna Group. It includes the rocks between the top of the Trimmers Rock Sandstone and the base of the Lower Mississippian Pocono Formation. These rocks are divided into members that are classified differently in the southern and northern parts of the area. In the southern part (GQ–689, 691, 698, 701) the formation is divided, in ascending order, into the Irish Valley, Damascus, Honesdale, Cherry Ridge, and Spechty Kopf Members. The Honesdale Sandstone Member, which tongues or wedges out northward in the subsurface of the area, separates the generally finer grained red beds of the Damascus from similar red beds of the Cherry Ridge. Where it is absent in the northern part of the area (GQ–692, 699, 700), the indistinguishable Damascus and Cherry Ridge Members are mapped together as the Buddys Run Member.

Mather (1840, p. 212–213, 227–233) named the Catskill for the mountains in Delaware, Greene, Schoharie, Sullivan, and Ulster Counties, New York. He included in it the rocks between the Onondaga Limestone and the Pottsville Formation. Vanuxem (1842, p. 186–194) redefined the Catskill to include only the rocks younger than the Chemung Formation. During the 20th century, many geologists who worked in New York discussed the Catskill, the strata it should and should not include, and its age relations at various places.

Meanwhile, I. C. White (1883, p. 49–75) solved the problem of the base of the Catskill in Pennsylvania to the satisfaction of most geologists who have studied the formation there by placing the lower contact at the lowest red bed. He assigned the marine strata below the lowest red bed to the Chemung (Trimmers Rock Sandstone of this report) and the transitional red- and olive-hued alternating marine and continental beds above it to his transitional Catskill-Chemung Group (Irish Valley Member of this report).

Lesley (in Lesley and others, 1892–95, p. 1601–1607), in his summation of the geology of Pennsylvania, included White's Catskill-Chemung Group in the Catskill, which he reduced in rank from a group to a formation. He followed White in placing the base of the Catskill at the base of the lowest red bed.

The problem of the upper contact of the Catskill was solved by Lesley (1876) and Franklin Platt and W. G. Platt (1877, p. xxvi) when they named the Pocono Sandstone (Pocono Formation of this report) and stated that the gray rocks of this unit overlay red rocks of the Catskill. The precise position of this contact at many localities has been the subject of much discussion; however, Lesley's and the Platts' definition of the Catskill and Pocono contact generally has been accepted. I. C. White (1883, p. 49–52) states that his Pocono-Catskill Transition Group (probably the Spechty Kopf Member of this report) was a transitional gray and red rock sequence between the gray Pocono and the red Catskill. Lesley, in 1892 (Lesley and others, 1892–95, p. 1601–1607), included this unit in his Catskill Formation, thus placing the top of the Catskill at the top of the uppermost red bed, the contact recognized in this report.

Williams (1887, p. 27) and Hartnagel (1912, p. 86) recognized that from east to west the basal beds of the Catskill overlay successively younger beds in the stratigraphic sequence. However, it remained for Chadwick (1933a, b; 1935a) and Willard (1933, 1935b, 1936, 1939a) to comprehend and describe the intertonguing of the Catskill with strata ranging in age from Middle Devonian in New Jersey and eastern New York to latest Devonian in western Pennsylvania and southern New York.

IRISH VALLEY MEMBER

The Irish Valley Member was named by Arndt, Wood, and Trexler (1962, p. C35) for a valley near Shamokin, Northumberland County, Pa. The member includes the strata between the Trimmers Rock Sandstone and either the Damascus or Buddys Run Member of the Catskill Formation.

Geologists who have worked in eastern Pennsylvania classified the strata assigned to the Irish Valley Member as follows: I. C. White (1883, p. 63–67) assigned them to his Catskill-Chemung Group. A few years later Lesley (in Lesley and others, 1892–95, p. 1594–1607) placed them at the base of the Catskill. Barrell in 1913 (1913–14, p. 455–458) assigned the upper beds of the unit to his Catskill-Chemung transition beds and the lower beds to his Chemung Group. Willard (1939a, p. 257–307) apparently did not recognize the strata included in the Irish Valley Member. According to his figure 72 and the text of his report (p. 291–295), continental red beds of his Damascus Red Shale (Damascus Member of the Catskill in this report) rest upon gray beds of the marine Trimmers Rock Sandstone in the vicinity of the area. It, therefore, appears that he overlooked the interbedded red and olive-gray marine and continental rocks of the Irish Valley Member.

The Irish Valley Member crops out on the flanks and crest of the Broad Mountain anticlinorium in the Klingerstown and Valley View quadrangles (GQ–699, 700). A second, less extensive, belt of outcrop lies north of the Sweet Arrow fault zone in the Pine Grove and Swatara Hill quadrangles (GQ–689, 691). A third belt is on the limbs and trough of the Pine Grove syncline in the eastern part of the Swatara Hill quadrangle (GQ–689).

Exposures in the various belts of outcrop of the Irish Valley are usually limited to natural cuts along steep valley walls and to manmade cuts along roads and railroads. Despite the scarcity of outcrops, the belts are easily mapped at most places by tracing alternating bands of red and olive-gray varicolored float. Delineation of these bands during fieldwork made it possible to determine the structural features and stratigraphic changes that are shown on quadrangles GQ–689, 691, 692, 699, 700.

The member generally underlies a subdued rolling upland that has shallow steep-sided valleys. The amount of relief locally is as much as 300 feet.

The lower and upper beds of the Irish Valley are conformable with underlying and overlying strata. The member intertongues with the underlying Trimmers Rock Sandstone and the overlying Buddys Run or Damascus Members at a scale large enough to be shown on 1 : 24,000 maps. The stratigraphic relations between the Irish Valley and the Trimmers Rock are described in the chapter about the latter formation. The only part of that description reemphasized here is that the contact between these units is at the base of the lowest red bed in the Upper Devonian rock sequence. The contact of the Irish Valley with overlying rocks is at the horizon above which the red beds dominate and below which red and olive-gray beds are intercalated the percentage of red beds decreasing stratigraphically downward.

Both the upper and lower contacts of the Irish Valley Member gradually rise westward in the Devonian sequence by intertonguing. It is possible that much, if not all, of the member in the southeastern part of the area is the time equivalent of the Trimmers Rock Sandstone in the northwestern part. Similarly, the lower beds of the Damascus Member in the southeastern part may be time equivalents of the Irish Valley of the northwestern part.

The Irish Valley has been measured and described at a locality a short distance north of Pine Grove on the west side of the Reading Railroad tracks and Swatara Creek (fig. 21, section 1). It has also been measured east of the area at localities north of Cressona (fig. 21, section 2) and south of Pottsville (fig. 21, section 3). An isopach map of the member, constructed from these measurements and from many other thicknesses that were calculated across the outcrop belts (fig. 22), indicates that the Irish Valley thickens northwestward from a minimum of 185 feet near Pine Grove to a maximum of 2,250 feet in western Schuylkill and southern Northumberland Counties and then thins in the same direction to about 1,800 feet north of Hooflander Mountain. The member is 900± feet thick south of the Sweet Arrow fault zone and only 185 to 300 feet thick a short distance north of the zone. This great difference in thickness across the zone suggests that faulting has brought close together rock sequences that originally lay far apart. A similar difference exists across the zone in the Trimmers Rock Sandstone and also suggests considerable foreshortening. The Irish Valley thickens northward at an average rate of about 180 feet per mile north of the

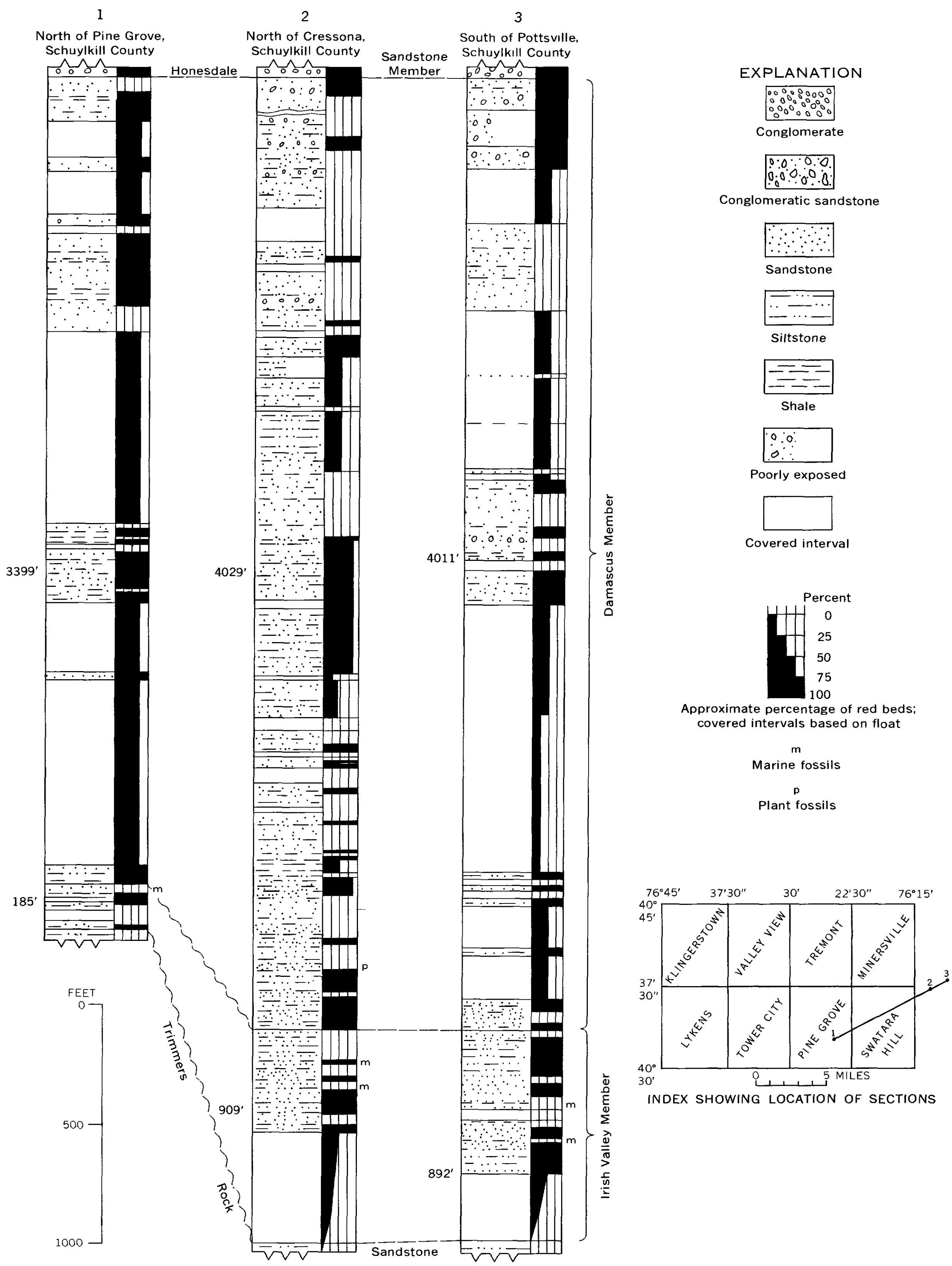


FIGURE 21.—Columnar sections of the Irish Valley and Damascus Members of the Catskill Formation.

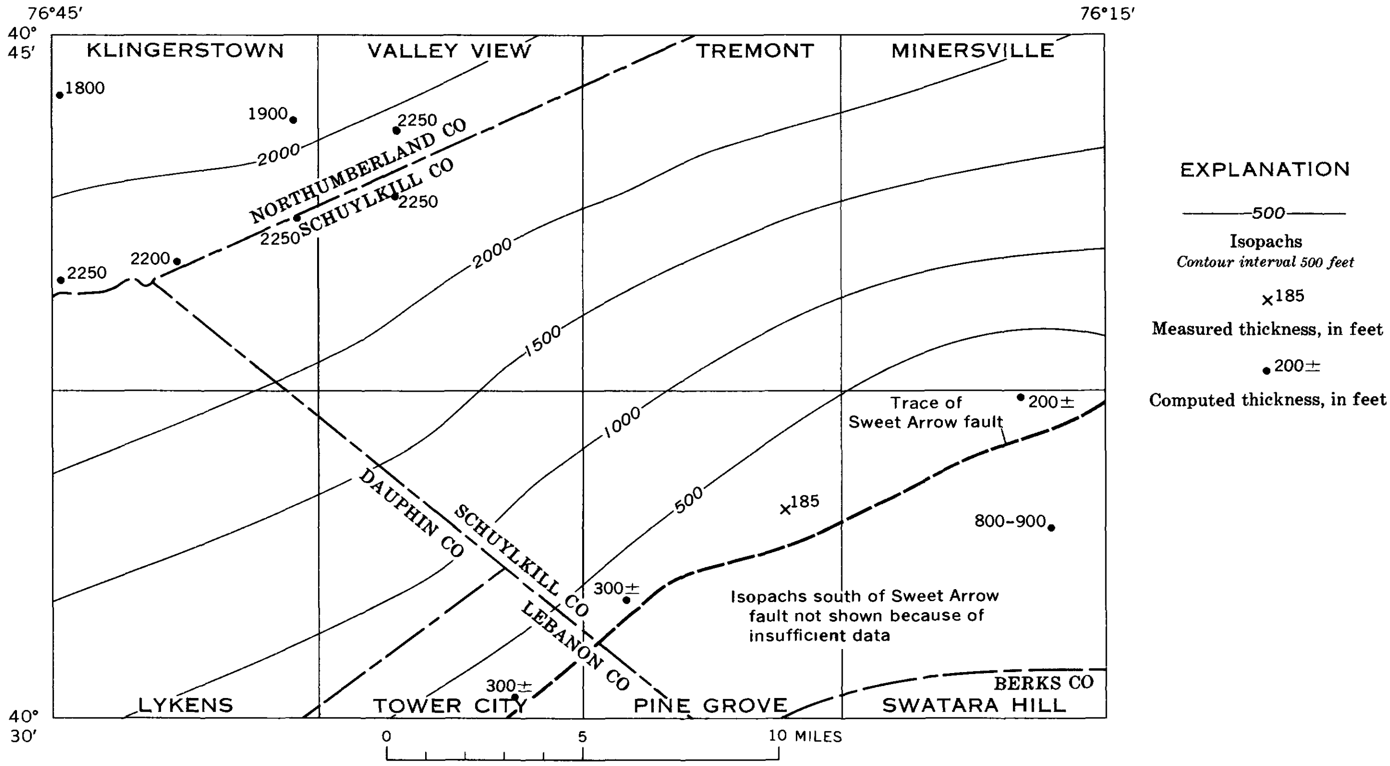


FIGURE 22.—Isopach map of the Irish Valley Member of the Catskill Formation.

fault zone in Schuylkill and Dauphin Counties. Before the folding and faulting, which shortened the Devonian rocks of the area about 30 percent, the member thickened northwestward at a rate of about 125 feet per mile.

Fresh and weathered sandstones in the Irish Valley Member are mainly light olive gray and pale brown, but light greenish gray, greenish gray, light brownish gray, brownish gray, pale red, and dusky yellows are common. The siltstone is predominantly grayish red, pale brown, and light olive gray, but some is light to medium gray, light greenish gray, and yellowish gray. The shale is pale to olive gray, greenish gray, yellowish gray, yellowish brown, and pale red.

Beds of the Irish Valley are 2 inches to 50 feet thick and average 3 feet. Most are tabular; but a few are irregular and wedge shaped, and a very few are cross stratified with small-scale planar crossbeds.

The coarser grained rocks of the Irish Valley are chiefly protoquartzite and subgraywacke. They are composed in large part of grains of common and vein quartz and fragments of quartzite with lesser amounts of biotite, muscovite, sericite, tourmaline, and zircon grains; rock fragments of schist and slate; unidentified clay minerals; and a very small amount of andesine and orthoclase grains. Grain size ranges from very fine to fine and averages very fine. Cementing materials are silica, limonite, hematite, and clay. The sorting is poor to excellent but is usually good. The shale and siltstone contain the same mineral assemblages as the coarser grained rocks but the amount of mica minerals and unidentified clay minerals increases as the grain size decreases.

The depositional environment of the Irish Valley was discussed in the description of the Trimmers Rock Sandstone and will not be discussed at great length here.

The rocks of the Irish Valley are chiefly red and olive gray. They are believed to have accumulated in part, in a westward regressing sea; in part, at the margins of the sea; and perhaps in part, in deltas and on flood plains. Most of the sediment that accumulated at or below sea level probably was red when it arrived at the strand and retained its color because deposition and burial were rapid; however, when deposition and burial were slow, reduction changed the red to olive gray.

Marine fossils in the olive-gray beds and in some red beds of the member suggest that the sea bottom supported prolific life. The interbedding of marine and continental rocks further suggests that this marine life flourished in shallow waters near the shore. These waters were generally regressing to the west because of a gradual filling of the depositional basin with sediment from a rising landmass east of the area. Minor transgressions took place from time to time and probably were caused by temporary decreases in the supply of detritus brought into a generally sinking basin. As time went on, however, the supply of detritus increased intermittently, so that by the end of Irish Valley sedimentation, the sea was largely forced west of the area.

Most geologists have believed that wholly marine Devonian deposition ended in the vicinity of the area with the end of Trimmers Rock deposition. Similarly, even though marine tongues are present in the Buddys Run and Damascus Members, the top of the Irish Valley Member is believed generally to mark the final regression of the Devonian sea.

The fauna of the Irish Valley Member has not been studied, but because the member tongues with the overlying Buddys Run and Damascus Members and with the underlying Trimmers Rock Sandstone, it can be assumed that the age is also early Late and middle Late Devonian (Cooper, G. A., and others, 1942, chart 4).

Regional correlations of the Irish Valley into other parts of Pennsylvania are uncertain because of the intertonguing relations of the member and because the stratigraphic relations of the lower part of the Catskill Formation elsewhere in the Commonwealth are imperfectly understood. The Irish Valley, therefore, is correlated only tentatively with part and perhaps all of the Chemung Formation of central Pennsylvania, with the Dellville Sandstone and Kings Mill Sandstone and Shale of Claypole (1885) of Perry County, and with the Shohola Formation of Willard (1936) of northeastern Pennsylvania and southeastern New York. The local discontinuous tongue of the Irish Valley near the middle of the Trimmers Rock Sandstone in the Pine Grove syncline (GQ–689) may be laterally correlative with the Walcksville Sandstone Member of the Catskill in Carbon County (Klemic and others, 1963, p. 26–27), which, in turn, may be correlative with the Analomink Red Shale of Willard (1935d) in Monroe County.

DAMASCUS MEMBER

The lowest persistent red division of the Catskill Formation on the Pocono Plateau of northeastern Pennsylvania was named the Damascus Red Shale by Willard (1936, p. 571, 584–585) for exposures near Damascus. After an extensive field reconnaissance on the Pocono Plateau (Harry Klemic, oral commun., 1959), Klemic, Warman, and Taylor (1963, p. 28–29) accepted Willard's correlation of the Damascus into the Lehighton quadrangle at the east end of the Southern Anthracite field. The authors also tentatively accept the Damascus as a member of the Catskill Formation in the southern part of the area. The term "red shale" is abandoned as a lithologic modifier because the member at many localities consists in large part of gray, olive, green, and brown sandstone, siltstone, and shale. As used in this report, the Damascus Member includes all strata lying between the Irish Valley and Honesdale Members.

The Damascus Member crops out in two belts in the area. The larger of these lies south of Second Mountain (GQ–689, 690, 691, 698); the smaller is in the upper reaches of Powell Creek valley (GQ–701). The lower part of the Damascus in both belts underlies a gentle rolling subdued topography. Stream valleys are not dissected deeply, and this part of the unit is commonly farmed intensively. The upper part of the member in the belt south of Second Mountain underlies steep heavily wooded slopes, which are dissected by transverse valleys only at wind and water gaps. The upper part of the unit in Powell Creek valley underlies gentle to steep moderately dissected heavily wooded slopes.

The lower contact of the Damascus with the Irish Valley Member is difficult to pick because it is in a zone of intertonguing. Individual tongues of Damascus or Irish Valley strata at many places are large enough so that they can be shown on 1:24,000-scale maps (GQ–689, 690, 691, 698). In general, the contact is arbitrarily placed at the horizon above which the main red bed sequence of the Damascus predominates over the red and olive-gray beds of the Irish Valley. The stratigraphic position of this contact rises gradually from east to west by intertonguing.

The upper boundary of the Damascus on the southern slope of Second Mountain (GQ–689, 690, 691, 698) is covered at most localities, but is placed where red and gray fine-grained rocks of the member are overlain by red and gray rock of the basal conglomeratic sequence of the Honesdale Sandstone Member. In the southern part of the Lykens quadrangle (GQ–701) this contact is not exposed and is arbitrarily placed at the base of a sequence of medium-grained to very coarse grained sandstone of the Honesdale and at the top of grayish-red fine-grained to very fine grained sandstone and shale of the Damascus.

The Damascus Member is best exposed within the area north of Pine Grove (fig. 21, section 1). Several miles east of the area and north of Cressona on the west branch of the Schuylkill River it is relatively well exposed (fig. 21, section 2), and a short distance farther east and south of Pottsville on the main branch of the Schuylkill River it is less well exposed (fig. 21, section 3).

In addition to the measured sections, the thickness has been calculated at many places in the member. These calculated and measured thicknesses and the thickness at places outside the area were used to compile figure 23, an isopach map of the member. The Damascus is not shown northwest of a line where the Honesdale Sandstone Member is inferred to wedge out, or south of the Sweet Arrow fault zone where the member has been removed by erosion. Between the fault zone and the wedge

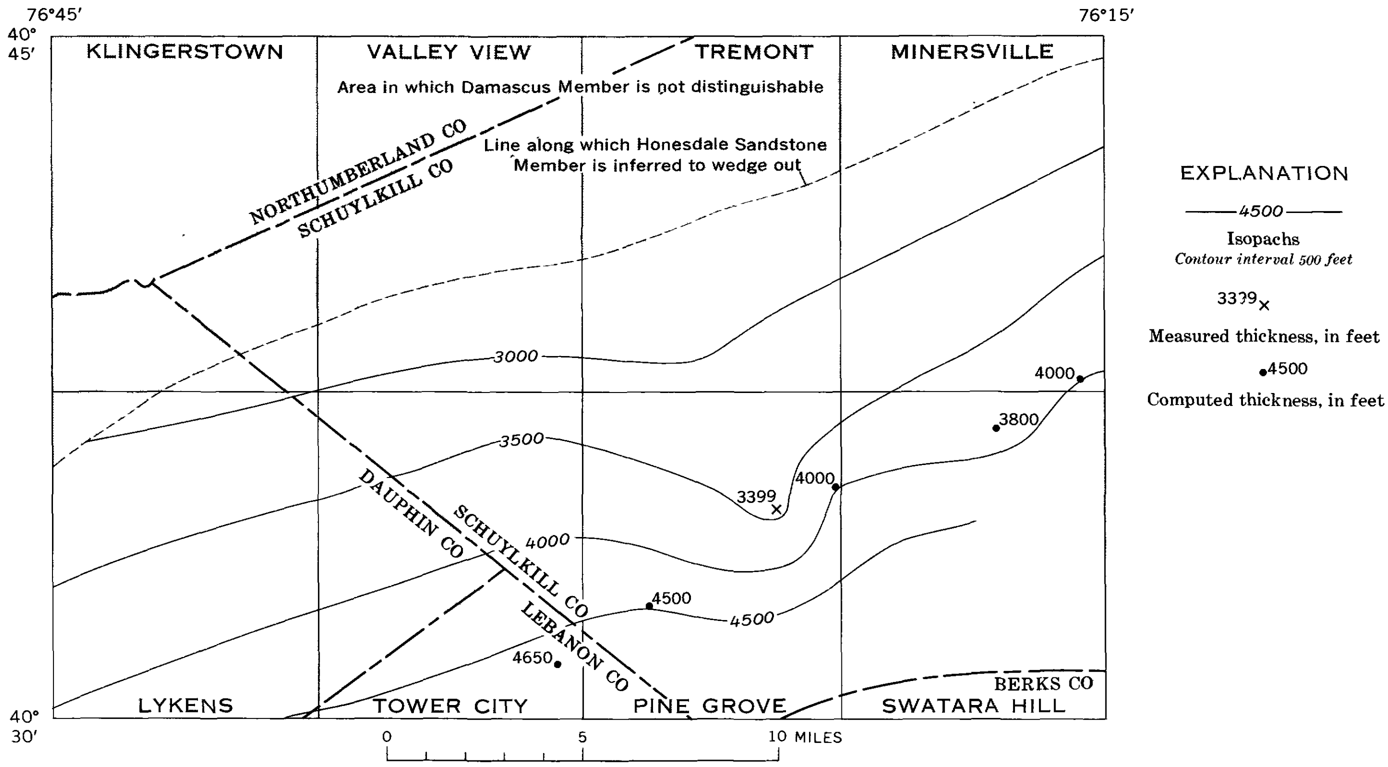


FIGURE 23.—Isopach map of the Damascus Member of the Catskill Formation.

out of the Honesdale, the Damascus thins northward from 4,650± feet near the zone to slightly more than 2,500 feet at the wedge out at an average rate of 190 to 220 feet per mile. The rocks of the member were telescoped about 30 percent during folding and faulting, and before deformation they probably thinned northward at an average rate of about 135 to 155 feet per mile.

The sandstone of the Damascus Member is largely grayish pink to grayish red, light olive to olive gray, pale brown, and dusky yellow. Some is light to dark brownish gray, pale to dark yellowish brown, greenish gray to dark greenish gray, and greenish to moderate yellow. Conglomerate is pale to dark yellowish brown, grayish red, and dusky yellow. The siltstone, in order of decreasing abundance, is grayish red, pale brown, light to olive gray, medium gray and light to brownish gray. The shale is largely light to olive gray and grayish red, a few beds being pale brown, light to brownish gray, and light to dark greenish gray, and dusky yellow. Grayish-red and pale-brown rocks compose about 50 to 75 percent of the member, and rocks of other colors constitute the remainder. The percentage of red beds in the member decreases eastward from about 75 percent in the southeastern part of the Tower City quadrangle (GQ–698) to about 50 percent at the east border of the Swatara Hill quadrangle (GQ–689), and is only about 40 percent east of the area near Cressona (fig. 21, section 2) and Pottsville (fig. 21, section 3).

The beds of the Damascus range in thickness from $\frac{1}{4}$ inch to 15 feet and average about 3 feet. Most sandstone beds are about 3 feet thick, but many are much thicker. The average siltstone and shale bed is about 1 foot thick, but many are as thin as $\frac{1}{8}$ inch or as thick as 15 feet. Slightly more than 50 percent of the member is tabular bedded, and the remainder is wedge bedded. Many beds are cross stratified with simple and planar cross-strata, which generally are less than 1 foot long but which locally may be as long as 5 feet. Cross-stratification is more common in the upper and middle parts of the member than in the lower. Graded bedding is common and reverse grading is rare.

Many shale and siltstone beds are internally contorted. In most places the contortions developed after consolidation and were caused by structural adjustments of more competent strata moving differentially against the less competent shale and siltstone. In a few localities, however, the contortions developed after deposition and before consolidation.

Plant fragments, raindrop impressions, and mud cracks are preserved in many sandstone and siltstone beds. Ripple marks, whose crests generally trend northeastward, are common in the upper parts of some siltstone and fine-grained sandstone beds.

Rock types of the Damascus Member are, in decreasing order of abundance, shale, sandstone, siltstone, and conglomerate. The sandstone is predominantly fine grained, but much is very fine to medium grained. Coarse-grained and very coarse grained sandstone is present throughout the unit, whereas conglomerate is confined to the upper part. The average diameter of the pebbles in the conglomerate is about three-eighths of an inch, and the maximum is as large as 2 inches.

Although most of the sandstone is subgraywacke, a considerable amount is protoquartzite and graywacke. The mineral assemblages of the three rock types are similar; the principal difference between protoquartzite and graywacke is an increase in the amount of feldspar, clay minerals, micaceous minerals, and rock fragments—such as quartzite, schist, phyllite, and slate in the graywacke—and a proportionate decrease in vein and common quartz. The principal mineral constituents of all types of sandstone are grains of vein and common quartz and fragments of quartzite. Accessory minerals are grains of muscovite, tourmaline, zircon, ilmenite, magnetite, leucoxene, sericite, unidentified clay minerals, and andesine and orthoclase altering to clay. In addition, there are considerable but greatly variable quantities of rock fragments of chert, schist, phyllite, and slate. Feldspar is present in small quantities throughout the member but is more common in the middle part. The matrix consists chiefly of sericitic material and unidentified clay minerals with varying amounts of hematite and limonite. Many red sandstone beds contain as much as 15 percent hematite and limonite. The sorting is commonly fair to good in the protoquartzite but is generally poor in the subgraywacke and graywacke. Most quartz grains are subangular to subrounded, but well-rounded quartz grains are relatively common. The cements in the nonred sandstone are largely silica and clay, but in the red sandstone the cement is largely hematite.

The mineral assemblages of the siltstone and shale of the member correspond to those of the three types of sandstone. The essential difference is an increase in the relative percentage of clay and micaceous minerals as the grain size decreases. Many red siltstone and shale beds contain as much as 15 percent hematite and limonite as cementing and interstitial material.

The pebbles of the conglomerate are largely white and gray vein quartz with considerable amounts of quartzite and minor amounts of schist, phyllite, slate, and shale.

Most of the lower part of the Damascus Member accumulated above sea level, but a few transgressions of the sea are recorded by thin tongues of olive-gray and red beds containing marine fossils. The upper part was deposited under wholly continental conditions.

Barrell (1913–14, p. 464–466) stated that the Catskill was deposited on a subaerial plain where deltaic conditions prevailed. His conclusion was based upon sedimentary features, such as mud cracks, raindrop impressions, plant rootlets, and the lack of marine fossils in most beds. He recognized, however, that the sea transgressed and regressed many times upon the delta in early Catskill time, but from the uniformity of red shale (red beds?) in the formation, he concluded that the delta was well drained when not beneath marine waters. He also believed that the red color of most Catskill rocks was due to sea-level oxidation during and after deposition and that the rocks were not red before that time. In addition, Barrell postulated that the gray beds of the Catskill accumulated in the sea.

When the large number of gray beds in the Damascus Member are considered, the theories advanced by Barrell demand that the sea fluctuate back and forth across the delta many hundreds of times. Barrell's concept of sea-level fluctuation probably is incorrect because most of the gray beds contain plant fossils and only a few contain marine fossils. An alternate postulate, here offered may be more nearly correct. Much of the sediment of the Damascus was deposited on a well-drained coastal plain or above sea level on coalescing deltas or alluvial fans and most probably was red at the time of deposition. When the sediment was deposited and buried rapidly, it retained its red color; but if deposition was followed by slow burial and if the area of deposition was poorly drained, the sediment underwent reduction and the color was changed to gray. The following illustrates the alternate postulate. In the south-central part of the area, where the member is composed of about 75 percent red beds, sediment probably accumulated in large part on a well-drained coastal plain, on coalescing deltas or on alluvial fans where reducing conditions, possibly in swamps or lakes, prevailed only about 25 percent of the time. In the eastern part of the area, where the member consists of about 50 percent red beds and farther east where it consists of about 40 percent red beds, sediment probably accumulated about 50 to 60 percent of the time in a poorly drained environment, possibly in swamps or lakes.

The marine fossils in some red and many gray beds in the Damascus show that Barrell was not correct in stating that the red versus gray color also indicates in all places the dividing line between subaerial or marine deposition. Because the red and gray colors were not precisely related to subaerial or marine deposition, it is necessary to rely upon sedimentary features and fossils as well as color when attempting to determine the environments of Damascus deposition.

The sedimentary features and fossils of the lower part of the Damascus indicate deposition upon a coastal plain in coalescing deltas, across which the sea occasionally transgressed. Sedimentary features and plant fossils in the upper part of the member indicate accumulation on a coastal plain dotted with swamps, lakes, and meandering rivers.

The exact age of the Damascus Member is not known because its fossils have not been studied, but its stratigraphic relations with other members of the Catskill suggest that it is of middle Late Devonian age (Cooper, G. A., and others, 1942, chart 4).

Correlations of the Damascus are difficult. Because of intertonguing, the lower beds of the member are laterally equivalent in westerly directions with the upper beds of the Irish Valley. For the same reason, they may be correlative with parts of the Chemung Formation and the Trimmers Rock Sandstone farther west in central Pennsylvania. The rest of the member is correlative with the lower part of the Buddys Run Member. Other correlations are uncertain, but Willard (1939a, p. 251–252, 292) states that the Damascus is equivalent to the Canadaway Group of Chadwick (1935b) in Bradford and Tioga Counties, Pa.

HONESDALE SANDSTONE MEMBER

The Honesdale Sandstone Group was named by I. C. White (1881, p. 66–69, 132, 140) for exposures at Honesdale in Wayne County. Willard (1939a, p. 288–291), many years later, reduced the Honesdale Sandstone to a formation in his Catskill Continental Facies. Both Willard and White correlated the unit with beds in the vicinity of the report area. Harry Klemic Warman, and Taylor (oral commun., 1959), during an extensive field reconnaissance on the Pocono Plateau, tentatively accepted the correlation of the Honesdale into the Lehighton quadrangle at the eastern end of the Southern Anthracite field (Klemic and others, 1963, p. 30). The Honesdale is here accepted tentatively as the Honesdale Sandstone Member of the Catskill Formation in the southern part of the area. It includes the strata lying between the Damascus and Cherry Ridge Members.

The Honesdale crops out in two belts in the area. The larger of these belts underlies the southern crest of Second Mountain in the Tower City, Pine Grove, Swatara Hill, and Minersville quadrangles (GQ–689, 690, 691, 698). The smaller belt occupies a part of the upper headwaters of Powell Creek in the southern part of the Lykens quadrangle (GQ–701).

In the larger belt the Honsedale forms the southern crest of Second Mountain. The southern crest commonly consists of two subcrests with an intervening swale. The subcrests are underlain by resistant and laterally persistent ledges of conglomerate and conglomeratic sandstone. The intervening swale is underlain by less resistent conglomerate, conglomeratic sandstone, sandstone, siltstone, and shale. The topography of the crestal area locally is rough and steep, although regionally its altitude is remarkably even. The two resistent ledges commonly form rock walls that rise as much as 50 feet above the adjacent less resistent rocks.

In the southern part of the Lykens quadrangle the member is considerably thinner than on Second Mountain. It underlies a rough and steep terrain, but because it is thinner than on Second Mountain and because the dip is gentle, the relief is considerably less.

The lower contact of the Honesdale on Second Mountain is covered at most places. It is at the horizon below which the red and gray sandstone, siltstone, and shale of the Damascus predominates and above which the basal red and gray conglomeratic sandstone and conglomerate of the Honesdale dominates.

The upper contact on Second Mountain is covered at most places, but where exposed, it is at the top of the uppermost resistent conglomeratic sandstone and is overlain by less resistent beds of red and gray sandstone, shale, and siltstone of the Cherry Ridge Member.

The Honesdale is not fully exposed anywhere in the area. The upper part crops out at Mill Creek Gap (fig. 24, section 1) in Second Mountain, and the upper and lower parts are exposed in Swatara Creek Gap in Second Mountain (fig. 24, section 2). Two other sections of the member are partly exposed to the east of the area. The closest of these is at West Branch Gap in Second Mountain (fig. 24, section 3); the other is about 1 mile farther east at Schuylkill River Gap (fig. 24, section 4). An isopach map of the member (fig. 25) was constructed from the thicknesses measured at these localities and from thicknesses calculated at many places in the area and the surrounding region. The Honesdale thins or wedges out in the subsurface of the northern part of the area. The line of zero thickness shown in figure 25 was drawn as far northward as possible in the subsurface to keep the rate of northward thinning to a minimum. As shown, it lies just south of the Catskill outcrop belt on Mahantango Mountain. The Honesdale thins northward at average rates of about 60 to 100 feet per mile in the western and central parts of the area, respectively. The Acadian and Appalachian orogenies are believed to have shortened the Devonian rocks of the area about 30 percent. Before deformation, the member apparently thinned northward at average rates of about 40 to 70 feet per mile in the same parts of the area.

Most freshly exposed rocks of the Honesdale are pale red; others are grayish red and pale brown; and a few are grayish pink, light to dark gray, brownish gray, greenish gray, olive gray, grayish black, grayish green, pale to dark yellowish brown, and pale to yellowish orange. Weathered rocks are predominantly pale red, but a few are grayish red, grayish pink, light to dark

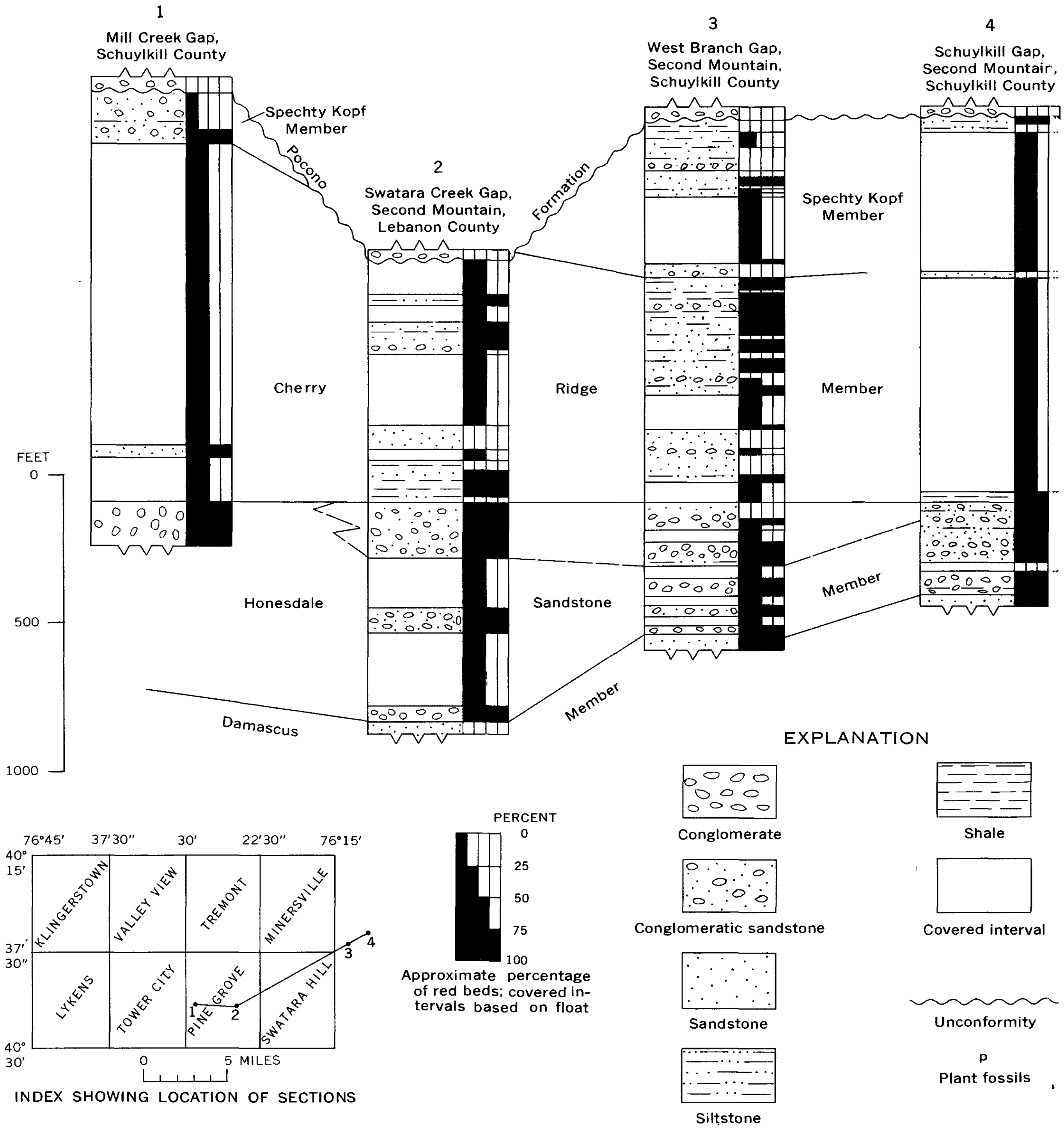


FIGURE 24.—Columnar sections of the Honesdale, Cherry Ridge, and Spechty Kopf Members of the Catskill Formation.

gray, greenish gray, grayish green, olive gray, brownish gray, pale to dark yellowish brown, and grayish black.

Many beds of the Honesdale are crossbedded with simple and planar cross-strata 6 inches to 20 feet long. These crossbeds and sparse ripple marks indicate that the depositing currents flowed from the sourtheast and east. This current pattern disagrees slightly with the thickness pattern shown in figure 25 which seemingly indicates that the source lay somewhat more to the south. Bedding forms are tabular and wedge shaped in about equal proportions. Bedding thicknesses range from about 2 inches to 10 feet, and the average bed is about 2 feet 6 inches thick.

The rocks of the Honesdale are predominantly con-

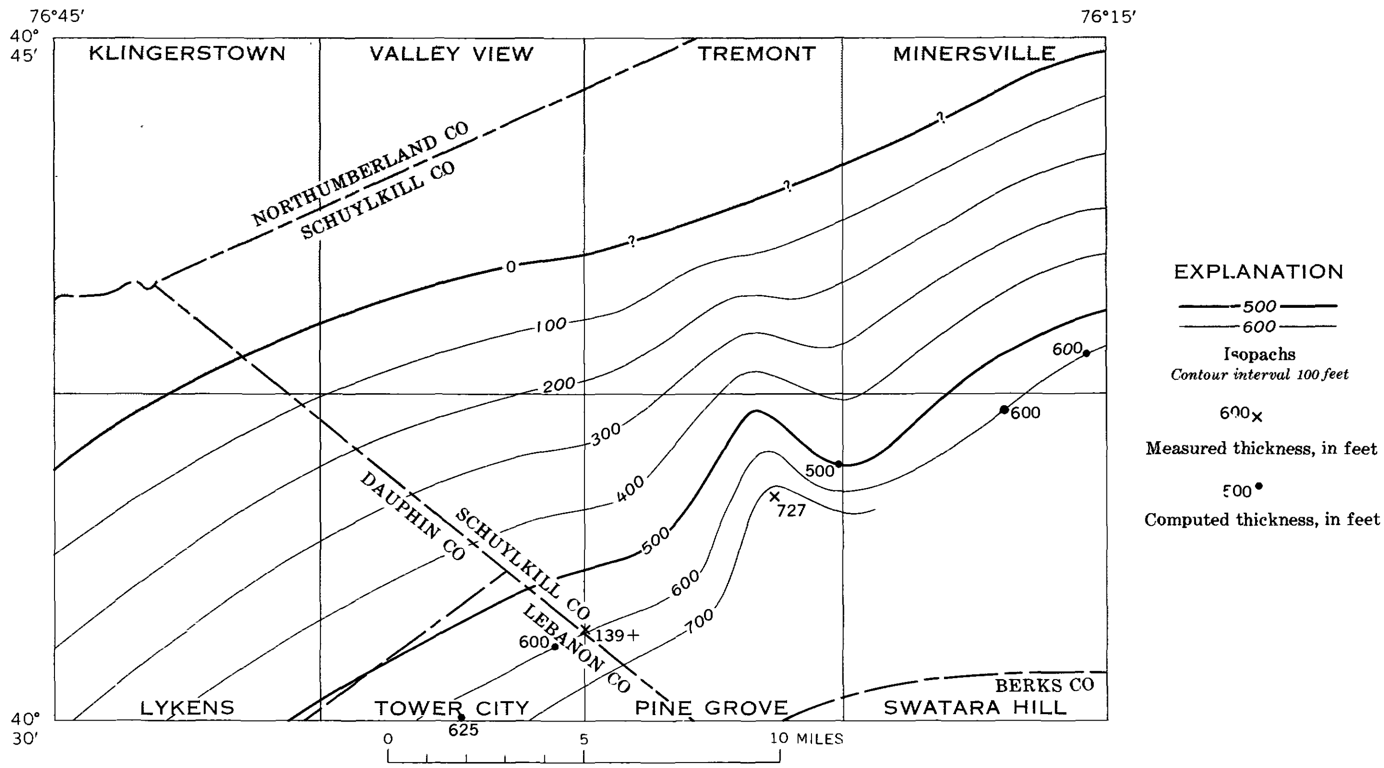


FIGURE 25.—Isopach map of the Honesdale Sandstone Member of the Catskill Formation (shaded where absent).

glomerate, conglomeratic sandstone, and coarse sandstone. Fine-grained sandstone, siltstone, and shale constitute only a small part of the member.

Conglomerate is concentrated largely in the upper and lower parts of the member. Many conglomerate beds are graded and contain pebbles whose maximum diameters range from ½ to 2 inches. These clasts consist largely of milky and gray vein quartz with lesser quantities of gray and varicolored quartzite, gray and black chert, schist, phyllite, and slate. The matrix of the conglomerate is largely coarse to very coarse grained; in some places, however, it is fine to medium grained. The basal few inches of some conglomerate beds contain numerous clay galls and shale fragments.

The sandstone in the Honesdale ranges from very coarse grained to very fine grained with the predominant grain size being fine to medium. Much of the siltstone is sandy, and the shale is commonly silty.

The rocks of the Honesdale are similar mineralogically, varying only as to the amounts of specific minerals, grain size, and iron content. They range in composition from rocks allied with the graywacke clan to rocks allied with the protoquartzite clan. Most of the coarser grained rocks are subgraywacke, and the finer grained ones are chiefly protoquartzite. The principal mineral grains in the three types of rock are common quartz, vein quartz, and quartzite. Accessory minerals are chert, leucoxene, hematite, and muscovite and fragments of schist, phyllite, and slate. Most of the rocks contain considerable quantities of sericite and unidentified clay minerals and small amounts of zircon and tourmaline.

Generally, sorting in the Honesdale is poor to fair. Grains range from angular to well rounded and average subangular. Some quartz and chert grains are exceptionally well rounded and may represent detritus from preexisting sedimentary rocks. In the coarser clastics of the member, the margins of a large percentage of the quartz grains are deeply sutured, probably because of pressure solution. In the finer clastics suturing is conspicuously absent, probably because individual grains commonly are surrounded completely by a matrix of clay, hematite, and sericitic material. The cements are silica, hematite, sericitic material, and unidentified clays. Many of the sandstone and conglomerate beds are slightly to moderately ferruginous. Almost all the shale and siltstone beds are highly ferruginous, containing 5 to 15 percent hematite.

The sediments of the Honesdale are believed to have accumulated a short distance downstream from a fall line where steep gradient streams flowing from a highland area became low-gradient streams flowing upon a flood plain. The rapid northward thinning of the Honesdale indicates that during deposition the fall line did not shift a great deal geographically.

The conglomerate in the Honesdale represents the coarse detritus eroded from a highland source known as Acadia, to the east and southeast of the area. The scattered conglomerates in the upper part of the Damascus indicate that orogenic movements marking the first major pulsation of the Acadian orogeny probably uplifted Acadia during the later part of Damascus sedimentation. Acadia had existed as a low source area for a considerable time before orogenic uplifting and had contributed sediments to rocks as old as the Trimmers Rock Sandstone. During Honesdale sedimentation it was so strongly uplifted that streams draining from it could transport pebbles and cobbles. The abundance of these clasts in the upper and lower parts of the Honesdale suggests that the uplift consisted of two phases or periods with an intervening period of relative quiescence. The abrupt transition from the sandstone, siltstone, shale, and fine conglomerate in the underlying Damascus Member to the coarse conglomerates in the Honesdale indicates that development of a mountainous terrain in Acadia was relatively rapid. The rapid change from coarse Honesdale detritus to fine Cherry Ridge detritus also indicates that the uplifted part of Acadia apparently was eroded rather quickly. The relatively small size of the Honesdale wedge of sediments in eastern Pennsylvania suggests, in addition, that the size of the uplifted mountain region probably was rather small.

The only fossils preserved in the rocks of the member are fragments of plants that have not been studied. The age of these rocks is believed to be middle Late or late Late Devonian because of the stratigraphic position of the member in the Catskill Formation and because Willard (1939a, p. 281) correlated the Honesdale with part of the Canadaway Group of Chadwick (1935b, p. 351). Within the area the member is correlative with the middle part of the Buddys Run Member; to the west it is correlative with the Clarks Ferry Member of Dyson (1963).

CHERRY RIDGE MEMBER

The name Cherry Ridge Group was applied by I. C. White (1881, p. 64–66) to a sequence of five stratigraphic units exposed near Cherry Ridge in Wayne County. As defined by White, the group included the strata lying between his Honesdale Sandstone Group and his Elk Mountain Lower Sandstone. Willard (1936, p. 577–578) changed the name from Cherry Ridge Group to Cherry Ridge Red Beds. Both White and Willard correlated the Cherry Ridge from the type locality into the vicinity of the Southern Anthracite field. Harry Klemic and others (oral commun., 1959) in a reconnaissance study of the Pocono Plateau substantiated the correlation, but they indicated uncertainty by only tentative acceptance of the Cherry Ridge in the Lehighton quadrangle (Klemic and others, 1963, p. 33). It is here tentatively accepted as a member of the Catskill Formation in the area. The Cherry Ridge Member includes the strata between the Honesdale Sandstone Member and Spechty Kopf Member. Where the latter is absent, the top of the Cherry Ridge is the unconformity at the base of the Pocono Formation.

The Cherry Ridge Member crops out in two belts where it is poorly exposed. The larger of these belts lies between the northern and southern crests of Second Mountain in the southeastern part of the area (GQ–689, 690, 691, 698). The smaller belt is in the upper headwaters of Powell Creek valley in the southern part of the Lykens quadrangle (GQ–701).

Along Second Mountain the outcrop belt of the member is a narrow shallow topographic swale that is largely surfaced by talus from the Honesdale and Spechty Kopf Members and the Pocono Formation. The swale deepens near each wind and water gap and locally becomes a deep narrow valley that parallels the trend of the mountain. It also deepens greatly where the Beuchler and Blackwood faults offset the mountain.

In the southern part of the Lykens quadrangle, the Cherry Ridge underlies a topographic swale on the flanks and crest of the Joliet anticline between a low ridge held up by the Honesdale and a high ridge formed by the Spechty Kopf Member and the Pocono Formation. The topography near the ridge of Honesdale Sandstone Member is characterized by moderate to gentle slopes, but is steep, rough, and talus covered on hillsides lying beneath the Spechty Kopf.

The contact between the Cherry Ridge Member and the underlying Honesdale is covered at most places along Second Mountain. Where exposed, it is at the top of the uppermost resistant conglomerate or sandstone ledge of the Honesdale and at the base of much less resistant beds of red and gray sandstone, siltstone, and shale of the Cherry Ridge. In the southern part of the Lykens quadrangle the contact is covered, but it is believed to be where resistant red conglomerate and sandstone beds are overlain by less resistant red beds

The Cherry Ridge Member in the Lykens quadrangle and on Second Mountain west of the Beuchler fault is overlain with apparent conformity by the Spechty Kopf Member. The contact is not well exposed at any point, but it is gradational and is arbitrarily placed where red beds of the Cherry Ridge are succeeded by gray and olive beds of the Spechty Kopf.

The Cherry Ridge is overlain unconformably by the Pocono Formation in the southeastern part of the report area. The unconformity is an erosion surface that developed in Early Mississippian time during a late phase of the Acadian disturbance (Trexler and others, 1961). This ancient erosion surface truncates the Spechty Kopf Member a few hundred feet west of the Beuchler fault (GQ–691). From the point of truncation eastward on Second Mountain, the Cherry Ridge is unconformably overlain by the Pocono Formation. Beds near the unconformity are generally covered by talus and soil mantle, and as a result the amount of angularity is not measureable. In the 2 miles between Beuchler fault and Swatara Creek Gap, however, the unconformity cuts out about 400 feet of the Cherry Ridge Member which indicates an angularity of about 2°.

The Cherry Ridge Member has been measured at two localities in the area and at one locality a short distance to the east. These localities are at Mill Creek Gap (fig. 24, section 1), Swatara Creek Gap (fig. 24, section 2), and West Branch Gap, about 2 miles east of the area (fig. 24, section 3). Other thicknesses have been computed across the belts of outcrop at many places in and adjacent to the area. A map showing the thickness of the member (fig. 26) has been constructed from the measured sections and computed thicknesses. The Cherry Ridge is not recognized in the northern part of the area because the Honesdale Sandstone Member which separates the similar rocks of the Cherry Ridge and Damascus Members wedges out. The thickness of the Cherry Ridge in the subsurface, except near points of measurement, may be in considerable error because of uncertainty as to where the pre-Pocono erosion surface truncated the Spechty Kopf and began eroding into the Cherry Ridge. If the general thickness pattern is approximately correct, the Cherry Ridge thickens northwestward at an average rate of about 75 feet per mile in the western part of the area and at an average rate of about 125 feet per mile in the eastern part. The Devonian rocks of the area were shortened about 30 percent during the Acadian and Appalachian orogenies; thus, the Cherry Ridge thickened northwestward before deformation at about 50 feet per mile in the western part of the area and about 90 feet per mile in the eastern part.

Most rocks in the Cherry Ridge Member are grayish red, pale yellowish brown, and pale red. The rest are light red, light to brownish gray, light to medium dark gray, light to olive gray, very pale to pale orange, grayish orange, yellowish brown, yellowish gray, grayish yellow gray, pale to moderate olive brown, dusky yellow, dusky yellowish green, and dusky yellowish brown.

The strata in the member range in thickness from one-eighth inch to 15 feet. The average sandstone bed is about 3 feet thick, and the average siltstone and shale bed is about 1 foot 6 inches thick. Most beds are tabular, but some are wedge shaped. Simple and planar crossbeds 2 inches to 15 feet long are present in many sandstone beds. Most shale and siltstone beds are structure-

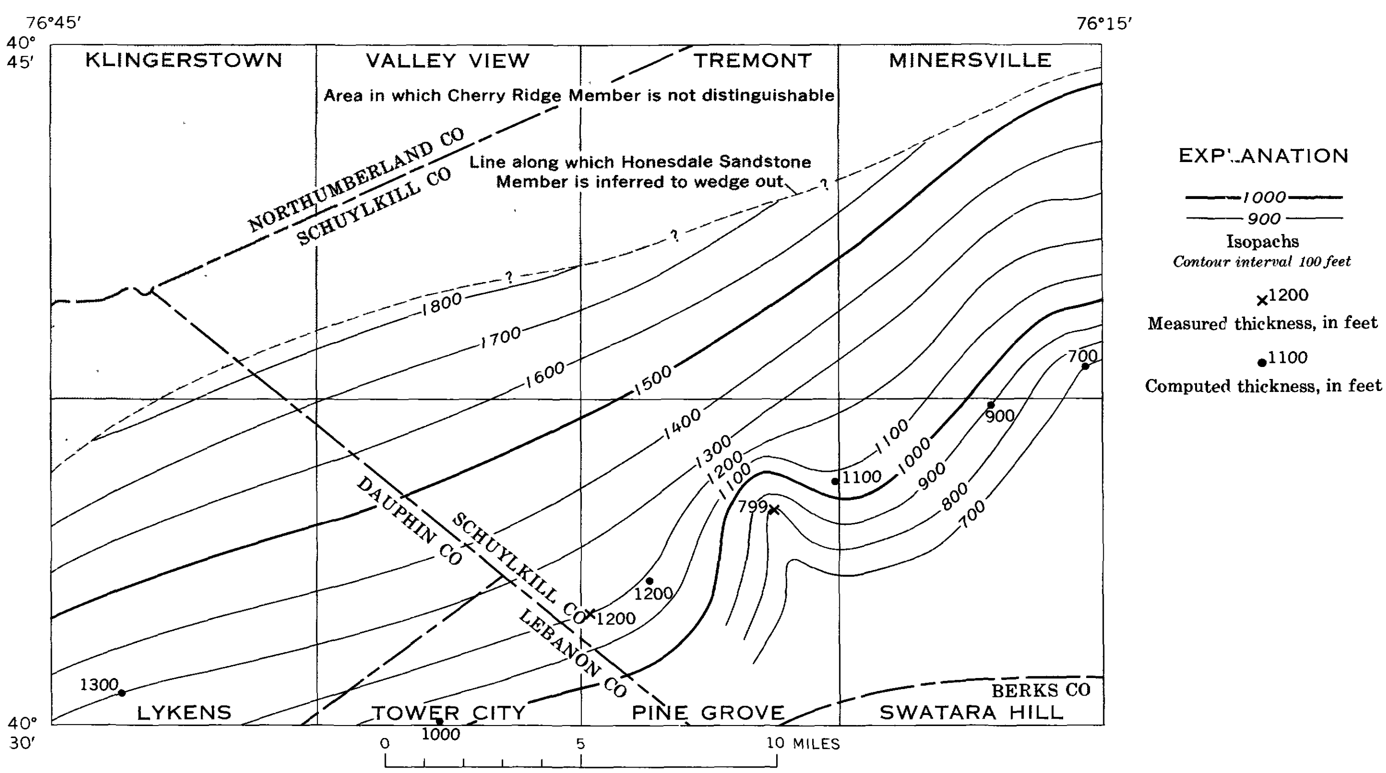


Figure 26.—Isopach map of the Cherry Ridge Member of the Catskill Formation.

less but some are laminated and fissile and a few are contorted. Some sandstone and conglomerate units are graded; others are reverse graded in their upper parts.

The Cherry Ridge is composed of about equal amounts of shale, siltstone, and sandstone. A few beds of conglomerate are scattered throughout the member.

The conglomerate in the Cherry Ridge is much less resistant to erosion than that of the underlying and overlying members. Some pebbles reach a maximum diameter of 1 inch, but most are about a quarter of an inch in diameter. They are composed mostly of milky and gray vein quartz with some milky, gray, and varicolored quartzite; gray, brown, and black chert; and varicolored schist, phyllite, and slate. The matrices are principally medium to coarse grained, but they range from fine to very coarse grained.

The sandstone is very fine to very coarse grained and averages fine to medium grained. Most siltstone is sandy, and the shale is chiefly silty.

The mineral assemblages of the rocks of the member are similar and closely resemble those of the Honesdale, except for a general decrease in grain size. These rocks, regardless of grain size, are compositionally allied with the graywacke, subgraywacke, and protoquartzite clans. Most of the sandstone and conglomerate is subgraywacke, and the finer clastics are largely protoquartzite. Common quartz, vein quartz, and quartzite are the principal mineral grains. Other common grains are chert, ilmenite, magnetite, leucoxene, muscovite, and fragments of schist, phyllite, and slate. All rock types contain considerable percentages of sericite and unidentified clay minerals and small percentages of tourmaline and zircon. Sorting in the coarser clastics is poor to fair; rounding ranges from angular to well-rounded and averages subangular. Many of the larger quartz grains have been deeply sutured by pressure solution. Most smaller quartz grains are not deeply sutured because they are surrounded by a protective matrix of clay, hematite, and sericitic material. Cementing media in all rock types are silica, hematite, and clay. Many of the red beds contain 5 to 15 percent hematite.

Sedimentary structures and plant fossils suggest that the rocks of the Cherry Ridge were deposited on a flood plain lying west and northwest of the highland area known as Acadia. The generally fine-grained character of these rocks indicates that Acadia was low lying during most, if not all, of their deposition, but the scattered conglomerate beds may indicate either slight rejuvenations of Acadia or climatic changes.

The mineral assemblages of the nonred and red beds of the Cherry Ridge are similar, except for variations in ferric versus ferrous iron content and differing percentages of clay matrix. Most of the nonred beds are coarser grained and contain less interstitial clay than do the red beds. The slightly coarser grain size and the smaller quantity of clay matrix in the nonred beds probably allowed a greater circulation of fluids than was possible in the finer textured and clayier red beds. Thus, the color of the nonred beds may have developed during diagenesis and consolidation because of the circulation of oxygen-poor water which reduced the ferric iron to ferrous iron. If such circulation occurred, reducing conditions prevailed in the coarser grained rocks and nonreducing conditions in the finer grained ones.

The only fossils of the Cherry Ridge are fragmentary fossil plants that have not been studied at present. Within the area the member seems to be late Late Devonian in age, as determined from stratigraphic position and from a correlation by Willard (1939a, p. 287), who stated that this unit correlates with the Conewango of western Potter and McKean Counties.

The Cherry Ridge Member of the southern part of the area is correlative with the upper part of the Buddys Run Member of the northern part.

BUDDYS RUN MEMBER

The Buddys Run Member of the Catskill Formation was named by Arndt, Wood, and Trexler (1962, p. C35) for a small creek in the southern part of Shamokin quadrangle, Northumberland County. It includes the strata lying between the Irish Valley and Spechty Kopf Members.

The Buddys Run crops out on the flanks and the crest of the Broad Mountain anticlinorium in the Klingerstown, Valley View, and Tremont quadrangles (GQ–692, 699, 700). The lower part of the member commonly underlies a subdued rolling upland that has 100 to 250 feet of relief. The upland is characterized by a trellis drainage pattern consisting of valleys that are eroded parallel and perpendicular to the strike of the outcrop belt. The upper part underlies gentle to steep, little dissected slopes that rise above the rolling upland to near the crests of Line and Mahantango Mountains.

The contact of the Buddys Run with the underlying Irish Valley Member is in the upper part of an intertonguing sequence of red beds and gray and olive beds. It is at the base of the main red-bed sequence of the Catskill Formation and above gray and olive beds of the Irish Valley that replace the red beds downward. The scale of intertonguing is so great that it can be shown on 1:24,000 maps.

The upper contact of the Buddys Run with the Spechty Kopf Member is nowhere well exposed but is believed to be comformable. Where observed, it is gradational and is placed where gray and olive beds of the Spechty Kopf predominate over underlying red beds of the Buddys Run.

The member is not completely exposed at any place in the area. Thickness variations, therefore, must be determined entirely from computations based upon structural attitudes and the width of the outcrop. Figure 27, an isopath map of the member, was constructed from many thicknesses computed in and adjacent to the area. The Buddys Run is not recognized south of a northeastward-trending line that is the approximate northwest limit of the Honesdale Sandstone Member. South of this line, strata correlative with the Buddys Run are divided into the Damascus, Honesdale Sandstone, and Cherry Ridge Members.

The Buddys Run ranges in thickness from 4,200± to 4,800± feet. The member is thickest in the western part of the Lykens and Klingerstown quadrangles and thins northeastward at an average rate of about 40 feet per mile. Deformation shortened the Devonian rocks of the northern part of the area about 15 percent. Before deformation the member thinned northeastward at an average rate of about 34 feet per mile.

Fresh and weathered rocks of the member are predominantly grayish red, pale red, and pale brown, many being grayish pink, light olive to olive gray, dusky yellow, and pale yellowish brown. A few are light to dark brownish gray, dark yellowish brown, greenish gray to dark greenish gray, light to medium gray, greenish yellow, moderate yellow, pale orange, and dusky yellowish brown.

The beds are ¼ inch to 15 feet thick and average about 3 feet. Most strata are tabular bedded; wedge-shapped beds, however, are relatively common. Simple and planar crossbeds 2 inches to 5 feet long are common in the upper part of the member but are rare in the lower. Many shale and siltstone units are internally contorted, some because of structural adjustments and others because of slumping during deposition.

Plant fragments, raindrop impressions, and mud cracks are preserved in many of the finer grained rocks, as are ripple marks, most of whose crests strike northeastward.

Sandstone is present throughout the Buddys Run. It ranges from very fine grained to very coarse grained, but is chiefly fine to medium grained. Thin conglomerate and conglomeratic sandstone beds are present locally in the middle and upper parts of the member. The sandstone and conglomerate is largely subgraywacke, but protoquartzite and graywacke are relatively common. The mineral constituents of these three rock types differ only in the comparative percentage of quartz to other constituents. They consist of vein and common quartz, muscovite, chlorite, biotite, tourmaline, zircon, ilmenite, leucoxene, plagioclase, orthoclase, sericite, unidentified clay minerals, and rock fragments of quartzite, chert, shale, slate, schist, and phyllite. Pebbles in the conglomerate are composed largely of white and gray vein quartz, quartzite, schist, phyllite, and slate.

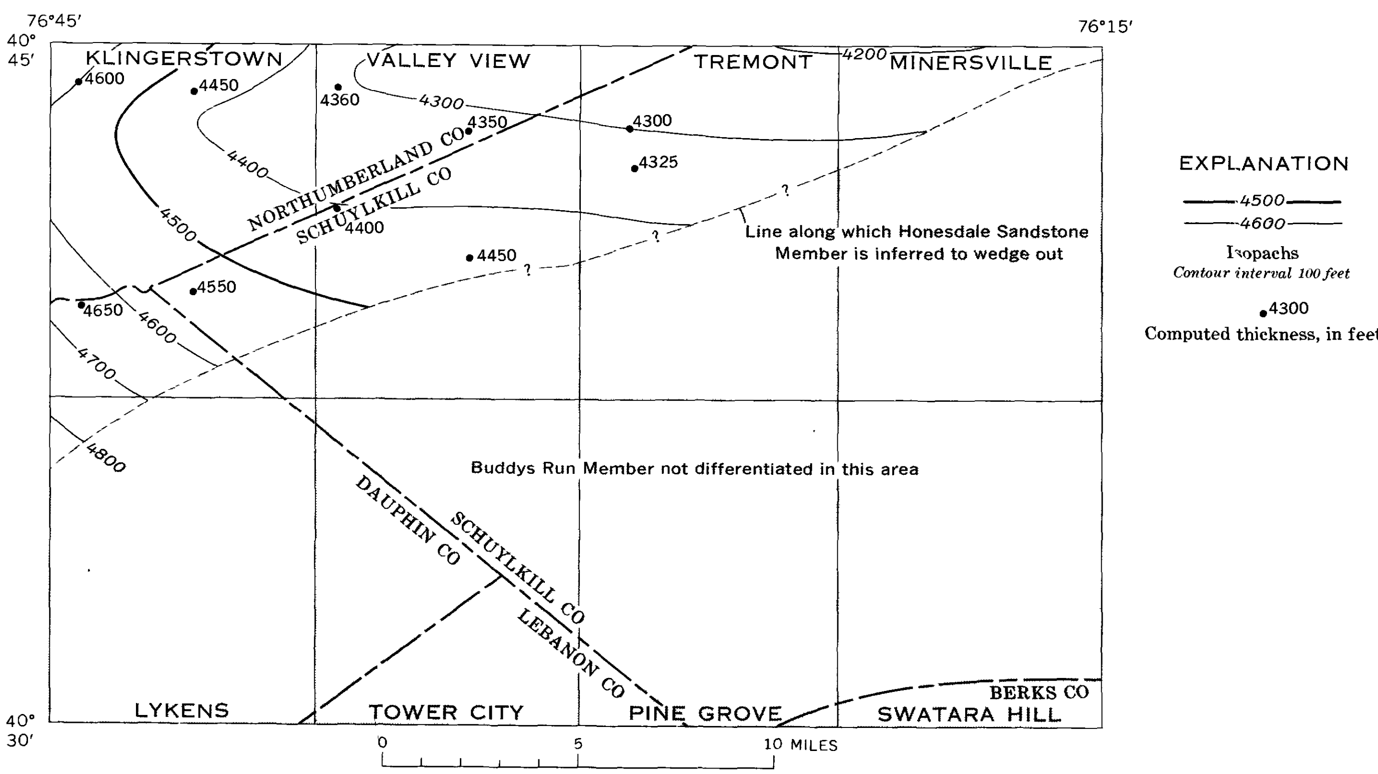


FIGURE 27.—Isopach map of the Buddys Run Member of the Catskill Formation.

The matrices and cement of the nonred sandstone and conglomerate consist largely of silica, unidentified clay minerals, and sericite, whereas in the red sandstone they are composed of these minerals and hematite and limonite. Sorting ranges from fair to good in protoquartzite but is chiefly poor to fair in subgraywacke and graywacke. Although most quartz grains are subangular to subround, angular grains are common.

The siltstone and shale of the Buddys Run is compositionally similar to the sandstone and conglomerate, differing only by increased percentages of clay and mica minerals.

Most of the lower 1,000 to 2,000 feet of the Buddys Run accumulated in subaerial environments but several thin marine tongues record transgressions of the sea. The rest of the member was deposited on flood plains.

The depositional environments of the Damascus, Honesdale, and Cherry Ridge Members, which are correlative with the Buddys Run Member, have already been discussed extensively in this report. Sediments of the Buddys Run are believed to have accumulated under conditions similar to those described and postulated for these members.

The fossils of the Buddys Run have not been studied, but the stratigraphic position of the member indicates that it is probably middle Late and late Late Devonian in age. The basal beds of the member tongue into the upper beds of the Irish Valley in the area and the member as a whole is correlative with the Damascus, Honesdale, and Cherry Ridge Members of the southern part of the area.

SPECHTY KOPF MEMBER

The Upper Devonian and Lower Mississippian Spechty Kopf Member of the Catskill Formation was named by Trexler, Wood, and Arndt (1962, p. C36–37) for a hill in the Lykens 7½-minute quadrangle of Dauphin County (GQ–701). It includes the strata lying between the Cherry Ridge or Buddys Run Member and the Pocono Formation.

The Spechty Kopf crops out on the slopes and crests of Line and Mahantango Mountains in the Klingerstown, Valley View, and Tremont quadrangles (GQ–692, 699, 700) on Broad and Peters Mountains in the southern part of the Lykens quadrangle (GQ–701) and west of the Beuchler fault on Second Mountain in the Pine Grove and Tower City quadrangles (GQ–691, 698). The upper part of the member also is exposed in the canyon of Rattling Creek in the central part of the Lykens quadrangle (GQ–701).

The member underlies a gentle to steep terrain on Broad and Peters Mountains. Elsewhere, it underlies rough, steep, and talus-littered slopes.

The lower contact of the Spechty Kopf with the underlying Buddys Run or Cherry Ridge Member is not well exposed in the area but is believed to be conformable. Where exposed, it is gradational and is arbitrarily placed where gray and olive beds of the Spechty Kopf predominate over red beds of the two lower members.

The contact between the Spechty Kopf and the overlying Pocono Formation is an angular unconformity (Trexler and others, 1961). The greatest observed angularity between these units is about 75° in Rattling Creek canyon south of Lykens (GQ–701). In localities where angular relations are not well developed, or where red beds of the Spechty Kopf are absent or covered, geologists have commonly assigned all or part of the member to the Pocono.

The Spechty Kopf is poorly exposed at most localities in the area; therefore, data concerning its thickness and lithology have been obtained from isolated outcrops and by computing thicknesses at many places across the belts of outcrop. It is moderately well exposed in the area at Mill Creek Gap (fig. 24, section 1) in Second Mountain and east of the area at West Branch Gap (fig. 24, section 3).

Figure 28 is an isopach map of the Spechty Kopf in the area as compiled from the measurements at Mill Creek and West Branch Gaps and from many computed thicknesses. The member is absent in the southeastern part of the area, may have been absent before modern erosion on the crest of Broad Mountain in the northwestern part, and reaches a thickness of about 2,400 feet in the southwestern part.

Rocks in the lower part of the Spechty Kopf are principally yellowish gray, brownish gray, and olive gray. In contrast, rocks in the upper part are chiefly grayish red and medium light gray to light olive gray. A few are light red, moderate red, pale yellowish brown, and pale brown throughout the member.

Most of the gray and olive rocks of the Spechty Kopf are wedge bedded and cross stratified, and the red beds are chiefly tabular bedded. Beds are ¼ inch to 15 feet thick; the gray and olive beds average about 3 feet thick and the red beds, slightly less than 2 feet. Simple and planar crossbeds as much as 5 feet long are common.

The Spechty Kopf is composed of sandstone, siltstone, shale, conglomerate, and coal in decreasing order of abundance. Sandstone and conglomerate predominate in the lower part of the member and siltstone and sandstone in the upper part.

The conglomerate is composed largely of gray, milky, and light-olive-gray vein quartz pebbles and cobbles that commonly are eratically distributed in an olive matrix. The gray and olive sandstone and the matrix of the conglomerate are mainly medium to coarse

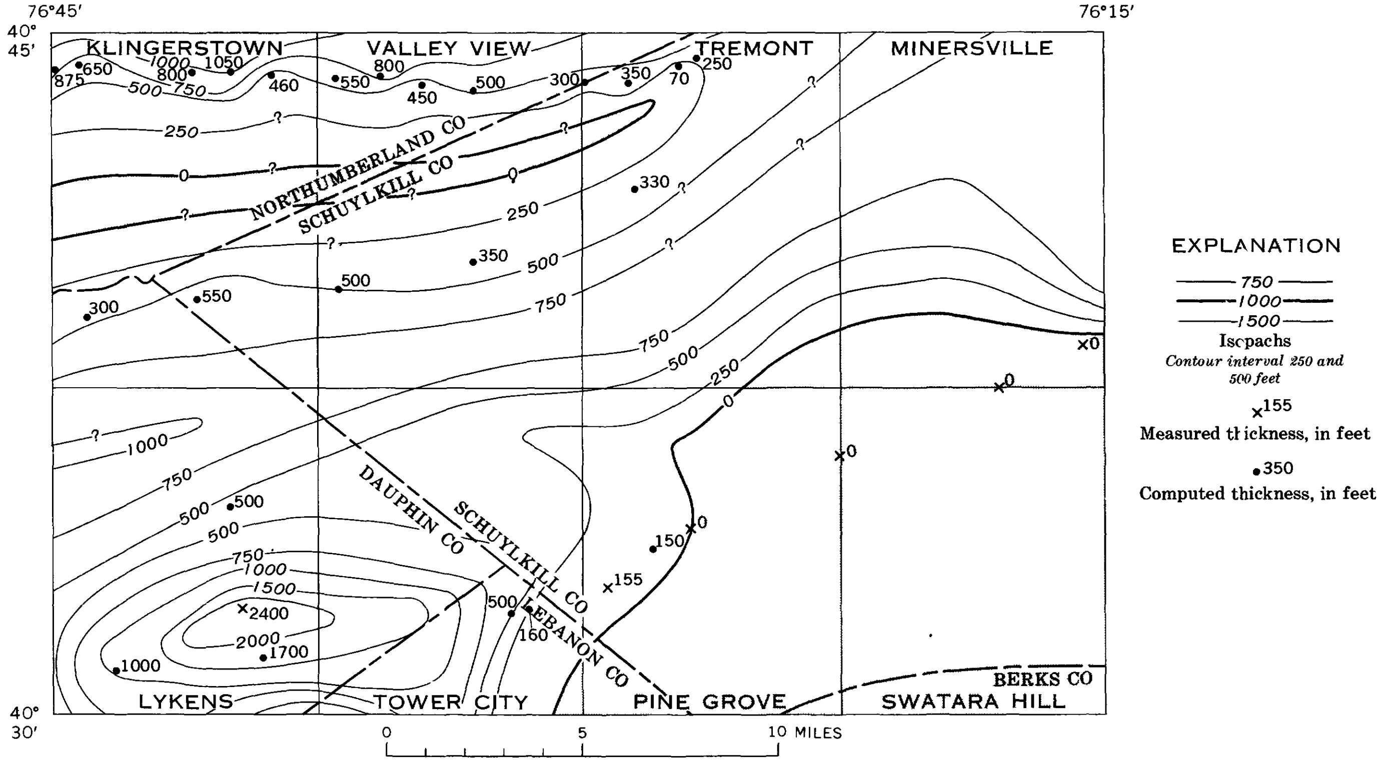


FIGURE 28.—Isopach map of the Spechty Kopf Member of the Catskill Formation (shaded where absent).

grained, some being very fine to very coarse grained. Although most of the red sandstone is fine to medium grained, much is very fine grained.

Regardless of their grain size and regardless of whether they are gray and olive or red, the rocks of the Spechty Kopf, in decreasing order of abundance, are subgraywacke, protoquartzite, and graywacke. The principal mineral constituents of all rock types are grains of vein and common quartz with subordinate amounts of biotite, muscovite, sericite, chlorite, unidentified clay minerals, plagioclase, orthoclase, leucoxene, ilmenite, zircon, and tourmaline, and rock fragments of quartzite, schist, phyllite, chert, slate, and shale. The grains are subangular to subround in the protoquartzite and angular to subangular in the subgraywacke and graywacke. Sorting is poor to fair in the protoquartzite and poor in the subgraywacke and graywacke. Cementing materials in the gray and olive rocks are silica, unidentified clay minerals, and sericite; but in the red beds they are chiefly limonite, hematite, and clay.

The sediments of the Spechty Kopf were eroded from an ancient highland that lay southeast and east of the area. These sediments are the sedimentary record of the second uplifting of Acadia during the Acadian orogeny.

The coarse sandstone and conglomerate in the basal and upper parts of the Spechty Kopf suggest that these parts of the member were deposited a short distance downstream from piedmonts of Acadia. In contrast, the finer grained sandstone, siltstone, and shale in the middle part of the member suggest that the source either was lower or farther away during their deposition.

At the end of Spechty Kopf deposition the effects of the third episode of the Acadian orogeny were felt farther to the northwest and deformed the rocks of the area. Subsequently, the rocks of the member were completely removed by erosion in some places, and then the area was covered by the basal sediments of the Beckville Member of the Pocono Formation.

The rocks of the Spechty Kopf are believed to have accumulated on a flood plain in environments similar to those described for the upper parts of the Damascus and Buddys Run Members and for the Honesdale Sandstone Member and Cherry Ridge Member.

The Spechty Kopf Member is of Late Devonian and Early Mississippian age. This assignment is based in part on plant fossils and in part on stratigraphic position. The assemblage of plant fossils from a coal bed on Line Mountain and from another on Berry Mountain contain *Adiantites* spp. (S. H. Mamay, oral commun., 1960), which is indicative of an Early Mississippian age (Read, 1955, p. 8). The stratigraphic position of the Spechty Kopf between the Lower Mississippian Pocono Formation and the Upper Devonian Cherry Ridge or Buddys Run Member of the Catskill Formation substantiates the age determined from the fossil plants.

Correlation of the Spechty Kopf with other stratigraphic units in Pennsylvania is difficult because Upper Devonian and Lower Mississippian rocks are rather unfossiliferous, because the pre-Pocono unconformity has not, as yet, been completely delineated, and because rocks of this age have not been studied adequately in most areas. The member may be partly or wholly correlative with the Elk Mountain Sandstone and Mount Pleasant Red Shale of Willard (1939a, p. 282–283) of eastern and northern Pennsylvania, with the marine Oswayo Formation of north-central Pennsylvania, and with the Pocono-Catskill Transition Group of I. C. White (1883, p. 49–52) of northeastern Pennsylvania. The lower part of the Pocono Formation in central and eastern Pennsylvania, as recognized in many published reports, may contain rocks that would be assigned to the Spechty Kopf by the authors.

THICKNESS OF THE CATSKILL FORMATION

The Catskill Formation ranges in thickness from 5,100 feet in the southeastern part of the area to 9,400± feet in the southwestern part (fig. 29); the latter thickness is the greatest recorded for Catskill rocks in Pennsylvania.

The effects of the Acadian orogeny and pre-Pocono erosion on the thickness of the Catskill are determinable by comparing figure 30, a map showing the thickness of the pre-Spechty Kopf rocks of the Catskill, with figures 28 and 29. During the orogeny, the rocks of the area were folded into several northeastward-trending anticlines and synclines that were subsequently beveled by the pre-Pocono erosion surface. The geographic positions of these ancient folds are approximately delineated in figure 28 by the northeastward-trending thin to thick belts of the Spechty Kopf Member. A comparison of figures 28 and 29 also indicates that these folds are partly recorded by thickness variations of the entire Catskill Formation.

PRE-POCONO UNCONFORMITY

For many years, geologists have speculated whether the Acadian orogeny extended into eastern Pennsylvania. Since the latter part of the 19th century there has been little detailed geologic mapping that could be used to solve this problem. C. D. White (1934) and Willard (1936, p. 599–600) believed that a disconformity separated the Pocono and Catskill Formations. It may be deduced from the outcrop patterns on Willard's geologic map (fig. 72) and from the text of his report that the region northeast of the report area was tilted down to the northwest and then eroded before the deposition of the Pocono Formation. In 1961, Trexler, Wood, and Arndt described the unconformity in the area and the surrounding region.

The unconformity between the Spechty Kopf Member of the Catskill and the Pocono Formation is best

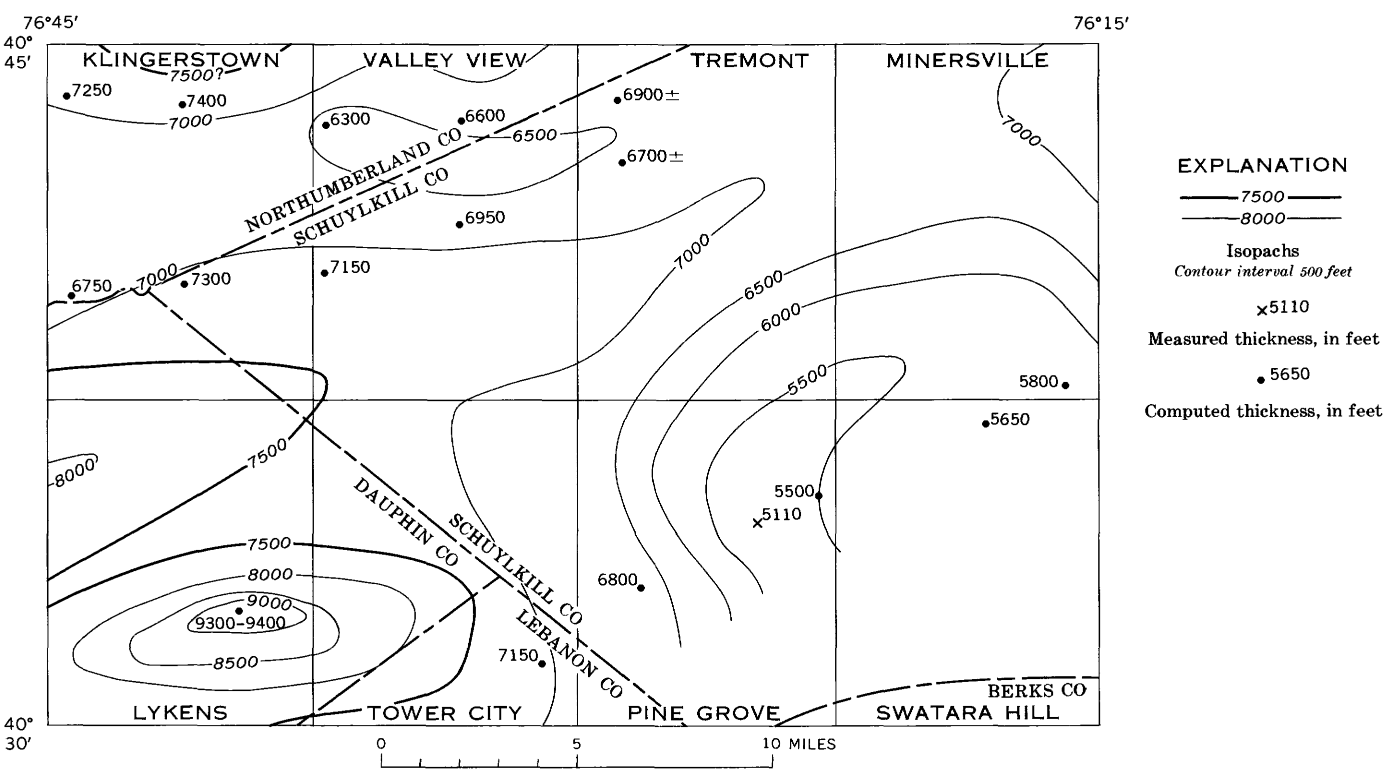


FIGURE 29.—Isopach map of the Catskill Formation.

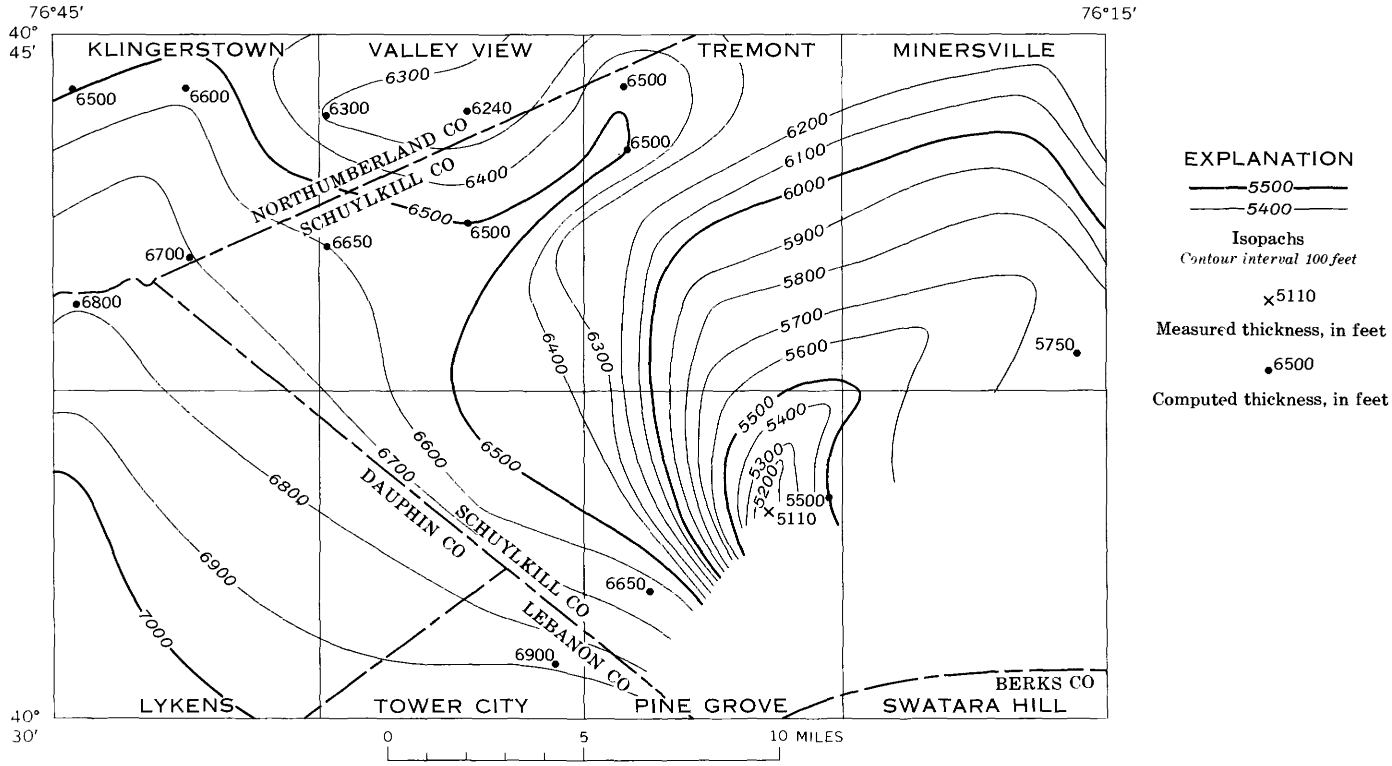


Figure 30.—Isopach map of members of the Catskill Formation older than the Spechty Kopf Member.

preserved on Broad and Peters Mountains in the southern part of the Lykens quadrangle (GQ–701). There, the upper 2,000 to 2,300 feet of the member is truncated by an erosion surface. Beds of the Spechty Kopf dip 10° to 30° more steeply northward on the north limb of the Joliet anticline than do overlying beds of the Pocono, and they dip as much as 25° less steeply southeastward on the south limb. These angular relations indicate that the Spechty Kopf locally was folded or tilted about 30° northward and northwestward and, subsequently, eroded before accumulation of the Pocono.

The angularity between bedding planes of the Spechty Kopf and the Pocono is as great as 75° in Rattling Creek canyon about 1 mile south of Lykens (GQ–701). Structural attitudes of Spechty Kopf rocks on Berry, Broad, and Peters Mountains indicate that the rocks were folded into a broad open syncline during the Acadian orogeny. Subsequently, during the Appalachian orogeny, the broad open syncline was further folded and became the Joliet anticline, the Rattling Creek syncline, and the Berry Mountain anticline (Trexler and others, 1961, fig. 38.2).

Detailed mapping on Line and Mahantango Mountains in the Klingerstown, Valley View, and Tremont quadrangles (GQ–692, 699, 700) has shown that the Sprechty Kopf Member varies greatly in thickness in very short distances and that it is separated from the Pocono by an erosion surface. The Spechty Kopf may have been completely removed by erosion between these mountains near the axis of the present Broad Mountain anticlinorium (fig. 28).

Pre-Pocono erosion also removed the Spechty Kopf Member from Second Mountain between a point a few hundred feet west of the Beuchler fault in the Pine Grove quadrangle and the east border of the Swatara Hill quadrangle (GQ–689, 691). In the area, where the Spechty Kopf is removed, the Pocono rests unconformably upon beds of the Cherry Ridge Member of the Catskill Formation at a divergent angle of about 2°.

MISSISSIPPIAN SYSTEM

Mississippian rocks range in thickness from about 5,000 feet on the Broad Mountain anticlinorium to about 8,500 feet in the Minersville synclinorium.

Much of the Spechty Kopf Member of the Catskill Formation is of Early Mississippian age. This member is described in the chapter on Devonian and Mississippian rocks. It is overlain unconformably by the Pocono Formation, which is divided, in ascending order, into the Beckville and Mount Carbon Members. The Mauch Chunk Formation of Late Mississippian and Early Pennsylvanian age rests conformably on the Mount Carbon Member and is divided into three informal members.

POCONO FORMATION

In 1876, Lesley (p. 221–227), in his description of the Boyd's Hill well at Pittsburgh, introduced the name Pocono Sandstone for the sequence of gray sandstone and conglomerate that lies between the red beds of the Mauch Chunk and Catskill Formations. The succeeding year, Franklin Platt and W. G. Platt (1877, p. XXVI), in describing the new formations proposed by Lesley, stated that the gray sandstone and conglomerate between the Mauch Chunk and Catskill on the Pocono Plateau "should be called the Pocono Formation, for it forms the mass of the great mountain plateau between the Delaware and Lehigh Rivers."

A few years later I. C. White (1882) proved that most of the Pocono Plateau was surfaced by the Catskill Formation. He (1883, p. 49) recommended that his Mount Pleasant Conglomerate and 300 to 500 feet of overlying red rock be reassigned to his Pocono-Catskill Transition Group. Lesley (in Lesley and others, 1892–95, p. XV) in his redefinition of the Pocono, ignored White's recommendation by stating: "Therefore, to find any Pocono on the Pocono Plateau, one must go a number of miles to the north of the front edge of the plateau, where ridges of the *lowest Pocono* rock, the Mount Pleasant Conglomerate, remain uneroded." Willard (1936, p. 597–598) accepted White's recommendation by removing White's Mount Pleasant Conglomerate and the overlying red beds from the Pocono and assigning them to the Catskill Formation. He, however, accepted Lesley's definition of a type area.

The Pocono is classified as a formation by the U.S. Geological Survey. It includes the rocks between the underlying Catskill Formation and the overlying Mauch Chunk Formation as defined by White and Willard. The base of the formation in the area is the pre-Pocono unconformity, and the top is the lowest red bed of the Mauch Chunk. The formation, in ascending order, consists of the Beckville and Mount Carbon Members in the report area.

The Pocono is the principal ridge former of the area, underlying from north to south: Line, Mahantango, Berry, Broad, Peters, and Second Mountains. It generally crops out on narrow and precipitous ridge crests, except on Broad Mountain, in the area where Line and Mahantango Mountains join, and in the area where Peters and Broad Mountains join. Dip slopes of the formation on these mountains are eroded either into subdued and poorly exposed cuestas and hogbacks or are covered by large talus fields.

BECKVILLE MEMBER

The Beckville Member was established by Trexler, Wood, and Arndt in 1962 (p. C38). It was named for a village a short distance south of the type section at West Branch Gap in Second Mountain. This gap is several miles east of the report area.

The member is poorly to moderately well exposed at Mill Creek and Swatara Creek Gaps in Second Mountain (GQ–691), is moderately well exposed at Rattling Creek Gap in Berry Mountain (GQ–701), and is largely covered at Pine Creek Gap in Mahantango Mountain (GQ–700). It underlies Second, Peters, Broad, Berry, Mahantango, and Line Mountains (GQ–689, 690, 691, 692, 698, 699, 700, 701). The Beckville commonly crops out on the crests of ridges. The slopes beneath these crests are generally rough and steep and are characterized by a series of subdued cliffs, cuestas, and hogbacks. In those localities where the crests are underlain by either the Spechty Kopf or Mount Carbon Member, the outcrop belt of the Beckville is largely covered by talus from those members.

The Beckville unconformably overlies the Spechty Kopf and Cherry Ridge Members of the Catskill Formation. The unconformity is beneath a widespread basal conglomerate. East of the Beuchler fault on Second Mountain, the Beckville overlies the Cherry Ridge, but elsewhere it rests on the Spechty Kopf. At most localities the basal conglomerate rests on red beds, but at a few places it overlies gray and olive conglomerate and sandstone. The position of the unconformity in these latter places can only be established with certainty by careful lateral tracing, and in a few localities, by angular discordance between underlying and overlying beds.

The upper contact of the Beckville is placed at the bottom of the basal conglomerate of the Mount Carbon Member. At most places in the area the upper part of the Beckville is sandstone, shale, siltstone, and coal; therefore, the contact between the finer grained clastics of the Beckville and the basal conglomerate of the Mount Carbon generally is easy to locate. Near the northwest corner of the area, however, the basal conglomerate grades westward into very coarse sandstone, and the position of the contact becomes hard to determine. For this reason, it is not advocated that the Pocono be divided into members to the north and west of the area.

Sections of the Beckville were measured at Mill Creek Gap (fig. 31, section 1); Swatara Creek Gap (fig. 31, section 2); West Branch Gap (fig. 31, section 3), which lies several miles east of the area; and Schuykill Gap (fig. 31, section 4), which is a short distance farther east.

Figure 32 is an isopach map of the Beckville Member. It was compiled from thicknesses at these sections and from thicknesses calculated across the outcrop belts in and adjacent to the area. The member is thickest on Broad and Peters Mountains and is thinnest in the sub-

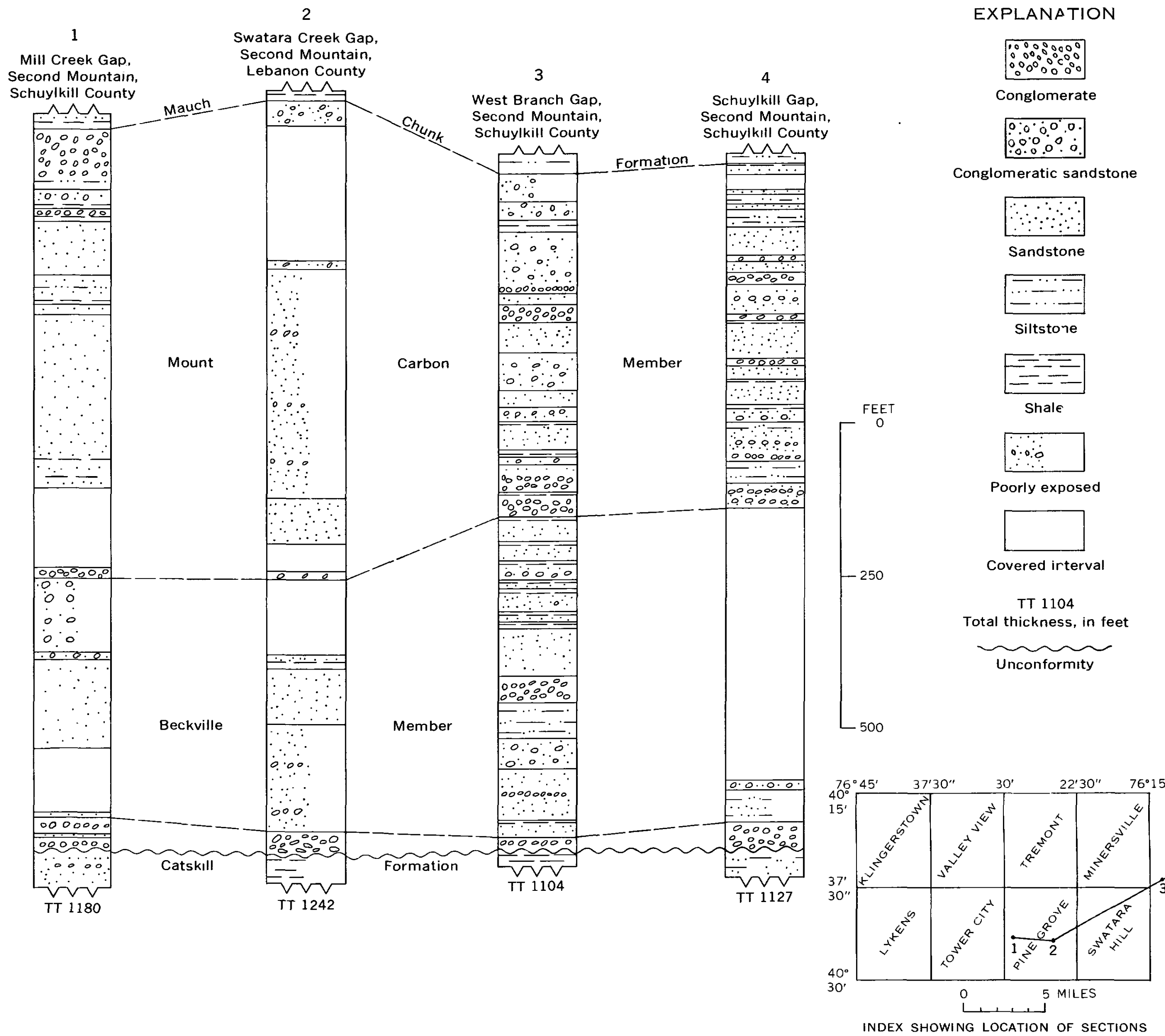


FIGURE 31.—Columnar sections of the Pocono Formation.

surface of the central and northern parts. The thickness trends of the member approximate those of the underlying Spechty Kopf Member (fig. 29), which suggests that the Acadian orogeny lasted into Beckville time.

Freshly exposed rocks of the member are mainly very light to medium gray, some being pale to olive gray, dark gray, greenish gray, brownish gray, yellowish gray, pale red, pale to dark yellowish orange, and grayish pink. Weathered rocks are commonly similar in color, but a few are moderate reddish orange, dark greenish gray, light brownish gray, olive black, and light brown.

The Beckville is composed of conglomerate, sandstone, siltstone, shale, and coal. Conglomerate is largely concentrated in lower part, and sandstone predominates in the upper part.

The conglomerate is in beds 1 foot to 25 feet thick, but most beds average about 4 feet. The average bedding unit is cross stratified and wedge shaped; the rest are tabular and lenticular. Simple and planar cross-strata range from small to as long as 20 feet.

The basal conglomerate is persistent throughout the area, is more extensive than any other conglomerate, and may be correlative with the Griswold Gap Conglomerate of I. C. White (1881, p. 56–57) of northeastern Pennsylvania. Pebbles and cobbles in this and other conglomerates range in diameter from 1/4 inch to 4 inches. Those

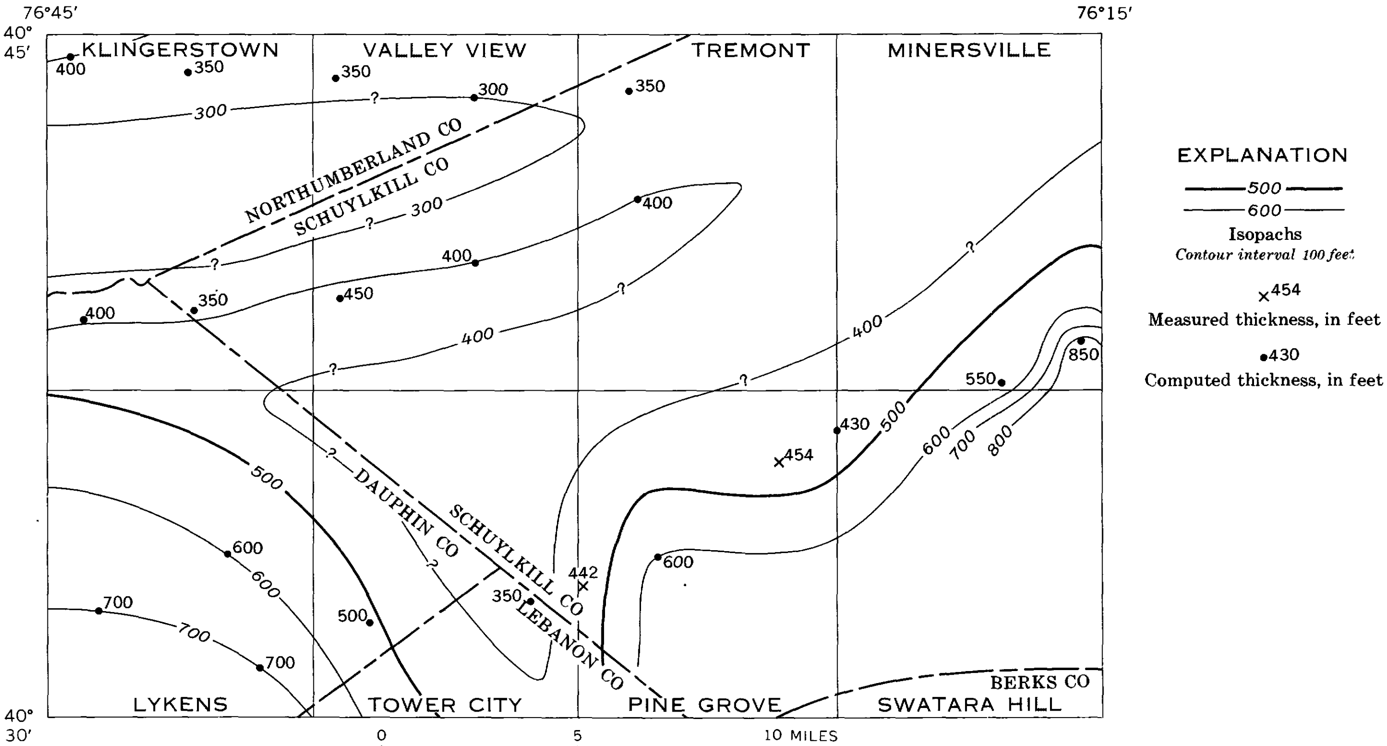


FIGURE 32.—Isopach map of the Beckville Member of the Pocono Formation.

of vein quartz and quartzite are commonly equidimensional to ovoid, are less commonly elongate, and are chiefly rounded to well rounded. The few clasts of other rock types generally are elongate and flattened. The long axis of most of the elongate and ovoid clasts are oriented parallel to bedding.

The sandstone beds of the Beckville range in thickness from 1 foot to about 10 feet and average about 3 feet. Most are tabular and cross stratified; the rest are wedge shaped and parallel stratified. Abundant simple and planar cross-strata are mostly less than 5 feet long, but some are as long as 25 feet.

The siltstone is mainly in beds about 3 feet thick, but ½-inch to 10-foot beds are common. The thicker beds are concentrated largely in the lower part of the member where thin beds of fine-grained sandstone and shale are intercalated at many places. Much of the siltstone contains scattered fine to medium grains of well-rounded quartz.

Most shale beds are 1 to 2 feet thick, but some are as thick as 8 feet. Thin units of fine-grained sandstone and siltstone are interbedded at many localities. Some shale is sandy, some contains abundant plant remains, and some contains thin lenses of coal. Elongate plant remains are generally oriented parallel to the strike of nearby crossbedding and ripple marks.

Compositionally, the rocks of the Beckville are allied with the orthoquartzite, protoquartzite, and subgraywacke clans. Regardless of differences in grain size, the mineral assemblages of these rocks are similar and vary only in the proportion of quartz to rock fragments, mica minerals, and unidentified clay minerals.

The pebbles and cobbles in the conglomerate consist of milky, white, light-gray, and medium-gray vein quartz; light- to dark-gray, brownish-gray, pale-red-purple, and medium-bluish-gray quartzite; and dark varihued sandstone, siltstone, schist, gneiss, slate, and chert. Most of these clasts are well lithified and rounded to varying degree. Poorly lithified angular shale chips and clay galls are present in the basal parts of many conglomerate and sandstone beds.

The siltstone, sandstone, and the matrices of the conglomerate consist of detrital grains of vein and common quartz; fragments of quartzite, sandstone, schist, slate, phyllite, and chert; varying amounts of chlorite, biotite, muscovite, and sericite; and small amounts of leucoxene, ilmenite, magnetite, rutile, sphene, tourmaline, and zircon; sparse grains of andesine partly altered to clay and sericite; and films of hematite and limonite. The matrices of much of the protoquartzite and subgraywacke are composed largely of crushed rock fragments, bent mica plates, quartz fragments, sericite, and unidentified clay minerals. Cementing media are silica, clay, sericite, and a binder of silt-sized rock and quartz fragments.

The borders of many quartz and quartzite grains have been sutured by pressure solution, and a few exhibit secondary overgrowths. Much of the quartzite consists of large angular to rounded fragments of metamorphosed sandstone; the rest consists largely of small irregular elongate fragments of metamorphosed siltstone. Most chert grains are well rounded. Many fragments of quartzite schist, slate, and phyllite are elongate, but others seem to have been crushed and to have flowed into the interstitial pore space.

Sorting is moderate to good in the orthoquartzite of the Beckville, poor to moderate in the protoquartzite, and chiefly poor in the subgraywacke. Although medium-grained sandstone predominates over other sandstone, very fine grained, fine- and coarse-grained, and very coarse grained sandstone is common. Most grains in the sandstone and in the matrices of the conglomerate are subangular to subround, but angular and rounded grains are rather common.

MOUNT CARBON MEMBER

Trexler, Wood, and Arndt (1962, p. C39) named the Mount Carbon Member of the Pocono Formation for a village that lies about 2.5 miles east of the area.

The member generally crops out in dip slopes on Second, Peters, Broad, Berry, Mahantango, and Line Mountains, but locally, it underlies the crests (GQ–689, 690, 691, 692, 698, 699, 700, 701). On Second, Mahantango, and Line Mountains it is largely covered by talus fields. Localities underlain by the Mount Carbon are either steep, rough, and talus-covered featureless slopes or poorly formed, imperfectly exposed, and partly talus-covered hogbacks and cuestas.

The lower contact of the Mount Carbon is at the base of a conglomerate that rests conformably upon finer grained beds of the Beckville Member. On Line Mountain near the west border of the area (GQ–701), the conglomerate grades westward into a very coarse grained sandstone. West and north of that border, the Beckville and Mount Carbon Members merge, and the Pocono is undivided.

The upper contact of the Mount Carbon Member is at the base of the lowest red sandstone, siltstone, or shale of the overlying Mauch Chunk Formation. The rocks in the uppermost part of the Mount Carbon are chiefly gray sandstone, but locally, are gray shale or siltstone. A sharp topographic break has been etched out at most places along the contact because of the difference in erosional resistance between the competent rocks of the Mount Carbon and the less resistant rocks of the Mauch Chunk.

Sections of the Mount Carbon were measured at Mill Creek Gap (fig. 31, section 1) and Swatara Creek Gap (fig. 31, section 2) in the Pine Grove quadrangle (GQ–691). Two other sections were measured east of the area at West Branch Gap (fig. 31, section 3) and Schuylkill Gap (fig. 31, section 4).

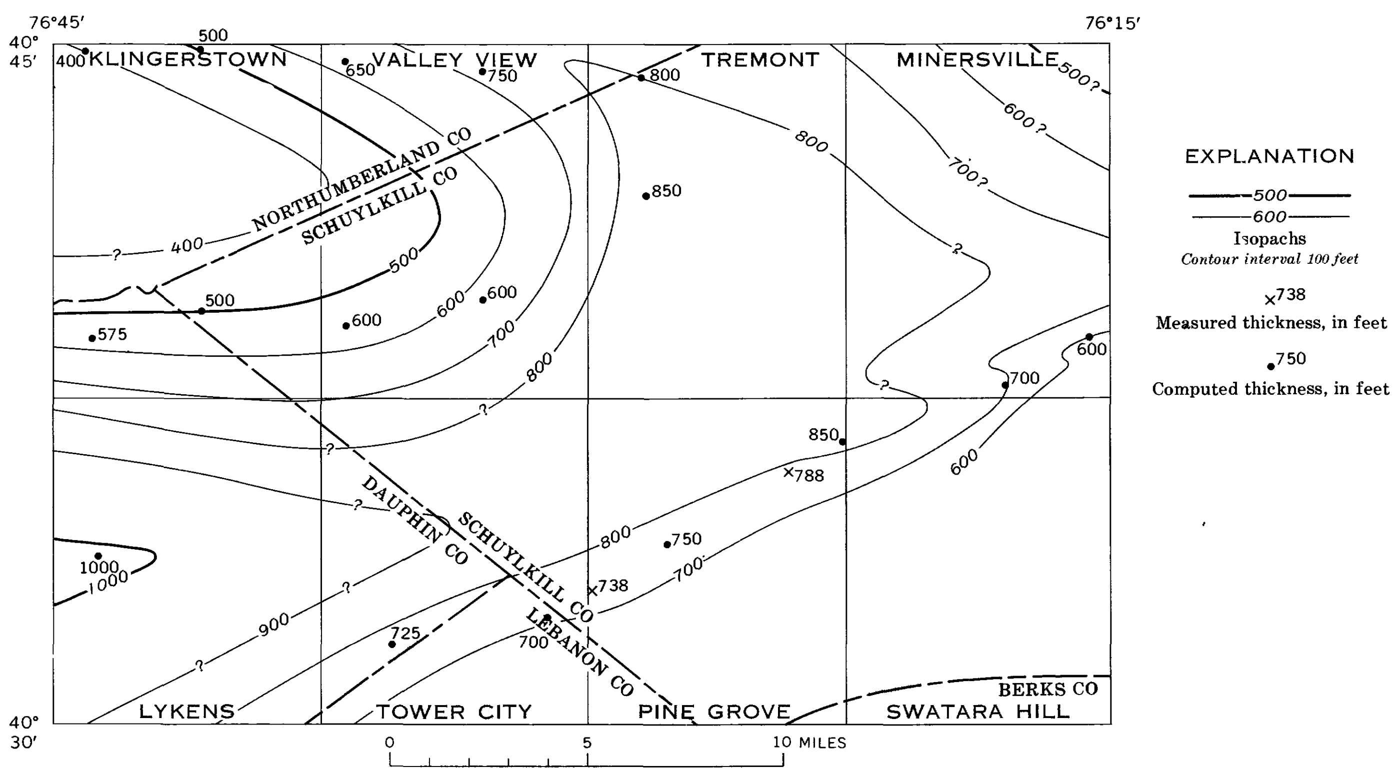


FIGURE 33.—Isopach map of the Mount Carbon Member of the Pocono Formation.

The thickness of the member is illustrated in figure 33. It is thickest in the southwestern part of the area, where it exceeds 1,000 feet, and is thinnest in the northwestern part, where it is slightly less than 400 feet thick. The thickness trends of the Mount Carbon approximate those of the underlying Beckville Member (fig. 32), the Spechty Kopf Member of the Catskill Formation (fig. 28), and the Catskill Formation (fig. 29), which suggests that folding as a result of the Acadian orogeny was still affecting the area slightly.

Fresh rocks of the member are chiefly light to dark gray, light to olive gray, and light to dark greenish gray; a few are light bluish gray, light to brownish gray, pale to moderate yellowish brown, very pale to grayish orange, grayish pink, grayish orange pink, grayish olive, dusky yellow, greenish black, and black. Weathered rocks are principally light to dark gray and light to olive gray. Other weathered rocks are the same color as the fresh rock, and are also dark yellowish brown, moderate olive brown, brownish black, olive black, yellowish gray, and moderate yellow green.

Individual beds are 2 inches to 40 feet thick and average about 3 feet. Wedge-shaped and lenticular beds are common, but most are tabular. Many sandstone and conglomerate beds exhibit simple and planar cross-stratification of 2 inches to 25 feet.

The member consists, in decreasing order of abundance, of sandstone, conglomeratic sandstone, conglomerate, siltstone, shale, and coal.

The conglomerate is composed of pebbles and cobbles of milky, white, and gray vein quartz and varicolored quartzite, schist, slate, shale, and sandstone. These clasts are set in a matrix of very fine to very coarse sand. Most vein quartz and quartzite clasts are equidimensional to ovoid and rounded to well rounded; and those composed of other rock types are chiefly elongate.

Much of the siltstone and shale is arenaceous and contains thin intercalated beds of very fine grained and fine-grained sandstone. Abundant plant fossils and thin lenses of coal are common in some shale beds.

Regardless of grain size, the rocks of the Mount Carbon Member are compositionally allied to the orthoquartzite, protoquartzite, and subgraywacke clans. The mineral assemblages of these rock types are similar and differ only as to the percentage of specific minerals. The principal constituents are grains of vein and common quartz and fragments of quartzite. Accessory constituents are fragments of schist, slate, phyllite, and chert; plates of chlorite, biotite, muscovite, and sericite; grains of leucoxene, ilmenite, magnetite, sphene, tourmaline, and zircon; traces of altered and unaltered andesine; and films of hematite and limonite. Even though the margins of most grains of quartz and quartzite have been deeply sutured by pressure solution, many seem to have been rounded to well rounded before suturing. All rock fragments except chert and quartzite are characteristically elongate and range from angular to well rounded. Most chert grains are rounded. Many schist, shale, and phyllite fragments seem to have been crushed and to have flowed into interstitial pore space. The matrices of much of the subgraywacke and protoquartzite are composed of these crushed fragments, bent biotite and muscovite plates, small fragments of quartz, sericite, and unidentified clay minerals. Cementing minerals in these rocks are silica, clay, sericite, crushed rock fragments, and silt-sized quartz grains. The principal cement in the orthoquartzite is silica.

Sorting in the subgraywacke is poor to fair, in the protoquartzite it is poor to moderate, and in the orthoquartzite it is moderate to good. Most of the sandstone is medium grained, but very fine grained, fine-grained, coarse-grained, and very coarse grained sandstone is common.

THICKNESS OF THE POCONO FORMATION

Figure 34 is an isopach map of the Pocono Formation constructed from thickness data in and adjacent to the report area. The formation is thinnest in the northwestern part, where it is probably less than 700 feet, and is thickest in the southwestern part, where it exceeds 1,700 feet. The area of greatest thickness coincides partly with the area of greatest thickness of the Spechty Kopf Member and the Catskill Formation. This approximate superposition suggests that the Acadian disturbance continued during deposition of the Pocono.

DEPOSITION OF THE POCONO FORMATION

Pocono rocks seem to have accumulated principally as flood-plain deposits, to a lesser extent as swamp deposits, and perhaps locally, as lacustrine deposits. They were deposited in an asymmetric northeastward-trending basin whose axis lay close to its southeastern margin and the source area. The sediment that formed these rocks was eroded from a mountainous region southeast of the report area that was uplifted during the final phase of the Late Devonian and Early Mississippian Acadian orogeny.

Sedimentation began when coarse clastics eroded from the mountainous region were distributed by streams as broad sheets upon the pre-Pocono erosion surface. As fluvial deposition continued, the mountain area was uplifted slightly from time to time. Near the middle of Pocono deposition the source was strongly rejuvenated, and the basal conglomerate of the Mount Carbon Member was deposited. During the remainder of Pocono time, the source area was eroded so that by the begin-

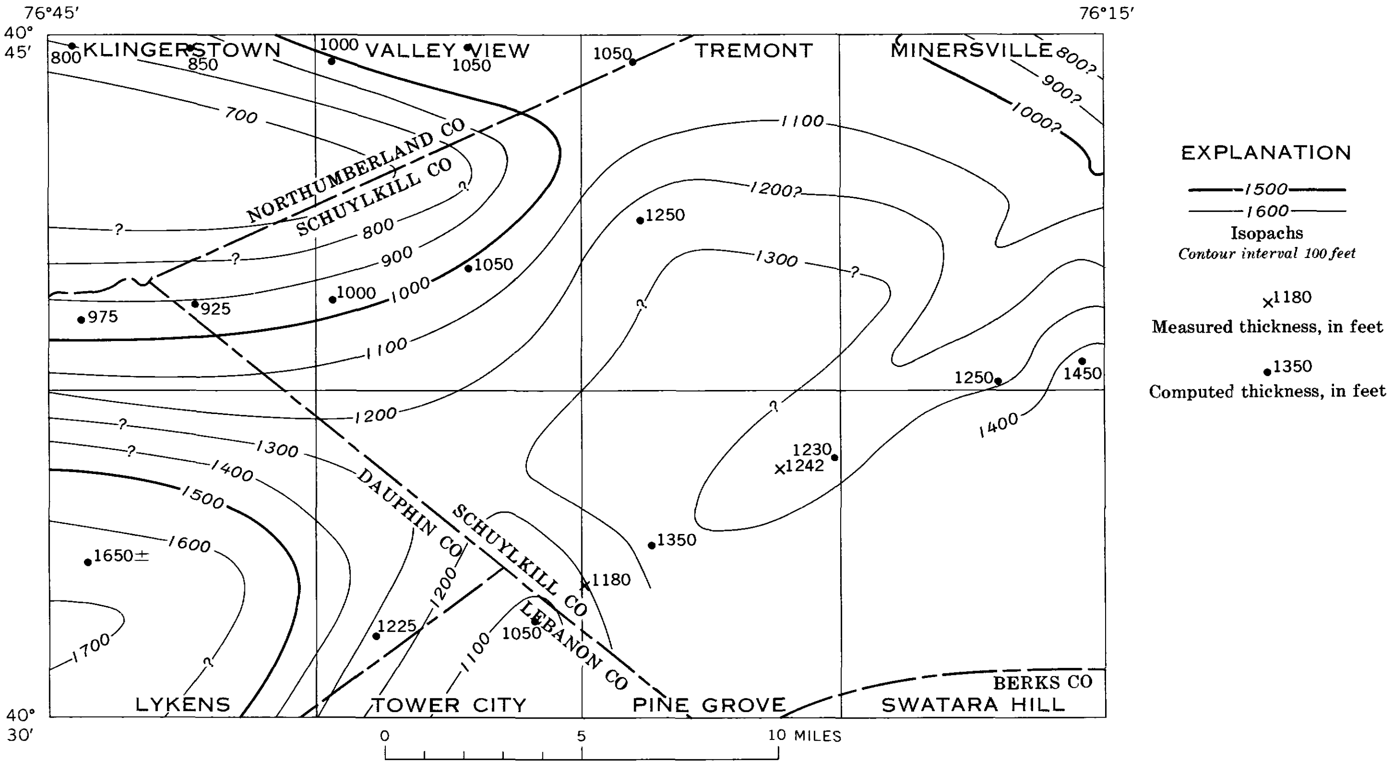


FIGURE 34.—Isopach map of the Pocono Formation.

ning of Mauch Chunk sedimentation it no longer contributed coarse detritus.

Thin lenticular coal beds, formed from the plant remains preserved in local swamps, are present at places. Because these coal beds are thin and lenticular, it seems likely that they were deposited in small shallow swamps. The coal beds and the gray color of the rocks of the Pocono indicate that reducing conditions prevailed during accumulation, which contrasts strongly with the general oxidizing conditions of Catskill and Mauch Chunk times.

Near Taylorsville, in the Minersville quadrangle (GQ–690), and in the northwestern part of the Klingerstown quadrangle (GQ–700), the pebbles in the rocks of the Pocono reach a maximum diameter of 2 inches and 1¼ inches, respectively. The maximum diameter increases gradually southeastward to 3 inches on Second Mountain (GQ–689, 690, 691, 698).

Pelletier (1958, fig. 16) used the southeastward increase in the maximum size of clasts in the Pocono at any particular locality as the principal evidence for concluding that the fall line of the Pocono basin of deposition lay about 25 miles northwest of Atlantic City, N.J. The method used to determine the location of the fall line was to plot a location graph transverse to the observed increase in maximum pebble and cobble diameter. He removed the effects of deformation by using values for crustal shortening that were calculated by Cloos (1940, p. 847). According to Pelletier (1958, p. 1056), 6-inch cobbles were probably the maximum size transported at those points where Pocono streams crossed the fall line into the basin of deposition.

Structural analyses made by the authors indicate that the amount of crustal shortening of Lower Mississippian rocks in the report area was about 25 percent, or 13 percent greater than the average calculated by Cloos. Recent structural studies in the Reading Prong and the Great Valley (A. A. Drake, Jr., oral commun., 1962) indicate that crustal shortening there was more than 50 percent, a minimum of 10 percent greater than the value calculated by Cloos.

Cobbles as much as 8 inches in diameter are rather common in the upper member of the Mauch Chunk Formation on Sharp Mountain, and cobbles of similar size are present in the Pottsville Formation in the southern part of the Southern Anthracite field. These 8-inch cobbles certainly did not accumulate above a fall line. Their occurrence in the Late Mississippian and Pennsylvanian basin of deposition argues strongly against using 6-inch cobbles to demarcate the Pocono fall line. Wherever observed by the authors, the present-day fall line of the Atlantic Coastal Plain is marked by boulders that greatly exceed the 6-inch limitation imposed by Pelletier. Much more complete data than are now available would be required to determine the average maximum size of these boulders. Casual observations by the authors, however, suggest that the average is between 10 and 12 inches. If stream velocities and gradients of the Pocono source area were similar to those of the present Piedmont, the Pocono fall line would have been marked by boulders of similar size.

Assuming that boulders as much as 10 to 12 inches in diameter marked the Pocono fall line, and using the values for crustal shortening determined in the area and those advocated by Drake to the south, the fall line lay about 40 to 50 miles southeast of the area. Six-inch cobbles, under this supposition, were deposited about 30 miles southeast of the area.

AGE AND CORRELATION OF THE POCONO FORMATION

The Pocono Formation is of Early Mississippian age as determined from plant fossils (Read, 1955, p. 9–15). It is correlative with the Price Sandstone, Big Stone Gap Member of the Chattanooga Shale, and Maccrady Shale of Virginia; with parts of the Chattanooga Shale, Fort Payne Chert, New Providence Shale, and Maury Formation of Alabama, Tennessee, and Georgia; and with the Knapp Formation, Berea Sandstone, Bedford Shale, and Cuyahoga Group of Western Pennsylvania and Ohio (Read, 1955; Weller and others, 1948, chart 5).

MISSISSIPPIAN AND PENNSYLVANIAN ROCKS

MAUCH CHUNK FORMATION

The Mauch Chunk Red Shale was named by Lesley (1876, p. 221–222, and chart opp. p. 224) in his description of the Boyd's Hill well near Pittsburgh in western Pennsylvania. The name was derived from the village of Mauch Chunk (now Jim Thorpe), in eastern Pennsylvania, near which the formation crops out. The unit continued to be designated as a red shale until Rothrock, Wagner, Haley, and Arndt (1953) classified it as a formation because of the large amounts of sandstone, siltstone, and conglomerate.

The U.S. Geological Survey recognizes the Mauch Chunk as a formation that includes the strata between the Pocono and Pottsville Formations. In this report the Mauch Chunk Formation is divided into three informal members.

The Mauch Chunk Formation is one of the principal "valley formers" of the area. From north to south it underlies the valleys and adjacent mountain slopes of Mahanoy Creek, Deep and Pine Creeks, Wiconisco, Clark Creeks, and the valley between Sharp and Second Mountains (GQ–689, 690, 691, 692, 698, 699, 700, 701). The lower and upper parts of the formation underlie moderate to steep slopes, are covered at many places by extensive deposits of talus and soil, and commonly are heavily forested. In contrast, the middle part generally underlies areas where the relief does not exceed 200 feet. Rocks in this part commonly are covered by a thin to thick mantle of soil and talus.

The stratigraphy of the Mauch Chunk has been studied more closely than that of any other rock unit, except the Pottsville and Llewellyn Formations. Despite this study, many facets of the stratigraphy are poorly understood or unknown because of structural complexities and the lack of stratigraphic markers in a thick monotonous red-bed sequence. In order to trace structural features and to discern stratigraphic changes, many 100- to 500-foot thick lithologic units were mapped in the Mauch Chunk. These units consist of resistant sequences of red sandstone interbedded with lesser amounts of red siltstone and shale, or nonresistant sequences of red siltstone and shale interbedded with lesser amounts of red sandstone. Most of the units are laterally persistent within a single belt of outcrop. This persistency aided greatly in the tracing of faults and folds, but because of soil cover at many localities, it was not of material aid in the understanding of stratigraphic changes.

LOWER MEMBER

The lower member of the Mauch Chunk Formation is sporadically exposed on the lower slopes of Mahantango, Berry, and Peters Mountains (GQ–692, 699, 700, 701), and is poorly exposed at several localities on Line Mountain (GQ–692, 699, 700), and at Lebanon Reservoir (GQ–691).

The contact of the member with the underlying Pocono Formation is placed at the base of the lowest red bed above gray beds of the Pocono. It is usually marked by a sharp topographic break. The top of the lower member is at the top of the uppermost gray, light-brown, or orange sandstone or conglomerate.

The thickness of the lower member is difficult to determine at most places because of cover and structural complications, but it ranges from 400 to 800 feet and probably averages about 600 feet. Available data are not sufficient to establish thickness trends.

In most localities the lower member consists, in ascending order, of a 200- to 500-foot unit of grayish-red and pale-brown shale, siltstone, and sandstone; a 10- to 75-foot unit of light- to medium-gray, light- to olive-gray, light-brown, and grayish-orange to dark-yellowish-orange sandstone and local conglomerate; a 200- to 300-foot unit of grayish-red and pale-brown shale, siltstone, and sandstone; and a 0 to 75-foot unit of sandstone and conglomerate similar to that of the underlying 10- to 75-foot unit. The gray-hued rocks of the member resemble those in the underlying Pocono

Formation, except that mica minerals are more abundant and rock fragments are less abundant.

In the few localities where beds of the lower member are well exposed they are generally tabular and are 2 inches to 2 feet thick. However, near Taylorville (GQ–692) the beds of the lower and upper gray sandstone units are 1 foot to 6 feet thick. Small-scale simple and planar crossbeds are present at a few places but are uncommon.

Rocks of the lower member consist of gray conglomerate and gray and red sandstone, siltstone, and shale. The red sandstone is mainly very fine to fine grained. The gray sandstone is chiefly medium grained, but it ranges from fine grained to conglomeratic and has vein quartz pebbles as much as 1 inch in diameter.

The mineral assemblages of the rocks of the lower member are similar and, regardless of grain size, are allied with the protoquartzite and subgraywacke clans. They consist principally of grains of common and vein quartz with fragments of quartzite, schist, shale, and chert; plates and irregular masses of biotite, muscovite, chlorite, sericite, and unidentified clay minerals; and small amounts of leucoxene, ilmenite, magnetite, zircon, tourmaline, and andesine. The red rocks contain considerable percentages of hematite and limonite as interstitial material and as films on grains of other minerals, whereas the gray rocks contain only small amounts of these minerals.

Cementing media of the red rocks are hematite, limonite, clay, and silica; in the gray rocks they are largely clay and silica. Sorting in both the red and gray rocks is poor to fairly good and averages fair, but the gray rocks are generally less well sorted. Rounding of grains for all rock types ranges from angular to well rounded and averages subangular.

MIDDLE MEMBER

The outcrop belts of the middle member are largely covered by talus and soil, and as a result, only the belt that underlies the valleys of Deep, Pine, and Mahanoy Creeks in the Lykens, Tower City, Klingerstown, Valley View, Tremont, and Minersville quadrangles (GQ–690, 692, 698, 699, 700, 701) was mapped in detail.

The basal contact of the middle member is at the top of the uppermost gray, light-brown, or orange sandstone or conglomerate bed of the lower member. The contact of the middle member with the upper member is arbitrarily placed at the base of the lowest gray conglomerate, sandstone, siltstone, or shale that resembles the overlying Pottsville Formation. The middle member locally intertongues with, and elsewhere grades laterally into, the upper member.

The wide range in apparent thickness of the middle member is largely due to low-angle thrust faults. Therefore, ideas concerning original thicknesses are only as reliable as the structural interpretation and the stratigraphic correlations and are not sufficient to establish trends. At the outcrop the member seems to be 2,000 feet to 6,000 feet thick and to average 5,000 feet. It is thinnest in the valley of Mahanoy Creek and in the valley between Sharp and Second Mountains and is thickest in Deep Creek and Pine Creek valleys. In the subsurface of the Minersville synclinorium, it may reach a structurally increased thickness of 7,000 feet, as is suggested by structural reconstructions and stratigraphic correlations.

Rocks in the middle member are chiefly grayish red, but a considerable number are pale red or pale brown. A few are light olive gray to olive gray, pale olive to grayish olive, greenish gray to dark greenish gray, brownish gray, dusky yellow green, and dusky red to moderate red.

Individual beds of the member are ⅛ inch to 16 feet thick and average about 1 foot 6 inches thick. Most seem to be tabular in small outcrops, but in larger outcrops wedge-shaped and lenticular beds are more common. Some siltstone and shale is fissile, but platy- and massive-bedded units are more numerous at most places. Small-scale simple and planar crossbedding is present locally and microcrossbedding and laminations are exceedingly common. Many microcross beds seem to have been deformed during deposition by slumping of partly consolidated material. Greenish-gray reduction spots and streaks are concentrated at many places along joints, crossbedding planes, and bedding planes. Mud cracks, ripple marks, and raindrop impressions are common in many siltstone and shale beds.

The middle member is composed of sandstone, siltstone, shale, and clay gall conglomerate, all of which are allied with the graywacke, subgraywacke, and protoquartzite rock clans. The red sandstone is predominantly very fine grained, and the gray sandstone is largely medium to coarse grained.

The rocks of the middle member are similar mineralogically, differing only in the percentages of the various constituents. The principal minerals are grains of common and vein quartz. Accessory minerals are fragments of quartsite, phyllite, slate, schist, and chert; variable amounts of grains and masses of unidentified clay minerals, biotite, chlorite, muscovite, sericite, leucoxene, ilmenite, magnetite, and specularite; and small amounts of grains of andesine, orthoclase, epidote, tourmaline, and zircon. Some red sandstone in the upper part of the member contains as much as 20 percent andesine. Cementing media are silica, limonite, hematite, unidentified clay minerals, sericite, and calcite.

Sorting ranges from poor to good but is somewhat better in protoquartzite than in subgraywacke and graywacke. Subangular to subround grains predominate in all rock types, but angular, rounded, and well rounded grains are not uncommon.

UPPER MEMBER

The upper member of the Mauch Chunk Formation is largely covered. Its belts of outcrop underlie the steeper slopes of Mahanoy, Broad, Bear, Short, Coal, Big Lick, Stony, and Sharp Mountains (GQ–689, 690, 691, 692, 698, 699, 701).

The lower contact of the upper member is placed at the base of the lowest gray conglomerate, sandstone, siltstone, or shale of Pottsville type lithology. It is commonly at or near a sharp topographic break in slope. The upper contact of the member is at the top of the uppermost red bed of Mauch Chunk-type lithology. The rocks bounded by these contacts constitute a transition zone (Smith, 1895, p. 1921; O. D. White, 1900, p. 763; Barrell, 1907, p. 453) where red rocks typical of the Mauch Chunk Formation alternate and tongue with gray rocks typical of the Pottsville Formation.

Detailed stratigraphic and mapping studies in the western part of the Anthracite region have demonstrated that beds of the upper member intertongue with, and grade laterally northward into, beds of the underlying middle member of the Mauch Chunk and southward into beds of the overlying Tumbling Run Member of the Pottsville Formation. These stratigraphic relations are difficult to demonstrate conclusively in the area because of cover and because the outcrop belts of the members strike parallel or subparallel to their depositional strike. Regionaly, however, the coarse-grained gray rocks of the Tumbling Run Member on the south margin of the report area tongue locally with, and grade laterally northward into, gray and red rocks of the upper member in the Western Middle and Eastern Middle Anthracite fields. They then seem to tongue and grade into the upper red beds of the middle member on the south margin of the Northern Anthracite field. Thus, it seems that the Tumbling Run of the area is equivalent to the upper part of the middle member of the Mauch Chunk on the south margin of the Northern Anthracite field.

The thickness of the upper member is difficult to determine because of cover, because the basal beds are truncated by faulting at many places, and because the member tongues with the underlying middle member and the overlying Pottsville Formation. The apparent thickness ranges from 0 to 900 feet and averages about 600 feet. Reliable thickness data are in sufficient for the member to be isopached or its thickness trends determined.

Rocks of the upper member consist of a Mauch Chunk- and a Pottsville-type facies. The Mauch Chunk-type facies consists of conglomerate, sandstone, siltstone, and shale that is predominantly grayish red and, to a lesser extent, pale red, pale reddish brown, and pale to moderate brown. The Pottsville-type facies includes conglomerate, sandstone, siltstone, and shale that is medium to dark gray, light to olive gray, pale to grayish olive, greenish gray, yellowish gray, brownish gray, dusky yellow, dusky yellowish green, and pale yellowish brown.

Most of the finer grained rocks of the upper member are tabular bedded, but about 50 percent of the coarser grained rocks are wedge bedded, and the remainder are tabular and lenticular bedded. Although the beds range in thickness from 1/8 inch to 30 feet, most beds are less than 5 feet thick and average about 2 feet. The coarser grained sandstone and conglomerate are commonly crossbedded with small- to medium-scale simple and planar cross-strata. Some beds are graded and others are reverse graded, but most are not graded. Many finer grained rocks are ripple marked and contain macerated plant fragments.

The rock composition of both the Pottsville- and Mauch Chunk-type focus is protoquartzite and subgraywacke, regardless of grain size. Sandstone of the Mauch Chunk-type facies is largely very fine to fine grained and that of the Pottsville-type facies is generally fine to coarse grained.

Generally, most of the clasts in the conglomerate of the upper member are about three-quarters of an inch in diameter, but cobbles as much as 3 inches in diameter are present at many places and 8-inch cobbles are present along Sharp Mountain. The average diameter decreases northwestward from a maximum 2½ inches on Sharp Mountain to a minimum of three-quarters of an inch in Mahanoy Valley. More than 50 percent of the clasts are vein quartz; the remainder, in decreasing order of abundance, are quartzite, chert, schist, phyllite, slate, sandstone, shale, and gneiss. Much of the conglomerate is gray and is similar to that in the overlying Pottsville Formation. However, grayish-red to pale-brown conglomerate is present at many localities. Beds of both the gray and grayish-red conglomerate commonly contain irregularly shaped granule- to cobble-sized fragments of red shale, siltstone, and sandstone in their basal parts. These fragments clearly indicate that the conglomerate beds accumulated near previously deposited finer grained red rocks which were being eroded or cannibalized.

The rocks of the upper member are composed principally of common and vein quartz and quartzite. Accessory constituents are fragments of chert, schist, slate, phyllite, shale, and sandstone; grains and irregular masses of biotite, muscovite, sericite, chlorite, leucoxene, magnetite, ilmenite, and unidentified clay minerals; and small amounts of andesine, epidote, tourmaline, rutile, sillimanite, sphene, and zircon.

The cementing media in the rocks of the Pottsville-type facies are silica, sericite, unidentified clay minerals, and calcite. The same cements are present in the Mauch Chunk-type facies, but hematite and limonite predominate. Sorting in both facies ranges from poor to good, but in general, it is better in rocks of the Mauch Chunk-type facies. The degree of rounding in either facies is angular to rounded, subangular and subround grains dominating. Quartz grains and quartzite fragments in the coarser grained sandstone and in the matrix of the conglomerate of the Pottsville-type facies commonly are deeply sutured; whereas in the finer grained rocks of both facies, the amount of mica and unidentified clay minerals is greater and the amount of suturing is much less. This fact suggests either that suturing is related to the amount of mica and clay separating the quartz and quartzite grains or to a more even distribution of pressure by the fine material over the surface area of each grain.

SECTIONS OF THE MAUCH CHUNK FORMATION

The outcrop belts of the Mauch Chunk Formation are so poorly exposed in the area that it was impractical to measure sections. Much of the formation, however, is well exposed at two localities a short distance to the east of the area in the Pottsville 7½-minute quadrangle. Because the thickness and lithology of the formation in the southwestern part of the Anthracite region have not been well documented previously, sections were measured at these places. The sections are on the east side of West Branch Gap and the east side of Schuylkill Gap (fig. 35, sections 1, 2).

DEPOSITION OF THE MAUCH CHUNK FORMATION

The Mauch Chunk Formation accumulated in a subsiding asymmetric basin, probably as flood-plain deposits. The thickest part of the formation in this basin is in and near the report area (Rogers, 1858b, p. 830–831; Barrell, 1907, p. 451).

Rogers (1858b, p. 794) believed that Mauch Chunk sediments accumulated in a sea that was foul with poisonous sediments; that the ripple marks, mud cracks, and raindrop impressions indicated nearby land; and that the sea frequently regressed, leaving the sediments exposed to terrestrial conditions. About a half century later, Lesley, D'Invilliers, and others (1892–95, p. 1806–1807) stated that these sediments were laid down on a broad shore-bordered lowland occupied by marshes, pools, and lagoons.

Barrell (1907, p. 474–475) concluded that the Mauch Chunk was deposited upon a great well-drained delta inundated by seasonal rains. He believed that animal and plant life flourished during the rainy seasons and either retreated from the delta or withered and died there during dry seasons. He also postulated that sandstone was laid down during wet seasons when streamflow was greatest and that shale was deposited during dry seasons when streamflow was minimal.

The stratigraphy and lithology of the Mauch Chunk in the area suggest that the rocks of the formation accumulated as flood plain deposits upon a broad well drained coastal plain which was not part of a delta. The abundance of mud cracks, animal tracks, raindrop impressions, and plant debris in all types of rocks indicates that the coastal plain was generally suitable for animal and plant life and was not subjected to great seasonal variations of wet and dry weather, which is inconsistent with Barrell's correlation of sandstone with wet seasons and shale with dry seasons.

The red color of the rocks of the Mauch Chunk probably was primary, and the coastal plain where they were deposited probably was well drained as suggested by Barrell. It seems likely that a poorly drained plain would have enhanced the reduction of the abundant iron oxide in the rocks of the formation and would have produced gray rocks. In this context, the gray beds in the lower and upper members probably record periods when the drainage was restricted and reducing conditions prevailed.

The sediments in the lower and middle members were eroded from the stumps of the Acadian Mountains and were transported by streams northwestward to the coastal plain where they were deposited. During deposition the mountains were uplifted slightly several times. The coarse-grained gray beds of the lower member record these uplifts and also record periods when drainage on the coastal plain was restricted and reducing conditions prevailed.

Near the middle of Mauch Chunk time, an andesitic or a dioritic mass that contained andesine was exposed in the source area. The quantity of andesine indicates that the extent of this mass was probably much greater than that of any other mass of similar composition that had been exposed previously.

Calcite as cementing and interstitial material also appeared suddenly during middle and late Mauch Chunk time. This calcite indicates that a small part of the source area may have been underlain by limestone, dolomite, or marble. Although it is conceivable that the calcite was derived from rocks other than these, it is unlikely because the mineral assemblages of all other upper Paleozoic sedimentary rocks in the area lack

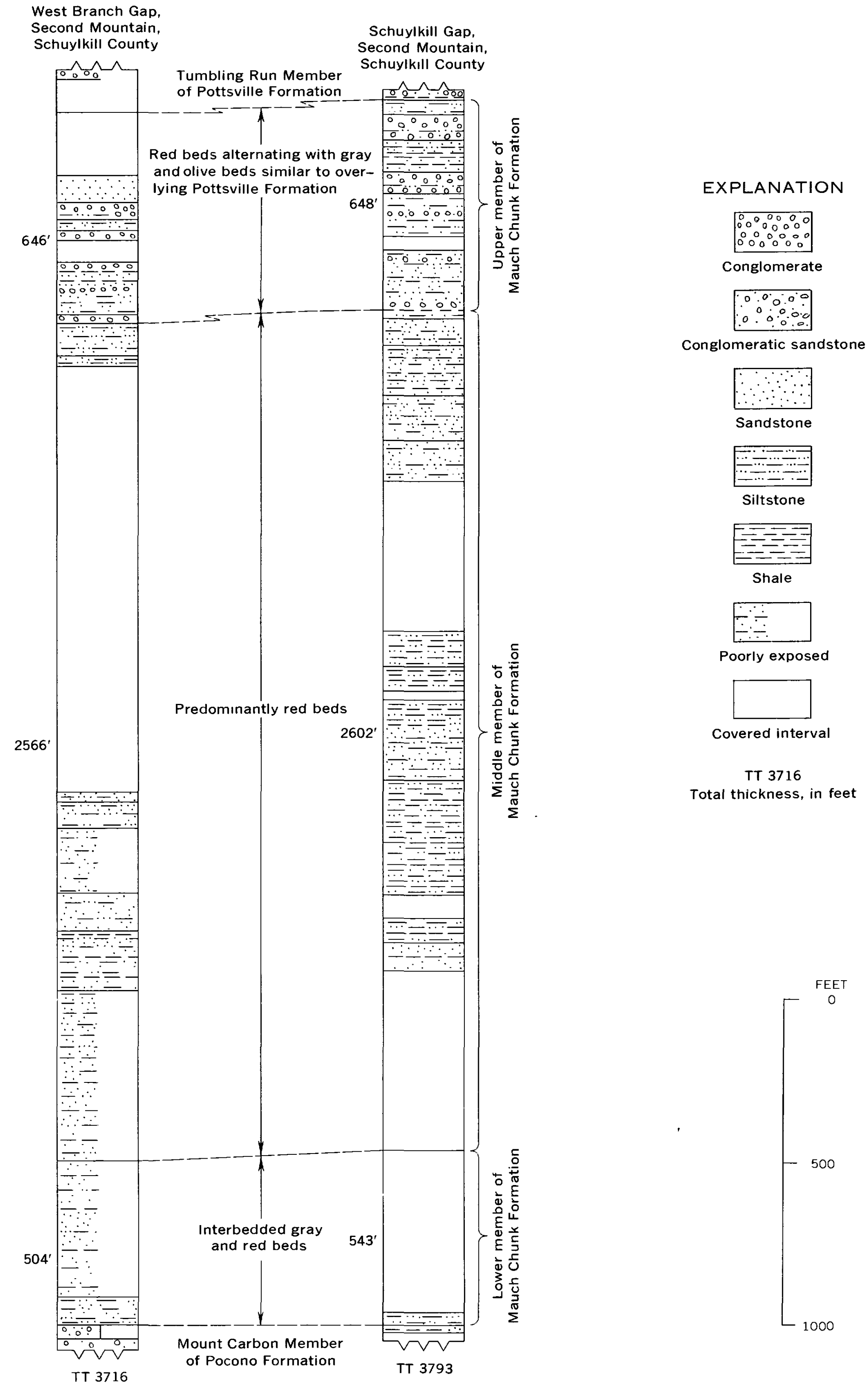


FIGURE 35.—Columnar sections of the Mauch Chunk Formation. See figure 31 for location of sections.

calcite but are otherwise similar to these calcite-bearing rocks.

During late Mauch Chunk time, a large region east and southeast of the area was uplifted by the first pulsation of the Appalachian orogeny. Abundant pebbles and cobbles of vein quartz, quartzite, mica, schist, phyllite, slate, and gneiss in the upper member show that the uplifted region was underlain principally by low-rank metamorphic rocks. Other pebbles and cobbles of sandstone, shale, and siltstone, however, indicate that a part was underlain by sedimentary rocks; common or igneous quartz fragments suggest that a small part was underlain by igneous rocks; and the presence of traces of sillimanite indicates that high-rank metamorphic rocks were locally exposed.

AGE OF THE MAUCH CHUNK FORMATION

The Mauch Chunk Formation in the area is of Late Mississippian and Early Pennsylvanian age. The Late Mississippian age of the bulk of the unit has been known and accepted for many years, but the Early Pennsylvanian age of the upper beds was recognized only recently (Wood and others, 1962, p. C40–C41).

The contact between the Mauch Chunk and the overlying Pottsville Formation at the type section of the latter unit is also the type locality where the time boundary between the Mississippian and Pennsylvanian periods was defined (Wood, and others, 1956, p. 2670–2671; Moore and others, 1944, p. 665). The type section is a short distance east of the report area at Schuylkill Gap in Sharp Mountain on the south edge of the Southern Anthracite field. The upper member of the Mauch Chunk intertongues with the Tumbling Run Member of the Pottsville, and the contact rises stratigraphically northward. Thus, at the type section of the Pottsville at Schuylkill Gap, the Pennsylvanian and Mississippian time boundary lies at the contact between the members, but in the northern part of the area, the time boundary lies at an indeterminate position in the upper member of the Mauch Chunk because of the northward rise.

The Mauch Chunk correlates not only with the lower part of the Pottsville Formation but also with many other Upper Mississippian and Lower Pennsylvanian units. It is correlative with parts or all of the Greenbrier Limestone of western Pennsylvania, Maryland, Virginia, and West Virginia; with the Bluefield Shale, Hinton Formation, Bluestone Formation, and Princeton Sandstone of West Virginia; with the Glen Dean Limestone, Pennington Formation, and Bluestone Formation of Virginia; and perhaps with the basal parts of the Lee and Pottsville Formations of western Pennsylvania, West Virginia, Virginia, Kentucky, and Tennessee (Weller and others, 1948, chart 5).

PENNSYLVANIAN SYSTEM

Pennsylvanian rocks in the area consist of an undetermined part of the Mauch Chunk Formation, probably not more than 200 feet; the Pottsville Formation, 800 to 1,500 feet thick; and the Llewellyn Formation, as much as 3,500 feet thick. The Pennsylvanian sequence ranges in total thickness from about 4,300 to 5,000 feet.

The Pottsville Formation of Early and Middle Pennsylvanian age consists, in ascending order, of the Tumbling Run, Schuylkill, and Sharp Mountain Members. Each member is composed of conglomerate, conglomeratic sandstone, sandstone, siltstone, shale, and anthracite.

The Llewellyn Formation of uppermost Middle and Late Pennsylvanian age consists of sandstone, siltstone, and shale; numerous thin to thick beds of anthracite; and a considerable amount of conglomerate and conglomeratic sandstone.

POTTSVILLE FORMATION

The Pottsville was named by Lesley in 1876 (p. 221–227) in the description of the Boyd's Hill well near Pittsburgh in western Pennsylvania. It was defined in this description as a conglomerate overlying the Mauch Chunk Red Shale of Lesley and underlying the Lower Productive Coal Measures (Allegheny Formation of modern usage).

C. D. White in 1900 (p. 755–756) ignored the existence of the original type section in the Boyd's Hill well and stated that the type section of the Pottsville, as commonly recognized by previous workers, lay south of the city of Pottsville in Schuylkill Gap. He measured a section on the east side of the gap along the tracks of the Pennsylvania Railroad and partitioned the formation into four paleobotanic divisions (p. 773–775). Since 1900, White's section has been considered the type for the formation. Recently, a better exposed reference section was established about 150 feet east of the type section on the eastern side of Pennsylvania Route 61 (Wood and others, 1956, p. 2671–2673).

The Pottsville has been classified as a conglomerate, a sandstone, a formation, a group, and a series. In the Anthracite region, the U.S. Geological Survey classifies it as a formation, which is divisible at many places into the Tumbling Run, Schuylkill, and Sharp Mountain Members.

TUMBLING RUN MEMBER

The lowest division of the Pottsville Formation in the Anthracite region was named the Tumbling Run Member in 1956 for a small stream southeast of Schuylkill Gap (Wood and others, 1956, p. 2671). It includes the Lower Lykens Division and the lower part of the Lower

Intermediate Division of C. D. White (1900, p. 773–775).

The Tumbling Run Member crops out on the slopes and crests of Bear, Big Lick, Broad, Coal, Little, Mahanoy, Sharp, Sherman, Short, and Stony Mountains (GQ–689, 690, 691, 692, 698, 699, 701). Slopes underlain by the members are commonly steep and rough and are generally covered by a thick mantle of talus; in contrast, where the member underlies the crests of the mountains, the surface is smooth and featureless and is characterized by small relief. All belts of outcrop are overgrown with a dense scrub forest.

The lower contact of the Tumbling Run Member is at the top of the uppermost red bed of the Mauch Chunk Formation. The upper contact is placed at the bottom of the basal pebble conglomerate of the Schuylkill Member. The rocks in the upper part of the Tumbling Run are generally finer grained than the overlying basal conglomerate and are light to dark gray, olive gray, greenish gray, grayish orange, moderate yellowish brown, and dusky yellow, whereas the overlying basal conglomerate is generally light to dark gray. Pebbles in the upper part of the Tumbling Run are sparse, are scattered at random in a sandstone matrix, are as much as 1½ inches in diameter, commonly are fractured and generally are rounded. They consist of a variety of rock types, such as vein quartz, quartzite, chert, schist, gneiss, sandstone, siltstone, and shale. The clasts in the basal conglomerate of the Schuylkill Member are more abundant, are evenly distributed in a sandstone matrix, are as much as 3 inches in diameter, are not fractured, and are generally well rounded. They are composed principally of vein quartz and quartzite.

It is difficult to determine the thickness of the Tumbling Run Member at most places because of structural complications, a dense forest cover, and a mantle of talus. Plates 1–3 show three lines of stratigraphic sections that cross the area from west to east. These lines illustrate the variations in lithology and thickness of parts or all of the Pottsville Formation and the lower and middle parts of the Llewellyn Formation as measured in boreholes, in mines, and on the surface. Twenty-three of these stratigraphic sections were measured, partly in beds of the Tumbling Run.

The Tumbling Run Member ranges in thickness from about 275 feet in the western and northwestern parts of the area to about 600 feet in the southeastern part (fig. 36). It thickens irregularly to the south and southeast at 25 to 30 feet per mile. Structural calculations indicate that the Pennsylvanian rocks of the area were telescoped about 45 percent during the Appalachian orogeny. Before deformation the Tumbling Run thickened southward and southeastward at about 14 to 17 feet per mile.

Fresh and weathered rocks of the member exhibit a considerable range in color. Conglomerate, conglom-

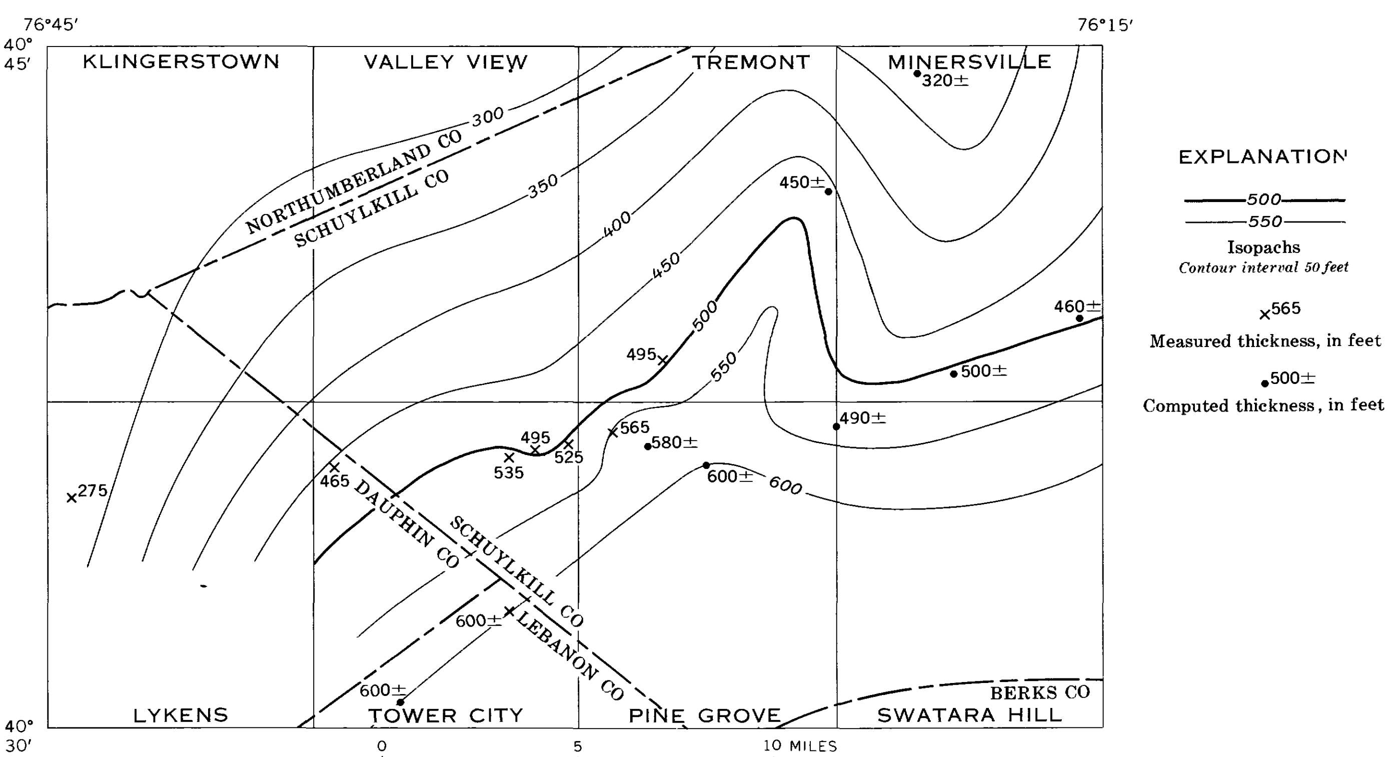


FIGURE 36.—Isopach map of the Tumbling Run Member of the Pottsville Formation.

eratic sandstone, and sandstone are light to dark gray, light to olive gray, yellowish gray, dusky yellow, greenish to dark greenish gray, grayish orange, yellowish brown, and moderate olive brown. Shale and siltstone are medium to dark gray, light to olive gray, dusky yellow, greenish gray, and black.

Conglomerate and conglomeratic sandstone compose about 55 percent of the Tumbling Run and are concentrated largely in the lower and middle parts. The cobbles and pebbles in these parts of the member have an average diameter of about 1 inch, but locally are as much as 8 inches in diameter. In the upper part of the member, the average diameter is about ½ inch, and the maximum is about 1½ inch. Cobbles and pebbles throughout the member are composed largely of vein quartz and quartzite and lesser amounts of chert, schist, gneiss, sandstone, shale, siltstone, slate, phyllite, and granite. Most are subrounded to well rounded and average subrounded to rounded.

The conglomerate of the Tumbling Run generally is poorly sorted. The cobbles and pebbles commonly are scattered in the matrix and exhibit no apparent lineation, other than the more ovoid ones being orientated parallel to bedding surfaces. The matrix is largely moderate olive brown, greenish gray, and olive gray, which contrasts with the light-to dark-gray matrix of the conglomerate of the Schuylkill and Sharp Mountain Members.

At natural exposures many cobbles and pebbles separate from the matrix of weathered conglomerate in the Tumbling Run Member in a broken or fractured state. At manmade exposures in roadcuts and strip-pit faces unfractured cobbles and pebbles separate from the same conglomerate. Cobbles and pebbles of the Schuylkill and Sharp Mountain Members are rarely fractured at weathered exposures. The general fractured condition of the clasts at weathered outcrops of the Tumbling Run, therefore, provides one of the more reliable means of distinguishing conglomerate of this member from that of the two overlying members. This difference in fracturing of clasts suggests that the conglomerate of the Tumbling Run was more firmly cemented than were the conglomerates of the other members. When deformation took place, incipient fractures developed in the clasts of the more firmly cemented Tumbling Run rocks with the result that weathering cause them to disintegrate. In the less firmly cemented rocks of the other members, adjustments in the cement protected the clasts from incipient fracturing.

Fine- to coarse-grained sandstone composes about 30 percent of the Tumbling Run, but is largely concentrated in the upper part. Most sandstone beds are 2 to 15 feet thick, an appreciable number being 1 inch to 1½ feet thick. Tabular bedding predominates over all other types, but lenticular and wedge bedding are common. Small- to medium-scale simple and planar crossbedding is prevalent in all bedding types.

Shale and siltstone compose about 15 percent of the Tumbling Run. Many siltstone and shale beds are ½ inch to 2 feet thick, the remainder being 2 to 20 feet thick. Most are tabular, but lenticular and wedge-shaped beds are common. Although some of the shale and siltstone is cross laminated, most is parallel laminated.

Six beds of anthracite are present in the member in the area. In ascending order, these are the Lykens Valley Nos. 7, 6, 5, 4, 3⅞, and 3¾ coal beds. Three of these, the Lykens Valley Nos. 6, 5, and 4 coal beds are persistent and have been traced throughout the area, except for the easternmost part (Wood and Trexler, 1968b, c). The others are less persistent and are confined to the western part of the area on the limbs of the Minersville synclinorium. On the north limb of the north trough of the synclinorium, however, they are too thin to be shown on the coal maps (Wood and Trexler, 1968b). The lowermost of the less persistent beds, the Lykens Valley No. 7, lies 50 to 110 feet below the Lykens Valley No. 6 coal bed and 50 to 210 feet above the base of the member. At several localities, the Lykens Valley No. 3¾ coal bed lies directly beneath or within 10 to 15 feet of the bottom of the basal conglomerate of the Schuylkill Member (pl. 1, secs. 7, 11, 14) and at one locality the Lykens Valley No. 3⅞ coal bed lies directly beneath the conglomerate (pl. 1, sec. 3).

Most of the conglomerate, sandstone, siltstone, and shale of the Tumbling Run is compositionally subgraywacke but some beds are orthoquartzite. The constituents of these rocks are similar, regardless of grain size, and differ only in the percentages of the various minerals present.

The mineral assemblage of the subgraywacke consists of grains of common and vein quartz, ilmenite, magnetite, tourmaline, zircon, sphene, and leucoxene; rock fragments of chert, schist, phyllite, slate, shale, sandstone, and quartzite; extremely small amounts of andesine, oligoclase, kyanite, staurolite, garnet, epidote, sillimanite, limonite and hematite; plates and irregular masses of biotite and muscovite; irregular masses and blebs of chlorite, sericite, and unidentified clay minerals.

Cementing media are silica, sericite, unidentified clay minerals, and a clastic binder of clay- and silt-sized quartz fragments, and films of hematite, and limonite. Most very coarse grains are rounded to well rounded. Very fine to coarse grains range from angular to rounded and average subangular to subrounded.

SCHUYLKILL MEMBER

The Schuylkill Member, medial division of the Pottsville Formation, was named in 1956 (Wood and others, p. 2671) for the Schuylkill River. The member includes the upper part of the Lower Intermediate Division and all the Upper Lykens Division of C. D. White (1900, p. 794–799).

The Schuylkill crops out in a long sinuous belt underlying the crests or upper slopes of Bear, Big Lick, Broad, Coal, Little, Mahanoy, Sharp, Sherman, Short, and Stony Mountains and Mine Hill (GQ–689, 690, 691, 692, 698, 699, 701). The surface of most of this belt is smooth; but in a few places it is rough and steep, and locally, it is mantled by talus and covered by a heavy forest.

The basal conglomerate of the Schuylkill conformably overlies finer grained rocks of the Tumbling Run Member.

The upper contact of the Schuylkill is at the bottom of the basal conglomerate of the Sharp Mountain Member which rests conformably upon finer grained rocks in the upper part of the Schuylkill. The pebbles and cobbles in the basal conglomerate of the Sharp Mountain Member are much coarser than those in the underlying Schuylkill. This characteristic coarseness and the resistance of this conglomerate to erosion aid greatly in distinguishing the boundary between these members.

Plates 1–3 illustrate three lines of stratigraphic sections in and adjacent to the area. These lines show the variations in lithology and thickness of the Pottsville and Llewellyn Formations. Of the 46 illustrated sections, 24 include part or all of the Schuylkill Member.

The Schuylkill is about 300 feet thick in the northeastern part of the area on Broad Mountain and is about 700 feet thick in the southwestern part near Lykens and Williamstown (fig. 37). The abundant thickness data (pls. 1–3; fig. 37) indicate that the member thickens southward at 30 to 60 feet per mile. Much structural information also indicates that the Pennsylvanian rocks of the area were shortened an average of about 45 percent during the Appalachian orogeny. Thus, before the orogeny, the Schuylkill Member thickened southward at about 16½ to 33 feet per mile.

Weathered and fresh rocks of the member are light to dark gray, the average being medium gray. At fresh manmade outcrops, many rocks are lightly stained by limonite. This staining is absent at most natural exposures.

Most of the beds in the Schuylkill are tabular, but a few are wedge shaped and irregular. Parallel-laminated beds predominate over cross-laminated beds. The cross-stratification is in general simple and planar and ranges from small to large. Most conglomerate and sandstone beds are 2 to 25 feet thick, and most shale

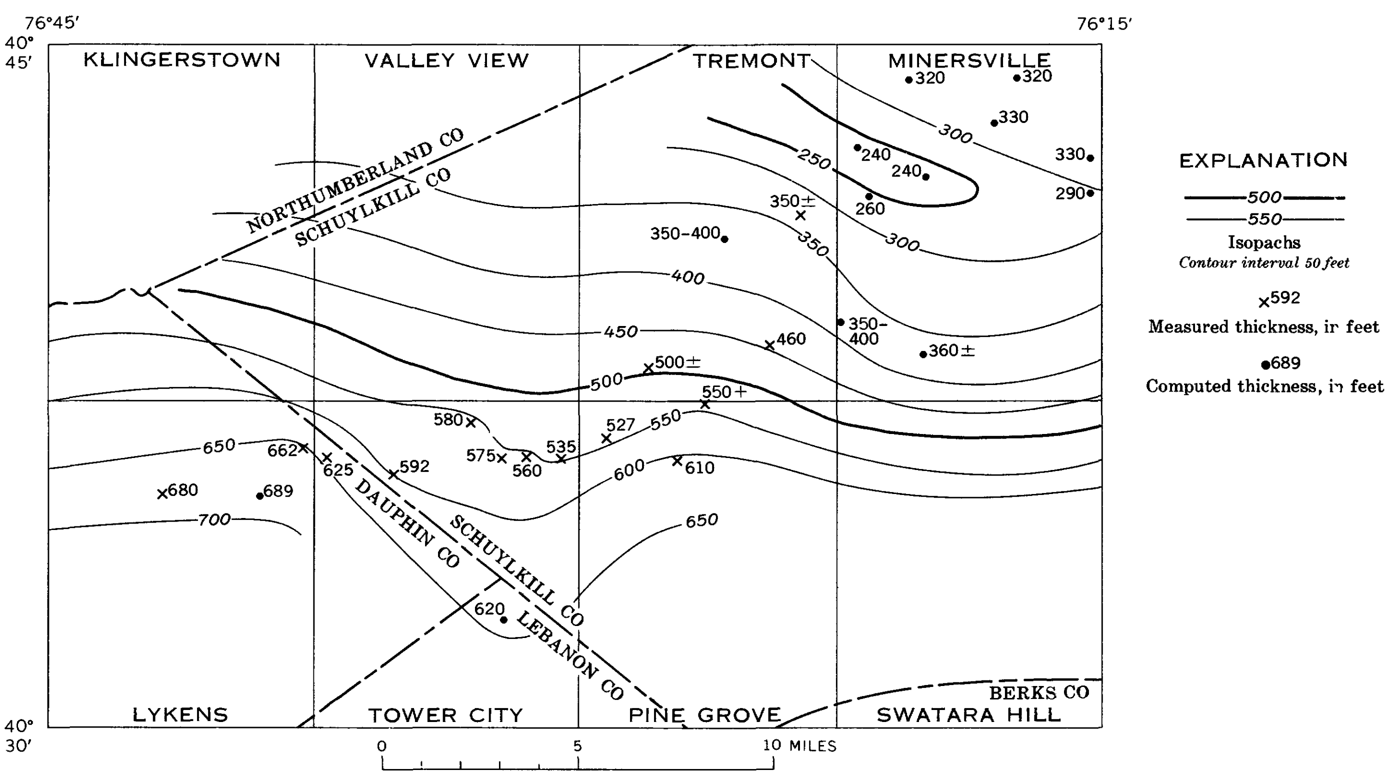


FIGURE 37.—Isopach map of the Schuylkill Member of the Pottsville Formation.

and siltstone beds are ½ inch to 1¾ feet thick. Graded and reverse graded beds are relatively common in the coarser grained rocks. Carbonized plant fragments are abundant in many shale units and are relatively common on bedding planes in the coarser grained rocks.

Conglomerate and conglomeratic sandstone compose about 50 percent of the member. The pebbles in these rocks are composed largely of vein quartz and quartzite. Chert is the next most common constituent. Schist, phyllite, gneiss, sandstone, shale, siltstone, and granite pebbles are present, but in general, are sparse. Most of these clasts are well rounded and equidimensional to ovoid. They are generally scattered evenly throughout a light- to dark-gray matrix consisting of very fine to very coarse grains of quartz, quartzite fragments, other rock fragments, and accessory mineral grains. The largest clasts are usually concentrated in the basal conglomerate of the member, where they locally are as much as 3 inches in diameter. Pebbles in the overlying conglomerate beds rarely exceed 1 inch in diameter, and average ¼ to ½ inch.

Much of the conglomerate and conglomeratic sandstone in the Schuylkill weathers into a pebble gravel that is distinctive because of the small diameter, the well-rounded shape, and the unfractured condition of the pebbles.

Sorting in the conglomerate and conglomeratic sandstone of the member is poor to moderate. Lineation at most localities is nil, except that the long axis of the more elongate and ovoid pebbles commonly parallels the bedding. Cementation in these rocks is weak to quartzitic but generally is moderate.

The sandstone in the Schuylkill ranges from very fine to very coarse grained and averages medium to coarse grained. It constitutes about 30 percent of the member and is distributed evenly throughout. Most sandstone beds are 2 to 15 feet thick, but some are as thin as ½ inch. Graded bedding is more common in sandstone than in the finer or coarser grained rocks. Sorting ranges from moderate to excellent and averages good. Cementation generally is moderate but ranges from weak to quartzitic.

Four relatively persistent beds of anthracite are present in the member. In ascending order, these are the Lykens Valley Nos. 3, 2, 1½, and 1 coal beds (Wood and Trexler, 1968b, c). Several thin and nonpersistent coal beds locally lie below the Lykens Valley No. 3, and others are present between the Lykens Valley Nos. 2 and 1½ coal beds. Only one of these, the Lykens Valley No. 3½ coal bed, underlies an area large enough to warrant naming.

The Lykens Valley Nos. 3, 2, 1½, and 1 coal beds are recognized at many places in the area. The No. 3 and No. 1 beds underlie most of the part of the Southern Anthracite field that is within the report area. The No. 1 bed is generally thin. However, it and the underlying Lykens Valley No. 4 bed of the Tumbling Run Member are perhaps the most widespread of the coal beds of the Pottsville Formation. The consistent stratigraphic position of the No. 1 bed at 5 to 50 feet below the basal conglomerate of the Sharp Mountain Member supports the concept that the Schuylkill and Sharp Mountain are conformable.

The Lykens Valley No. 2 and No. 1½ coal beds thin rapidly east of the latitude of Sherman and Little Mountains and are not recognized in the eastern part of the area.

The rocks of the member are compositionally protoquartzite and subgraywacke regardless of differences in grain size. The principal mineral constituents of both types of rock are grains of vein and common quartz and fragments of quartzite. Accessory minerals are varying amounts of ilmenite, magnetite, zircon, sphene, tourmaline, rutile, and leucoxene; traces of andesine and oliogoclase; varying amounts of fragments of chert, schist, phyllite, slate, siltstone, and shale; irregular masses and blebs of sericite, chlorite, biotite, muscovite, hematite, limonite, and unidentified clay minerals; and varying amounts of clay- to silt-sized quartz fragments. Cementing media are silica, sericite, clay minerals, a binder matrix of clay- to silt-sized quartz fragments, and films of hematite and limonite. Most of the coarse to very coarse grains are well rounded, whereas the very fine to medium grains are generally subrounded. The margins of many quartz grains and quartzite fragments have been deeply sutured by pressure solution.

SHARP MOUNTAIN MEMBER

The Sharp Mountain Member, youngest division of the Pottsville Formation, was named in 1956 (Wood and others, p. 2671) for Sharp Mountain at Schuylkill Gap. C. D. White (1900, p. 801) assigned the beds included in this member to his Upper Intermediate Division of the Pottsville. He and Read (in Moore and others, 1944, p. 680) recognized the persistence of the basal conglomerate of this member throughout the Southern Anthracite field.

The member crops out at or near the crests of Bear, Big Lick, Broad, Coal, Little, Sharp, Short, and Stony Mountains and Mine Hill (GQ–689, 690, 691, 692, 698, 699, 701). It is the principal "ridge former" in the part of the coal field that is in the area. At most places it underlies cuestas or hogbacks covered by dense forests. Elsewhere, as on Broad Mountain, the member underlies a smooth rather featureless terrain.

The contact between the Sharp Mountain and Schuylkill Members is at the bottom of the basal conglomerate of the Sharp Mountain Member, which rests conformably upon conglomeratic sandstone, sandstone, siltstone, and shale in the upper part of the Schuylkill. The basal conglomerate commonly weathers into a ledge, a series of ledges, or a stone wall that rises above the finer grained beds of the Schuylkill. In some localities it weathers into a gravel that mantles the surrounding terrain.

The upper contact of the Sharp Mountain with the Llewellyn Formation is at the base of the underclay or shale that underlies the Buck Mountain (No. 5) coal bed. Where the shale is absent, the contact is at the base of the coal bed.

The Sharp Mountain Member is the most distinctive stratigraphic unit in the Pottsville and Llewellyn Formations. Its widespread distribution and characteristic conglomerate lithology are helpful in determining the base of the Llewellyn and in delineating structural features that cross the outcrop belt.

Plates 1–3 illustrate three lines of stratigraphic sections within and adjacent to the area. These lines of sections show 46 localities where the thickness and lithology of the Pennsylvanian rock sequence has been measured and studied. Of the 46 sections, 31 include part or all of the Sharp Mountain Member. The member is thickest, 315± feet, in the vicinity of Tower City No. 1 Tunnel (pl. 1, section 10; Wood and Trexler, 1968b, sheet 4) and is thinnest, 120± feet, north of Colket Water Level Tunnel (pl. 1, section 17; Wood and Trexler, 1968b, sheet 3). The abundant thickness data indicate that the member thins northwestward at about 40 feet per mile in the northeastern part of the area, at 50 feet per mile in the central part, and at about 20 feet per mile in the southwestern part (fig. 38). Before the crustal shortening resulting from the Appalachian orogeny, which is estimated to have been about 45 percent, the member thinned northwestward at 22 feet, 27.5 feet, and 11 feet per mile, respectively.

The beds of the Sharp Mountain Member accumulated in a series of north- and northwest-trending thick and thin lobes. The conglomerate in the thick lobes contains cobbles as much as 8 inches in diameter, whereas the cobbles in the thin lobes reach a maximum diameter of about 5 inches (fig. 39). The greater size of the cobbles in the thick lobes suggests that these lobes were the sites of large alluvial fans formed by streams capable of transporting much larger and heavier detritus than the streams that deposited the material in the thin lobes. The thin lobes, in fact, may represent interfan alluviation from small streams.

Fresh and weathered rocks in the Sharp Mountain Member are principally medium gray, but they range from light to dark gray.

Tabular and lenticular beds are about equally abundant in the member. Wedge-shaped and irregular beds are uncommon. Most strata are cross laminated, but many are parallel laminated. The attitudes of the small-

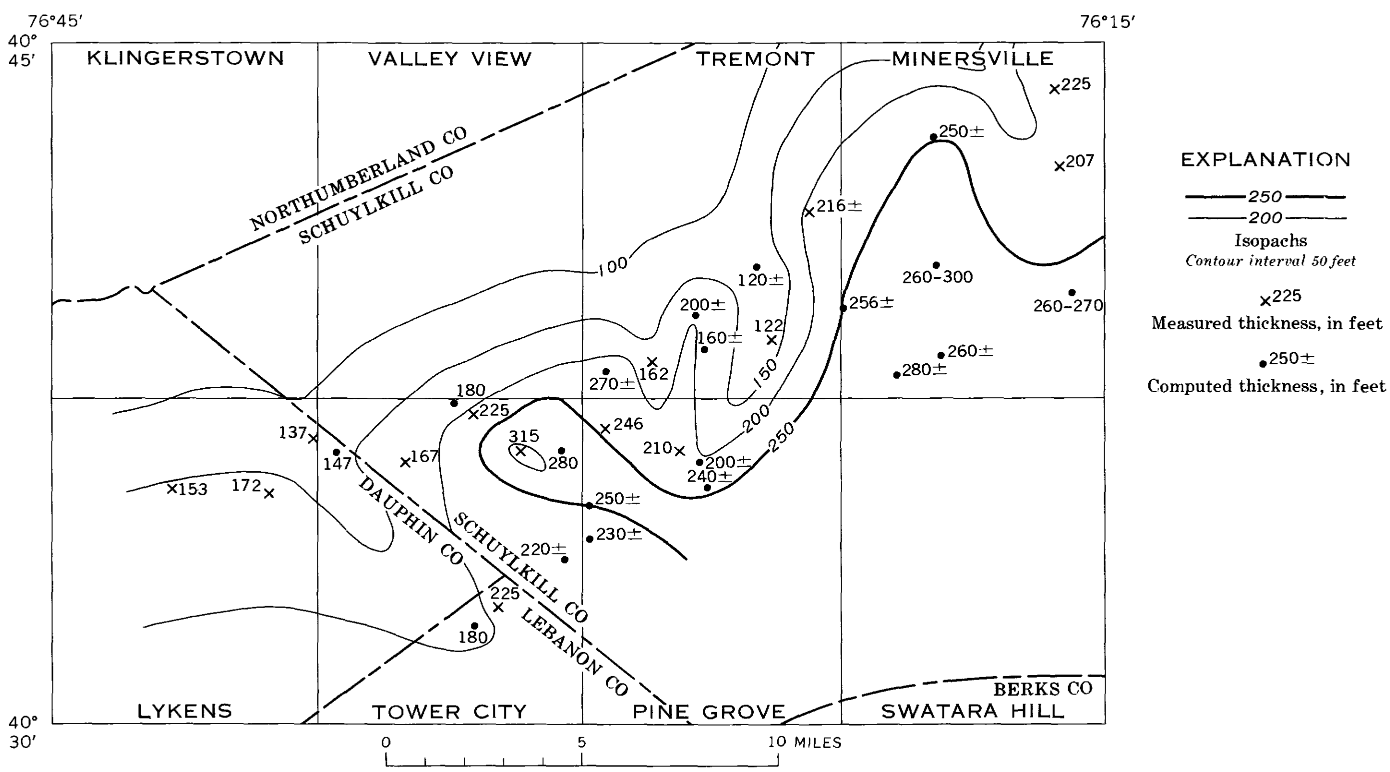


FIGURE 38.—Isopach map of the Sharp Mountain Member of the Pottsville Formation.

to large-scale simple and planar cross-strata and the orientation of ripple marks generally indicate deposition from northwest-flowing currents. Most beds are 4 to 10 feet thick; the rest range in thickness from 4 inches to 4 feet. Conglomerate and sandstone beds are generally thicker than adjacent siltstone and shale beds. Graded bedding is common, and reverse grading, although well developed locally, generally is rare. Plant debris is abundant in all rock types, but is well preserved only in siltstone and shale.

Conglomerate is largely concentrated in the basal part of the member, and finer grained rocks predominate in the upper part. Over most of the area, the Sharp Mountain is composed of about 45 percent conglomerate, 25 percent conglomeratic sandstone, 15 percent sandstone, 5 percent siltstone, 9.5 percent shale, and 0.5 percent anthracite. From south to north across the area, the percentage of conglomerate decreases slightly and the percentage of sandstone increases proportionately.

At places on Sharp and Broad Mountains, cobbles and pebbles in the basal conglomerate of the Sharp Mountain Member reach a maximum diameter of 8 inches. The average diameter of the clasts in this unit is about three-quarters of an inch for the area as a whole. Above the basal conglomerate the clasts reach a maximum diameter of about 2 inches, but generally are less than half an inch in diameter. They are composed of 60 to 85 percent vein quartz, 15 to 35 percent quartzite, and small percentages of chert, schist, gneiss, sandstone, slate, siltstone, and granite. Most are rounded to well rounded and are scattered evenly in a very fine to very coarse grained light- to dark-gray matrix.

The conglomerate of the Sharp Mountain Member weathers at many places into a gravel that mantles the outcrop belt and adjacent rock units. This gravel is distinctive because of the coarseness, rounded and well-rounded shape, and unfractured condition of the cobbles and pebbles.

Sorting in the conglomerate of the Sharp Mountain ranges from poor to excellent and averages moderate to good. There is little apparent lineation of the clasts, except that the long axis of the more elongate and ovoid ones commonly parallels the bedding. Most of the conglomerate is well cemented, but locally, the bonding ranges from weak to quartzitic.

Fine-grained to very coarse grained sandstone is largely concentrated in the upper part of the member. Sandstone beds are 2 inches to 20 feet thick, but most are between 2 and 5 feet thick. Tabular bedded units predominate, lenticular units are relatively common, and wedge-shaped and irregular units are rare. Many beds are crossbedded, and the rest are parallel laminated.

Most of the shale and siltstone is in the upper part of the Sharp Mountain Member and is in beds 1 inch to 2 feet thick. Locally, however, some beds are as thick as 10 feet. Nearly all the shale and siltstone is carbona-

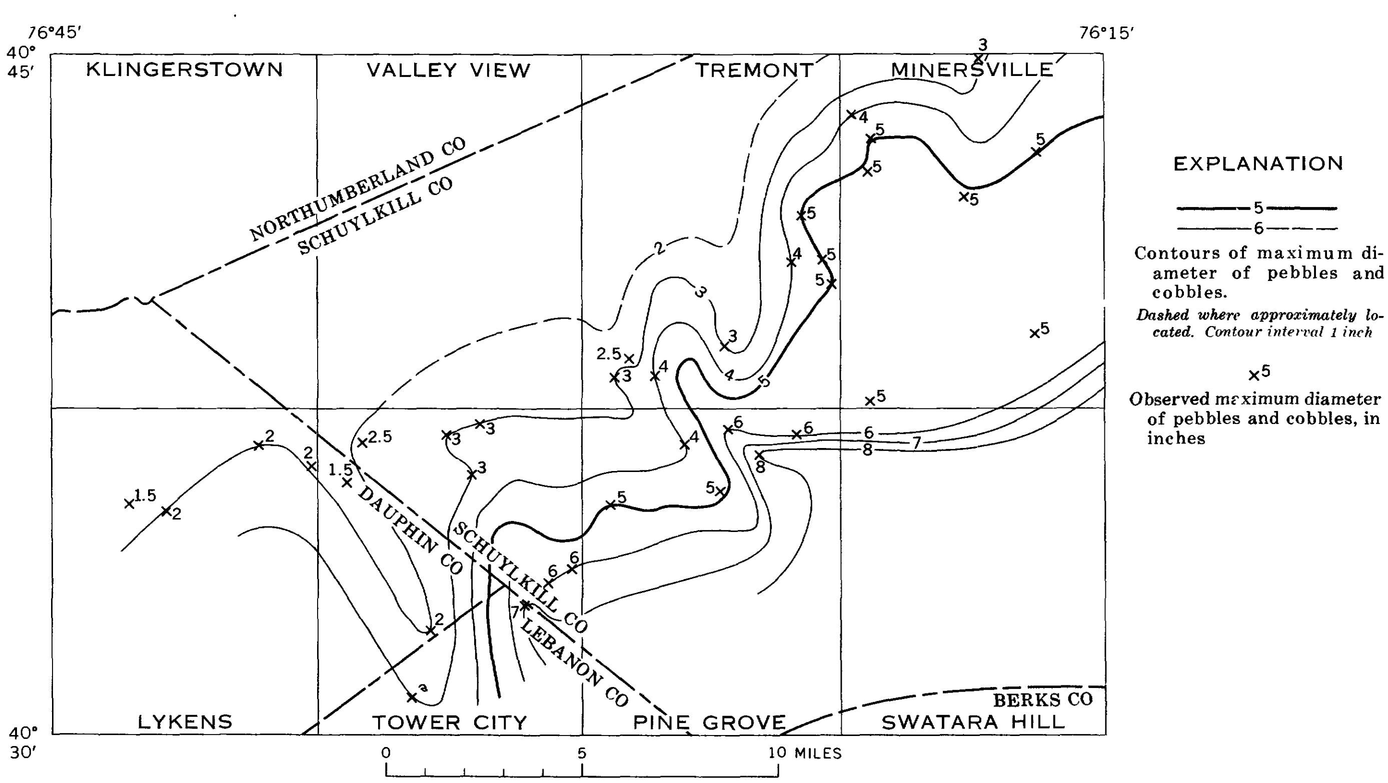


FIGURE 39.—Contour map of maximum diameter of pebbles and cobbles in the Sharp Mountain Member of the Pottsville Formation.

ceous and contains much macerated to well-preserved plant debris.

Three anthracite beds are present in the upper part of the member. In ascending order, these are the Scotty Steel Nos. 2 and 3 coal beds and the Little Buck Mountain (No. 4) coal bed. The Scotty Steel beds underlie much of the area (Wood and Trexler, 1968b, c). The No. 3 bed is slightly thicker and more widespread than the No. 2 bed. The Little Buck Mountain coal bed is known only in the northern and central parts of the Minersville quadrangle (Wood and Trexler, 1968b, sheets 1, 2).

The conglomerate, sandstone, and shale of the Sharp Mountain Member are compositionally protoquartzite and subgraywacke. The principal constituents of both rock types are grains of vein and common quartz and fragments of quartzite. Other constituents are fragments of chert, schist, gneiss, slate, siltstone, and sandstone; plates, blebs, and irregular masses of muscovite, biotite, chlorite, sericite, and unidentified clay minerals; grains of ilmenite, leucoxene, magnetite, sphene, tourmaline, and zircon; sparse grains of andesine, staurolite, epidote, and rutile; films of hematite and limonite; and varying amounts of silt- and clay-sized quartz fragments. Grains of quartz, chert, and quartzite range from subrounded to well rounded but average rounded. In contrast, other types of grains are generally subrounded. The margins of many quartz and quartzite grains are commonly sutured deeply by pressure solution, but where they are surrounded by a matrix of mica and clay minerals, their original shapes are preserved and suturing is negligible.

THICKNESS OF THE POTTSVILLE FORMATION

The Pottsville Formation thins northwestward from about 1,500 feet on Sharp Mountain to about 800 feet on Broad Mountain (fig. 40). Thickening of the formation southeastward at the southernmost outcrops on Sharp Mountain indicates that the axis of the Pottsville depositional basin lay south of this mountain, but the distance to the axis is unknown because of erosion.

DEPOSITION OF THE POTTSVILLE FORMATION

Sediments of the Pottsville Formation accumulated on a broad flood plain that overlay a slowly subsiding asymmetric depositional basin. They were laid down as fluvial and swamp deposits. The gray and olive colors of the rocks, the preservation of carbonaceous material in rocks of all types, and the presence of anthracite all suggest that deposition was in a reducing environment. Many swamps existed during deposition, and the vegetable matter that accumulated in them was later transformed into coal in the numerous anthracite beds.

The end of Pottsville deposition was marked by the formation of a very large swamp in which the vegetable matter of the Buck Mountain (No. 5) and correlative

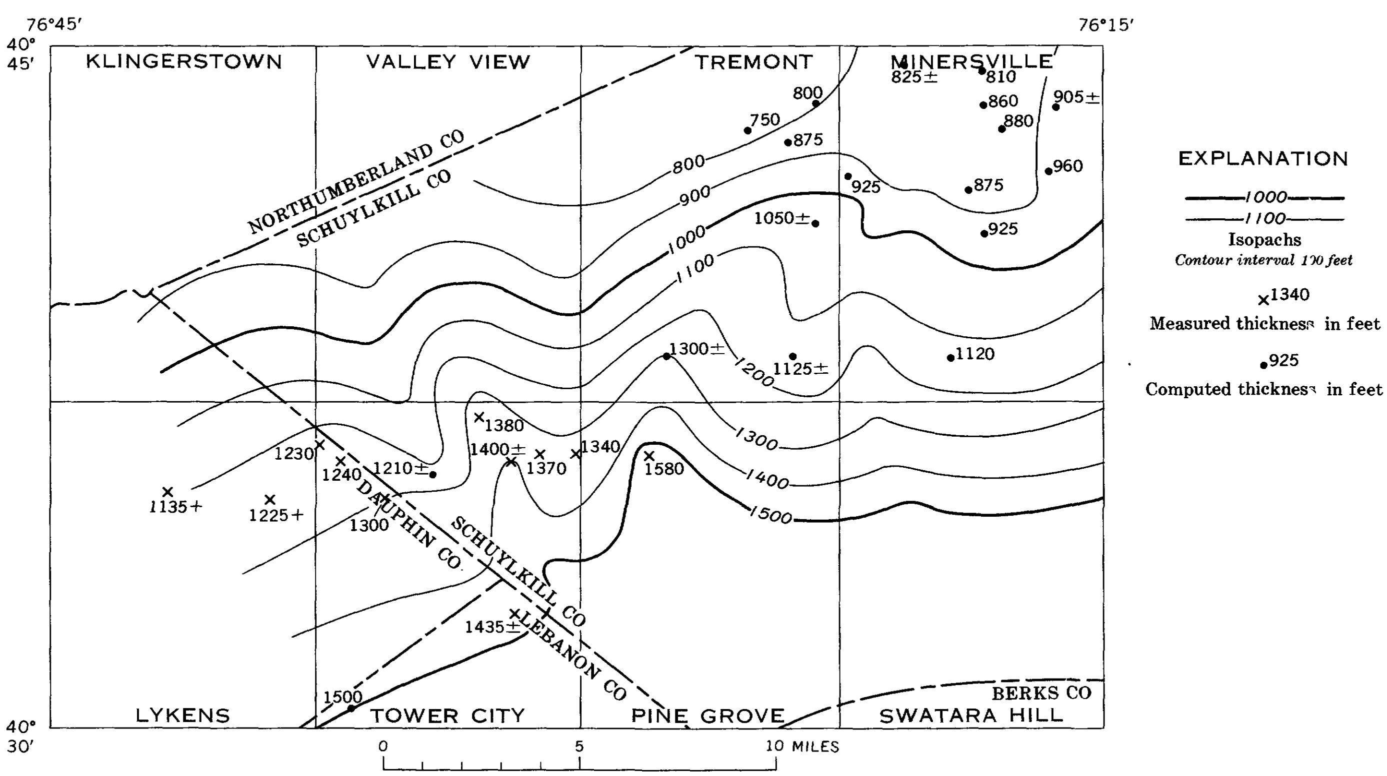


FIGURE 40.—Isopach map of the Pottsville Formation.

coal beds accumulated. This swamp apparently extended over the entire Anthracite region and may have extended into parts of western Pennsylvania, West Virginia, and Ohio. The concept that this swamp extended over such a large area is based upon the paleobotanical equivalency of the Buck Mountain coal bed of the Anthracite region and the Lower Kittanning coal bed of the Allegheny Formation of western Pennsylvania (C. B. Read, oral commun., 1954).

Sedimentary structures, thickness patterns, and the direction of increase in grain size indicate that the rocks of the Pottsville were derived from a source southeast and east of the area. During late Mauch Chunk time, a large part of the source region was uplifted by the first pulsation of the Appalachian orogeny. Coarse detritus from this uplifted part of the source region accumulated in the area during deposition of the upper member of the Mauch Chunk and the lower part of the Tumbling Run Member of the Pottsville. The rocks of the upper part of the Tumbling Run and the Schuylkill Members are finer grained than those of the lower part. This decrease in grain size probably records the erosional reduction of the source region and marked the end of the first pulsation of the orogeny.

Shortly before the end of Schuylkill deposition, the second pulsation of the Appalachian orogeny rejuvenated the source. A large volume of coarse detritus was eroded from the uplifted region and is preserved in the lower part of the Sharp Mountain Member. During deposition of this member, the second pulsation lessened in intensity, and finer detritus was eroded thereafter from the source. This finer detritus is preserved in the rocks of the upper part of the Sharp Mountain Member and the lower part of the Llewellyn Formation.

Pelletier (1958, p. 1059–1060) postulated that the fall line of the Pottsville Formation lay near Philadelphia. He determined the fall line by preparing a location graph of maximum pebble and cobble sizes from which the effects of deformation were removed by using crustal shortening data calculated by Cloos (1940, p. 841). His fall line was based upon the location where 6-inch cobbles theoretically would have been deposited.

Crustal shortening in the Pennsylvanian rocks of the area averages 45 percent, or 15 percent more than the value determined by Cloos (1940, p. 847). South of the report area, in the Great Valley and Piedmont regions, crustal shortening was a minimum of 50 percent (A. A. Drake, Jr., oral commun., 1962), or 10 percent more than the value calculated by Cloos.

Cobbles as much as 8 inches in diameter are present in the Pottsville Formation on Sharp Mountain (GQ–689, 690, 691). Obviously, these cobbles were deposited below a fall line. Therefore, 6-inch cobbles did not mark the Pottsville fall line as advocated by Pelletier. Casual observations by the authors indicate that 10- to 12-inch boulders mark the fall line of the Atlantic Coastal Plain. Many more data would be required than are now available to determine accurately the average maximum size of these boulders. Nevertheless, the size is much greater than that advocated by Pelletier. If stream velocities and gradients in the Pottsville source area were similar to those of the modern Piedmont, it seems likely that the Pottsville fall line would also have been marked by 10- to 12-inch boulders.

When the effects of crustal shortening have been nullified by using the data determined by the authors and Drake, a Pottsville fall line based upon 10- to 12-inch boulders lay at about the position of the modern Reading Prong, or about 40 miles northwest of the 6-inch fall line determined by Pelletier. The Reading Prong is underlain by metamarphic and igneous rocks similar to those believed to have supplied much of the detritus preserved in the Pottsville.

The Tumbling Run and Sharp Mountain Members of the Pottsville Formation and the upper member of the Mauch Chunk Formation are the only stratigraphic units in the area in which garnet, epidote, staurolite, kyanite, and sillimanite and clasts of gneiss and granite have been found. These minerals and clasts suggest that an igneous mass and a high-temperature metamorphic mass were exposed in the source region.

AGE AND CORRELATION OF THE POTTSVILLE FORMATION

The Pottsville Formation is of Early and Middle Pennsylvania age (Read and Mamay, 1960, table 176.1). For many years the formation at its type section has been considered the type basal unit of the Pennsylvanian System.

The Tumbling Run Member of early Early Pennsylvanian age contains fossil plants of the zone of *Neuropteris pocahontas* and *Mariopteris eremopteroides* and of the overlying zone of *Mariopteris pottsvillea* and *Aneimites* spp. These plants indicate that the member correlates with the Pocahontas Formation and the lower parts of the Lee and New River Formations of southern West Virginia, southwestern Virginia, Kentucky, and Tennessee.

The Schuylkill Member of late Early Pennsylvanian age contains plant remains from the zone of *Mariopteris pygmaea* and *Neuropteris tennesseeana*. These fossils indicate that the Schuylkill correlates with the upper part of the New River Formation, the Bloyd Shale of the Morrow Series (Read and Mamay, 1960, table 176.1), and with the upper part of the Lee Formation of southwestern Virginia, Kentucky, and Tennessee (Read, 1947, table 1). Bode (1958, p. 254) assigned the flora of the Lykens Valley No. 1 coal bed to the Westphalian zone A of Europe.

The upper part of the Sharp Mountain Member contains a fossil flora of the zone of *Neuropteris rarinervis* and the upper part of the zone of *Neuropteris tenuifolia* (Wood and others, 1956, table 2) or Westphalian zone D of Europe (Bode, 1958, p. 254). The lower part of the zone of *Neuropteris tenuifolia* and the zone of *Megalopteris* spp., and their European equivalents Westphalian zones B and C, may be present in the lower coarsely conglomeratic sequence of the member, but they have not been identified (Wood, and others, 1956, p. 2684–2685).

The upper beds of the Sharp Mountain Member correlate with the lower part of the Allegheny Formation of western Pennsylvania, Ohio, and West Virginia, and with the lower part of the Des Moines Series of the midcontinent region. The lower part of the member probably correlates with the Kanawha Formation of West Virginia, Virginia, and Kentucky.

Read and Mamay (1960, table 176.1) classified the Sharp Mountain as late Middle Pennsylvanian. The Buck Mountain (No. 5) coal bed at the base of the Llewellyn Formation contains latest Middle Pennsylvanian plants of the zone of *Neuropteris rarinervis;* the authors, therefore, tentatively suggest that the member may represent all but the uppermost part of Middle Pennsylvanian time.

LLEWELLYN FORMATION

The Llewellyn Formation was named in 1962 (Wood and others, p. C41–C42) for a town in the Minersville quadrangle. The rocks assigned to this formation were originally referred to an informal stratigraphic unit known as the "coal measures" (Smith, 1895, p. 1920). Geologists later assigned them to the Allegheny and Conemaugh Formations (Willis, 1912, 439–442; Lohman, 1937, p. 47; Rothrock and others, 1950) and to an informal unit known as post-Pottsville rocks (Wood and others, p. 2678, fig. 3, and table 2).

Much of the complexly folded and faulted Minersville synclinorium is underlain by the Llewellyn Formation (GQ–689, 690, 691, 692, 698, 699, 701; Wood and Trexler, 1968b, c). The west-central part of the Southern Anthracite field is underlain by this formation and by the Pottsville Formation.

There are few natural exposures of the rocks of the formation, but there are many manmade outcrops in strip pits and in underground mines. Millions of tons of debris from these strip pits and underground mines have been dumped between the strip pits; therefore, much is known about the rocks in the pits, but relatively little is known about the rocks between them.

The Llewellyn underlies a series of gentle to steep, rolling to rough parallel ridges and valleys that have been eroded into a trellis drainage pattern. The drainage system has been greatly modified at many hundreds of places by strip and underground mining for anthracite.

The contact between the Llewellyn Formation and the underlying Sharp Mountain Member of the Pottsville Formation is placed at the base of the underclay or shale of the Buck Mountain (No. 5) coal bed. In a few localities the shale is absent, and the contact is placed at the base of the coal bed. The upper limit of the Llewellyn is the present erosion surface at most places, but in a few places the formation is overlain by Quaternary alluvium.

The Llewellyn reaches a maximum thickness of about 3,500 feet in the central part of the Minersville quadrangle near the village for which it was named. It is impractical to construct a thickness map of the formation because of structural complications and the small area in which the youngest beds are preserved.

The thickness and lithology of the lower and middle parts of the Llewellyn Formation are illustrated at 39 localities (pls. 1–3). The upper part is so poorly exposed that its variations in thickness and lithology could not be determined and studied. The lower and middle parts of the formation are characterized by many lateral variations in rock type. Despite these variations, the continuity and equivalency of thin to thick rock sequences over large areas is indicated by the associated anthracite beds. The persistency of the anthracite beds has been proven by mapping and by underground and strip mining. It is impossible, however, to show many of the data that prove this persistency on the coal maps and coal structure sections prepared simultaneously with this report (Wood and Trexler, 1968b, c).

The Llewellyn is characterized by sharp thickness changes between datum horizons that are only a few feet to several hundred feet apart. It is also characterized by rather even or uniform thicknesses between datum horizons that are much farther apart. For example, local thickness variations are numerous and sharp between the Buck Mountain (No. 5) coal bed and the coal beds of the Mammoth coal zone, between individual coal beds of the Mammoth coal zone, and between the uppermost coal bed of the coal zone and the Diamond (No. 14) or Peach Mountain (No. 18) coal beds (pls. 1–3). However, the average thickness between the Buck Mountain and Diamond coal beds is about 900 feet, and the greatest variation in thickness between these beds is 775 to 950 feet. Similarly, the average thickness between the Buck Mountain and Peach Mountain coal beds is about 1,400 feet, and the variation is only 1,300 to 1,475 feet. The sharp local variations between nearby datum horizons probably resulted from unequal volumes of sediment of different grain size being deposited from place to place at the same time. The overall uniformity in thickness between datum horizons that are many hundreds of feet apart seems to be due to each part of the basin of deposition being filled by approximately equal volumes of each textural size.

Freshly broken rocks of the Llewellyn Formation are very light to dark gray, grayish black, light brownish gray, greenish gray, and grayish orange. Weathered rocks are commonly the same colors, but are also pale to light brown, pale to dark yellowish brown, and light olive gray. Many weathered rocks have a salt-and-pepper appearance where hues of gray are combined with the colors listed above.

Tabular, lenticular, and wedge-shaped beds are common in the Llewellyn. Most beds are 1 to 2 feet thick, but some are as thin as half an inch, and others are as thick as 20 feet. Thin- to thick-channel sandstone deposits are present at many places but are sparse enough to be noteworthy. Many beds are graded, but reverse grading is rare. Although parallel laminations predominate over cross-laminations, small to large simple and planar cross-strata are abundant. The cross-laminations, ripple marks, oriented plant fragments, and the direction of increase in pebble diameters generally indicate deposition by streams flowing from the southeast. The ripple marks and oriented plant fragments are preserved principally in very fine grained to medium-grained sandstone, siltstone, and silty shale. Well-preserved plant fossils are most common in shale adjacent to anthracite beds, but macerated plant debris is abundant in all rock types.

Conglomerate and conglomeratic sandstone are scattered at random in the Llewellyn. Most are in small lenticular bodies, but several conglomerate lenses between the Buck Mountain (No. 5) coal bed and the Bottom Split (No. 8) coal bed are widespread in the Minersville and Tremont quadrangles (GQ–690, 692). A conglomerate bed overlies the roof shale of the Primrose (No. 11) coal bed and is traceable over much of the Tower City (GQ–698), Tremont, Pine Grove (GQ–691), and Minersville quadrangles. A series of widespread conglomerate beds crop out in the southern part of the Minersville, Pine Grove, and Swatara Hill (GQ–698) quadrangles between the 22–25 coal beds.

The conglomerate beds below the Peach Mountain (No. 18) coal bed contain pebbles as much as 2 inches in diameter and average half an inch. Conglomerate above this coal bed commonly contains pebble averaging about three-quarters of an inch in diameter and locally contains cobbles as much as 3 inches in diameter. These large clasts are composed mainly of vein quartz with substantial amounts of quartzite and lesser amounts of chert, schist, gneiss, slate, sandstone, siltstone, and shale. The percentage of clasts derived from metamorphic terranes seems to increase stratigraphically upwards.

Most of the conglomerate in the Llewellyn is poorly to moderately sorted. However, locally, some is well sorted and closely resembles the basal conglomerate of the Sharp Mountain Member of the Pottsville Formation.

Fine- to coarse-grained sandstone is present throughout the Llewellyn. The bedding ranges in thickness from ½ inch to 15 feet and averages about 1 foot 6 inches. Tabular beds predominate, but lenticular beds are common, and wedge-shaped and irregular beds are abundant locally. Although most strata are parallel laminated, small- to large-scale simple and planar cross-strata are common at many places.

Most siltstone and shale beds are 1 inch to 2 feet thick, but many sequences of these beds are 20 to 50 feet thick. Much of the siltstone and shale is carbonaceous and commonly contains large amounts of plant debris and intercalated beds of anthracite at many places. The underclay or shale of some anthracite beds is characterized by abundant *Stigmaria* impressions, and a few beds contain upright tree trunks. Well-preserved plant leaves and stems are common in the roof shales of many anthracite beds.

Forty beds of anthracite are economically and stratigraphically important enough to be identified in this report by name and number or only by number (Wood and Trexler, 1968b, c). The beds have been numbered from the base of the formation upward, starting with 5 and ending with 29.

The more widely persistent and thicker beds of anthracite are the Buck Mountain (No. 5) coal bed, Skidmore (No. 7), Bottom Split (No. 8), Middle Split No. 8½), Top Split (No. 9), Holmes (No. 10), Primrose (No. 11), Diamond (No. 14), Tracy (No. 16), and Peach Mountain (No. 18) coal beds. These beds are confined to the lower 1,475 feet of the formation, and most lie in the lower 650 feet.

Eleven other, generally thinner, slightly less persistent, and somewhat more lenticular coal beds are in the lower 1,600 feet of the Llewellyn. They are the Seven Foot (No. 6), Lower Four Foot (No. 9½), Orchard (No. 12), Little Orchard (No. 13), Little Diamond (No. 15), Clinton (No. 15¼), Little Clinton (No. 15½), Upper Four Foot (No. 16½), Little Tracy (No. 17), Tunnel (No. 19), and Rabbit Hole (No. 20) coal beds.

Numerous, local, lenticular, and generally thin beds of anthracite are present at many places in the area. From the base upward the economically and stratigraphically more important of these anthracite beds are the 6L, 7L, 8¼, 10L, 13½, Diamond Leader (14L), 14¼, 14½ Faust (No. 20), 22, 22L, 23, 23½, and 24–29 coal beds.

The conglomerate, sandstone, siltstone, and shale in the Llewellyn is predominantly subgraywacke. Some of the coarser grained sandstone and conglomerate approaches protoquartzite in composition and is here classified as "clean" subgraywacke. The rocks in the lower part of the formation contain higher percentages of common and vein quartz and smaller percentages of rock fragments, unidentified clay minerals, sericite, and other mica minerals than do those in the upper part. The principal constituents of Llewellyn rocks are common and vein quartz, fragments of quartzite, and blebs and plates of sericite and muscovite. Other constituents are fragments of chert, schist, slate, and siltstone; blebs of chlorite; grains of ilmenite, leucoxene, magnetite, rutile, tourmaline, zircon, and sphene; films of limonite and hematite; a few grains of andesine; rare interstitial fillings of halloysite; and varying amounts of unidentified clay minerals, silt- and clay-sized fragments of quartz, and carbonaceous material.

The rocks of the Llewellyn are cemented principally by silica, sericite, and silt- and clay-sized quartz fragments. Other cements are limonite, hematite, clay minerals, and carbonaceous material. Grains of quartz and fragments of chert and quartzite range from angular to well rounded and average rounded. Other rock fragments and grains range from angular to rounded and average subrounded. Chert fragments are generally better rounded than are other constituents. Many quartz grains and quartzite fragments are deeply sutured, but where they are embedded in a fine-grained matrix of nonquartz minerals, suturing generally is absent.

Llewellyn sediments accumulated in an oxygen-poor or reducing environment upon a broad flood plain as fluvial and swamp deposits. The flood plain was bounded to the southeast by a nearby highland area that had been uplifted during late Mauch Chunk and Pottsville time. It was bounded to the west and northwest by a distant fluctuating coastline. A slowly subsiding asymmetric depositional basin underlay the flood plain. The short southeastern limb of the basin probably extended only a few miles southeast of the report area. The much broader northwestern limb may have extended into central and north-central Pennsylvania.

The scattered conglomerate, conglomeratic sandstone, and coarse-grained sandstone of the Llewellyn indicate that the highland area southeast of the report area was uplifted intermittently. The small areal extent of the coarse detrital deposits and the small average size of the clasts in the lower and middle parts of the formation suggest that uplifted parts of the highland were eroded quickly during early and middle Llewellyn time. In contrast, the greater areal extent of the coarse detrital deposits and the greater average diameter of the clasts in the upper part of the formation suggest that the uplifted parts of the highland were considerably larger than they had been earlier or that they had advanced northwestward towards the report area.

The lithology, sedimentary features, and plant fossils of the finer grained rocks of the Llewellyn indicate that these rocks were carried onto the flood plain by meandering sluggishly flowing streams. The flood plain apparently had little topographic relief and was easily flooded by slight changes in base level, precipitation, or depositional habits of streams. Such changes resulted in repeated flooding of parts of the flood plain and the development of local and regional swamps that, in some instances, covered hundreds of square miles. Abrupt lithologic changes from conglomerate and coarse-grained sandstone to carbonaceous shale, siltstone, and anthracite indicate that in many places the transition from fluvial to swamp deposition, and the reverse was rapid. Vegetation grew rapidly and profusely in the swamps. The 20 to 40 feet or more of anthracite preserved locally in coal beds of the Mammoth coal zone (pls. 1–3) indicate that such growth apparently continued for long periods of time. Calculations based upon Ashley's (1907, p. 42) rates of deposition of vegetable material converted to old peat and coal indicate that the 40 feet or more of anthracite locally present in individual beds of this zone accumulated during periods of as long as 12,000 years. Most beds of anthracite in the area are considerably less than 40 feet thick and therefore, took far less time to accumulate, the numerous anthracite beds, however, undoubtedly indicate that many swamps existed for hundreds, and in some instances for thousands, of years during Llewellyn time.

The swamps of the Llewellyn were each successively destroyed by blankets of fine to coarse detritus from the highland southeast of the flood plain. No regional blankets of coarse sand and gravel seem to have been deposited during accumulation of the lower and middle parts of the Llewellyn, but many extensive coarse sandstone and conglomerate sequences were deposited during accumulation of the upper part. These more extensive coarse deposits record uplift and erosion of the highland mass that lay southeast of the area during the third major pulsation of the Appalachian orogeny. The oregony may have consisted of more than three major pulsations, but this is uncertain because Paleozoic rocks younger than the Llewellyn have been removed by erosion. If there were only three pulsations, the coarse detritus in the upper part of the formation presumably records the beginnings of the pulsation that later advanced northwestward and deformed the area. However, several pulsations unrecorded by sedimentation may have taken place before the orogeny drew to an end.

The Llewellyn Formation is of latest Middle Pennsylvanian and Late Pennsylvanian age. Its plant fossils were originally studied by Lesquereux (1880–84). Subsequently, Read (in Wood and others, 1956, table 2; in Wood and others, 1958, table 1) concluded that the rocks of the formation contained fossil plants from the upper part of his zone of *Neuropteris rarinervis*, and from his zone of *Neuropteris flexuosa* and *Pecopteris* spp. C. D. White (1900, p. 830) stated that the lower unit of the formation, the Buck Mountain (No. 5) coal bed, correlates with the Lower Kittanning coal bed in the middle part of the Allegheny Formation of central and western Pennsylvania. Read (oral commun., 1954, 1955; written commun., 1955; in Wood and others, 1956, table 2; in Wood and others, 1958, table 1) confirmed this correlation and stated that these coal beds contain a flora characteristic of the upper part of his zone of *Neuropteris rarinervis*. In 1960, Read and Mamay (p. B381) assigned the zone of *Neuropteris rarinervis* to the Middle Pennsylvanian but incorrectly indicated that the basal part of the Llewellyn equivalent to the post-Pottsville rocks of Wood, Trexler, Arndt, Yelenosky, and Soren (1956, p. 2678), which includes the Buck Mountain coal bed, was entirely in the zone of *Neuropteris flexuosa* and *Pecopteris* spp. Thus, despite Read and Mamay's 1960 placing of the rocks equivalent to the Llewellyn in the Upper Pennsylvanian, Read's 1954–58 identifications indicate that the basal part of the formation is of latest Middle Pennsylvanian age.

Information is scant concerning the fossil plants in Pennsylvanian rocks younger than the Pottsville Formation. Table 1 shows the observed range of fossils in the area. It was compiled from field and office identifications by Read, Mamay, and Bode and from published identifications by Bode (1958, p. 253–254). Fossils were not collected from above the Faust (No. 21) coal bed because they are so poorly preserved and fragmental that they are largely unidentifiable. Some species may have actual ranges, therefore, which extend above the Faust. All fossils listed are from the floor or roof shales of coal beds.

According to Read (in Wood and others, 1956, table 2) and Read and Mamay (1960, table 176.1), the rocks included in the Llewellyn Formation correlate with the upper part of the Allegheny Formation and the lower part of the Conemaugh Formation of central and western Pennsylvania, Maryland, Ohio, and West Virginia and with the upper part of the Des Moines Series of the midcontinent region. Bode (1958, p. 254 and fig. 13) agreed with Read and Mamay that the Llewellyn correlates with the upper part of the Allegheny and the lower part of the Conemaugh Formation and, in addition, stated that the youngest rocks of the Llewellyn contain fossil plants characteristic of the D zone of the Westphalian Stage of Europe.

QUATERNARY DEPOSITS

The youngest naturally occuring sedimentary deposits of the report area are Pleistocene and Recent talus and alluvium. Recent manmade deposits of mine waste resulting from the intensive mining have been indiscriminately deposited during the last 150 years on the Pottsville and Llewellyn Formations as well as on the talus and alluvium.

TALUS

Extensive talus fields rest unconformably on the underlying rocks and blanket the slopes of many of the mountain ridges in the area. The larger of these fields are on the slopes of Bear, Big Lick, Blue, Broad, Coal, Line, Little, Mahantango, Second, Sharp, Sherman, Short, and Stony Mountains (GQ–689, 690, 691, 692, 698, 699, 700, 701). Many other smaller fields are present in the area, but are not shown on the maps of this report. The talus fields generally are only a few tens of feet thick and are composed of small to large rectilinear blocks of sandstone and conglomerate mixed with smaller fragments derived from siltstone, shale, and anthracite beds.

At many places talus deposits have hampered the discovery and exploitation of coal beds in the Pottsville and Llewellyn Formations. Numerous deposits mantle considerable areas in the west-central part of the Southern Anthracite field.

The talus fields appear to have been forming since at least the beginning of Pleistocene time. This age assignment is based on the fact that glacial deposits in nearby areas are adjusted to preexisting topography similar to that in the area, which suggests that the relief of the area was partially developed before glaciation. Some talus deposits, therefore, must have been forming before the beginning of glaciation.

ALLUVIUM

Alluvial deposits, which rest unconformably upon underlying rocks, are shown only on the quadrangle maps (CQ–689, 690, 691, 692, 698, 699, 700, 701) along Deep, Little Mahantango, Lower Little Swatara, Mahanoy, Mahantango, Mill, Pine, Swatara, Upper Little Swatara, and Wiconisco Creeks. Most of these deposits are thin, but they commonly mask the outcrops of stratigraphic units, folds, and faults in the valleys of these creeks.

Table 1.—Observed ranges of plant fossils in the Llewellyn Formation and upper part of the Pottsville Formation

[Plant identifications by C. B. Read, S. H. Mamay (U.S. Geological Survey) and H. Bode (West German paleobotanist, Munster) stratigraphic assignment by G. H. Wood, Jr.]

Fossil plant species	Pottsville Formation	Llewellyn Formation: Coal bed and number — Buck Mountain (No. 5)	Bottom Split (No. 8)	Holmes (No. 10)	Orchard (No. 12)	Diamond (No. 14)	Tracy (No. 16)	Upper Four Foot (No. 16½)	Peach Mountain (No. 18)	Faust (No. 21)
Alethopteris lonchitica	—	—	—	—	—	—	—	—	—	—
A. serlii	—	—	—	—	—	—	—	—	—	—
Annulari sphenophylloides							—	—	—	—
A. stellata	—	—	—	—	—	—	—	—	—	—
Asterophyllites equisetiformis										—
Aulacotheca sp		—								
Bowmanites sp									—	
Callipteridium sp								—	—	
C. pteridium		—								
Cordaites sp			—	—	—	—	—	—	—	—
Linopteris rubella	—	—	—	—						
Mariopteris cordata-ovata		—	—	—	—	—	—	—	—	—
M. cf. *M. muricata*		—								
M. cf. *M. sphenopteroides*		—	—	—	—					
M. cf. *M. nervosa*					—					
Neuropteris clarksoni		—	—	—	—	—	—	—	—	—
N. falcata						—				
N. flexuosa		—	—	—	—	—	—	—	—	—
N. cf. *N. grangeri*							—			
N. cf. *N. heterophylla*							—			
N. missouriensis	—	—	—	—	—	—	—	—	—	—
N. ovata	—	—	—	—	—	—	—	—	—	—
N. rarinervis	—	—	—							
N. scheuchzeri	—	—	—	—	—	—	—	—	—	—
N. tenuifolia		—								
N. vermicularis		—	—	—	—					
Odontopteris spp										—
Pecopteris arbarescens	—	—	—	—	—	—	—	—	—	—
P. candolliana		—	—	—	—	—	—	—	—	—
P. feminaeformis										—
P. lamuriana		—								
P. unita	—	—	—	—	—	—	—	—	—	—
P. vestita	—	—								
Sigillaria sp					—	—	—	—	—	—
Sphenophyllum emarginatum		—	—	—	—					
S. majus										—
S. oblongifolium										—

Deposits of alluvium are present along most of the permanent stream courses in the Southern Anthracite field, but they are not shown on the coal maps or quadrangle maps because they would make the maps unnecessarily difficult to read.

In addition to the naturally developed alluvial deposits, many hundreds of large to small deposits of mine waste have been derived from strip pits, underground mines, and breaker plants. Most underground mine-waste deposits were accumulated during the period from 1870 to 1945 and the strip-pit waste deposits, since 1930.

The natural alluvial deposits are Pleistocene and Recent. This age assignment is based on the assumption that most stream channels probably had their greatest development during the advances and retreats of the glaciers of Pleistocene time. Therefore, it seems probable that part of the detritus in these deposits accumulated during these advances and retreats and that the rest was laid down in Recent time.

STRUCTURE

The rocks of the area have been folded and faulted during three episodes of mountain building: the Taconic, Acadian, and Appalachian orogenies. Structural features that formed during the Appalachian orogeny are superimposed on and obscure those formed during the earlier orogenies. Consequently, much more is known and described below about the structural features formed during the Appalachian orogeny than about those of the earlier events.

The report area lies in the southeastern part of a large poorly defined north-northeast-trending structural depression or sag that occupies much of east-central and northeastern Pennsylvania (Gray and others, 1960). The depression has been largely obscured by a multitude of superimposed east-northeast-trending faults and folds. The trough of the depression, which is interrupted at many places by the superimposed faults and folds, is sinuous and trends about N. 20° E. from the vicinity of Pottsville to the vicinity of Wilkes-Barre, both of which lie east and northeast of the area. Structurally, the boundary of the depression is difficult to define; but to the north it is at the structural front of the Appalachian Mountains, to the east it is at the west edge of the Pocono Plateau, to the west it lies in the contorted outcrop belt of the Middle Devonian rocks, and to the south it is in the outcrop of the Tuscarora Sandstone.

Other than the great depression, the principal structural elements of the area are the Broad Mountain and New Bloomfield anticlinoria and the Minersville synclinorium. Superimposed upon these structural elements and their subsidiary folds are numerous small to large thrust, underthrust, high-angle reverse, bedding-plane, and tear faults.

STRUCTURAL FEATURES FORMED DURING THE TACONIC OROGENY

The Taconic orogeny of Late Ordovician time profoundly affected the region south, southeast, and east of the area. Its effects are poorly known in the area because rocks of Odrovician age are largely covered. However, these rocks were folded during the disturbance, then eroded, and finally, covered unconformably by the Tuscarora Sandstone of Early Silurian age. The unconformity is not exposed in the area, but is believed to be angular as it is to the east at Schuylkill Gap in Blue Mountain. Some field evidence observed on Blue Mountain and to the south in the southern part of the Swatara Hill quadrangle (GQ–689) supports this belief. In these localities the Tuscarora Sandstone dips 35° to 45° N., and the Martinsburg Shale, about 4,000 feet to the south, dips 10° to 45° S. There is no indication of a modern anticline between these exposures; therefore, when the Tuscarora Sandstone was deposited, the Martinsburg beds probably dipped 45° to 90° S.

STRUCTURAL FEATURES FORMED DURING THE ACADIAN OROGENY

A few structural features that formed during the Acadian orogeny of Late Devonian and Early Mississippian time have been identified in the area.

The Spechty Kopf Member of the Catskill Formation is overlain angularly by the Beckville Member of the Pocono Formation on Broad and Peters Mountains in the southern part of Lykens quadrangle (GQ–701), (Trexler and others, 1961, fig. 38.2). Structural attitudes indicate that rocks of the Spechty Kopf dipped an average of 20° N. when the Beckville was deposited.

Several miles north of Broad and Peters Mountains in Rattling Creek gorge through Berry Mountain (GQ–701), a generalized anticinal feature consisting of two anticlines and a syncline in the Spechty Kopf Member of the Catskill is overlain by a single anticline in the Beckville Member of the Pocono. Locally, the angularity between the beds of these members is as great as 75°.

About 2,400 feet of Spechty Kopf rocks underlie the southern part of Broad Mountain, but only about 450 to 500 feet of the lower part of these rocks is preserved in the core of the complex anticlinal and synclinal feature at Rattling Creek. This thickening indicates that an Acadian syncline underlies the area between Rattling Creek and Broad Mountain. This syncline and the anticlinal feautre in Rattling Creek gorge are substantiated on the isopach map of the Spechty Kopf (fig. 28).

Two other Acadian folds can be inferred from thickness variations of the Spechty Kopf Member. An anticline seems to have formed in Northumberland County at the approximate position of the crest of the modern Broad Mountain anticlinorium (GQ–692, 699, 700). The postuated crest and trend of this Acadian anticline are depicted on figure 28 by the elongate area where the member is absent. In addition, a syncline south of this anticline seems to have formed in Dauphin and central Schuylkill Counties.

Acadian anticline and synclines seem to have had nearly the same trend as those formed later during the Appalachian orogeny. This similarity in trend probably means that some Acadian folds have not been identified. The modern Broad Mountain anticlinorium and Berry Mountain anticline appear to be rejuvenated Acadian folds. Although other modern folds may also have been rejuvenated, definite evidence is lacking.

Faults that formed during the Acadian orogeny have not been identified, but some of the faults that are confined to pre-Pocono rocks may originated during this orogeny.

STRUCTURAL FEATURES FORMED DURING THE APPALACHIAN OROGENY

Many thousands of folds and faults were formed in the area during the late Paleozoic Appalachian orogeny. Only the larger of these are shown on the quadrangle, coal, and tectonic maps and the structure sections (GQ–689, 690, 691, 692, 698, 699, 700, 701; Wood and Trexler, 1968b, c).

The principal structural features formed during the Appalachian orogeny were two anticlinoria and a synclinorium. Superimposed upon these complex fold systems are a multitude of low-angle, high-angle reverse, underthrust, tear, and bedding-plane faults.

FOLDS

The largest structural features in the area from south to north are: the south, or main, trough of the Minersville synclinorium, the New Bloomfield anticlinorium, the north trough of the Minersville synclinorium, and the Broad Mountain anticlinorium. Each of these complex fold systems is composed of many en echelon subsidiary folds which trend east-northeast. The crests and troughs of subsidiary folds large enough to depict at scales of 1:12,000 and 1:24,000 are shown on the maps and structure sections (GQ–689, 690, 691, 692, 698, 699, 700, 701; Wood and Trexler, 1968b, c). Many hundreds of other folds too small to be shown are present both at the surface and in the subsurface.

Individual folds range in length from less than 1 inch to about 27 miles and range in amplitude from a fraction of an inch to about 16,000 feet. None extend across the full width of the area from west to east; but several cross two to three quadrangles, and many cross a single quadrangle.

Typically, the folds of the area are doubly plunging, but because of the regional northeast plunge, more folds terminate in that direction than to the southwest. All folds, however, die out eventually to the southwest, whether in the area or west of it.

The anticlines and synclines that formed during the Appalachian orogeny trend N. 55° E. to N. 85° E. and generally plunge northeast at 4° to 8°. Locally, where folds are dying out, however, the plunge may be as great as 30° NE. or 30° SW. The rate of regional northeast plunge and the maximum rate of plunge near fold terminations were determined from extensive mine and field data.

Most large anticlines north of the Branchdale anticline (GQ–690) are simple open folds with broad crestal areas; in contrast, most large synclines are acute or tight folds. In their midparts the larger synclines generally are V-shaped in cross section, whereas near their terminations they are U-shaped. This change from V- to U-shaped cross sections does not seem to be related to rock competency. Locally, where anticlines and synclines are closely spaced and smaller, the anticlines commonly are tightly folded but less so than the adjacent synclines. Southeast of the Branchdale anticline and northwest of the Sweet Arrow fault zone (GQ–689), all folds are tight, many are broken by high-angle reverse faults that truncate the crests and troughs, and many have overturned limbs.

Parallel folding is the principal style of fold deformation. Mining data indicate that many folds approach parallelism so closely that their shapes, as illustrated on the structure sections (GQ–689, 690, 692; Wood and Trexler, 1968c), could have been reconstructed mechanically with little error. Most folds die out at depth and show little stratigraphic thinning on their limbs. Some larger synclines locally are similar folds that exhibit pronounced thickening in their troughs and thinning on their limbs. Many folds, even though nearly parallel, are locally disharmonic. Disharmonic relations appear invariably to be associated with incompetent sequences of shale or coal and are most common in the troughs or on the lower limbs of synclines.

BROAD MOUNTAIN ANTICLINORIUM

The Broad Mountain anticlinorium is one of the largest structural features in the area. To the southwest it extends at least as far as the Maryland-Pennsylvania boundary, and to the northeast it dies out against the

Pocono Plateau. It trends N. 65° to 80° E. across the central and northern parts of the Klingerstown, Valley View, and Tremont quadrangles (pl. 4; GQ–692, 699, 700). The crest of the anticlinorium in the Klingerstown and western Valley View quadrangles is the Hooflander Mountain anticline, and in the eastern Valley View and Tremont quadrangles it is the Frackville anticline.

The principal subsidiary anticlines and synclines of the Broad Mountain anticlinorium are the Locust Mountain, Mahanoy Creek, North Hooflander Mountain, Eisenhuth Run, Powder Hill, Hans Yost, Mine Hill, West West Falls, Peaked Mountain, Crystal Run, and Swatara anticlines and the Mahanoy Basin, Hooflander Mountain, New Boston, Beury, Jugular, Heckscherville, Little Mountain, North Little Mountain, Deep Creek, Dam, and Forestville synclines. Each of these anticlines and synclines is several miles long and has a maximum amplitude of several thousand feet.

Outcrops of the axial sections of the subsidiary folds of the Broad Mountain anticlinorium are uncommon. However, the axis of the Jugular syncline is well exposed in strip pits on the Lykens Valley Nos. 4 and 5 coal beds near the Hans Yost and Dyer Run faults in the northeastern part of the Tremont quadrangle (GQ–692). The axis of the Heckscherville syncline is also well exposed in strip pits on several coal beds west of Mount Pleasant in the Minersville quadrangle (GQ–690). The axes of the Powder Hill anticline and Beury syncline are poorly exposed in the canyon of Dyer Run in the same quadrangle. Several strip pits on the Buck Mountain (No. 5) and Seven Foot (No. 6) coal beds have exposed the axis of the Dam syncline in the eastern part of the Tremont quadrangle. In addition, the axes of the Forestville syncline and Swatara anticline are exposed in the canyon of Swatara Creek and in strip pits on the Buck Mountain, Bottom Split (No. 8), and Middle Split (No. 8½) coal beds in the western part of the Minersville quadrangle. The axis of the Crystal Run anticline is well exposed in the canyon of Crystal Run north of Forestville (GQ–690).

The dips of the flanks of the Broad Mountain anticlinorium are usually less than 45° near the crest. Lower on the flanks where subsidiary folds are common, however, dips probably average about 50° to 60° and may be as steep as 90°.

Rocks as old as the Keyser Limestone of Silurian and Devonian age and as young as the Pocono Formation of Early Mississippian age crop out along the crest of Broad Mountain anticlinorium, whereas on the flanks, the rocks are as young as the Middle and Upper Pennsylvanian Llewellyn Formation.

The crestal folds of the Broad Mountain anticlinorium plunge east-northeast at an average rate of 6° to 6½° or about 600 feet per mile. The amplitude of the anticlinorium as measured on the base of the Mauch Chunk Formation ranges from 13,000 to 18,000 feet in the western part of the area and is about 10,000 feet in the eastern part.

NEW BLOOMFIELD ANTICLINORIUM

The New Bloomfield anticlinorium is one of the larger structural features in the area. It trends N. 60° to 75° E. across the southern part of the Lykens quadrangle, the central part of the Tower City quadrangle, and the northern part of the Pine Grove quadrangle and plunges out near Tremont in the southeastern part of the Tremont quadrangle (GQ–691, 692, 698, 701). It extends many miles southwest of the area to the vicinity of New Bloomfield, for which it was named (Miller, J. T., 1961, p. 35).

The crestal fold of the anticlinorium in the area is the Joliet anticline. This anticline dies out west of the area where the crestal fold is the Berry Mountain anticline. Subsidiary folds of this anticlinorium are less numerous than they are on the Broad Mountain anticlinorium. The principal folds are the West Big Lick Mountain, Georges Head, and East Georges Head anticlines and the western part of Big Lick Mountain anticline and Rattling Creek, Tremont, and Lorberry synclines. Each of these folds is several miles long and has an amplitude ranging from 0 to 1,000 feet or more.

The axis of the Joliet anticline is well exposed in Rausch Creek and in numerous strip pits between this creek and the point where the fold terminates near Tremont (GQ–691). Axial sections of the subsidiary folds of the anticlinorium are also relatively common. The axes of the Rattling Creek syncline and Berry Mountain anticline are well exposed south of Lykens in the canyon of Rattling Creek (GQ–701). In addition, the axes of the Tremont syncline and Big Lick Mountain anticline are well exposed in many strip pits in the Tower City (GQ–698), Pine Grove (GQ–69[illegible]), and Tremont quadrangles. Other strip pits have exposed the axes of the Lorberry syncline, Georges Head anticline, and East Georges Head anticline in the Tower City and Pine Grove quadrangles.

Dips of the flanks of the New Bloomfield anticlinorium commonly are between 20° and 60°, but locally, are nearly horizontal and in some places are overturned.

The oldest rocks exposed in the crest of the New Bloomfield anticlinorium are in the Honesdale Sandstone Member of the Catskill Formation of Late Devonian age, and the youngest are in the Llewellyn Formation of Middle and Late Pennsylvanian age. The intensity of deformation seems to increase stratigraphically upwards and eastward on the crest and flanks of the anticlinorium, and the folding is most complex where the Joliet anticline dies out.

The crestal and subsidiary folds of the anticlinorium plunge northeastward at 9° to 16° or at a rate of 850 to 1,500 feet per mile. The structural relief of the anticlinorium as measured on the base of the Pottsville Formation ranges from 0 feet where the Joliet anticline dies out near Tremont to about 12,000 feet in the southwestern part of the Lykens quadrangle.

East of the termination of the Joliet anticline (pl. 4; GQ–692), the New Bloomfield anticlinorium loses its identity as it merges into the broad downwarp of the Minersville synclinorium. Thus, the anticlinorium rises westward out of the medial part of the Minersville synclinorium, splitting that structural feature into a north and south part. As it rises, subsidiary folds form on its flank in a complex structural pattern.

MINERSVILLE SYNCLINORIUM

The Minersville synclinorium probably is the most important structural feature in the area. East of Tremont (GQ–692) it consists of a complexly folded generally downwarped area whose broad trough area trends N. 50° to 60° E. West of Tremont, the broad downwarped area is split by the eastward-plunging New Bloomfield anticlinorium. The northern split trends S. 80° W. from near Tremont to the west border of the area near Lykens (GQ–701) and is here named the north trough of the Minersville synclinorium. The southern split trends S. 50° to 60° W. from near Tremont to the southwest corner of the Tower City quadrangle (GQ–699) and is here named the south trough of the Minersville synclinorium.

The main trough and its southwestward continuation, the south trough of the synclinorium, extends about 40 miles southwest of the area to near Blaserville and a similar distance northwest of the area to near Jim Thorpe. Within the area this trough of the synclinorium is everywhere the axis of the Dauphin syncline. East of Baird Run in the Pine Grove quadrangle (GQ–691), it is overlain by the upper plate of the Blackwood thrust fault.

The north trough of the Minersville synclinorium extends about 80 miles west of the area, where it terminates in the vicinity of Fort Loudon. In the area, the north trough is the Donaldson syncline from the vicinity of Tremont to the vicinity of Lykens. Farther west to the border of the area, Shiro syncline occupies the trough.

The principal subsidiary folds of the Minersville synclinorium are the New Mines, Phoenix Park, Middle Creek, Big Lick Mountain, Branchdale, Georges Head, and East Georges Head anticlines and Phoenix Park, Tremont, Fisher, and Llewellyn synclines. Each of these folds is several miles long and has an amplitude measureable in hundreds or thousands of feet.

Only a few axes of the subsidiary folds of the Minersville synclinorium are exposed. The axis of the Fisher syncline is visible in several strip pits on the Bottom Split (No. 8) and Middle Split (No. 8½) coal beds in the western part of the Minersville quadrangle (GQ–690). The axis of the Branchdale anticline has been exposed in several strip pits on the Peach Mountain (No. 18) and Tunnel (No. 19) coal beds in the vicinity of Branchdale in the Minersville quadrangle. In addition, the axes of the Fisher syncline and Middle Creek anticline are exposed in the gorge of Swatara Creek near Zerbe.

North of the north trough of the Minersville synclinorium, the flanks of the subsidiary folds generally dip about 45°, but locally may be vertical. South of this trough, the flanks of the folds are commonly overturned northward as much as 50°–60°.

The south limb of the south or main trough of the Minersville synclinorium is overturned at most localities. This limb embraces 15,000± feet of rocks ranging in age from Late Devonian (Trimmers Rock Sandstone) to Middle and Late Pennsylvanian (Llewellyn Formation). Structurally, the limb is bordered to the north by the trough of the Dauphin syncline and the Blackwood thrust fault and to the south by the Sweet Arrow fault zone. At most places the rocks of this limb dip 60° to 89° S. overturned, but locally they dip as much as 30° S. overturned, are vertical, or dip slightly north.

At most places the Llewellyn Formation is exposed in the axial part of the Minersville synclinorium, but northwest of Lykens the upper part of the Pottsville Formation occupies the axial part of the North trough.

Regionally, the troughs of the Minersville synclinorium plunge gently northeast at 1° to 3°. The amplitude of the main, or south, trough is unknown because the south limb of the synclinorium is overturned and may be the limb of a nappe that before modern erosion structurally overlay the trough and north limb, but the 15,000± feet of overturned rocks exposed on the south limb indicates the minimum amplitude.

The structural relief on the base of the Pottsville Formation between the south trough of the synclinorium and the crest of the New Bloomfield anticlinorium ranges from 0 to 12,000 feet. The amplitude of the north trough, also on the Pottsville Formation, ranges from 13,000 to 18,000 feet.

ANTICLINES

Many hundreds of anticlines and associated synclines are present in the area. Only the larger of the anticlines are shown on the maps and sections (pl. 14; GQ–689, 690, 691, 692, 698, 699, 700, 701; Wood and Trexler, 1968b, c), and only the largest or structurally more important of these are described.

HOOFLANDER MOUNTAIN ANTICLINE

The easternmost 13 miles of Hooflander Mountain anticline lies in the area and strikes about N. 80° E. across the central part of the Klingerstown quadrangle and the west-central part of the Valley View quadrangle, where it is the crestal fold of the Broad Mountain anticlinorium (GQ–699–700). Near its eastern termination, it splits into two anticlines and an intervening syncline. This furcation takes place in the incompetent upper shale member of the Mahantango Formation and appears to be the result of disharmonic folding of shale above a single parallel fold in the competent Montebello Sandstone Member of the Mahantango Formation.

Rocks ranging in age from Silurian (lower part of Keyser Limestone) to Late Devonian (Trimmers Rock Sandstone) crop out on the crest of the Hooflander Mountain anticline, and rocks as young as the Mauch Chunk Formation are exposed on the limbs. The dips on the limbs range from about 1° to 75° and average about 40°. The anticline is asymmetric, and the south limb is generally slightly steeper than the north limb. The amplitude of the anticline on the base of the Irish Valley Member of the Catskill Formation at the west border of the area is about 16,000 feet.

FRACKVILLE ANTICLINE

The Frackville anticline trends N. 65° to 80° E. from the eastern part of Lykens quadrangle (GQ–701) to the northeast corner of Tremont quadrangle (GQ–692). Its extent in the area is about 17 miles. Throughout all but the westernmost 5 miles, it is the crestal fold of the Broad Mountain anticlinorium and is generally the principal structural feature separating the Southern and Western Middle Anthracite fields. It is en echelon with the Hooflander Mountain anticline to the west and is separated from that fold by a shallow syncline. The two anticlines plunge in opposite directions and probably formed as pressures were transferred from one weak zone to another during deformation.

Rocks that range in age from Late Devonian (Trimmers Rock Sandstone) to Early Mississippian (Pocono Formation) crop out on the crest of the Frackville anticline, and rocks as young at Late Pennsylvanian (upper part of Llewellyn Formation) crop out on the flanks.

The limbs of Frackville anticline dip from about 1° to 70° and average about 30°. The north limb of the anticline is generally somewhat steeper than the south limb, in contrast to the en echelon Hooflander Mountain anticline with its steeper south limb. The amplitude of the Frackville anticline on the base of the Pocono Formation ranges from about 6,000 feet in the northeastern part of the area to about 12,000 feet in the north-central part.

WEST WEST FALLS ANTICLINE

The West West Falls anticline trends N. 80° to 85° E. for about 19 miles from the west border of the northern part of the Lykens quadrangle (GQ–701) to the east-central part of the Tremont quadrangle (GQ–692) where it divides into the Crystal Run anticline, Forestville syncline, and Swatara anticline. These three folds in the east half of the Minersville quadrangle (GQ–690) become a part of the complexly deformed north limb of the Minersville synclinorium. The locality where the West West Falls anticline splits has been rather well exposed by erosion, strip mining, and underground mining. The reason for splitting is not understood, but the incompetent beds of the Mauch Chunk Formation were flexed into a single anticline, whereas the overlying more competent beds of the Pottsville Formation and the less competent beds of the Llewellyn Formation were flexed into a compound fold system whose structural relief decreases eastward and stratigraphically upward into the generally downwarped north limb of the Minersville synclinorium.

The West West Falls anticline is the southernmost of the large upwarp structure lying between the crest of the Broad Mountain anticlinorium and the north trough of the Minersville synclinorium. Rocks of the Mauch Chunk, Pottsville, and Llewellyn Formations crop out on the crest and limbs. The dips of the limbs range from about 1° to 80° and average about 45°. The fold is asymmetric; the south limb generally is slightly steeper than the north limb. The amplitude of the anticline on the base of the Mauch Chunk Formation seems to range from 0 feet to about 6,000 feet and averages about 3,500 feet.

BRANCHDALE ANTICLINE

The Branchdale anticline is about 6 miles long. It trends N. 60° to 80° E. from Newtown in the western part of the Minersville quadrangle (GQ–690) to the east border of the area. The north limb is overturned from near Steins to the east border. The anticline is the northernmost fold of a group in which the north limb of each fold is overturned for a considerable distance. This overturning approximately demarcates the boundary between open folds to the north and tight folds to the south.

The Branchdale anticline is strongly asymmetric; rocks on the upright south limb dip an average of 35° S., whereas rocks on the north limb are overturned and have dips as low as 57° S. Thus, rocks on the north limb have been rotated as much as 158° beyond the 35° average dip of the south limb. The anticline has an average amplitude of about 1,000 feet on the Primrose (No. 11) coal bed. Crestal sections are well exposed in strip pits on the Peach Mountain (No. 18) and Tunnel (No. 19) coal beds near Branchdale (Wood and Trexler, 1968b, sheet 2).

BIG LICK MOUNTAIN ANTICLINE

The Big Lick Mountain anticline is about 11 miles long and trends N. 60° to 75° E. from the northeastern part of the Tower City quadrangle (GQ–698) to where it terminates against the Newtown fault in the south-central part of the Minersville quadrangle (GQ–690). Locally, between Tremont and Donaldson (GQ–692) and near Newtown the north limb is overturned. Its dips range from 75° N. to 80° S. overturned, whereas the dips of the upright south limb range from 50° to 60° S. The anticline is strongly asymmetric to the north and plunges east-northeast. Its amplitude averages about 1,000 feet on the Primrose (No. 11) coal bed. Sections of the limbs and crest are well exposed in many strip pits in the Tower City, Pine Grove, Tremont, and Minersville quadrangles (Wood and Trexler, 1968b, sheets 2–4; 1968c, sheet 1).

The Newtown fault and Big Lick Mountain anticline separate major structural provinces (provinces 3 and 4, Arndt and Wood, 1960, p. B183) in the Minersville quadrangle (GQ–690). North of this fold and fault the Minersville synclinorium is characterized by upright folds, numerous short high-angle reverse faults, and low-angle folded thrust faults. In contrast, to the south, the synclinorium is characterized by few complete folds, by numerous long high-angle reverse faults, by an apparent lack of low-angle folded thrust faults, and by alternating fault-bounded belts of upright and overturned rocks. A short distance west of the Minersville quadrangle, the boundary between structural provinces 3 and 4 leaves the crest of the Big Lick Mountain anticline, trends southwestward across several minor folds and faults that underlie the borough of Tremont, and intersects the trough of Llewellyn syncline (pl. 4; GQ–692; Wood and Trexler, 1968b, sheet 3). The boundary follows this trough westward to Lower Rausch Creek (Wood and Trexler, 1968b, sheet 4). At the latter point, the boundary is arbitrarily placed along Lower Rausch Creek transverse to the structural grain to where the creek intersects the Blackwood fault. It then follows Blackwood fault westward to the trough of the Dauphin syncline. West of the intersection of the Dauphin syncline and Blackwood fault, the boundary follows the trough of the syncline (pl. 4; GQ–692; Wood and Trexler, 1968b, sheet 4; 1968c, sheets 1, 2).

JOLIET ANTICLINE

The Joliet anticline is about 20 miles long and trends N. 55° to 75° E. from the south border of the Lykens quadrangle to the southeast corner of the Tremont quadrangle. It is the crestal fold of the New Bloomfield anticlinorium and is the largest anticline in the southwestern part of the area. The amplitude on the base of the Pottsville Formation ranges from 0 feet near Tremont to about 12,000 feet in the southwestern part of the Lykens quadrangle.

West of Lower Rausch Creek (GQ–691, 698, 701) the Joliet anticline is a broad relatively symmetrical flexure, but to the east, it gradually becomes sharper and asymmetrical, the north limb being slightly steeper. The dips of both limbs range from about 1° to 70° and average about 30°. Crestal sections are well exposed in the valley of Lower Rausch Creek and in strip pits a short distance to the east on the Bottom Split (No. 8), Middle Split (No. 8½), Holmes (No. 10), and Primrose (No. 11) coal beds (Wood and Trexler, 1968b, sheet 4). Between Joliet and Lower Rausch Creek the axis is well known in the subsurface because of mining on the Lykens Valley Nos. 1–4 coal beds.

ROEDERSVILLE ANTICLINE

The Roedersville anticline is the southernmost large upwarp in the area. It trends about N. 75° E. for nearly 8 miles across the eastern part of the Pine Grove and Swatara Hill quadrangles (GQ–691, 689). The fold is not well exposed, but structural attitudes and the stratigraphic succession prove its existence.

Rocks as old as Late Silurian (Bloomsburg Red Beds) and as young as Late Devonian (Trimmers Rock Sandstone) crop out on the crest and flanks of the Roedersville anticline. The structural relief of the anticline in the upper plate of the Sweet Arrow fault zone averages about 1,500 feet.

The Roedersville anticline, Sweet Arrow fault zone, Blackwood fault, the overturned south limb of the Dauphin syncline, and the Pine Grove syncline probably formed at approximately the same time. As shown on the structure sections (GQ–689, 691), a part of the Roedersville anticline is believed to lie in the subsurface in the lower plate of the Sweet Arrow fault zone about 1 mile south of the trace of the zone. The placement of a part of the anticline in the upper plate and a part in the lower plate of the Sweet Arrow fault zone probably formed as follows. During the Appalachian orogeny, the area between the present trough of the Minersville synclinorium and Blue Mountain was subjected to gradually increasing deformation. The Roedersville upwarp, flanked to the north by the Dauphin downwarp, formed gradually. The limb between the anticline and syncline was uplifted gradually and oversteepen to the north. During the oversteepening, the major low-angle Blackwood thrust fault crosscut the limb, the trough of the syncline, the crest of the anticline, and the south limb of the anticline. The trough of the syncline deepened as compression and faulting continued. The anticline grew in amplitude, and the plane of the fault was warped gently over the crest of the anticline. Movement on the fault, therefore, was impeded, and the deforming forces sought release through the northward-spooning low-angle Sweet Arrow fault zone which formed at the crest and on the south limb of the Roedersville anticline. As the fault zone formed, rocks in the upper plate slid northwestward from their original position on the crest and south limb of the Roedersville anticline to their present position above the overturned rocks of the common limb of the anticline and Dauphin syncline in the lower plate.

SYNCLINES

Many synclines are present in the area, but only the larger are shown on the maps and sections (pl. 4; GQ–689, 690, 691, 692, 698, 699, 700, 701; Wood and Trexler, 1968b, c). In addition, only the larger or structurally more important of these are described.

HECKSCHERVILLE SYNCLINE

The Heckscherville syncline trends N. 75° E. for about 6 miles from a point about 3 miles west of Buck Run in the Minersville quadrangle to the east border of the area. It is the northernmost heavily mined syncline in the west-central part of the Southern Anthracite field.

Rocks of the Pottsville and Llewellyn Formations crop out on the limbs and in the trough of the syncline, both of which have been exposed at many places, particularly west of Mount Pleasant in strip pits on the Bottom Split (No. 8), Top Split (No. 9), Holmes (No. 10), and Primrose (No. 11) coal beds (Wood and Trexler, 1968b, sheet 1). The axis and limbs are also well known in the subsurface from workings of the Glendower, Thomaston, Payne, and Pine Knot mines.

The amplitude of the Heckscherville syncline averages about 2,000 feet, and the geometry of the fold is probably better known than that of any other syncline in the area. Dips on the limbs range from about 10° to 70° and average about 50°. Mining data indicate that the syncline is nearly symmetrical and that it plunges gently east at 1° to 5°. The geometric configuration, the sequence of coal beds, and the relationship of the syncline to nearby folded low-angle thrust faults are shown on the quadrangle and coal maps (GQ–690; Wood and Trexler, 1968b, sheet 1) and on the structure sections (Wood and Trexler, 1968b, sheet 5, sections *A–E*).

Underground in the Thomaston mine near Heckscherville, the trough of the syncline has been disharmonically folded into two synclines and an anticline (Wood and Trexler, 1968b, sheet 5, section *B*). The disharmonic relations seem to have been caused by differential bedding-plane slippage between coal, shale, and sandstone sequences. Locally, the thickness of coal beds and intervening rocks have been increased several times by the differential slippage.

LITTLE MOUNTAIN SYNCLINE

The Little Mountain syncline trends N. 55° to 70° E. for about 5 miles from the central part of the Tremont quadrangle to the west-central part of the Minersville quadrangle (GQ–692, 690; Wood and Trexler, 1968b, sheets 1–3). The Heckscherville syncline is en echelon with the Little Mountain syncline to the east. Mine workings indicate that as the Heckscherville syncline dies westward, Little Mountain syncline dies eastward. The forces causing deformation apparently were transferred betweent the two major folds across two minor intervening anticlines and a syncline.

Rocks on the limbs of Little Mountain syncline dip from about 10° to 80°. The amplitude of the fold on the Buck Mountain coal bed ranges from 0 feet at its terminations to 2,500 feet in its midpart. Strata from the middle part of the Tumbling Run Member of the Pottsville Formation to the Top Split (No. 9) coal bed of the Llewellyn Formation crop out in the trough of the syncline, and rocks of the middle member of the Mauch Chunk Formation are intermittently and poorly exposed on the limbs. Axial sections are exposed in several strip pits along the fold.

The Little Mountain syncline plunges eastward at 2° to 4°. Near the common border of the Tremont and Minersville quadrangles, it is linked structurally to the Peaked Mountain anticline by a subsidiary anticline and syncline. The linking folds plunge westward in the opposite direction from the plunge of the major folds.

NORTH LITTLE MOUNTAIN SYNCLINE

The North Little Mountain syncline trends N. 60° to 80° E. from the east-central Valley View quadrangle (GQ–699) to the south-central Tremont quadrangle (GQ–692). It is en echelon with the Little Mountain syncline to the east and the Deep Creek syncline to the west. It is about 5 miles long and has an amplitude of 0 to 2,500 feet on the base of the Pottsville Formation. The limbs of the syncline dip from 1° to 50° and average about 40°.

An axial section of the North Little Mountain syncline is exposed where the paved road between Hegins and Weishample crosses Little Mountain. Rocks as old as the middle member of the Mauch Chunk Formation and as young as the Tumbling Run Member of the Pottsville Formation crop out in the trough and on the limbs of the syncline.

DEEP CREEK SYNCLINE

The Deep Creek syncline is about 13 miles long and trends N. 60° to 85° E. from the west border of the Klingerstown quadrangle (GQ–700) to the southeastern part of the Valley View quadrangle (GQ–699). It is en echelon with the North Little Mountain syncline. This latter fold is en echelon with the Little Mountain syncline, which is, in turn, en echelon with the Heckscherville syncline. The Deep Creek-Heckscherville synclines and the Shiro-Donaldson synclines are the only fold systems that extend completely across the area from west to east.

The middle member of the Mauch Chunk Formation is at the surface throughout the length of the Deep Creek syncline. Rocks of this member are relatively soft and easily eroded. As a result, the syncline underlies a wide valley, in contrast to most other synclines in the area.

The Deep Creek syncline is one of the more tightly folded flexures north of the Minersville synclinorium. Its limbs and trough have been wrinkled into many subsidiary folds, some of which are large enough to be shown on maps and structure sections (pl. 4; GQ–699, 700).

The trough of the Deep Creek syncline is not well exposed at any locality, but the existence of the fold is authenticated by numerous outcrops on the limbs. Dips at these outcrops range from 10° to 80° and average about 20°. The amplitude of the syncline on the base of the Mauch Chunk Formation seems to range from 0 feet to about 2,500 feet and to average about 2,000 feet.

DONALDSON SYNCLINE

The Donaldson syncline trends N. 65° to east-west for about 27 miles from the northwestern part of the Lykens quadrangle to the east-central border of (pl. 4; GQ–690, 691, 692, 698, 701; Wood and Trexler, 1968b, c) Minersville quadrangle and extends an unknown distance east of the area. It is the north trough of the Minersville synclinorium between the northwestern part of the Lykens quadrangle and the southeastern part of the Tremont quadrangle. East of the Tremont quadrangle, the Donaldson syncline and the north trough merge imperceptibly into the series of folds on the north limb of the main trough of the synclinorium as the structural relief is reduced gradually.

The axis of the Donaldson syncline is visible only in the northeastern part of the Tower City quadrangle several miles east of Rausch Creek Gap, where an axial section is exposed in a strip pit on the Tracy (No. 16) coal bed (Wood and Trexler, 1968c, sheet 1). Many hundreds of manmade and natural outcrops are present on the limbs. The trough generally has been eroded into a series of valleys and is surfaced by rocks of the Llewellyn Formation, except in the westernmost 1 mile where rocks of the Pottsville Formation crop out (GQ–701; Wood and Trexler, 1968c, sheet 3). The limbs are composed of rocks ranging stratigraphically from the middle member of the Mauch Chunk Formation to the 23½ coal bed of the Llewellyn Formation.

West of the John Veith shafts in the central part of the Minersville quadrangle (Wood and Trexler, 1968b, sheet 2), Donaldson syncline is generally an isoclinal fold. To the east it is asymmetrical, and its south limb is overturned. This limb is also overturned near Donaldson in the Tremont quadrangle (GQ–692). The dip of the north limb ranges from about 10° to 70° S. and averages about 45°; the dip of the south limb ranges from about 5° N. to 57° S. overturned.

The Donaldson syncline is the deepest downwarp in the area, with the exception of the Dauphin syncline. Its amplitude on the Buck Mountain (No. 5) coal bed ranges from 0 feet at its west termination to about 7,800 feet at the east border of the Lykens quadrangle (GQ–701). It maintains an amplitude of 4,000 to 7,800 feet between the east border of the Lykens quadrangle and Donaldson in the Tremont quadrangle (GQ–692). East of the Donaldson, the amplitude decreases rapidly to less than 1,000 feet over most of the Minersville quadrangle (GQ–690) and is about 700 feet at the east border of the area.

The near-surface configuration of Donaldson syncline is rather well known from much strip and underground mining. At depth, however, the axis has not been crossed by mining at any locality. The configuration at depth, therefore, has been extrapolated from near-surface data on the limbs and in the axial area (Wood and Trexler, 1968b, sheets 5, 6, sections *B–P*; 1968c, sheet 4, sections *A–P*).

TREMONT SYNCLINE

The Tremont syncline trends N. 75° to 85° E. for about 12 miles from the north-central part of Tower City quadrangle (GQ–698) to the southwestern part of the Minersville quadrangle (GQ–690). It terminates to the west on the south slope of Big Lick Mountain and is truncated to the east by the South Newtown fault. Rocks of the Pottsville Formation crop out along the axis in the western 2 miles, and those of the Llewellyn Formation crop out in the eastern 10 miles.

The axis and limbs of the Tremont syncline are well exposed at many places in the Tower City quadrangle (Wood and Trexler, 1968c, sheet 1) where coal beds from the Skidmore (No. 7) to the Orchard (No. 12) have been extensively strip and underground mined. The limbs crop out in many strip pits in the Pine Grove quadrangle (Wood and Trexler, 1968b, sheet 4) but rarely are exposed in the Tremont and Minersville quadrangles (sheets 2, 3).

The subsurface configuration of the western part of the Tremont syncline, like that of the Heckscherville and Forestville synclines, is well known. Extensive underground mine workings in the Westwood, Keefer, Tower City, Porter, and Brookside mines and numerous surface outcrops provide structural control that is exceptional for the area (Wood and Trexler, 1968c, sheet 4, sections *A–G*). Structure sections *A* through *G* illustrate the westward change of the syncline from a V-shaped fold in the midpart to a U-shaped fold near the end.

The limbs of the Tremont syncline dip from 8° to 80° and average about 60°. The fold plunges eastward at 3° to 5°, and the amplitude ranges from 0 feet to about 5,000 feet and averages about 2,500 feet.

The Tremont syncline is the southernmost syncline in the coal field whose south limb is not overturned. At its eastern termination the trough is truncated by the South Newtown fault. East of this cutoff, the rocks dip south in both fault plates, which suggests that faulting replaced folding as the stress-releasing mechanism. Structure sections (Wood and Trexler, 1968b, sheet 5, sections *E–G*) west of the cutoff indicate that the syncline loses amplitude eastward and that the trough probably did not exist in the upper plate of the fault.

DAUPHIN SYNCLINE

The Dauphin syncline is the most deeply downwarped and one of the more important synclines in the Anthracite region, although others are longer. It forms the south, or main, trough of the Minersville synclinorium and extends at least 40 miles southeast of the area, and an unknown distance eastward beneath the Blackwood fault. It strikes N. 50° to 65° E. from the southwestern part of the Tower City quadrangle to the west-central part of the Pine Grove quadrangle (pl. 4; GQ–691, 698; Wood and Trexler, 1968b, sheet 4; 1968c, sheet 2). The trough and south limb are truncated by the Blackwood fault near the junction of West Branch Fishing Creek and Baird Run (pl. 4; GQ–691; Wood and Trexler, 1968b, sheet 4). East of this point to the east border of the area, the trough lies below the fault plane; the north limb is continuous, whereas the south limb is broken by several large tear faults.

The Llewellyn Formation is exposed in the trough of Dauphin syncline, and rocks ranging in age from Middle Devonian (Mahantango Formation) to Late Pennsylvanian (upper part of Llewellyn Formation in part) form the limbs. The axis is everywhere concealed but numerous strip pits and extensive underground mine workings on the limbs show the near surface configuration of the syncline and indicate an eastward plunge of ¼° to 3°. The dip of the north limb ranges from 30° to 85° S. and averages about 50° S. At most localities the south limb is overturned northward, and the dip is 30° to 89° S. and averages about 75° S.; locally however, it is vertical or inclined steeply north.

The amplitude of the Dauphin syncline is unknown because the south limb has been overturned and thrust faulted, but it is at least 15,000 feet in the eastern part of the area. The structural relief between the trough and the crest of the adjacent Joliet anticline is about 12,000 feet in the western part of the area (GQ–701).

The subsurface configuration of the Dauphin syncline is illustrated on structure sections (GQ–690, 691, 698; Wood and Trexler, 1968b, sheets 5, 6, sections *B–P*; 1968c, sheet 4, sections *A–V*). West of the intersection of the trough and Blackwood thrust fault, the fold is asymmetric, the rocks in the south limb being slightly overturned northward. East of the intersection, the rocks in the upper and lower plates of the fault are overturned. Because the rocks in the upper plate represent a truncated part of the south limb of the Dauphin syncline which has been displaced northward, it is believed that the syncline is strongly overturned beneath the fault. As shown on the structure sections (GQ–690, 691, 698), the axial plane may approach recumbency beneath Sharp and Second Mountains.

PINE GROVE SYNCLINE

The Pine Grove syncline is the southernmost large downwarp north of the Great Valley in the Ridge and Valley province. It trends N. 55° to 80° E. for about 16 miles from the southeast corner of the Tower City quadrangle to the east-central border of the Swatara Hill quadrangle (GQ–689, 691, 698). East of the area, it extends at least 12 miles to the vicinity of New Ringgold. The trough and north limb are truncated in the southeastern part of the Tower City quadrangle by the Sweet Arrow fault zone, and the extent of the south limb to the southwest is uncertain.

The trough of Pine Grove syncline is underlain by rocks of the Trimmers Rock Sandstone and Irish Valley Member of the Catskill Formation. The limbs contain rocks ranging in age from Early Silurian (Tuscarora Sandstone) to Late Devonian (Irish Valley Member of the Catskill Formation). The trough line is more sinuous than that of most other synclines, reflecting the comparative flatness of the limbs near the axis. Dips near the axis commonly are less than 15° and rarely exceed 20°, but farther from the axis they range from 20° to 65° and generally are about 25°.

Axial sections of the Pine Grove syncline are not exposed. Several faults and smaller folds have broken and wrinkled the syncline limbs. The larger of these are the Applebee, Suedberg, Swope, and White Horse faults; the Mill and Irving synclines; and the Outwood, Suedberg, and Swope anticlines (GQ–689, 691). These structural features are believed to be confined to the upper plate of the Sweet Arrow fault zone.

The Pine Grove syncline plunges eastward at 3° to 5°. The amplitude is unknown but seems to increase eastward from a few hundred feet in the Tower City quadrangle to at least 5,000 feet in the eastern part of the Swatara Hill quadrangle.

Structurally, the Roedersville anticline and Pine Grove syncline form an isolated bulb-shaped area of apparently slightly deformed upright rocks lying above the Sweet Arrow fault zone and between overturned highly deformed rocks of the south limb of the Minersville synclinorium (south limb of the Dauphin syncline) and the generally overturned and recumbent more highly deformed rocks of the Great Valley (see Gray and others, 1960). The bulbous shape of this area is reflected by a corresponding northward curvature of the south limb of the synclinorium and the trace of the Sweet Arrow fault zone and by a southward curvature of the north side of the Great Valley.

The presence of an area of upright rocks lying in the midst of an overturned area is anomalous. It seems probable that the Pine Grove syncline and Roedersville anticline are the uneroded remnants of a lower upright limb of a nappe or recumbent fold that was moved to its present structural position along the Sweet Arrow thrust fault zone. The inverted limb and upper upright limb of the nappe have been removed by erosion in the area, but their roots may be represented by overturned and upright rocks in the Great Valley to the south. The Sweet Arrow fault zone truncates the axis of the Pine Grove syncline in the southeastern part of the Tower City quadrangle. In the Indiantown quadrangle, southeast of this point, the fault zone gradually begins to truncate overturned rocks of the south limb of the syncline or the inverted limb of the nappe. East of the report area the Roedersville anticline and Pine Grove syncline plunge out near New Ringgold. As these folds terminate, the fault zone truncates the lower upright limb and gradually cuts into the inverted limb. The lower upright limb of the nappe, or the bulbous area formed by the upright rocks of the anticline and syncline, is, structurally eliminated southwest and northeast of the report area by the Sweet Arrow fault zone cutting upward into the inverted limb.

FAULTS

Numerous thrust, high-angle reverse, underthrust, bedding-plane, and tear faults fracture the rocks of the area. Many of these are shown on the maps and sections (pl. 4; GQ–689, 690, 691, 692, 698, 699, 700, 701; Wood and Trexler, 1968b, c), but by far the majority are not shown because they are too small to depict at scales of 1:12,000 and 1:24,000, or they were not exposed.

The density of mapped faults is greatest in that part of the area underlain by anthracite. Exposures in strip pits and in the many square miles of underground mine workings have made possible the identification and tracing of faults in this part of the area. Elsewhere, the density of mapped faults is appreciably lower. In the past, this difference in density has been attributed to the coal-bearing rocks being unusually susceptible to faulting but it is probably because the faults can be identified and traced more completely in mine workings. The rocks in the larger natural exposures outside the coal-bearing area generally are cut by a multitude of faults. Most of these are small, but many that appear to be large cannot be traced beyond the limits of an outcrop.

One of the primary difficulties in fault terminology is that faults are arbitrarily termed "high" or "low angle" with reference only to their attitudes at present. The commonly accepted division between high and low angle is 45° to the horizontal. This arbitrary separation of faults at 45° is usable and meaningful only if the faults have not been folded subsequently. In this report, therefore, high- and low-angle faults are classified on the basis of whether they are believed to have dipped more or less than 45° at the time of faulting. If a fault is believed to have had an initial dip different from the present dip, reasons are presented for that belief.

Faults that trend parallel or subparallel to the regional structural grain, whose hanging wall moved upwards, and whose attitudes at the time of faulting are believed to have been less than 45°, are termed "thrusts" in this report. Many thrust faults have been folded, and their original attitudes and structural parallelism have been obscured.

Faults whose hanging walls moved upwards and whose attitudes at the time of faulting are believed to have been more than 45° are termed "high-angle reverse faults." Some have been folded or tilted; but most are not folded, or are at most, slightly warped.

Faults that trend parallel or subparallel to the regional structural grain and whose footwalls moved upward are termed "underthrusts," irrespective of their angle of dip. Underthrusts formed during all stages of the deformation of the area. Therefore, those that formed in the early stages of deformation were folded in later stages.

Faults that trend at considerable angles to the regional structural grain and whose primary movements were in general parallel to fault traces are termed "tear faults." The larger tear faults are the terminations of or the imbrications from thrust faults. Near the termination of some thrusts, the strike swings more to the south, and the slip gradually shifts from upward, to the northwest, on the thrust part to horizontal on the tear part. The other tear faults are considerably smaller and are the result of differential structural adjustments within individual thrust blocks. Most tears dip steeply to the east near the surface, but at depth the dip of the larger tears is believed to decrease gradually to nearly horizontal, as would be expected on combined thrust and tear planes.

Faults with planes parallel or nearly parallel to bedding attitudes of adjacent rocks are termed "bedding-plane faults." They seem to have formed during all stages of deformation and are, therefore, high angle and low angle and folded and nonfolded. Their parallelism to bedding and their infrequent exposure makes them difficult to date with reference to other fault types or to attempt to classify their angular relations to the horizontal at the time of faulting. The Short Mountain fault is the only bedding-plane fault shown on the maps of this report. It is identified as a thrust (GQ-701; Wood and Trexler, 1968c, sheet 3) because its plane locally cuts the bedding. Other faults of this type may be present but could not be recognized.

LOW-ANGLE FOLDED THRUST AND UNDERTHRUST FAULTS

Low-angle folded thrusts and underthrusts generally are the largest and most complex faults in the area. Some extend over and through several anticlinal crests and synclinal troughs. They are difficult to trace on the surface because of extensive cover and vegetation, the low angle between their planes and truncated strata, and the general similarity of offset strata at many localities. It is doubtful that most of the mapped folded faults would have been discerned without substantiating evidence from mine workings.

The larger folded faults are confined to the Mahantango, Mauch Chunk, and Pottsville Formations, the Trimmers Rock Sandstone, and the lower part of the Llewellyn Formation. They are confined to these formations probably because of differences in competency between sequences dominated by sandstone and conglomerate and sequences dominated by shale and coal.

In many localities where competent rocks overlay incompetent rocks, the folded low-angle faults formed as thrusts on which the upper more competent plate tore loose from the more incompetent lower plate and overrode it for varying distances. Elsewhere in those localities where incompetent rocks overlay competent rocks, the folded low-angle faults formed as underthrusts, the lower competent plate moving away from the source of compression at a faster rate than the upper incompetent plate. The incompetent plate generally crumbled into a series of folds.

FISHER RIDGE FAULT

The Fisher Ridge fault, the westernmost folded thrust fault, is on the crest and flanks of the Broad Mountain anticlinorium in the central part of the Klingerstown quadrangle (GQ-700). It seems to have originated in the upper beds of the upper shale member of the Mahantango Formation and to have cut upward into the basal and medial parts of the Trimmers Rock Sandstone. The plane of the fault is not exposed, but its existence is proven by beds both above and below being truncated and offset along a persistent discordancy which is traceable on both limbs of the Hooflander Mountain anticline and across the crest of the North Hooflander Mountain anticline as well as across the trough of the Hooflander Mountain syncline.

HEGINS FAULT

The Hegins fault may not be a single continuous fault. A fault is exposed at three localities along its hypothetical trace: north of Spring Glen in the Valley View quadrangle (GQ-699), about half a mile northwest of Coleman Church in the Klingerstown quadrangle (GQ-700) and about 0.2 mile south of Klingers School in the Tremont quadrangle (GQ-692). Elsewhere, the fault separates structurally discordant strata or is completely undiscernible between parallel beds above and below. Therefore, it has been shown as a doubtful fault throughout most of its length.

The trace of the Hegins fault is sinuous and trends generally northeastward. It enters the area in the northwest corner of the Lykens quadrangle (GQ-701) and trends about N. 80° E. for 8 miles to a point where it crosses the northeast-plunging West West Falls anticline (GQ-699). On the north flank of the anticline it trends nearly due west for slightly more than 7½ miles to where it crosses the axis of the Deep Creek syncline as a sharp horseshoe bend (GQ-700). Thence, it trends about N. 80° E. on the north flanks of the Deep Creek, North Little Mountain, and Little Mountain synclines for about 20 miles to the axis of the Hans Yost anticline in the east-central part of the Tremont quadrangle (GQ-692), where its trace reverses sharply to about N. 80° W. for about 1 mile to the axis of the Jugular syncline (GQ-692). After folding around this axis, it continues a sinuous trend of N. 30° E. for about 4 miles to the northeast corner of the Tremont quadrangle (GQ-692), where it joins the underlying Mauchono fault.

The entire trace of the Hegins fault is in the middle member of the Mauch Chunk Formation and generally separates older strata above from younger strata below. In the three places where the fault is exposed, drag folds, slickensides, offset beds, and overturned beds were observed. Elsewhere, mappable sequences in the formation that consist largely of shale or sandstone have been truncated both at the top of the lower plate and the base of the upper plate.

From its northeast end across the Tremont quadrangle and almost to the west side of the Valley View quadrangle, the Hegins fault lies at or near the base of red sandstone sequence B (GQ–690, 699) in the upper plate and several hundred feet above the same sequence in the lower plate. Between the west side of the Valley View quadrangle and a point near where the fault crosses the Deep Creek syncline (GQ–700) it separates strata several hundred feet above red sandstone unit B in the lower plate from strata of an older sandstone, unit C, in the upper plate, except for a short distance in the southeast corner of the Klingerstown quadrangle where the fault rises stratigraphically into sandstone sequence B in the upper plate. From a point on the north flank of Deep Creek syncline near where the Hegins fault crosses the axis and thence on the south flank nearly to the southeast corner of the Klingerstown quadrangle, the fault separates a younger sandstone sequence A in the lower block from strata several hundred feet below sandstone sequence C in the upper block. On the north flank of the West West Falls anticline in the southeast corner of the Klingerstown quadrangle, around the nose of the plunging anticline, and along the south flank to the west side of the Lykens quadrangle (GQ–701), the Hegins fault is principally between sandstone sequence B in the lower plate and sandstone sequence C in the upper plate.

The Hegins fault cuts somewhat irregularly through the beds of the middle member, but about 1,000 feet of strata are added to the base of the upper plate between its northeast end and the Lykens quadrangle as the fault cuts downward stratigraphically from the lower part of sandstone sequence B into the upper part of sandstone sequence C (GQ–692, 699, 700). The fault also cuts downward into the lower plate about 800 feet between its northeast end (GQ–692) and a point on the north flank of the Deep Creek syncline about 1,500 feet from where it crosses the axis (GQ–700). In this last 1,500 feet, the fault rises stratigraphically towards the axis, and about 1,000 feet of beds is added to the top of the lower plate. On the flanks of the West West Falls anticline the fault again cuts downward into the lower plate and truncates 1,500 to 2,000 feet of strata. Thus, the top of the lower plate has a stratigraphic range of at least 2,000 feet.

At most places the Hegins fault is believed to cut the bedding at an angle of 10° to 20° (GQ–690, 692, 698, 699, 701, structure sections), but locally, as in the trough of the Deep Creek syncline, the angle appears to be as great as 45° (GQ–700, structure sections).

Genetically, the Hegins fault appears to be a low-angle or slightly discordant bedding-plane thrust that passes into the subsurface to the southeast where its exact nature is concealed. The fact that the west and central parts of the fault trend from N. 80° E. to east-west, generally pass beneath the surface to the south-southeast, and generally rise above the surface to the north-northwest suggests that the upper plate of the fault along these parts of the fault moved chiefly northward. However, the upper plate along the eastern part of the fault (GQ–692) may have moved more to the west because that part of the fault trends about N. 20° E. and generally passes into the subsurface to the east-southeast.

The junctioning of the Hegins and Mauchono faults in the northeast corner of the Tremont quadrangle (GQ–692) resulted from an upward truncation of the intervening beds by the Upper Mauchono fault whereas the Hegins remained approximately parallel to bedding.

The amount of displacement on the Hegins fault is unknown. However, the minimum stratigraphic displacement is about ½ mile, and with the generally low angle of truncation of beds, the dip slip necessary to produce that displacement would be 1 to 3 miles. The actual displacement may be much greater than this minimum amount of dip slip.

MAUCHONO FAULT

The Mauchono is the only folded underthrust fault identified in the area. It is not exposed, but its presence is revealed at many places by the strong discordance of beds above and below. Elsewhere, it separates slightly discordant strata or is undiscernible in the midst of apparently concordant strata where stratigraphic units are known to be missing. Locally, because of doubt as to its existence, the Mauchono is classified as a doubtful or probable fault.

The Mauchono fault is a detachment between competent rocks below and incompetent rocks above. The detachment plane in general parallels the underlying rocks and truncates the overlying rocks. Truncated ends of the overlying rocks have not been found below the detachment. The parallelism of underlying rocks and the lack of correlative truncated ends clearly show that the Mauchono fault is not a conventional thrust. This failure to conform to conventional characteristics has baffled the authors for years. If the fault is a thrust, correlatives of the truncated ends must lie below the detachment plane. Structural relations across the fault indicate that the only places where this could occur is in the trough of the Deep Creek syncline and the north trough of the Minersville synclinorium. Surface geologic relations, however, indicate that the correlative ends are not in these troughs. The authors, therefore, reluctantly have been forced to classify the Mauchono fault as an underthrust where underlying competent beds moved toward the axis of the Broad Mountain anticlinorium, pierced deeply into the overlying incompetent beds, and at the same time crumbled and attenuated the overlying beds. If this interpretation is incorrect, the authors are at a loss to explain how the Mauchono detachment formed.

The Mauchono fault enters the area in the southeast corner of the Klingerstown quadrangle (GQ–700) and trends N. 80° to 85° E. to near the west border of the Tremont quadrangle (GQ–692). At that point the trend changes, and the fault strikes N. 50° E. to where it leaves the area at the northwest corner of the Tremont quadrangle. Throughout this distance the Mauchono lies on the south slope of Mahantango Mountain. North of the area, it folds across the crest of the east-plunging Broad Mountain anticlinorium and then trends slightly south of west on the north slope of Line Mountain to where it intersects the northern boundary of the Tremont quadrangle about midway between the east and west borders. From this point it trends approximately west on the north slope of the mountain across the northern part of the west half of the Tremont quadrangle and the northern parts of the Valley View and Klingerstown quadrangles (GQ–699, 700).

Near Erdman in the Klingerstown quadrangle the Mauchono splits into two branches for about 2,000 feet. Across much of the Tremont quadrangle it consists of a lower main fault and an upper branch known as the Upper Mauchono fault. It crops out in a fenster, or structural window, on the crest of the West West Falls anticline near Gratz (GQ–701). The oldest beds of the middle member of the Mauch Chunk that are preserved in the upper plate of the fault crop out directly west of the fenster.

Except for three localities in the Tremont quadrangle (GQ–692), the Mauchono rests everywhere upon the lower mappable shale unit of the middle member of the Mauch Chunk. The three exceptions are 1.5 to 2.0 miles southwest of Weishample, near Salem Church, and 0.7 to 0.8 mile northwest of St. John's Church. At each of these localities the fault cuts stratigraphically downwards several hundred feet through the lower shale unit into the upper part of the lower member of the Mauch Chunk. Thus, beds below the Mauchono fault are at nearly the same stratigraphic horizon throughout its extent.

The relations between the Mauchono fault, its branches, and the rocks in the upper plate are more complex than those described for the lower plate. Locally, the fault plane rises or falls with reference to the base of the upper plate, but regionally, it rises stratigraphically upwards into the upper plate to the east and to the north. For example: on the south limb of the Broad Mountain anticlinorium the fault cuts off 2,400 to 2,500 feet of the middle member of the Mauch Chunk eastward from the base of the upper plate. Also, in the fenster near Gratz (GQ–701), about 1,500 feet of strata is truncated eastward. Similarly, the fault cuts 2,700 to 2,900 feet of the middle member from the base of the upper plate northward across the anticlinorium. In fact, the amount of truncation in this direction in the eastern part of the Tremont quadrangle (GQ–692) is so great that red sandstone sequence A, which lies directly below the upper member of the Mauch Chunk on Sherman and Little Mountains, is in contact with the Mauchono fault to the north of Line Mountain.

The amount of slip on the Mauchono fault is unknown. On structure sections (GQ–690, 692, 698, 699, 700, 701) the fault is shown as an underthrust which truncates the bedding of the upper plate at angles of 5° to 25° and is virtually parallel to the bedding of the lower plate.

Genetically, the Mauchono underthrust formed in the following manner. Paleozoic rocks in and adjacent to the area were torn loose from the underlying Precambrian basement complex and transported an unknown distance to the northwest during the Appalachian orogeny. Many small to large folds formed in the Paleozoic rocks above a smooth southeast-sloping upper surface of the basement complex. The rocks in the larger folds could only move upward or laterally above the smooth surface of the complex. As a result, rocks in the cores and on the flanks of the larger anticlines moved upward toward the axial areas of the anticlines instead of downward into the trough of the adjacent synclines. The Mauchono fault formed on the crest and flanks of the Broad Mountain anticlinorium as a detachment separating less competent rocks of the middle part of the Mauch Chunk Formation in the upper plate from more competent rocks of the lower part of the formation in the lower plate. Because the more competent rocks in the lower plate could not move downward into the troughs of the adjacent synclinoriums, they moved upward to accommodate spatially to the rising anticlinorium. Inasmuch as they were the transmitters of stress, they tore loose from the overlying less competent rocks, moved upward relatively faster, and locally attenuated, pierced, and crumbled the less competent rocks. This mechanism of fault formation may have been extremely prevalent throughout the folded Appalachians, but it has not been described heretofore.

POTTCHUNK FAULT

The Pottchunk fault is believed to be the largest and most complex low-angle folded thrust in the area. On the geologic maps (pl. 4, GQ–690, 691, 692, 698, 699, 701) this fault is classified as "probable" because it has not been observed at the surface and because the regional discordancy at the Pottchunk horizon may be an angular unconformity that has been faulted at many localities.

Throughout the northern and western parts of the area, the upper member of the Mauch Chunk Formation rests discordantly upon different beds of the middle member. Although the discordancy is not exposed at the surface, it is exposed in tunnels and shafts of the Brookside mine and in the Big Lick Mountain, Gratz, Porter, and Shiro tunnels of the Tower City and Lykens quadrangles (GQ–698, 701). In each of these underground exposures the discordancy is a fault, and sedimentary structures that are suggestive of an angular unconformity are absent.

The Pottchunk fault between the south border of the Lykens quadrangle and the north border of the Minersville quadrangle lies successively on the lower and middle slopes of Stony, Big Lick, Short, Coal, Bear, and Broad Mountains (GQ–690, 701). In the northernmost parts of the Tremont and Valley View quadrangles, it lies on the south slope of Mahanoy Mountain whose crest lies north of the area (GQ–692, 699).

The trace of the Pottchunk is extremely sinuous. It zigzags back and forth with a general northeast trend between the south border of the Lykens quadrangle (GQ–701) and the north border of the Minersville quadrangle (GQ–690). Local trends of the trace, however, range from the extremes of due east to due west. The fault reenters the northern part of the area about midway along the north border of the Tremont quadrangle (GQ–692) trends slightly south of west, and leaves the area again in the northeast corner of the Valley View quadrangle (GQ–699), trending northwest. The sinuous trace results from the fault having been folded across 11 synclines and 11 anticlines. Principal among these folds are the Dauphin, Donaldson, Dam, North Little Mountain, Jugular, Beury, and New Boston synclines and the Joliet, West West Falls, Peaked Mountain, Hans Yost, Power Hill, Eisenhuth Run, and Frackville anticlines.

Throughout much of the northern and western parts of the area the plane of the Pottchunk fault lies at or near the base of the upper member of the Mauch Chunk Formation. There are three exceptions to this generalization. One is in the northwestern part of the Lykens quadrangle (GQ–701). There, the upper member thins gradually westward on Big Lick, Coal, and Short Mountains from an average thickness of about 500 feet at the east border of the quadrangle to 0 feet at Shiro tunnel. This westward thinning is due to truncation from the base up of the upper member by Pottchunk fault. However, a normal stratigraphic thickness of the member, about 500 feet, is present between splits of the fault in outcrops that lie 0.5 mile south of Shiro tunnel, and a similar thickness is preserved beneath the fault in the trough of Shiro syncline a few hundred feet west of the Lykens quadrangle. These outcrops prove beyond doubt that Pottchunk fault truncates about 500 feet of the upper member from the base of the upper plate. They also prove that, locally, the upper member lies beneath the plane of the fault. The second exception is in the northern parts of the Tremont and Valley View quadrangles (GQ–692, 699) where a few feet of the middle member overlie the fault. The third exception is in the east-central and southern parts of the Lykens quadrangle and the western and central parts of the Tower City quadrangle (GQ–698, 701). There, several hundred feet of the middle member are believed to overlie the fault.

The Pottchunk fault truncates the rocks of the lower plate irregularly at many places in the area. Although the localities where the truncations are the most pronounced invariably are obscured by talus deposits, the discordant relations between the fault and the rocks of the lower plate are determinable from the stratigraphy of the upper part of the middle member of the Mauch Chunk Formation. Red sandstone map unit A has been labeled on the maps and sections so that the discordant relations can be determined by using it as a stratigraphic marker. Unit A lies about 900 feet below the Pottchunk fault in the northwestern part of the Minersville quadrangle (GQ–690). From there to a point about 4 miles east of the west end of Little Mountain, the fault cuts downward more than 900 feet into the lower plate and through unit A (GQ–692). It then cuts upward so that it intersects unit A near the trough of the North Little Mountain syncline. From the trough to a point south of Hegins, the fault rests on unit A. From the latter point to a point midway across the Lykens quadrangle (GQ–698, 701), the fault generally overlies unit A, but locally, cuts upward or downward for short distances. West of the point midway across the Lykens quadrangle, the fault gradually cuts stratigraphically upward, so that more than 2,800 feet of the middle member overlies unit A at the west edge of the quadrangle. Thus, regionally from the northwestern part of the Minersville quadrangle to the west border of the Lykens quadrangle, more than 1,900 feet of the middle member are added to the lower plate of Pottchunk fault, and the total stratigraphic range of beds at the top of the lower plate is in excess of 2,800 feet.

The relationship between the Pottchunk fault and the lower plate is uncertain on Short, Big Lick, and Stony Mountains in the Tower City and Lykens quadrangles (GQ–698, 701).

At most places the Pottchunk fault cuts the bedding of the lower plate at an average angle of less than 10° and is parallel or subparallel to the bedding of the upper plate.

The slip of the Pottchunk fault is unknown and probably will remain unknown until the fault has been completely mapped and understood. Based on the large-scale truncation of the beds in the lower plate, the slip seems to be large and may be measureable in miles. If so, the Pottchunk is an overthrust. Within the area it is unique because it is the only large fault that has moved younger rocks in the upper plate over older rocks in the lower plate. The general northeast trend of the fault and similar trends on numerous imbricate faults suggest that the direction of movement was to the northwest.

The Pottchunk fault generally separates the incompetent middle member of the Mauch Chunk Formation from the overlying more competent upper member of the Mauch Chunk Formation and the Pottsville Formation. It seems probable that the fault formed near the horizon separating these competent and incompetent rocks because of the differential capabilities of these rocks to transmit compressive forces during the deformation of the area. The overlying competent sequence apparently tore loose or detached from the less competent sequence at the horizon where the differences in competency were the most pronounced and then moved northwestward, away from the source of compression. The underlying less competent sequence, being less capable of transmitting the compressional forces, therefore crumpled into folds which were not propagated above the fault. Assuming that this interpretation is correct, the Pottchunk fault is a detachment overthrust which moved a competent virtually unfolded sequence across an incompetent folded sequence.

UNCONFORMITY VERSUS STRUCTURAL ORIGIN OF HEGINS, MAUCHONO, AND POTTCHUNK DISCORDANCIES

Geologists visiting the Anthracite region commonly have questioned whether the Hegins, Mauchono, and Pottchunk discordances are faults, angular unconformities, or faulted angular unconformities. One of the authors of this report (Trexler, 1964) argued that the discordances are faulted unconformities. Subsequently, he and the other authors have reevaluated the discordancies and have concluded that they are faults. The following reasons led to this conclusion.

1. The discordancies are at faults wherever exposed.
2. Channel deposits, soil zones, and weathered zones in the rocks beneath the discordancies, which commonly are associated with unconformities, are absent.
3. Mauchono discordancy parallels or nearly parallels underlying strata, but is angular to overlying strata at many places. Locally, across the Board Mountain anticlinorium (GQ–692), overlying beds are truncated against the discordancy at a rate of 1 foot of stratigraphic thickness for each 4 feet of horizontal distance. The overlying beds that are angular to the discordance do not appear to be foreset beds or other types of beds deposited with an initial dip angular to the plane of discordance. The fact that this discordancy parallels underlying rocks, rather than being angular to them, and is strongly angular to overlying rocks, indicates that the Mauchono is a fault rather than an angular unconformity.
4. Some of the youngest rocks in the plate between the Hegins and Mauchono discordancies are preserved where these discordancies join in the northeastern part of Tremont quadrangle (GQ–692). About 1,200 feet of rock is added to the base of the section eastward above the Mauchono discordancy in a distance of 2½ miles west of the point of junction, and about 200 feet is added to the top of the section westward and below the Hegins discordancy. If the discordancies are eastward-merging angular unconformities, the opposing directions of erosional truncation indicate that during Mauch Chunk time a very small part of the area in the Tremont quadrangle underwent the following sequence of events: Deposition of the basal part of the middle member, erosion, tilting to the west; deposition of the medial part of the middle member, tilting to the east, erosion; and deposition of the upper part of the middle member. Such a complicated sequence of events is not recorded elsewhere in the area by these discordancies or by rocks of the middle member. The beds that terminate against the Mauchono discordancy below and that are truncated by the Hegins discordancy above are alternating units of red sandstone and red shale. The shale units are laminated and seem to

have been deposited as horizontal beds. Therefore, neither angular initial dip nor angular unconformities seem to explain the discordancies, whereas the observed relations are consistent with faulting along the discordancies.

5. The Mauchono and Pottchunk discordancies are regionally subparallel to each other but are irregularly angular to the intervening strata (GQ–690, 691, 692, 698, 699, 700, 701, structure sections). This angularity between the discordancies and intervening strata is most pronounced in the northeastern part of Tremont quadrangle (GQ–692). There, red sandstone map unit A and adjacent map units underlie the Pottchunk discordancy on Little and Broad Montains but rest on the subjacent Mauchono discordancy a few miles to the north on Line Mountain. Several thousand feet of the middle member of the Mauch Chunk that are older than unit A terminate northward against the Mauchono discordancy between Little and Line Mountains. Similarly, several thousand feet of rock younger than unit A that are not present on Little Mountain come in beneath the Pottchunk discordancy. If the discordancies are subparallel angular unconformities, these relations are difficult to understand without a southward tilting and then a northward tilting of the same small area which in reason 4 had a westward tilting followed by an eastward tilting. Obviously, the directions of tilting determined in reasons 4 and 5 are not compatible. In addition, the units that are inclined to the discordancies are too diverse lithologically and too uniform in bedding over large areas to have been deposited with pronounced initial dips. However, the relations between the subparallel discordancies and the inclined strata are expectable if they are the result of faulting and truncation of the intervening inclined strata.
6. Rocks of the middle member of the Mauch Chunk are truncated irregularly by the Pottchunk discordancy; but they do not have a distinctly different fold pattern, as would be expected if they had been deformed before the development of an unconformity. Instead, the pattern of truncation is that of a fault ridging and furrowing into its lower plate.
7. The Dyer Run, Hans Yost, Jugular, and Mine Hill faults are large well-authenticated low-angle folded thrusts whose displacements range from 500 to 4,500 feet (GQ–692, 690; Wood and Trexler, 1968b, sheets 1–3, 5). These faults extend downdip to the Pottchunk discordancy but do not extend below it. Because they do not extend below the discordancy, their large displacements must represent similar displacements along the discordancy. Thus, these faults are considered imbricate thrusts rising from a master fracture, the Pottchunk fault.
8. If the Pottchunk discordancy is an unconformity, developing at the end of the middle member time, beds of that member should not overlie the discordancy, and beds of the upper member should not underlie it. As indicated in the description of the fault, however, beds of the upper member underlie the discordancy in the western part of the Lykens quadrangle (GQ–701) and those of the middle member overlie it at places in the Lykens, Tremont, Tower City, and Valley View quadrangles (GQ–692, 698, 699, 701). These relations are consistent with a fault origin of the discordancy and argue against an unconformity.

The evidence in the preceding paragraphs indicates that the Hegins, Mauchono, and Pottchunk discordancies are faults. It also indicates that the Hegins and Mauchono faults were not emplaced on angular unconformities. However, the Pottchunk fault may have been emplaced on an unconformity that provided a convenient zone of structural weakness.

IMBRICATE FAULTS FROM THE POTTCHUNK FAULT

A series of folded low-angle imbricate faults that root in the Pottchunk fault crop out in the northeastern part of the area. From north to south they are the Hans Yost, Dyer Run, and Jugular faults and the numerous branches of the Mine Hill fault complex (pl. 4; GQ–690, 692; Wood and Trexler, 1968b, sheets 1–3). Mine and field data show that these faults break across the Pottsville and Llewellyn Formations and the upper part of the upper member of the Mauch Chunk Formation at angles of 5°± to 15°± with dip slips ranging from 500 to 4,500 feet. These faults have not been found in the middle member of the Mauch Chunk. Therefore, after cutting through the upper member they must merge into a master fault, the Pottchunk. The data also show that the dip slips on the imbricate fault decreases updip, probably because of an upward absorption of deformation by folding rather than by faulting. The locations and orientations of these faults do not seem to have been controlled in any manner by folds, which suggests that the faults formed either before or during the early stages of folding.

Several high-angle reverse faults are believed to intersect the Pottschunk fault at depth in the central part of the area and could be mistaken for imbricating fractures (GQ–690 691, 692, 698, 701). Principal among these are the Newtown, South Newtown, Red Mountain, and Tremont faults. They differ from the imbricate faults by cutting across the bedding at angles of 5° to 90° rather than at angles of 5°± to 15°± and by having slips that increase updip rather than downdip. These high-angle fractures formed at or near the crests and troughs of high folds as compression was released by faulting rather than by continued folding. Thus, they probably formed during the late stages of folding rather than before or during the early stages.

Hans Yost fault

The Hans Yost fault rises from the Pottchunk fault on the south limb of the Hans Yost anticline in the east-central part of Tremont quadrangle (GQ–692) and merges upward with Dyer Run fault on the common limb of the Jugular syncline and Powder Hill anticline about 1.8 miles east of the west border of the Minersville quadrangle (GQ–690). Between the points of separation and junction it is folded across the crest of the Hans Yost anticline and the trough of the Jugular syncline.

The Hans Yost fault is not exposed, but its trace can be mapped by scattered outcrops of displaced and deformed rocks. The greatest stratigraphic displacement is at the trough and on the north limb of the Jugular syncline in the east-central part of the Tremont quadrangle where the Lykens Valley Nos. 4 and 5 coal beds and contiguous rocks of the Tumbling Run and Schuylkill Members of the Pottsville Formation are repeated (GQ–692; Wood and Trexler, 1968b, sheet 1). The slip is difficult to determine at the surface because of the subparallel attitudes of strata across the fault. Combined drill-hole and surface geologic data, however, indicate the upper plate moved northwestward about 2,000 feet in the vicinity of the trough of the Jugular syncline. The slip at other localities is unknown, but it is believed to decrease gradually eastward to about 500 feet where the Hans Yost and Dyer Run faults join.

Dyer Run fault

The Dyer Run fault rises from the Pottchunk fault on the south limb of the Hans Yost anticline in the east-central part of the Tremont quadrangle (GQ–692). Between this point and the west border of the Minersville quadrangle (GQ–690), it is folded successively across the Hans Yost anticline and Jugular syncline and onto the south limb of the Powder Hill anticline. East of the border it lies on the south limb of the anticline and strikes about N.75°E. across the northern part of the quadrangle. The subjacent Hans Yost fault joins the Dyer Run fault about 0.5 mile northwest of Mount Pleasant.

Although the Dyer Run fault is not exposed, numerous stratigraphic offsets indicate its existence and location. The more pronounced of these offsets are at the trough of the Jugular syncline where the Lykens Valley Nos. 4 and 5 coal beds of the Tumbling Run Member of the Pottsville Formation are repeated and in the north-central part of the Minersville quadrangle where the Tumbling Run Member rests on the Schuylkill Member of the Pottsville (Wood and Trexler, 1968b, sheet 1).

The Tumbling Run overlies the Dyer Run fault everywhere except near the root, where the upper member of the Mauch Chunk locally overlies it. The fault rests successively on the upper member of the Mauch Chunk and the Tumbling Run, Schuylkill, and Sharp Mountain Members of the Pottsville in the lower plate between the root and a point about 0.5 mile northwest of Mount Pleasant (GQ–690). East of the point near Mount Pleasant, the fault cuts downsection in the lower plate into the underlying Schuylkill to a locality about one-half mile west of Dyer Run. From that point to the east border of the Minersville quadrangle, the fault rests on the Tumbling Run Member.

Surface truncations of stratigraphic units are of little value in determining the slip of the Dyer Run fault because of the nearly parallel attitudes between the fault and adjacent strata. However, in the vicinity of the trough of the Jugular syncline, combined drill-hole and surface geologic data indicate that the upper plate of the fault slipped northwestward about 2,000 feet. The drill holes that provided the subsurface data were not found in the field, but their coordinates indicate that they were in the outcrop belts of the Sharp Mountain and Schuylkill Members southeast of Mount Pleasant Fire Tower (Wood and Trexler, 1968b, sheets 1, 5). The slip of the Dyer Run fault west of the place where it joins the Hans Yost fault probably averages about 2,000 feet, but just east of that point it may reach a maximum of 2,500 feet. The slip seems to decrease gradually eastward from this maximum to about 200 feet at the east border of the Minersville quadrangle.

Jugular fault

The Jugular fault is one of the better known folded thrusts in the southern part of the Anthracite region. It was first recognized near the beginning of this century by the Reading Anthracite Co. of the Philadelphia & Reading Corp. and may have been the first thrust fault to have been recognized in the region. It rises from the Pottchunk fault on the south limb of the Hans Yost anticline in the east-central part of the Tremont quadrangle (GQ–692) and in a distance of about 2½ miles successively truncates, in the lower plate, the south limb and crest of the anticline, the south limb of the Jugular syncline, and the trough of the syncline. From the trough to the east border of the Minersville quadrangle (GQ–690) it strikes about N. 80° E., separating the north limbs of the Heckscherville and Jugular synclines.

Throughout its extent in the area, the Jugular fault has moved older rocks northward over younger rocks and it is in all respects a typical low-angle thrust. At the surface in the central and eastern parts of the Minersville quadrangle, it separates the Heckschersville syncline from the underlying Jugular syncline, and in the western part of the quadrangle and in the Tremont quadrangle it breaks the north limb of the Little Mountain syncline. In the subsurface south of these synclines, it is folded successively across the axes of the Peaked Mountain anticline, Rohresville syncline, Mine Hill anticline, Dam syncline, Crystal Run anticline, Forestville syncline, and New Mines anticline.

At the surface, the angle between the strike of the Jugular fault and the strike of adjacent rocks rarely exceeds 10°, and at depth the fault generally truncates the adjacent rocks at angles of 5° to 10° along strike and downdip. Exceptions to this general parallelism are found in the westernmost part of the fault where beds intersect the trace at nearly right angles, at several places adjacent to the fault where the upper plate has been intensively deformed, and between branches of the fault near Heckscherville and Coal Castle where the beds are overturned. Despite the inherent difficulty of determining slip where bedding and fault attitudes are nearly parallel, mine and surface geologic data indicate that the dip slip ranges from 3,000 to 4,500 feet (GQ–690; Wood and Trexler 1968b, sheet 5, sections *A–G*). North of the Heckscherville and Little Mountain synclines, the attitude, slip, and relations between the Jugular fault and truncated strata are well known from much mine and surface geologic data; but south of these synclines, the data are sparse or must be inferred. South of the synclines, therefore, the fault and its relations to truncated strata are largely projected according to the pattern of the well-controlled area.

Mine Hill fault complex

The Mine Hill fault complex consists at some places of a single fracture and at others of closely spaced branches which locally interlace (pl. 4). Considered as a whole, the complex is more widespread than any of the other imbricate faults from the Pottchunk.

In Tremont quadrangle (GQ–692; Wood and Trexler, 1968b, sheet 3) the branches of the Mine Hill complex are known from the base up as Lower Mine Hill, Hentzes, Middle Mine Hill, and Upper Mine Hill faults. These branches merge in the western and central parts of Minersville quadrangle (GQ–690; Wood and Trexler, 1968b, sheet 1) into a single fracture, the Mine Hill fault, which splits eastward in the eastern part of the quadrangle into a North, a Middle, and a South Branch.

About 2 miles northeast of Goodspring (GQ–692; Wood and Trexler, 1968b, sheet 3), on the north slope of Broad Mountain and on the south limb of the West West Falls anticline, Lower and Middle Mine Hill faults and Hentzes fault rise from the Pottchunk. They then trend generally northeastward and fold across the complex crest of the West West Falls anticline and the trough of the Dam syncline. About half a mile northeast of the trough, the Hentzes fault joins the Lower Mine Hill fault, which, with Middle Mine Hill fault, continues northeastward to a locality about half a mile east of the western boundary of the Minersville quadrangle. There, these faults join the Upper Mine Hill fault to form a single fracture, the Mine Hill fault.

The Upper Mine Hill fault does not connect with the Pottchunk at the surface, but it is interpreted to do so at depth because it joins with the other branches of the complex to form the Mine Hill fault in the western part of the Minersville quadrangle (GQ–690). It parallels the other branches on the south limb of the West West Falls anticline and across the crest of the Swatara anticline and the trough of the Forestville syncline. North of the trough, however, it diverges and is folded successively across the Crystal Run anticline and Dam syncline to the point at which it joins the other branches to form the Mine Hill fault.

Northeast of the place where the branches join, the Mine Hill fault trends along the common limb of the Peaked Mountain anticline and Dam syncline to a point about 0.4 mile southeast of Buck Run (GQ–690; Wood and Trexler, 1968b, sheet 1). There, the fault turns sharply southward and truncates in the lower plate the trough of the Rohresville syncline and the crest of the Mine Hill anticline; it also truncates several minor folds in the upper plate. It then trends generally eastward on the south limb of the Mine Hill anticline for about 1.4 miles and then it splits into a North and a South Branch. These branches continue eastward and parallel to each other to the east border of the Minersville quadrangle, a Middle Branch separating from the South Branch about 0.85 mile west of the border.

Even though the faults of the Mine Hill complex are sparsely exposed because of soil mantle, heavy forest cover, and mine waste, surface evidence of their existence is ample. Underground evidence from the Kemble, Lytle, Oak Hill, Pine Hill, and Wadesville mines and numerous smaller mines is even more abundant (Wood and Trexler, 1968b, sheets 1, 3). In the eastern part of the complex, these data provide a virtually complete three-dimensional picture of the complexities of the faulting (Wood and Trexler, 1968b, sheet 5, sections A–E). Mining activity has not been as great in the western part of the fault complex; therefore, the three-dimensional configuration of the faults is not as well authenticated (Wood and Trexler, 1968b, sheets 5, 6, sections F–M).

All branches of the Mine Hill complex except one are simple low-angle faults on which older strata have been thrust northward across younger strata. On the south limb of Dam syncline, Hentzes fault shoved younger rocks over older rocks (GQ–692; Wood and Trexler, 1968b, sheet 3). There, as shown by their lithologies and paleobotany, beds in the upper and middle parts of the Sharp Mountain Member of the Pottsville Formation rest upon beds in the upper part of the Tumbling Run Member, for the normally intervening beds of the lower part of the Sharp Mountain and all of the Schuykill Member are absent. The Hentzes fault at this locality cuts down into older beds in the upper plate; but it remains parallel to the bedding in the lower plate, so that the Schuylkill Member and part of the Sharp Mountain Member are locally eliminated rather than duplicated. Thus, the fault locally scalped the crest of the West West Falls anticline and the north limb of the Dam syncline, which strongly suggests that the fault formed after folding commenced but before folding had advanced to its present magnitude.

A klippe composed of the beds of the Schuylkill Member overlies the Upper Mine Hill fault and upright beds of the Llewellyn Formation in the core of the Dam syncline about 0.5 mile west of the place where the main trace of the fault folds across the trough. The logs of several old drill holes and isolated outcrops show that the rocks above the mined Seven Foot coal bed are characteristic of the Schuylkill rather than the Llewellyn. The old drill holes were not located, but mining company data indicate that they were situated 1,000 to 2,000 feet east of the place where structure section *H* crosses the syncline (Wood and Trexler, 1968b, sheets 3, 5). Although a fault is not exposed, the only way in which Schuylkill beds could overlie upright Llewellyn beds is to have been transported there by the Upper Mine Hill fault.

The angle between the strike of the faults of the complex and the strike of contiguous beds usually is less than 10°. Exceptions to this near parallelism are rare and occur only where the rocks have been severely deformed adjacent to faults. Because of the near parallelism the slip of the faults of the complex is difficult to determine from surface data; but the slip can be determined at most places from the abundant subsurface information. The slip of individual faults ranges from 50 to 3,000 feet, and the combined average slip of all faults generally is about 4,500 feet. Thus, if the slip of one fault decreases locally, the slip of the other faults of the complex commonly increases proportionately, so that the overall slip remains nearly constant.

Faulting of the Mine Hill complex was probably initiated on a single fracture represented by the Lower Mine Hill fault, Mine Hill fault, and North Branch of the Mine Hill fault. As faulting continued, the plane of this fracture was warped locally, and imbricate thrust faults formed near the crests of the upwarps. Additional warping produced successive imbricate faults and finally resulted in the series of closely spaced fractures of the Mine Hill complex.

LOW-ANGLE NONFOLDED THRUST AND UNDERTHRUST FAULTS

Low-angle thrust and underthrust faults that are not folded occur principally in the southeastern and northeastern parts of the area but are relatively rare. The principal ones are the Reservoir, Blackwood, Sharp Mountain, Lorberry, and Sweet Arrow faults. Each of these is younger than the folding that deformed adjacent rocks. The Sweet Arrow faults are slightly younger than the Blackwood, Lorberry, and Sharp Mountain faults, but their chronologic relationship to the Reservoir fault is unknown.

RESERVOIR FAULT

The Reservoir fault is a low-angle thrust that strikes at N. 75° to 90° E. across Broad Mountain on the common limb of the Eisenhuth Run anticline and Beury syncline in the northern part of the Minersville quadrangle (GQ–690; Wood and Trexler, 1968b, sheet 1). It splits into two branches separated by rocks of the Tumbling Run Member of the Pottsville Formation in the vicinity of Minersville reservoir.

Soil, talus, and a heavy forest obscure the Reservoir fault everywhere except in the gorge of Dyer Run near Minersville reservoir where the upper or south branch is poorly exposed. Elsewhere, the existence of the fault is inferred from anomalous thicknesses of the Tumbling Run Member of the Pottsville Formation and anomalous stratigraphic relations and structural attitudes between beds of the upper member of the Mauch Chunk Formation and the Tumbling Run.

Field data, stratigraphic data, and structural reconstructions indicate that the dip slip of the Reservoir fault ranges from 500 to 1,000 feet. The direction of net slip is unknown.

The Tumbling Run Member of the Pottsville and the upper member of the Mauch Chunk generally have been thrust northward upon themselves by the Reservoir fault in a consistent relationship of older rocks resting upon younger rocks. However, near Minersville reservoir, younger rocks rest on older rocks on the lower branch of the fault. This reversal in relationship may have originated as follows. Initial movement on Reservoir fault probably commenced with strata of the Tumbling Run being shoved northward over strata of the upper member of the Mauch Chunk along a plane that nearly paralleled bedding. Shortly after movement began, but before great displacement, an imbricate fault broke upward from the original fault near the present location of Minersville reservoir, thus forming an upper and lower branch. Movement then ceased on the lower branch but continued on the main fault and upper branch. The cessation of movement on the lower branch took place as younger rocks were being thrust over older rocks.

Several high-angle reverse faults are offset by the Reservoir fault in the canyon of Rattling Creek near Gordon reservoir (GQ–690; Wood and Trexler, 1968b, sheet 1). These faults, in turn, offset the New Boston syncline, Eisenhuth Run anticline, and Beury syncline and therefore had movement preceding the Reservoir fault and after formation of the folds. The Reservoir fault is offset by the Gordon fault a short distance east of Dyer Run on Broad Mountain. Thus, it formed after the associated folds and high-angle reverse faults and before the Gordon fault.

The Reservoir fault, if projected westward about 1,000 beyond its last point of surface control on the north slope of Broad Mountain, would intersect the Pottchunk fault acutely. It was not found on strike west of the Pottchunk fault, indicating either that it dies out in the 1,000 feet where control is lacking or that its slip of 500 to 1,000 feet was absorbed along a zone of preexisting structural weakness, the Pottchunk. It is unlikely that the Reservoir fault dies out within the 1,000 feet; therefore, it has been shown on the maps to intersect the Pottchunk. Because it is younger than the Pottchunk, however, it is not believed to split off the Pottchunk but to be a later fault whose slip was absorbed or originated in the Pottchunk zone of weakness.

BLACKWOOD FAULT

The Blackwood fault, one of the larger nonfolded thrusts in the area, lies on the north slope of Sharp Mountain and trends N. 65° to 70° E. across the Swatara Hill and Minersville quadrangles and a part of the Pine Grove quadrangle (pl. 4; GQ–689, 690, 691; Wood and Trexler, 1968b, sheets 2, 4). In Baird Run, in the central part of the latter quadrangle, it departs from this trend, gradually swinging towards the south and truncating numerous stratigraphic units as it crosses Sharp and Second Mountains. South of the latter mountain in the valley of Mill Creek, it locally strikes N. 45° W., which is the maximum departure from its regional strike. Southwest of Baird Run, the Blackwood fault is a steep southeast-dipping tear fault whose southeast side moved northeastward about 1,500 feet. East of Baird Run it is a simple southeast-dipping thrust whose upper plate slipped northwestward several thousand feet.

The trace of the Blackwood fault terminates to the southwest against the Sweet Arrow fault zone in the west-central part of the Pine Grove quadrangle (GQ–691). The extent of the fault east of the area is unknown. Even though the fault is not exposed, numerous stratigraphic truncations provide ample control for locating the trace and for proving the structural importance of the fault.

The slip of the Blackwood fault, which generally increases eastward, ranges from 600 to 5,000 feet. West of the point where it joins the Beuchler fault, the slip averages about 1,500 feet; but east of that point to the east border of the area, it increases to an average of about 3,000 feet.

The relatively straight trace of the Blackwood fault and its parallelism with slightly to moderately overturned strata in the upper plate indicate that it dips steeply south and southeast. However, it is believed to flatten to nearly horizontal at a depth of 3,000 to 6,000 feet (GQ–691, section *A–A′*). Such flattening cannot be proved but it provides the simplest explanation of the local and regional structure that is consistent with a coordinated picture of the surface and subsurface geology.

The Blackwood fault is believed to have formed simultaneously with, or shortly after, the overturning of the south limb of the Dauphin syncline, as shown by its truncation of this limb. The faults of the Sweet Arrow zone are believed to be imbricate slices that rise from the Blackwood fault as it arches over the crest of an overturned anticline which lies in the subsurface south of modern Dauphin syncline (GQ–689 and 691, structure sections). The anticline, during formation, warped the plane of the Blackwood fault and impeded movement, and as a result, the Sweet Arrow faults broke upwards.

SHARP MOUNTAIN FAULT

The Sharp Mountain fault is a simple south-dipping nonfolded thrust that lies on the south slopes of Sharp Mountain in the Pine Grove quadrangle (GQ–691). At most places the middle member of the Mauch Chunk Formation has moved northward a maximum of 2,500 feet over the upper member of the same formation and members of the Pottsville Formation.

The fault is not exposed, but its existence is inferred because the middle member of the Mauch Chunk lies adjacent to beds of the Tumbling Run, Schuylkill, and Sharp Mountain Members of the Pottsville. The upper member of the Mauch Chunk and the two lower members of the Pottsville are absent locally and must lie concealed at depth beneath the middle member and the fault.

The Blackwood, East Branch Blackwood, and Beuchler faults progressively offset the Sharp Mountain fault from west to east and are, therefore, considered to be younger than it is. As indicated by the elimination at the surface of a part of the stratigraphic sequence, the Sharp Mountain fault must dip southward less steeply than the overturned rocks that it truncates.

LORBERRY FAULT

The Lorberry fault, a small south-dipping underthrust 2.2 miles long, is situated on the north slope of Sharp Mountain near Panther Head in the north-central part of the Pine Grove quadrangle (GQ–691; Wood and Trexler, 1968b, sheet 4). The fault is not exposed, but numerous overturned anthracite beds and associated rocks of the Pottsville and Llewellyn Formations are truncated along a narrow linear zone that strikes about N. 70° E. About 1,100 feet of the Pottsville and Llewellyn Formations are truncated eastward from the lower plate, and about 300 feet are faulted off from the upper plate in the opposite direction. The stratigraphic throw on the Lorberry fault ranges from 200 to 1,500 feet, and the dip slip ranges from 700 to 2,000 feet.

The Lorberry fault terminates to the east against the Beuchler fault and to the west against the East Branch Blackwood fault, both of which are tear faults that rise from the Blackwood thrust. The footwall block of the Lorberry fault overlies the thrust. The dip slip of the thrust where overlain by the footwall block of the Lorberry fault is about 1,500 feet, which is the same as the dip slip of the thrust and East Branch Blackwood tear fault where the footwall block is absent. The similarity of slip and the confinement of the Lorberry fault by the tear faults and thrust indicate that it formed at the same time and in the same stress environment as the thrust and tear faults. It must also be a sliver from the thrust. However, the stratigraphic relations of the overturned rocks of the upper and lower plates of the Lorberry fault indicate that the lower plate was thrust northward and updip beneath the upper plate, which is the opposite of the slip of the plates of the thrust.

SWEET ARROW FAULT ZONE

The Sweet Arrow fault zone, consisting of the North Sweet Arrow and Sweet Arrow faults is the largest and structurally most important thrust zone in the region. Wood and Kehn (1961) determined that the zone is at least 80 miles long, extending from west of the Susquehanna River to near the Lehigh River. A small segment of the zone lies in the report area.

The faults of the Sweet Arrow zone are not exposed, but their existence is proved by many truncated stratigraphic units and structural features. The zone consists in some places of two faults, and at others of a single fault. The belt of truncation associated with the zone is narrow and strikes N. 50° to 75° E. across the southeastern part of the Tower City quadrangle, the central part of the Pine Grove quadrangle, and the northern part of Swatara Hill quadrangle (GQ–689, 691, 698).

The relatively straight trace of the Sweet Arrow zone and its parallelism with the moderately overturned rocks in the footwall indicate that the faults of the zone dip moderately southeastward at angles of 40° to 70°. The dip is believed to flatten to nearly horizontal at a depth of 5,000 to 7,000 feet (GQ–689, 691, structure sections). This flattening cannot be proved, but it represents the simplest explanation of structure that is consistent with an integrated concept of the surface geology.

The zone, in its westernmost 5 miles within the area, consists of a single fracture, the Sweet Arrow fault. There, the fault separates overturned strata of the Damascus Member of the Catskill Formation in the lower plate from generally upright strata of the Irish Valley Member of the Catskill and the Trimmers Rock Sandstone in the upper plate (GQ–691, 698). About 1 mile east of Mill Creek in the southwestern part of the Pine Grove quadrangle, the fault bends sharply southeastward for a short distance and cuts downsection in the upper plate through the Trimmers Rock. It then resumes the original strike to near Beuchler. In this segment it separates generally overturned beds of the Trimmers Rock in the lower plate from generally upright beds of the upper shale member of the Mahantango Formation in the upper plate. Near Beuchler, it splits into the North Sweet Arrow and Sweet Arrow faults which parallel each other to the east border of the area. The lower plate of this fault zone consists of overturned rocks of the Trimmers Rock Sandstone. The medial plate between the two faults is composed of crumpled rocks of the upper shale member of the Mahantango Formation. The upper plate of the zone is composed of many truncated stratigraphic units. Upright and overturned rocks of the Montebello Member of the Mahantango are added to the base of the upper plate between Oak Grove Union School in Pine Grove quadrangle (GQ–691) and Sweet Arrow Lake in Swatara Hill quadrangle (GQ–689). East of the lake to Conrad School, the lower shale member of the Mahantango Formation, the Marcellus Shale, the Selinsgrove Limestone, the Needmore Shale, the Ridgeley Sandstone, and the upper part of the Bloomsburg Red Beds are also added to the base of the upper plate. Bloomsburg Red Beds overlie the Sweet Arrow fault from Conrad School to a locality about half a mile west of the east border of the area where the Ridgeley Sandstone intervenes.

Many structural features are cut off by the Sweet Arrow fault zone. Principal among these are the Pine Grove syncline, Outwood anticline, Mill syncline, and Roedersville anticline in the upper plate and the Blackwood and Beuchler faults in the lower plate. Extensions of the latter faults are absent south of the zone in the upper plate. It is likely, therefore, that these faults originated shortly before or at about the same time as the Sweet Arrow zone and are part of a single fault system (GQ–691, section *A–A′*).

From the standpoint of relative age, the faults of the Sweet Arrow zone are the youngest of the large nonfolded low- to high-angle thrusts in the area. This age assignment is supported by the following facts: The faults are not folded, but they truncate many anticlines and synclines; they are younger than the overturning of the south limb of Dauphin syncline; and they are the same age or younger than the Blackwood and Beuchler faults.

The Sweet Arrow faults are believed to be imbricate thrusts from the Blackwood fault. Imbrication apparently occurred near the crest of an overturned anticline that lies in the subsurface south of Dauphin syncline.

The amount and direction of slip on the Sweet Arrow fault zone is unknown, but structural reconstructions indicate that an average minimum dip slip is about 3½ miles (GQ–689, 691, and 698, structure sections). Stratigraphic variations across the zone, such as the following example, may aid in a final determination of the net slip.

Within the area, the Irish Valley Member of the Catskill Formation is much thicker south of the Sweet Arrow fault zone than to the north, and the Trimmers Rock Sandstone contains a medial red sandstone south of the zone but not to the north. A similar thick section of the Irish Valley and a medial red sandstone in the Trimmers Rock is present north of the zone about 14 miles east of the area near New Ringgold. If the extremely different thicknessses and the medial red sandstones are correlative, the northeast-southwest slip on the zone may have been many miles. Thus, the average minimum dip slip of 3½ miles may be misleading as to the magnitude of the actual displacement on the Sweet Arrow fault zone.

HIGH-ANGLE REVERSE FAULTS

High-angle reverse faults are numerous in the area. Most dip steeply southward, but some dip steeply northward. Only the larger and structurally more important of these are herein described.

The Newtown, South Newtown, Tremont, and Red Mountain faults are high-angle reverse fractures that are believed to terminate at depth at the Pottchunk fault (GQ–690, 691, and 692, structure sections). Field and mining data indicate that these faults formed during or after the principal episode of folding. The extension of these faults downdip to the prefolding Pottchunk fault is based on the assumption that the latter provided a preexisting zone of structural weakness where later faults could root.

The high-angle reverse faults that probably root in the Pottchunk differ from the folded imbricate and nonfolded thrusts as follows:

1. They generally turncate the bedding at angles of 5° to 90° instead of at angles of 5° to 15°± as do the folded faults.
2. They commonly dip southward at angles of 60° to 85° instead of at angles less than 45°.
3. They generally crop out at or near the troughs and crests of tight folds and parallel axial planes closely at depth in contrast to the folded imbricates which were not influenced by preexisting flexures and the nonfolded thrusts which commonly cut across the limbs of folds and are not parallel to axial planes at depth.
4. Their slip increases updip, whereas the slips of the folded imbricate and nonfolded thrusts generally decrease updig or remain about the same.

The slip on numerous other high-angle reverse faults that also are associated with tightly folded synclines generally increases updip. These latter faults however, either die out before reaching the Pottchunk or cut across it at a large angle. The geometric relations between these faults and tight folds are especially well known in the Donaldson syncline in the Lykens and Tower City quadrangles (Wood and Trexler, 19[illegible]c, sheet 4, sections *A–P*), because of the extensive mining. There, the slip of most of these faults increases upward, but on a few it decreases. Trexler (1964) suggested that this reversal in slip was caused by a greater amount of folding and crumbling in the younger rocks, an upward absorption of deformation by branching of faults, and the partial formation of faults after folding had stopped. His suggestion explains adequately why the slip of most faults in the Donaldson syncline increases upward whereas that of others decreases. It also probably is applicable at many other places in the area.

Many south-dipping high-angle reverse faults break the north limb of an anticline and with increasing depth cut successively across the north limb, crest, and south limb of the anticline and finally terminate in the trough of a syncline lying to the south. In this southward crosscutting these faults generally truncate incompetent rocks of the upper and middle parts of the Llewellyn Formation near the surface; slightly more competent strata of the lower part of the Llewellyn at depths of hundreds of feet, and the much more competent rocks of the Pottsville Formation at depths of 1,000 feet or more. The troughs of the synclines where these faults terminate are usually occupied by incompetent rocks of the Llewellyn and competent rocks of the Pottsville. Initial fracturing on these faults probably began as the rocks on the flanks of the folds were attenuated by differential bedding-plane slippage during folding. As attenuation increased, faulting became more pronounced, dying out downward in thickened incompetent beds in the axial parts of synclines and increasing upward because rocks in the upper plates moved out of the compressed axial areas.

NEWTOWN FAULT

The Newtown fault trends sinuously east-northeastward across the southern half of Minersville quadrangle (GQ–690; Wood and Trexler, 1968b, sheet 2). It is not exposed, but many truncated stratigraphic units and structural features indicate that it is a south-dipping high-angle reverse fault that has a northward stratigraphic displacement of 0 to 2,600 feet and averages 2,000 feet. It dies out westward near Tremont, but east of the area its extent is unknown.

At most places the lower plate of Newtown fault is the south limb of the Branchdale anticline; the fault however, furrows deeply into the plate as it cuts westward successively across the crest of the anticline and onto the south limb of Donaldson syncline. The fault truncates in the upper plate the crest of the Big Lick Mountain anticline west of Steins, the trough of an unnamed syncline west of Newtown, and the crest of an unnamed anticline south of Minersville, and joins the South Newtown fault near Steins.

Because it truncates folds, the Newtown fault obviously formed during or after the principal episode of folding. However, it probably merges into the prefolding Pottchunk fault because its slip is too great to have been absorbed by folding or stratigraphic thickening above the horizon of that fault. The slip of the Pottchunk fault was probably increased locally by movement on the Newtown fault.

Along with other structural features, the Newtown fault divides the Minersville synclinorium into two structural subprovinces (GQ–689, 690, 691, 692; Wood and Trexler, 1968b, sheets 2–4). In the northern subprovince the folds are generally upright, folded thrust faults are relatively common, short high-angle reverse faults are common, and overturned limbs of folds are uncommon. In contrast, in the southern subprovince, complete folds are uncommon, alternating fault-bounded belts of upright and overturned rocks are common, long high-angle reverse faults near the crests and troughs of folds are common, and folded faults are sparse or absent. The boundary between the subprovinces west of the Newtown fault is discussed at length in the description of the Big Lick Mountain anticline (p. 89). The location of the boundary east of the area is uncertain.

SOUTH NEWTOWN FAULT

The South Newtown fault trends east-northeast across the southern part of the Tremont and Minersville quadrangles (GQ–690, 692; Wood and Trexler, 1968b, sheets 2, 3) between the Tremont and Newtown faults. It is not exposed, but many truncated stratigraphic units and structural features indicate that it is a south-dipping high-angle reverse fault with a northward displacement of 300 to 1,300 feet.

The lower plate of the South Newtown fault, at most places, is the north limb of the Tremont syncline. West of the village of Newtown, the trough of the syncline emerges from below the fault for a few thousand feet (GQ–690; Wood and Trexler, 1968b, sheet 2) and emerges again at Coal Run to continue westward uninterrupted (Wood and Trexler, 1968b, sheet 3). Southwest of Coal Run, the fault cuts across the south limb of the Tremont syncline and joins the Tremont fault. The upper plate of the South Newtown fault is everywhere the upright north limb of the Llewellyn syncline.

The crest, trough, and limbs of several folds are truncated by the South Newtown fault, which is therefore believed to be either younger than, or about the same age as, the folds. This fault may have formed during folding by a transfer of stress between the en echelon Llewellyn and Tremont synclines. If it formed in this manner, the South Newtown fault is contemporaneous with the folding.

Although the South Newtown fault is either contemporaneous with or postfolding in age, it probably merged at depth with the prefolding Pottchunk fault because its slip is too great to have been absorbed by folding or stratigraphic thickening above the horizon of the Pottchunk, and the slip of the Pottchunk fault probably was increased locally by the slip of the South Newtown fault.

TREMONT FAULT

The Tremont fault strikes east-northeast from Tremont to a point southeast of Steins (GQ–690, 692; Wood and Trexler, 1968b, sheets 2, 3) It is not exposed, but the relations of many truncated upright rocks in the lower plate and upright to overturned rocks in the upper plate indicate the existence of a south-dipping high-angle reverse fault. These relations indicate that the stratigraphic displacement of the fault ranges from 0 to about 1,000 feet and averages about 500 feet.

The South Newtown fault joins the Tremont fault in the borough of Tremont. East of the point of juncture, the Tremont fault separates upright rocks of the north limb of the Llewellyn syncline in the lower plate from upright and overturned rocks of the north and south limbs and trough of the syncline in the upper plate. West of the point of merger, it separates upright rocks of the south limb of the Tremont syncline from upright rocks of the north limb of an unnamed anticline.

RED MOUNTAIN FAULT

The Red Mountain fault trends east-northeast from near Lorberry in the northern part of the Pine Grove quadrangle across the southern parts of Tremont and Minersville quadrangles (pl. 4; GQ–690, 691, 692; Wood and Trexler 1968b sheets 2–4). Its extent east of the area is unknown. Although the Red Mountain fault is not exposed, many truncated stratigraphic units and structural features in both the upper and lower plates indicate that it is a south-dipping high-angle reverse fault whose stratigraphic displacement ranges from 0 to 3,000 feet and averages about 1,200 feet. Between the west end of the fault (GQ–691; Wood and Trexler 1968b, sheet 4) and the point where the crest of East Georges Head anticline is truncated (GQ–690; Wood and Trexler, 1968b, sheet 2), the lower plate is successively, from west to east, the north limb of the Georges Head anticline and the south limb of the East Georges Head anticline. East of that point to the east border of the area, it is the generally overturned south limb of the Llewellyn syncline. The upper limb is everywhere the generally upright north limb of the Dauphin syncline.

The Red Mountain fault is the same age as, or slightly younger than, the folds it truncates. Despite this age assignment it probably merged into the prefolding Pottchunk fault because its slip is too large to have been absorbed by folding above the horizon of the latter fault. Thus, the slip of the Pottchunk fault was probably increased locally by the slip of the Red Mountain fault.

ORWIN FAULT

The Orwin fault strikes east-northeast across the central parts of the Lykens and Tower City quadrangles (GQ–698, 701) and dies out in the northwestern part of Pine Grove quadrangle (GQ–691). It is not exposed in the area, but its presence is indicated by the local absence of stratigraphic units at many places along the north limbs of the Berry Mountain and Joliet anticlines. The absence of these stratigraphic units on the north limbs of these anticlines indicates that the Orwin fault is a south-dipping high-angle reverse fault whose northward stratigraphic displacement locally reached 1,000 feet but generally was about 400 feet.

The trough of a small syncline is cut off by the Orwin fault in the eastern part of the Tower City quadrangle. Other than this truncation, the only evidence that can be used to determine the age of the fault is the large angle between the fault and the strike of the lower and middle members of the Mauch Chunk Formation south and east of Tower City. The truncation and the large angle suggest that the Orwin fault deformed during or after folding.

RAUSCH CREEK FAULT COMPLEX

The Rausch Creek fault complex consists of the Rausch Creek fault and a series of anastomosing branches that strike east-northeastward across the northern part of the Lykens and Tower City quadrangles (GQ–698, 701; Wood and Trexler, 1968c, sheets 2, 3). The fractures of the complex are south-dipping high-angle reverse faults that break the north limb of the Donaldson syncline a short distance north of the trough.

The plane of the Rausch Creek fault is poorly exposed north of the water gap at Lykens on the lower slope of Coal Mountain in a strip pit on the Middle (No. 8½) and Top (No. 9) Split coal beds of the Mammoth coal zone. Elsewhere, the faults of the complex are covered by soil mantle, mine waste, and heavy forest. Extensive underground workings in the Short Mountain, Williamstown, and Valley View mines indicate that stratigraphic displacement of the faults ranges from 0 to 2,000 feet and averages about 200 and 400 feet on the larger branches and on the Rausch Creek fault, respectively. The slip of individual faults generally decreases upward, probably because the incompetent rocks of the Llewellyn Formation tended to yield by folding rather than by fracturing (Wood and Trexler, 1968c, sheet 4, sections *B–S*).

The south-dipping Rausch Creek fault complex truncates the axes of many small folds at the surface and in the subsurface; but it does not truncate the axis of the Donaldson syncline which is only a few hundred feet to the south. This fact suggests that the complex and syncline formed simultaneously and that the small folds are contemporaneous drag phenomena.

COAL MOUNTAIN FAULT COMPLEX

A series of closely connected faults strike east-northeast across the northern part of the Lykens quadrangle on the slopes of Coal Mountain and on the north limb of the Donaldson syncline (GQ–701, Wood and Trexler, 1968c, sheet 3). The North Coal Mountain fault is exposed on the north slope of the mountain in strip pits on the Lykens Nos. 4 and 5 coal beds. Elsewhere, this fault and the other faults of the complex are covered by soil mantle, mine waste, and heavy forests.

Mine records of the Short Mountain, Williamstown, and Gratz mines show that the combined displacement of the faults of the Coal Mountain complex ranges from 0 to about 2,100 feet and that the displacements of North and South Coal Mountain faults reach maximums of 500 and 1,000 feet, respectively. Records also show that the displacement of the complex increases gradually westward, reaches a maximum near Short Mountain tunnel, and then dies rapidly westward (Wood and Trexler, 1968c, sheets 3, 4).

The age of the Coal Mountain fault complex is not known with certainty because it does not truncate any major fold axes. The faults of the complex, however, do cut off the axes of several small folds which seem to be contemporaneous drag phenomena.

BIG LICK MOUNTAIN FAULT COMPLEX

The Big Lick Mountain fault complex strikes east-northeast from the western part of the Lykens quadrangle to the western part of the Tower City quadrangle on the north-dipping south limb of the Donaldson syncline (GQ–698, 701; Wood and Trexler, 1968c, sheets 1, 3). The faults of the complex are not exposed. Records of the Williamstown mine show that the faults are generally north dipping, but that in some localities they are overturned near the surface (Wood and Trexler, 1968c, sheet 4). The records also show that the displacement of the faults of the complex ranges from 0 to 2,000 feet and increases either updip or downdip. The displacements of the North and South Branches of the complex average about 200 and 600 feet, respectively.

Two small folds are truncated by a branch of the Big Lick Mountain fault complex in the upper part of the water gap north of Lykens (Wood and Trexler, 1968c, sheet 3). These flexures are undoubtedly drag folds that formed during faulting and are thus useless in determining the relations between major folding and faulting. Although the dip of the faults of the complex is generally opposite that of the faults of the Rausch Creek and Coal Mountain complexes, the close relationship between the three complexes and the core of the Donaldson syncline can be interpreted to indicate that the complexes and the syncline are for the most part contemporaneous.

APPLEBEE FAULT

The Applebee fault trends east-northeast from the southern border of the Pine Grove quadrangle to the east border of the area (GQ–689, 691). The fault follows the crest and north slope of Blue Mountain and the south limbs of the Swope anticline and Pine Grove syncline. It is not exposed, but the westward truncation and the local structural elimination of rocks of the Clinton Formation in both the foot and hanging walls indicates that the Applebee is a north-dipping high-angle reverse fault whose southward stratigraphic displacement may have been as great as 600 feet (GQ–689, and 691, structure sections).

WHITE HORSE AND ROUND HEAD FAULTS

The White Horse and Round Head faults extend from the south-central part of the Swatara Hill quadrangle to the east border of the area (GQ–689), following the upper south slope of Blue Mountain. Neither fault is exposed, but the repetition of the Tuscarora Sandstone and the Clinton Formation indicates that they are north-dipping high-angle reverse faults. The dip on these faults is uncertain; therefore, the slips are also uncertain. Structural reconstructions, however, indicate that the stratigraphic displacement of White Horse fault increases gradually eastward and may range from about 800 to 1,800 feet. They also indicate that the slip of the Round Head fault may be about 500 feet.

HIGH-ANGLE FAULTS IN THE NORTHERN PART OF THE AREA

A series of north-striking, east-dipping, high-angle reverse faults cut across the subsidiary folds of the south limb of the Broad Mountain anticlinorium (GQ–690; Wood and Trexler, 1968b, sheet 1). These faults are not exposed, but their presence is proved by offsets of stratigraphic units and fold axes. The dip slips of these faults range from 0 to about 400 feet and average about 200 feet.

Although the faults in the northeastern part of the Minersville quadrangle cut many other faults and folds, three are truncated by the Reservoir fault. Thus, they are younger than the folding of the area but are older than nonfolded thrusts, such as the Reservoir fault. Structurally, they are important because they indicate that the northeastern part of the area was compressed from east to west during the time that intervened between regional folding and latest thrust faulting. This episode of east-west compression is discernable only by these faults and by the possible westward slip of the Sweet Arrow fault zone (p. 105).

TEAR FAULTS

Tear faults are common throughout the area, but those large enough to show at scales of 1:12,000 and 1:24,000 are present only in the central and southern parts. The principal tears are the Beuchler and East Branch Blackwood faults and the southwestern part of the Blackwood fault (GQ–691). Four other smaller tear faults offset the south limb of the Dauphin syncline, striking nearly at right angles to the regional structural grain. In addition, a small tear fault is present between the

Hentzes and Middle Mine Hill faults near Donaldson in the Tremont quadrangle (GQ–692; Wood and Trexler, 1968b, sheet 3). The Blackwood fault is described in the chapter on nonfolded thrust faults. The Beuchler and East Branch Blackwood faults are described below. The other tear faults are not described because their relations are adequately shown on the geologic and coal maps.

Without exception, the larger tear faults break the overturned south limb of the Minersville synclinorium (south limb of the Dauphin syncline) and are thus younger than the overturning. Geologic relations with other structural features indicate that these faults are also, without exception, imbricate tears from the post-overturning Blackwood thrust and are the same age or slightly older than the Sweet Arrow faults. All smaller tear faults, except the one near Donaldson, also break the south limb of the synclinorium. Thus, most tear faulting took place after the overturning of this limb and at the same time, or slightly before, the last episode of thrust faulting.

BEUCHLER FAULT

The Beuchler fault is a large northwest-trending tear that merges to the north into the Blackwood thrust and ends or joins the Sweet Arrow thrusts to the south (pl. 4; GQ–691). The Beuchler breaks obliquely across a thick overturned sequence of rocks on the south limb of the Dauphin syncline in the central part of the Pine Grove quadrangle. Although the fault is not exposed, the strike slip is directly measurable on Second and Sharp Mountains where the northwest side of the fault has been displaced 2,500 to 3,500 feet northwestward.

East of the point at which the Blackwood thrust and the Beuchler tear merge, the slip of the thrust is greatly increased by the addition of the slip of the tear (Wood and Trexler, 1968b, sheet 6, sections *K* and *L*). This increase is proved by the much older beds of the Llewellyn Formation that overlie the thrust east of the intersection.

The Beuchler fault is inclined steeply to the southeast at the surface, as shown by the slight southeast curvature of the trace as it crosses the valleys adjacent to Sharp and Second Mountains. This steep southeast inclination of the fault probably decreases to near horizontal at depth as the planes of the Blackwood and Beuchler faults merge. Merging at depth must take place because of the surface merging and the increase of slip on the Blackwood thrust. In addition, because the latter fault is a tear in its southwestern part, the Beuchler fault must also be a tear imbricating from the Blackwood.

EAST BRANCH BLACKWOOD FAULT

The East Branch Blackwood fault is a northeast-trending tear in the west-central part of Pine Grove quadrangle (GQ–691; Wood and Trexler, 1968b, sheet 4) that extends from the upper headwaters of Baird Run to Fishing Creek valley. At those localities the tear merges with the Blackwood fault. The plane of the fault is not exposed, but offsets of overturned strata of the Pottsville and Llewellyn Formations on the slopes and crest of Sharp Mountain show that the fault dips steeply southeast and that the southeast side of the fault moved northeastward about 700 feet with respect to the northwest side.

Structurally, the East Branch Blackwood fault is an imbricate tear from the Blackwood fault. Southwest and northeast of the area where the East Branch Blackwood fault is present, the slip of the Blackwood fault is about 1,500 feet. About 700 feet of this 1,500-foot displacement occurs on the East Branch Blackwood tear and the remaining 800 feet on the Blackwood fault.

BEDDING-PLANE FAULTS

Bedding-plane faults are numerous in the area. Many outcrops contain one or more of these faults, some of which may have had large net slip but little or no stratigraphic throw. Whether large or small, most bedding-plane fractures cannot be correlated from locality to locality because of cover, approximate parallelism to bedding, and lack of stratigraphic displacement.

Bedding-plane faults formed throughout the long deformation of the area. Therefore, some are folded, others are slightly warped, and the rest are not folded. Many of the low-angle thrust faults of the area cut the bedding at acute angles and probably are bedding-plane faults near their roots. The net stratigraphic displacement of some of these low-angle thrusts is several thousand feet. A considerable number of faults that initially paralleled or nearly paralleled the bedding probably continued to follow differences in rock competency throughout their extent and did not, as a result, become recognizable as low-angle thrusts. These bedding-plane faults, whether large or small, thus became planes of structural detachment, generally between rocks of different competency. The net slip of some of these faults, therefore, may approximate the net slip of even the largest of the low-angle faults.

Only one bedding-plane fault, the Short Mountain fault, has been traced over an appreciable part of the area. It is shown on the maps as a thrust. No other fractures of this type could be traced.

SHORT MOUNTAIN FAULT

The Short Mountain fault is a folded bedding-plane fault on the north and south limbs of the Donaldson and Shiro synclines in the western part of Lykens quadrangle (GQ–701; Wood and Trexler, 1968c, sheet 3). It is confined between the Lykens Valley Nos. 4 and 5 coal beds of the Tumbling Run Member of the Pottsville Formation and at most places directly overlies the Lykens Valley No. 5 coal bed or is separated from this coal bed by only a few feet of rock.

The plane of the Short Mountain fault is discontinuously exposed in a series of strip pits on the south limb of the Shiro syncline for a distance of about 1 mile near the west border of Lykens quadrangle. It is also exposed on the north slope of Coal Mountain south and southwest of Gratz.

The lower 50 to 100 feet of the upper plate of the Short Mountain fault is composed of hard thick beds of sandstone and thinner beds of small-pebble conglomerate. The upper part of the lower plate consists in ascending order, of 12± feet of the Lykens Valley No. 6 coal bed; 12 feet of shale, siltstone, and sandstone; and 6± feet of the Lykens Valley No. 5 coal bed. Locally, the fault plane furrows through the Lykens Valley No. 5 coal bed or ridges, so that a few feet of carbonaceous shale are present between the coal bed and the fault plane.

The Short Mountain fault is well exposed in two strip pits a few hundred feet south of where the Donaldson syncline dies out in the western part of the Lykens quadrangle. In these pits the fault is folded over an unnamed anticline. The rocks in the upper plate dip gently southeastward on the south limb of the anticline and northward and northeastward on the north limb. In contrast, the rocks in the lower plate, although generally conforming to the anticline, have been contorted into a series of drag folds. In these folds, the Lykens Valley No. 5 coal bed commonly is partly or wholly truncated on the anticlinal crests and is fully preserved in the synclinal troughs. The trend of the axial lines of the drag folds indicates that the upper plate of the fault slipped northward, but slickensides show that the latest movement was west or northwest.

The amount of slip on the Short Mountain fault is unknown because of the lack of truncated stratigraphic or structural features that could be used to determine relative displacement. However, it seems probable that the slip does not exceed several hundred feet.

JOINTS

Joints are abundant at many outcrops, but are sparse or absent at others. They are most abundant and best formed in thin-bedded siltstone and thin- to medium-bedded sandstone, are less abundant and less well formed in thicker bedded siltstone and sandstone, and are poorly formed and widely spaced in thick-bedded conglomerate and shale.

Joints that trend nearly parallel to the strike of bedding, regardless of dip, are termed "strike joints"; those that trend nearly perpendicular to the strike of bedding and are nearly vertical are termed "dip joints." All other joints are termed "oblique."

Strike joints are present in most outcrops. They generally are planar in thin-bedded sandstone and siltstone, but are commonly irregular and curvilinear in thicker bedded conglomerate and shale.

Most dip joints are persistent smooth planes that slice cleanly through all rock constituents, such as concretions and pebbles. They are common in most exposures and tend to be slightly irregular and curvilinear in coarser grained rocks, but are less irregular and curvilinear than are either the strike or oblique joints.

Locally, oblique joints are numerous, but they are commonly less conspicuous, extensive, and consistent in strike than the other types of joints. Some are planar but most are irregular and curvilinear. They generally break around rock constituents, such as concretions and pebbles.

The authors measured 1,393 joints at 226 localities. The attitudes of the joints at these localities are shown on the tectonic map (pl. 4). These data clearly indicate that strike and dip joints dominate the joint system. However, the data are so obscured by other structural symbols on the map that a series of diagrams (fig. 41) were compiled to summarize some of the relations between joints, structural grain, and regional plunge.

Strike joints composed about 28 percent of the 1,393 measured, dip joints, about 25 percent, and oblique joints, about 47 percent. Strike and dip joints dominate the joint sets (fig. 41), except in the south half of the Pine Grove quadrangle and in the eastern half of the Tremont quadrangle. In these localities, oblique joints striking from N. 30° to 60° E. are most common.

The arithmetic mean strike of the strike joints range from N. 55° E. in the southern part of the Tower City quadrangle to N. 78° E. in the Klingerstown quadrangle and the northern half of the Tower City quadrangle; for the area as a whole, it is N. 72° E. About 65 percent of the strike joints form acute angles with overlying bedding planes in the downdip direction, about 10 percent are perpendicular to the bedding, and about 25 percent form obtuse angles with overlying bedding planes.

The average strike of bedding and the average strike of fold lines (structural grain) ranges from N. 52° E.

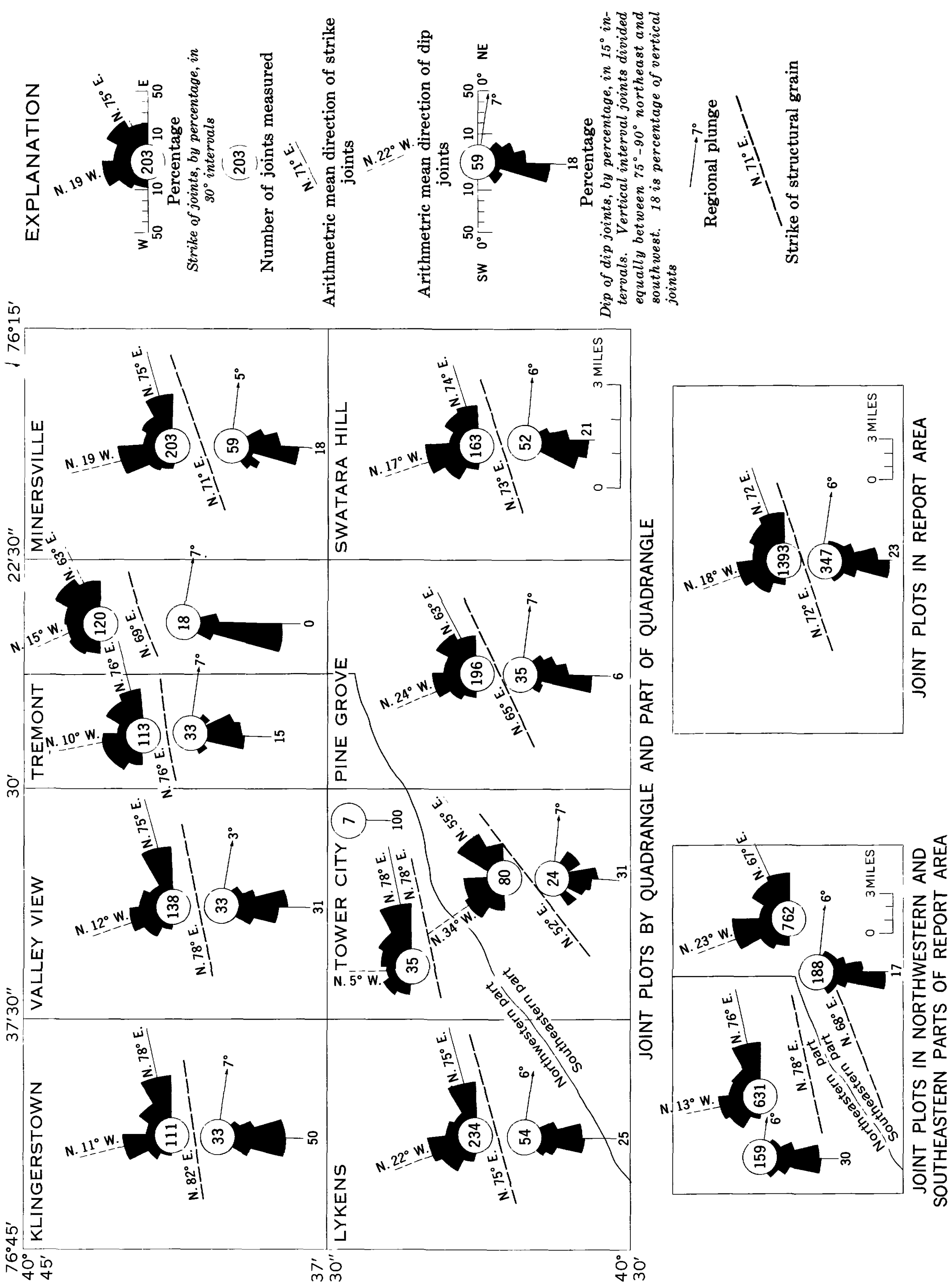


FIGURE 41.—Orientation of joints in the report area, with reference to regional plunge and strike of structural grain.

in the southern part of the Lykens and Tower City quadrangles to N. 82° E. in the Klingerstown quadrangle; it is N. 72° E. for the area as a whole. The angle between the arithmetic mean strike of the strike joints and the average structural grain ranges from 6° in the eastern half of the Tremont quadrangle to 0° in the western part and in the northern parts of the Lykens and Tower City quadrangles. The average angle between the strike joints and the structural grain is 2° if each quadrangle is considered individually, but it is 0° for the area as a whole.

The arithmetic mean strike of the dip joints ranges from N. 5° W. in the northern part of the Tower City quadrangle to N. 34° W. in the southern parts of this quadrangle and the Lykens quadrangle. It is N. 18° W. for the area as a whole, which is exactly perpendicular to the arithmetic mean strike of the strike joints and the average structural grain.

The average regional plunge of structural features is 6° in a N. 72° E. direction. Most dip joints are inclined slightly to the southwest, which is expectable, with a northeast regional plunge. Twenty-three percent of the dip joints are vertical, 29 percent hade 1° to 15° SW., and 16 percent hade more than 15° SW. In contrast, only 17 percent of these joints hade from 1° to 15° NE., and 15 percent hade more than 15° NE.

The exact parallelism of the structural grain and strike joints and the exact perpendicularity of the dip joints to the strike joints and grain indicate that both types of joints were formed by the orogenic forces that established the structural grain.

The orientations of the joints of the area do not seem to have been noticeably affected by their proximity to the low-angle folded thrust faults, but they have been greatly affected by the folding.

The age of the joints in the Appalachian Mountains and Allegheny Plateau has not been positively determined. Parker (1942, p. 392–393, 407) recognized joints sets that are similar in all respects to those of the area in eastern, central, and southern New York and northern Pennsylvania. He concluded that jointing clearly preceded folding and was the first structural effect of the Appalachian orogeny. This conclusion was based upon strike joints fanning the crests of anticlines, peridotite dikes intruding the dip joints, and minor faults offsetting both the dip and strike joints.

Sheppard (1956) reached a somewhat differing view as to the age of jointing in his study of the region between the Susquehanna River and Swatara Creek. The latter lies in the southern part of the area. He concluded:

After rotation of the bedding back to horizontal, no correlation of the joint sets was obtained, indicating that jointing had not existed prior to folding. In this area there is an absence of shear joints while a strike and dip set prevail. The attitudes of these sets suggest that jointing formed shortly after the folds were defined.

In the description of strike joints it was indicated that 65 percent form acute angles with overlying bedding planes in the downdip direction and only 25 percent form obtuse angles. This relationship indicates that these joints probably were rotated slightly parallel to their strikes by mass movement of material from the axial parts of synclines towards the axial parts of anticlines during folding. Thus, the strike joints must have been in existence before large-scale folding.

Similarly, in the description of the dip joints, it was shown that 45 percent are inclined to the southwest, whereas only 32 percent are inclined to the northeast. This suggests that these joints were rotated slightly to the southwest after formation. The average 6° NE. plunge of structural features records a regional tilting that is concordant with the degree of rotation of the joints. The dip joints, therefore, also seem to have been in existence before the folding.

The data from the area thus indicate that the strike and dip joints were in existence before large-scale folding and are unrelated to the earlier low-angle folded thrust faults. This conclusion largely supports that reached by Parker (1942) and disagrees with that of Sheppard (1956). However, it modifies Parker's conclusion as follows: the first effect of the Appalachian orogeny was not jointing; it was low-angle thrust faulting followed by jointing.

CLEAVAGE

Fracture and flow cleavage are common in incompetent rocks, but are largely absent in more competent rocks. Both are relatively common in shale and siltstone of the Mandata Member of the Helderberg Formation, the Needmore Shale, the Marcellus Shale, the lower and upper members of the Mahantango Formation, and in the more incompetent rocks of the Catskill, Mauch Chunk, and Llewellyn Formations.

Flow cleavage is best developed near the axes of folds in the southern part of the area, generally is much less well developed in the northern part, but locally, is well-developed near the axes of the more intensively deformed folds. In some outcrops it has obliterated all traces of bedding, other sedimentary features, and fossils. Throughout the area, flow cleavage generally is vertical or steeply inclined to the southeast, and at many places the intersections of the cleavage and the bedding form an excellent lineation that parallels the axes of adjacent folds. At other places the intersections are so closely spaced that linear rock fragments have formed whose bounding surfaces are the bedding and cleavage planes. These fragments are as much as 1 foot long and

1 inch thick and are commonly referred to as "pencil shale."

Most fracture cleavage is in shale and siltstone beds less than 2 feet thick, which lie between thicker more competent beds. Where the bedding is upright, it invariably dips more steeply than the bedding and parallels the axial planes of nearby folds; but where the rocks are overturned, it dips less steeply than the bedding. In the places where siltstone and shale beds are thicker than 2 feet, fracture cleavage generally is absent, and a sheeting fracture of unknown origin commonly parallels the bedding.

Much of the anthracite in the area breaks into fragments of various sizes along conchoidal fractures, although, locally, it is fracture cleaved. In the places where deformation was not too severe, the cleavage planes are widely spaced, but where deformation was more profound, the planes generally are closely spaced. Locally, where deformation was intensive, the coal lying between cleavage planes has been fractured into numerous thin conchoidal shelly fragments that have been rotated parallel to the bedding of adjacent more competent rocks, and in a few places the coal has been pulverized.

In the western and central parts of the Lykens quadrangle (Wood and Trexler, 1968c, sheet 3), the Lykens Nos. 4, 5, and 6 coal beds on the south limb of the Donaldson syncline have the characteristic blocky or cubical cleat of semianthracite. On the north limb they generally are intensively fracture cleaved and in a few places are pulverized, which suggest that there was more pressure and differential movement of beds on the north limb than on the south.

The presence or absence of fracture cleavage and cleat determines whether, during mining, anthracite breaks into blocks, large conchoidal fragments, smaller shelly fragments, or a powder. This, in turn, has greatly influenced the market value of anthracite, the procedures locally used in mining, and the relative safety of mining.

REGIONAL TECTONICS

The report area is a small part of the Appalachian Mountain system. In Pennsylvania this mountain system is divided into two structural provinces: the Folded Appalachians and the Piedmont. In general, the complexity of tectonism increases southeastward from the northwestern structural front of the mountains across the Folded Appalachians and the Piedmont to the Atlantic Coastal Plain. Igneous and metamorphic rocks of Precambrian and early Paleozoic age have been intricately deformed and altered in the Piedmont. Sedimentary rocks of Paleozoic age in the Folded Appalachians have been intensively folded and faulted, but have not been metamorphosed, except locally near the Piedmont.

The Anthracite region includes much of the northern and central parts of the Folded Appalachians in eastern Pennsylvania. Its northern boundary lies several miles north of the structural front of the mountains on the Allegheny Plateau. The southern boundary is Blue Mountain, the east border is approximately defined by the west margin of the Pocono Plateau, and the west border is roughly defined by the Susquehanna River. The report area lies in the southwestern part of the region and is characterized by a more complex structure than are most other parts.

The southeastward increase in structural complexity of the mountains is especially apparent in the Anthracite region because of the control provided by mining. Arndt and Wood (1960), after detailed mapping in the southwestern part of the region and reconnaissance elsewhere, divided the region into four structural provinces. This division is based upon the increasing structural complexity and was adopted after recognizing that structural features progressively formed in both temporal and geographic sequences.

Temporally, the structural features of the Anthracite region formed progressively in the following sequence of increasingly more complex stages as quoted from Arndt and Wood (1960, p. B182);

1. Folding of horizontal strata into broad anticlines and synclines.
2. Low-angle thrusting and imbricate faulting, followed by formation of subsidiary folds on the larger folds to develop anticlinoria and synclinoria. Additional low-angle thrusting followed by high-angle thrusting accompanied the subsidiary folding.
3. Folding of low-angle and high-angle thrusts, and offsetting of pre-existing structural features by high-angle thrusts.
4. Development of overturned folds, and offsetting of overturned folds by tear faults and high-angle thrusts.

In the sequential development of the structure of the Anthracite region, Arndt and Wood demonstrated that structural features characteristic of each temporal stage were also characteristic of geographic areas or structural provinces. They concluded that any area that contains structural features typical of a specific stage underwent deformation characteristic of each preceding stage. The provinces are as follows:

1. Northeastern part of the Anthracite region. This province is characterized by structural features typical of stage 1.
2. North-central part of the Anthracite region. This province contains structural features of stage 2 superimposed on those of stage 1.
3. South-central part of the Anthracite region. This province contains structural features of stage 3 superimposed on those of stages 1 and 2.

4. Southernmost part of Anthracite region. This province contains structural features of stage 4 superimposed upon those of stages 1–3.

The progressive temporal and geographic deformation described by Arndt and Wood is here modified so that the last episode of deformation of stage 4 is low-angle faulting rather than high-angle faulting. Wood and Kehn (1961, p. 263) incorrectly concluded that the Sweet Arrow fault zone was high angle. This conclusion guided Arndt and Wood in determining their sequence of development. Later work by Wood, Kehn, and Trexler convinced them that the fault zone was high angle near the surface but probably low angle at depth.

The report area includes parts of structural provinces 3 and 4 of Arndt and Wood. The boundary between these provinces, from east to west, across the area is successively the Newtown fault, the crest of Big Lick Mountain anticline, the structural maze beneath the borough of Tremont, the trough of the Llewellyn syncline, Rausch Creek, Blackwood fault, and the trough of the Dauphin syncline (pl. 4; GQ–690, 691, 692, 698, 701). Except for the areas at Tremont and Rausch Creek, the boundary is everyhere located at faults or at the crests and troughs of folds. In these localities, however, it is defined on the basis of geographic location rather than structural features. At most places the boundary separates upright strata to the north from alternating belts of overturned and upright strata to the south.

The controlling structural feature of the area was the Minersville synclinorium. The trough of the synclinorium, or nearby smaller structural features, separates generally upright strata from generally overturned strata. During the latter part of the orogeny it probably separated a foothills folded and faulted belt from a massif of largely overturned and intensively deformed strata. The boundary between structural provinces 3 and 4 lies at or near the trough and is one of the more important tectonic boundaries of the ancestral Appalachian Mountains, perhaps comparing locally in importance with the boundary between the Folded Appalachians and the Piedmont.

Structure sections (GQ–689, 690, 691, 692) show, with reference to the top of the middle member of the Mauch Chunk Formation, that the altitudes of the bottoms of the larger synclines descend evenly southward towards the trough of the Minersville synclinorium. Similarly, the sections show, with reference to the base of the middle member, that the altitudes of the tops of the larger anticlines descend evenly southward towards the trough. Although the rate of descent of the synclinal troughs and anticlinal crests are individually concordant, they are decidedly discordant each to the other. This discordancy is believed to have been caused by differential transmission of compressive forces through strata of differing competency. The following evidence illustrates this belief. During the early stages of the formation of the Minersville synclinorium, the deforming forces were transmitted northwestward by competent rocks above the middle member of the Mauch Chunk, by incompetent rocks of the member, and by competent rocks below the member. Because the competency of these rock units differed, they transmitted the forces northwestward differentially, and the competent rocks above the middle member tore loose from the incompetent rocks of the member along a fracture system that formed into the Pottchunk fault and its associated imbricates. Similarly, the competent rocks below the middle member tore loose along the Mauchono fault and other similar fractures. In addition, the Hegins fault formed in the medial and lower parts of the member as the result of differential adjustments.

As the deforming forces of the Appalachian orogeny increased in intensity, the Minersville synclinorium deepened, movement decreased or ceased on the Pottchunk and Mauchono faults, and the limbs of the synclinorium developed many subsidiary folds The crests of the subsidiary anticlines below the Mauchono fault and the troughs of the subsidiary synclines above the Pottschunk fault were controlled in their upward and downward formation by the preexisting structural discordencies at these faults. They also were controlled by the buffering action of the incompetent strata of the middle member. During the formation of the subsidiary folds, high-angle faulting took place, principally above the Pottchunk fault; at about the same time or shortly thereafter, the south limb of the synclinorium was overturned. This was followed by the formation of the Blackwood, Beuchler, and Sweet Arrow faults which crosscut the limb without regard to the preexisting structural grain.

Broad Mountain and New Bloomfield anticlinoria controlled the formation of many smaller structural features in the northern and western parts of the area. These large fold systems, however, do not seem to have played as important roles in the tectonic formation of the area as did the Minersville synclinorium. Nor do they appear to separate large tectonically different areas or provinces as does the synclinorium.

Reasons for the Minersville synclinorium being tectonically so important are not understood. However, as illustrated on structure sections (GQ–689, 690, 691, 692, 698, 701), the Mauch Chunk Formation is believed to thicken markedly in the trough. This thickening may have been responsible for the localization of the various types of structural features to the north and south. It is also possible that the trough was superimposed upon a structural depression or zone of weakness that formed during the Taconic orogeny.

A major tectonic fact determinable from the structure sections and geologic maps (GQ–689, 690, 691, 692) is that rocks older than the middle member of the Mauch Chunk Formation are discordant at many places with those younger than the middle member. This discordancy suggests that the middle member functioned during deformation as a tectonic buffer which separated discordantly transmitted sets of forces. This separation resulted in the formation of the Pottchunk and Mauchono faults.

DEPTH OF DEFORMATION

During the last 50 years a difference of opinion has become increasingly apparent among geologists studying the Appalachians as to whether the Precambrian basement complex is involved in the structural features present in the Paleozoic rocks. The available data are more easily interpreted to support the concept that the basement complex is not greatly involved.

The geometry of the folds that are depicted on structure sections (GQ–689, 690, 691, 692, 698, 699, 700, 701) indicates that all folds decrease in amplitude with increasing depth. Projection of the rate of decrease to depths greater than −12,000 feet illustrated on the sections shows that all but the largest folds apparently die out between 20,000 and 25,000 feet below the surface and that even these disappear at about 40,000 feet. According to Colton (1961, pl. 11), 35,000 to 40,000 feet of Paleozoic rocks underlie east-central Pennsylvania. If the concept of decreasing amplitude is correct, the Precambrian rocks probably were not involved in the folding, were not as greatly compressed laterally southeast to northwest as were the sedimentary rocks, and therefore, must be separated from the sediments by planes of detachment. Gray and others (1960, structure section *D–D′*), Gwinn (1964, p. 897), Woodward (1964, p. 351), Rodgers (1963, p. 1534–1535), and Perry (1964, p. 668) have postulated the existence of such planes in the lower part of the Paleozoic sequence in western and central Pennsylvania.

Nevin (1936, p. 65) and Ver Wiebe (1936, p. 933–936) argued that anticlines cannot form unless supported from below by the basement. According to their reasoning, folds could not form above detachment faults unless similar folds lay directly below, because voids would occur at the cores of the anticlines. Their reasoning is incorrect, because at many places rocks above and below thrust faults and detachment planes are known to be differentially folded. In many detached fault blocks, the troughs of the synclines seem to be largely limited in their downward development by detachment planes, and the strata above the planes on the limbs of anticlines seem to have flowed or to have been forced into the space previously occupied by the upward-moving anticlinal crests. The cores of the anticlines, therefore, were always supported, and voids did not exist at any time. The process of transferring strata from the limbs of anticlines to the cores logically occurred most commonly in areas where the amplitude of folding increased upward, as in the report area. The amount of increase in amplitude of an anticline is dependent upon the severity of differential compression, the amount of stratigrapic thickening in the core, and, lastly, upon the number of detachment planes existing within the confines of the anticline.

There is little positive geophysical evidence to show whether the Precambrian basement complex is involved in the deformation of the Folded Appalachians but what is known is discussed below.

Joesting, Keller, and King (1949, p. 1761) determined by an aeromagnetic survey of part of central Pennsylvania that the basement complex lies at a greater depth southeast of the Appalachian structural front than it does to the northwest. This relationship is anomalous because Lower Cambrian rocks, which normally lie only a short distance above the complex, crop out southeast of the front, and Pennsylvanian rocks, which lie thousands of feet above the Lower Cambrian rocks, crop out northwest of the front. Joesting, Keller, and King concluded that the "basement surface is not deformed concordantly with the highly folded Paleozoic rocks."

Five north-south aeromagnetic lines of the U.S. Geological Survey extend across the area from the New York-Pennsylvania border to Marietta, Pa. Magnetic anomalies due to the basement complex are prominent near the border. Depth determinations computed from these anomalies indicate that the complex lies about 13,000 to 14,000 feet below the ground surface (W. J. Dempsey, oral commun., 1960), which is in close agreement with a depth of 14,500 feet determined from stratigraphic data (Colton, 1961, fig. 4). About 30 miles south of the border these magnetic anomalies disappear, and a gentle smooth southward-inclined magnetic gradient prevails for about 90 miles to near Campbelltown. According to Dempsey, the smooth gradient may be due to any of several factors: (a) a basement complex that is not highly magnetic and does not contain units of greatly differing magnetic intensities; (b) a complex buried too deeply to influence the airborne magnetometer (more than 25,000 ft); or (c) an upper surface of the complex that slopes gently and smoothly southward and does not rise into the cores of anticlines or descend beneath synclines. These three possibilities are evaluated below.

Joesting, Keller, and King (1949, p. 1761 and fig. 7) found that in Pennsylvania the magnetic trend of the folded Appalachians is N. 60° E., which is in strong contrast to the N. 30° E. strike of Appalachian fold axes in the south-central part of the State and the N. 60°–80° E. strike in the central and eastern parts. They concluded that this divergence in trends indicates a corresponding divergence between the structural trends of the complex and the overlying Paleozoic rocks. This divergence probably indicates that the basement complex is significantly magnetic.

The five aeromagnetic lines cross several large anticlines where the basement complex, if in normal stratigraphic succession, should lie considerably less than 25,000 feet below the surface. Near Danville, on the Montour anticline, the complex should be about 13,000 feet below the surface (Colton, 1961, figs. 6, 7, 9). Near Elysburg, on the Selinsgrove anticline, the complex should be about 18,000 feet below the surface, and in the report area, near Line Mountain on the Frackville anticline, the complex should be well above the 25,000-foot limitation of the airborne magnetometer. However, the smooth gradient of the magnetics across these anticlines argues that the complex is not deformed upwards into the axial parts and strongly suggests that it is everywhere more than 25,000 feet below the surface.

It logically follows that if the complex is not upwarped in the cores of the anticlines, it is not downwarped below the synclines, and that the upper surface of the basement complex probably is relatively smooth.

The numerous folds in the Paleozoic rocks and the probable smooth upper surface of the basement complex are incompatible, unless one or more detachment planes exist. It seems highly likely, therefore, that such a detachment separates the main body of Paleozoic rocks from the basement. This is consistent with the existence of other detachment planes, such as the Mauch-ono and Pottchunk faults.

STRUCTURAL HISTORY

The first structural events recorded by the rocks of the area are of Late Ordovician age. During the Middle and Late Ordovician time the southern part and perhaps all of the area was deformed by forces of the Taconic orogeny which formed mountains to the south, southeast, and east. The structural features that formed in the area during the orogeny are largely unknown because they are buried by younger sediments.

In Silurian and Early and Middle Devonian time intermittent epeirogenic uplift extended into the area from the Auburn promontory. Recurrent erosion beveled the underlying strata and produced a series of unconformities that seem to coalesce or merge southward towards the promontory.

The region east, southeast, and south of the area was again uplifted during the Acadian orogeny, which began in latest Middle Devonian time and continued intermittently into Early Mississippian time. The pre-Pocono unconformity indicates that large-scale folding and perhaps faulting during this orogeny extended into the area in Early Mississippian time and also indicates that an ancestral Broad Mountain anticlinorium began to form.

The mountainous region east, south, and southeast of the area seems to have been low and tectonically quiescent during Middle Mississippian time. Near the end of the Mississippian, however, it was strongly uplifted and remained tectonically active with but minor periods of quiescence throughout the rest of Paleozoic time and perhaps early Mesozoic time. Mountain building caused by the Appalachian orogeny advanced northwestward into the area after the youngest preserved strata of the Llewellyn Formation of Late, but not latest Pennsylvanian age were deposited. Elsewhere in the central Appalachians, mountain building during this orogeny seems to have begun during latest Pennsylvanian or earliest Permian time (Berryhill and deWitt, 1955; Berryhill and Swanson, 1962) and to have ended by the end of Middle Triassic time.

Although the exact timing of the Appalachian orogeny is uncertain, the order of deformation of individual structural features or associated series of structural features can be determined.

The youngest sediments of Late Pennsylvanian age now present in the area were rather flat lying at the end of their deposition. Perhaps only an additional few hundred feet of sediments accumulated, but it seems more likely that many more thousands accumulated and have since been eroded off. Gentle folding occurred after these sediments were laid down either in latest Pennsylvanian or earliest Permian time. It seems likely that the earliest of these gentle folds were a rejuvenated ancestral Broad Mountain anticlinorium, the ancestral Minersville synclinorium, and the ancestral New Bloomfield anticlinorium.

The intensity of deformation gradually increased, and the ancestral Minersville synclinorium slowly formed into a large shallow open fold on whose northern limb a series of bedding-plane faults formed at a consistent horizon near the top of the Mauch Chunk Formation. As the deformation continued increasing in intensity, these bedding-plane faults finally merged into a major thrust, the Pottchunk fault. During the later stages of the formation of this thrust fault, compression was so great that the thrust plane was gradually warped into minor folds at many places, and the upper plate of the thrust fault was broken into a series of imbricate thrust slices that rose upward from the minor folds in the following order: Hans Yost, Dyer Run, and Jugular faults and the various branches of the Mine Hill fault complex.

The Mauchono fault formed penecontemporaneously with the Pottchunk on the flanks and crest of the growing Broad Mountain anticlinorium. Competent strata, which underlay the plane of the Mauchono, moved northward relative to less competent strata, which overlay the fault, and regionally pierced deeply into the less competent strata on the crest and on the upper limbs of the anticlinorium. Simultaneously with the formation of the Mauchono, a large thrust, the Hegins fault, formed in the medial part of the Mauch Chunk Formation.

After the formation of the Pottchunk, Hegins, and Mauchono faults, most of the anticlines and synclines of the area formed into virtually their present forms as subsidiary folds of the Minersville synclinorium and New Bloomfield and Broad Mountain anticlinoria. In addition, numerous high-angle reverse faults fractured the subsidiary folds, and the south limb of the synclinorium was overturned.

At about the time the Minersville synclinorium was overturned, the trough and south limb were fractured by the Blackwood, East Branch Blackwood, and Beuchler faults and by smaller tear faults. Shortly thereafter, the Sweet Arrow faults split off from the Blackwood fault near the crest of an overturned subsurface anticline and moved upright beds in the upper plate north and west across overturned beds in the lower plate.

The Blackwood, East Branch Blackwood, Beuchler, and Sweet Arrow faults are the youngest large faults. A series of small northward-trending high-angle reverse faults and the eastward-trending Reservoir fault in the northern part of the Minersville quadrangle, however, may be of approximately the same age. Although the slip of each of these high-angle reverse faults is negligible by comparison with those of the younger large faults, these and the Sweet Arrow faults show that the area was subjected to westward-directed compression during the later stages of its structural formation.

The Appalachian orogeny apparently ended in the area with the westward compression. Subsequently, the area seems to have been subjected to a long period of erosion that has lasted to the present.

ECONOMIC GEOLOGY

COAL

Anthracite and semianthracite have been the principal economic resources of the area for about 150 years, and much coal has been mined during this period.

HISTORY AND METHODS OF MINING

Early records report that anthracite was used by gunsmiths before 1755, by blacksmiths about 1769, and by nail manufacturerers in 1788 (Hudson Coal Co., 1932, p. 24–26). Despite this early utilization, it was 1808 before anthracite was successfully burnt in an open grate and 1812 before it was fired in a furnace (Hudson Coal Co., 1932, p. 26–27, 31).

A map prepared by a Mr. Schul in 1770 indicated the presence of coal near Mahanoy City and Shamokin, a short distance north of the area. However, it was 1790 before Necho Allen found anthracite in the northeastern part of the area on Broad Mountain (Hudson Coal Co., 1932, p. 29) and 1795 before a blacksmith named Whetstone used anthracite from the Southern Anthracite field in a forge. Other blacksmiths began using coal from the field in their forges about 1812. The first shipment of coal to Philadelphia was made in 1800, but profitable shipments were not possible until 1812.

Completion of Schuylkill Navigation Canal, parallel to Schuylkill River, in 1825 provided the first economical transportation to the Philadelphia market area. The canal was 108 miles long, cost $2,200,000 to build, and incorporated the first tunnel built for transportation in North America. The canal was operated at a profit for more than 50 years, but by the end of the 19th century it had fallen into disrepair and now is obliterated at most places.

The anthracite industry became economically important between 1825 and 1835. By 1837, the value of coal property in the Anthracite region had increased from practically nothing to several hundred dollars per acre, and mining and land speculation were rampant. As anthracite became increasingly more important, a railroad network gradually developed, and a lagging steel industry mushroomed, because it was for the first time independent of bituminous coal heretofore imported from Great Britain.

Railroad construction began in the Anthracite region in 1828 and soon spread throughout eastern Pennsylvania, New York, Delaware, New Jersey, and Maryland. By 1842 the final rail link with Philadelphia was completed, and the anthracite industry promptly took its place as one of the economic giants of the United States. From 1842 to 1880 the Anthracite region produced about 450 million tons of coal. The Southern Anthracite field produced about 67 million tons, or about 15 percent of this coal, and the report area produced about 27 million tons, or about 5.5 percent.

Most of the large mining companies of the Anthracite region came into existence between 1825 and 1875. The majority continued to operate until the 1930's, and several continued operations until nearly the present. All the original companies have now discontinued mining, most have liquidated their coal and mineral land holdings, and many have diversified into other types of business or industry.

Several techniques of underground mining have been used in the region. The basic mining techniques are the room-and-pillar, chute, and longwall methods. The first two of these have been used extensively in the area, and the longwall method has been used in the few places where beds are flat or gently inclined.

Most large mines employed the room-and-pillar method where the coal beds dip less than 20° and the chute method where they dip more than 20°. In either method, rooms and intervening pillars were excavated, and considerable timbering was necessary to support the roof. After the completion of a series of rooms above a haulageway, the pillars were gradually removed, and the roof rock was supported by additional timbering. If the roof rock could be adequately supported by timbering or could be induced to collapse gradually at a distance from individual pillars, all coal was removed. In most places, however, the roof was unstable and soon began to sag or crack, and mining was discontinued after removing only a part of each pillar. The stage of mining in which rooms and pillars were excavated by either method was commonly known as "first mining" and usually resulted in the extraction of about 50 percent of the coal in a bed. The stage of mining in either method during which pillars were partly or wholly removed was known as "second mining" or "robbing." This latter stage usually resulted in the extraction of 10 to 20 percent additional coal for a maximum average extraction of 60 to 70 percent. These extraction percentages are based on numerous mine maps from the area that show the dimensions of rooms and pillars during both phases of mining.

Longwall mining has been attempted only where the coal beds are gently inclined or flat. In this method a working place was extended for several hundred feet at right angles to a haulageway. Successive slices of coal were then removed from the entire length of the working place, and the roof was allowed to collapse gradually into the open space resulting from the removal of the coal.

Strip mining utilizes draglines, power shovels, bulldozers, and large dump trucks. The draglines are commonly employed to remove overburden, but in some places they have been used to excavate coal, particularly in the more steeply dipping beds. The capacities of presently operating draglines range from 1 to 85 cubic yards and average about 10 cubic yards. Power shovels with a capacity of 1 to 10 cubic yards are the principal coal-excavating machines in the larger strip pits.

Many strip pits consist of long narrow excavations dug along the strike of steeply dipping coal beds. Where a coal bed is exceptionally thick or where the dip is less steep, such as on the crests of anticlines, in the troughs of synclines, or on gently inclined limbs of folds, large strip pits are common. In both the long narrow strip pit and the large strip pit, the ratio of thickness of overburden or waste coal seldom exceeds 25 to 1. The walls of many strip pits are several hundred feet high.

EXTENT OF MINE WORKINGS

Hundreds of small to large underground anthracite mines have operated during the last 150 years. A large number of surface openings into these mines are shown on the coal maps (Wood and Trexler, 1968 b, c). However, hundreds of additional mine openings are not shown on the coal maps because they have been destroyed by strip-mining operations, have collapsed, or have been completely overgrown and concealed by vegetation.

The linear extent of underground haulageways, tunnels, and drifts is unknown, but probably amounts to at least 1,000 linear miles and may be as great as 2,000 to 3,000 miles. There is also a vast series of underground rooms in the numerous coal beds that are interconnected at many places and are isolated at others.

The original topographic configuration of some of the more intensively mined parts of the area has been completely changed by the combined results of strip mining, underground mining, and waste dumping. The topography has been altered so swiftly in some localities that relatively modern topographic base maps are useless for geologic mapping or determining locations. Aerial photographs commonly are usable for geologic mapping for only 3 to 5 years, and in areas of very active mining, photographs have become obsolete in 1 to 2 years.

CONDITION OF MINE WORKINGS

The drifts, shafts, slopes, tunnels, and other underground mine workings are deteriorating rapidly. Large-scale mining ceased at most places during the early 1930's and has ceased elsewhere in the last several years.

The workings of these abandoned large mines are generally unsafe because of caving, rotting timbers, threatening collapse, mine gas, and water. In the abandoned smaller mines, conditions are the same or worse. Despite this generally poor condition of abandoned mine workings, several score of closely inspected small mines operate intermittently.

Most abandoned and many operating underground mines contain pools of water. These mine-water pools either completely fill the workings or are isolated in parts of the workings. Many mine-water pools overlie other parts of the same mine or adjacent mines. The location and size of most mine-water pools are known, but many others are unknown. Modern underground mines occasionally penetrate these unknown pools, and the miners may be either severely injured or drowned.

The mine-water pools commonly are supplied from two sources: direct connections with the surface through the original mine openings and through strip pits, and percolating ground water which enters the pools from unmined coal beds and adjacent strata.

PRODUCTION

The production from the Anthracite region is known with considerable exactitude from 1882, when the Federal Government and the Commonwealth of Pennsylvania began keeping comprehensive records, to the present, but it is less well known for the period preceding 1882. Before 1882 the records were largely based on census reports supplemented by State and trade sources. Production apparently began about 1755, but it was 1824 before it exceed 10,000 net tons per year.

Pennsylvania anthracite was the chief mineral fuel of the United States from about 1824, when 13,685 net tons was produced, to 1870, when the nation's production of bituminous coal exceeded that of anthracite from Pennsylvania (Tryon and Mann, 1928, p. 528–533). Anthracite production has exceeded 1 million tons a year since 1839.

The amount of anthracite extracted from the area is difficult to determine because records of the Federal Government are maintained for the entire region, individual coal fields, commercial districts, counties, and not by mine or breaker plant. They are, however, based largely upon the amount of coal cleaned at breaker plants (F. T. Moyer, oral commun., 1963).

Records of the Pennsylvania Department of Mines and Mineral Industries are maintained for the entire region, mining or inspection districts whose numbers and boundaries have changed many times, commercial districts, and the larger mines. The records are based upon the amount of cleaned coal reported by mines after processing of raw coal at breaker plants. Thus, neither Federal nor State records are based upon the raw coal extracted (F. T. Moyer, oral commun., 1963) or upon arbitrary quadrangle borders such as those that bound the area and that have little relation to individual coal fields, commercial districts, counties, and mining or inspection districts.

Unfortunately, the production figures of the Federal Government and the State government differ. In some years this difference is as much as 8 million net tons; for most years it is negligible statistically, amounting to several hundred thousand tons or less. The difference is due to the inclusion of dredged coal in the Federal figures and the fact that the Federal tonnages are based upon cleaned coal reported by breaker plants, whereas the State tonnages are based upon cleaned coal reported by mines. The difference in total tonnage from 1882 to 1960, excluding dredged coal from the Federal tonnages, amounts to only 4,516,540 net tons, even though for some years differences are 3 to 8 million net tons.

Table 2 gives the total and yearly production of the Anthracite region from 1769 to 1960. Tonnages for the years 1807 to 1960 were tabulated from the U.S. Bureau of Mines (Moyer and others, 1961, p. 154–155; Tryon and Mann, 1928, table 9) and for the years 1769 to 1806, they were tabulated from "Pennsylvania's Mineral Heritage" (Pennsylvania Bur. Statistics and Information, 1944). Table 2 also gives the yearly production of the Southern Anthracite field as reported by the U.S. Bureau of Mines from 1924 to 1960 (U.S. Bur. Mines, 1928–32, 1933–60), U.S. Geological Survey from 1913 to 1923 (U.S. Geol. Survey, 1883–1927), and the Pennsylvania Second Geological Survey (1882–89, various reports on anthracite). The production of the Southern Anthracite field for all other years was estimated by the authors after determining the year-to-year production percentage of the Southern Anthracite field with reference to the total yearly production of the Anthracite region.

The production of the area was determined from records for the following years: 1882–89, 1910, 1915, 1920, 1925, 1930, 1935, 1940, 1945, 1950, 1955, and 1960. The tonnages for all other years were estimated by determining the average 5-year production percentage of the area with reference to the year-to-year production of the Southern Anthracite field.

The production of the Anthracite region from 1769 to 1960 was 5,227,392,894 net tons. The Southern Anthracite field in the same period is estimated to have produced about 755 million net tons and the report area about 318 million net tons. The value of coal produced in the Anthracite region from 1890 to 1960 was about $17.1 billion (Moyer and others, 1961, p. 154–155) and is estimated to have been about $1 billion from 1769 to 1890, for a total of about $18.1 billion from 1769 to 1960. During the same period the Southern Anthracite

TABLE 2—*Production of anthracite, in net tons, from the Anthracite region, Southern Anthracite field, and report area*

[Southern Anthracite field: Estimated tonnage from 1769–1880 based on average percentage of production for years 1880–1960 equals about 70 million. Report area: Estimated percentage from 1769–1880 based on average percentage of production for years 1880–1960 equals about 28 million.]

Year	Anthracite region [1]	Southern Anthracite field [2]	Report area [2]
1960	18, 817, 441	4, 530, 628	2, 856, 521
1959	20, 649, 286	5, 269, 930	[3] 3, 200, 000
1958	21, 171, 142	5, 086, 583	[3] 3, 000, 000
1957	25, 338, 321	6, 061, 879	[3] 3, 400, 000
1956	28, 900, 220	7, 425, 427	[3] 4, 100, 000
1955	26, 204, 554	5, 958, 776	3, 313, 008
1954	29, 083, 477	5, 952, 615	[3] 3, 100, 000
1953	30, 949, 152	7, 352, 940	[3] 3, 400, 000
1952	40, 582, 558	8, 979, 129	[3] 3, 300, 000
1951	42, 669, 997	8, 801, 942	[3] 3, 200, 000
1950	44, 076, 703	9, 100, 374	3, 264, 940
1949	42, 701, 724	8, 261, 266	[3] 2, 800, 000
1948	57, 139, 948	12, 118, 732	[3] 4, 100, 000
1947	57, 190, 009	11, 881, 102	[3] 4, 050, 000
1946	60, 506, 873	13, 203, 552	[3] 4, 500, 000
1945	54, 933, 909	12, 290, 357	4, 503, 257
1944	63, 701, 363	14, 285, 542	[3] 5, 150, 000
1943	60, 643, 620	12, 006, 749	[3] 4, 350, 000
1942	60, 327, 729	10, 123, 626	[3] 3, 700, 000
1941	56, 368, 267	8, 561, 966	[3] 3, 100, 000
1940	51, 484, 640	7, 437, 509	2, 841, 099
1939	51, 487, 377	7, 051, 710	[3] 2, 650, 000
1938	46, 099, 027	6, 073, 139	[3] 2, 300, 000
1937	51, 856, 433	6, 067, 922	[3] 2, 300, 000
1936	54, 579, 535	6, 877, 678	[3] 2, 600, 000
1935	52, 158, 783	6, 190, 511	2, 588, 155
1934	57, 168, 291	7, 467, 559	[3] 3, 300, 000
1933	49, 541, 344	6, 352, 024	[3] 2, 800, 000
1932	49, 855, 221	7, 030, 323	[3] 3, 100, 000
1931	59, 645, 652	7, 958, 000	[3] 3, 500, 000
1930	69, 384, 837	9, 570, 400	4, 989, 655
1929	73, 828, 195	10, 394, 720	[3] 4, 700, 000
1928	75, 348, 069	10, 987, 200	[3] 4, 900, 000
1927	80, 095, 564	11, 795, 840	[3] 5, 300, 000
1926	84, 437, 452	11, 428, 400	[3] 5, 150, 000
1925	61, 817, 149	8, 812, 160	3, 905, 869
1924	87, 926, 862	12, 055, 760	4, 800, 000
1923	93, 339, 009	13, 544, 160	[3] 5, 400, 000
1922	54, 683, 022	8, 076, 320	[3] 2, 100, 000
1921	90, 473, 451	12, 029, 620	[3] 4, 800, 000
1920	89, 598, 249	13, 227, 200	5, 404, 270
1919	88, 092, 201	12, 496, 960	[3] 5, 200, 000
1918	98, 826, 084	14, 252, 000	[3] 5, 900, 000
1917	99, 611, 811	14, 109, 760	[3] 5, 050, 000
1916	87, 578, 493	12, 406, 240	[3] 5, 200, 000
1915	88, 995, 061	12, 046, 400	5, 365, 475
1914	90, 821, 507	12, 433, 220	5, 400, 000
1913	91, 524, 922	12, 679, 530	[3] 5, 500, 000
1912	84, 361, 598	[3] 12, 500, 000	[3] 5, 450, 000
1911	90, 464, 067	[3] 13, 400, 000	[3] 5, 800, 000
1910	84, 485, 236	[3] 13, 900, 000	5, 838, 972
1909	81, 070, 359	[3] 12, 200, 000	[3] 5, 200, 000
1908	83, 268, 754	[3] 12, 500, 000	[3] 5, 400, 000
1907	85, 604, 312	[3] 12, 800, 000	[2] 5, 500, 000
1906	71, 282, 411	[3] 10, 700, 000	[3] 4, 600, 000
1905	77, 659, 850	[3] 10, 800, 000	[3] 4, 600, 000
1904	73, 156, 709	[3] 9, 500, 000	[3] 4, 100, 000
1903	74, 607, 068	[3] 9, 700, 000	[4] 4, 200, 000
1902	41, 373, 595	[3] 5, 500, 000	[3] 2, 400, 000
1901	67, 471, 667	[3] 8, 800, 000	[3] 3, 800, 000

See footnotes at end of table.

TABLE 2—*Production of anthracite, in net tons, from the Anthracite region, Southern Anthracite field, and report area*—Continued

Year	Anthracite region [1]	Southern Anthracite field [2]	Report area [2]
1900	57, 367, 915	[3] 6, 900, 000	[3] 3, 000, 000
1899	60, 418, 005	[3] 6, 600, 000	[3] 2, 900, 000
1898	53, 382, 645	[3] 5, 900, 000	[3] 2, 600, 000
1897	52, 611, 681	[3] 5, 800, 000	[3] 2, 600, 000
1896	54, 346, 081	[3] 6, 000, 000	[3] 2, 700, 000
1985	57, 999, 337	[3] 5, 800, 000	[3] 2, 000, 000
1894	51, 921, 121	[3] 4, 600, 000	[3] 2, 100, 000
1893	53, 967, 543	[3] 4, 800, 000	[3] 2, 200, 000
1892	52, 472, 504	[3] 4, 700, 000	[3] 2, 100, 000
1891	50, 665, 431	[3] 4, 600, 000	[3] 2, 100, 000
1890	46, 468, 641	[3] 4, 200, 000	[3] 1, 900, 000
1889	45, 546, 970	3, 703, 952	1, 703, 519
1888	46, 619, 564	3, 646, 251	1, 721, 195
1887	42, 088, 197	3, 971, 418	1, 903, 674
1886	39, 035, 446	3, 838, 727	1, 646, 157
1885	38, 335, 974	3, 970, 638	1, 807, 107
1884	37, 156, 847	3, 537, 407	1, 639, 234
1883	38, 456, 845	3, 541, 124	2, 085, 893
1882	35, 121, 256	3, 198, 974	1, 897, 761
1881	31, 920, 018	[3] 3, 150, 000	[3] 1, 700, 000
1880	28, 649, 812	[3] 2, 600, 000	[3] 1, 500, 000
1879	30, 207, 793		
1878	21, 689, 682		
1877	25, 660, 316		
1876	22, 793, 245		
1875	22, 485, 766		
1874	24, 818, 790		
1873	26, 152, 837		
1872	24, 233, 166		
1871	19, 342, 057		
1870	15, 664, 275		
1869	17, 083, 134		
1868	17, 003, 405		
1867	16, 002, 109		
1866	15, 651, 183		
1865	11, 891, 746		
1864	12, 538, 649		
1863	11, 785, 320		
1862	9, 695, 110		
1861	9, 799, 654		
1860	8, 115, 842		
1859	9, 619, 771		
1858	8, 426, 102		
1857	8, 186, 567		
1856	8, 534, 779		
1855	8, 141, 754		
1854	7, 394, 875		
1853	6, 400, 426		
1852	6, 151, 957		
1851	5, 481, 065		
1850	4, 138, 164		
1849	3, 995, 334		
1848	3, 805, 942		
1847	3, 551, 005		
1846	2, 887, 815		
1845	2, 480, 032		
1844	2, 009, 207		
1843	1, 556, 753		
1842	1, 365, 563		
1841	1, 182, 441		
1840	967, 108		
1839	1, 008, 322		
1838	910, 075		

See footnotes at end of table.

TABLE 2—*Production of anthracite, in net tons, from the Anthracite region, Southern Anthracite field, and report area*—Continued

Year	Anthracite region [1]	Southern Anthracite field [2]	Report area [2]
1837	1,071,151		
1836	842,832		
1835	690,854		
1834	464,015		
1833	600,907		
1832	447,550		
1831	217,842		
1830	215,272		
1829	138,086		
1828	95,500		
1827	78,151		
1826	59,194		
1825	42,988		
1824	13,685		
1823	8,563		
1822	4,583		
1821	1,322		
1807–20	[3] 12,000		
1769–1806	[3] 7,270		
Total 1960–1769	5,227,392,894	[3] 755,000,000±	[3] 318,000,000±

[1] Includes dredge coal.
[2] Dredge coal excluded.
[3] Tonnage estimated by authors. Estimates based upon average tonnage of preceding and succeeding years.

field is estimated to have produced coal valued at about $2.6 billion and the report area about $1.1 billion.

About 102 square miles of the report area is underlain by coal-bearing rocks of Pennsylvanian age, but most of the coal has been produced from an area of about 85 square miles. If all coal production came from this 85-square-mile area, each square mile would have produced about 3.7 million net tons valued at about $12.8 million.

A square-mile foot of anthracite weighs about 1.28 million net tons (Averitt, 1961, p. 18). Each of the 85 square miles has yielded an average of about 2.89 square-mile feet of anthracite. If only 50 to 70 percent of the original coal was recovered from the mined beds, an average of 1.23 to 2.89 square-mile feet of coal still remains in mine workings under each square mile. Thus, an average of 4.12 to 5.78 square-mile feet of coal was recovered and disturbed under each square mile. On the basis of a single coal bed, the number of square-mile feet recovered under each square mile amounted to the mining of a bed ranging in thickness from 4.12 feet for 70 percent recovery to 5.78 feet for 50 percent recovery.

The anthracite of the report area averages about 12,700, or 12.7×10^3, BTU (British thermal units) per pound. A net ton of anthracite thus averages about 25.4×10^6 BTU of energy. The report area has produced about 8.07×10^{15} BTU of energy, the Southern Anthracite field has produced about 19.2×10^{15} BTU, and the Anthracite region about 132.7×10^{15} BTU. To 1960, the United States had consumed about 1.5×10^{18} BTU of energy (U.S. Atomic Energy Comm. 1962, p. 18). The report area produced slightly less than 0.6 percent of the total United States energy consumption before 1960, the Southern Anthracite field slightly less than 1.4 percent, and the Anthracite region about 9.5 percent.

With respect to the 318 million net tons of anthracite production estimated from the area, hundreds and perhaps thousands of small underground mines have produced from a few tons to 5,000 net tons, several hundred underground mines of intermediate size have produced 5,000 to 50,000 net tons, and about 40 large underground mines and numerous strip mines have produced the remaining tonnage. Principal among the large underground mines were the Brookside, Blackwood, Branchdale, Buck Run, Colket, Glendower, Goodspring, Indian Head, Joliet, Kalmia, Kemble, Lincoln, Lone Eagle, Lorberry, Lytle, Marshfield, Monterey, Oak Hill, Otto, Pine Hill, Pine Knot, Phoenix, Primrose, Salem, Short Mountain, Thomaston, Tower City, Westwood, Williamstown, and Valley View mines. The principal surface openings of some of these large mines are shown on the coal maps (Wood and Trexler, 1968b, c) where they have not been destroyed by strip mining, by dismantling, or by passage of time.

COAL BEDS

Coal beds are in the Spechty Kopf Member of the Catskill Formation; the Beckville and Mount Carbon Members of the Pocono Formation; the Tumbling Run, Schuylkill, and Sharp Mountain Members of the Pottsville Formation; and the Llewellyn Formation. They are commercially minable only in the Pottsville and Llewellyn Formations, although prospectors have opened up a number of thin or shaly coal beds at places in the outcrop belts of the Spechty Kopf, Beckville, and Mount Carbon Members.

About 60 coal beds are known in the Pottsville and Llewellyn Formations, and 52 have been mined. Most of these coal beds are not persistent over the entire outcrop of these formations; many, however, underlie considerable parts. Only the more extensive stratigraphically and economically important coal beds are described below. In addition, only those beds more than 14 inches or about 1.2 feet thick are shown on the coal maps and structure sections (Wood and Trexler, 1968b, c).

The coal beds of the Pottsville and Llewellyn Formations are so similar in their characteristics that they are difficult to correlate by physical and chemical properties. Most consist of banded anthracite with alternating bright bands of vitrain and bright attritus and dull bands of fusian, dull attritus, and impure or bony coal (Wagner, 1953, p. 3–6).

In the northeastern and central parts of the area the coal beds are largely anthracite characterized by alternating bright jetlike layers and very fine granular dull layers. The bright layers commonly have a brillant vitreous luster and break into hackly surfaces along conchoidal fractures. The thicker bright layers generally are well jointed or cleated. The layers of finely granular dull coal commonly have a smooth matte surface and a nonvitreous luster except for isolated bright specks of vitrain. Thick finely granular layers are conspicuous in many outcrops of coal because of their relative lack of jointing or cleating.

West of Brookside mine in Tower City and Lykens quadrangles (Wood and Trexler, 1968c, sheets 1, 3) and west of Kalmia mine in Tower City quadrangle (Wood and Trexler, 1968c, sheet 2) the coal beds are largely semianthracite and consist of alternating layers of bright and fine granular dull coal. The bright layers are generally fractured by a closely spaced cubical or prismatic cleat which causes the coal to break into small fragments instead of the larger conchoidal fragments typical of the anthracite of the northeastern and central parts of the area.

Many of the coal beds south of the Joliet anticline and the Newton fault have been fractured into incoherent masses of "potato chiplike" fragments, the "shelly" coal of miners.

The floor rocks of the coal beds of the Pottsville Formation are mainly shale and siltstone, and the roof rocks are mainly sandstone and conglomerate. However, the type of floor and roof rock commonly is not consistent for more than a few hundred feet.

Roof and floor rocks of the coal beds of the Llewellyn Formation are chiefly shale and siltstone, but in a few places they are sandstone and conglomerate. The type of roof and floor rock is not consistent from place to place, however, and is described only if known to persist for considerable distance along strike or downdip.

COAL BEDS OF THE TUMBLING RUN MEMBER OF THE POTTSVILLE FORMATION

Six coal beds have been recognized in the Tumbling Run Member over considerable parts of the area. Three of these, the Lykens Valley Nos. 4, 5, and 6 coal beds underlie much of the area and are described hereafter. The others, the Lykens Valley Nos. 3¾, 3⅞, and 7 coal beds, are recognized only in the western part of the area and are not described. The stratigraphic position of each of these coal beds is described in the chapter on the Tumbling Run Member and illustrated on plates 14, 15, and 16.

LYKENS VALLEY NO. 6 COAL BED

The Lykens Valley No. 6, or "Little," coal bed has been mined on the south limb of the Donaldson syncline and on the north limb of the Dauphin syncline in the Lykens and Tower City quadrangles (Wood and Trexler, 1968c, sheets 1–3). It has been opened in many prospects, trenches, and small mines on Broad Mountain and in the Dam, Forestville, and Little Mountain synclines in the Minersville, Tremont, and Valley View quadrangles (Wood and Trexler, 1968b, sheets 1, 3; 1968c, sheet 1). West of the western part of the Tremont quadrangle the average thickness of the bed is 4.4 feet, of which 4.2 feet is coal (table 3). The average thickness of the bed and of the coal is much less elsewhere in the area.

The Lykens Valley No. 6 coal bed was mined from about 1,600 feet above sea level to a depth of about 1,000 feet below sea level in Williamstown mine. It was also mined extensively in the Short Mountain, Brookside, Big Lick, and Kalmia mines.

LYKENS VALLEY NO. 5 COAL BED

The Lykens Valley No. 5, or "Big," coal bed was mined intensively in the Donaldson syncline west of Tremont, in the Dauphin and Lorberry synclines west of Rausch Ceek, and across the crest of the Joliet anticline (Wood and Trexler, 1968b, sheets 3, 4; 1968c, sheets 1, 3). Where initensively mined, this bed averaged 9.3 feet in thickness and contained an average of 8.1 feet coal (table 3). Farther east the bed has been opened by many strip pits, small mines, and trenches on Broad Mountain and in the Dam, Forestville, and Little Mountain synclines. The thickness decreases rapidly eastward, and commercial mining seems to have been unprofitable east of the west boundary of the Minersville quadrangle (Wood and Trexler, 1968b, sheets 1, 2).

The roof rock of the Lykens Valley No. 5 coal bed is sandstone or conglomerate, and the floor rock is shale at most places. Coal from the bed commonly consists of layers of vitrain, bright attrital coal, impure coal, and dull attrital coal that average ¼ to 1 inch in thickness. Thin to thick partings of siltstone and shale are common at many places in the Lykens Valley No. 5.

Much of the coal produced in the western part of the area was obtained from this bed at the Short Mountain, Big Lick, Brookside, Tower City, Valley View, Porter, and Williamstown mines. In the Williamstown mine, workings on this bed extend from 1,600 feet above sea level to about 2,200 feet below sea level, the deepest known mining in the Southern Anthracite field.

TABLE 3.—*Average thickness and range in thickness of Llewellyn and Pottsville coal beds based on 5,030 observations*

[Thicknesses obtained from mining company data]

Member, and coal bed	Bed thickness (feet)			Coal thickness (feet)			Average amount of refuse (percent)
	Average	Maximum	Minimum	Average	Maximum	Minimum	
Llewellyn Formation							
No. 29							
28							
27							
26							
25							
24	3.0	3.5	0.1				
23½	5.0	7.0	.1				
23	4.0	5.0	.1				
22	3.0	5.0	.1				
22L							
Faust (21)	5.8	10.0	.1	4.0	6.0	0.1	31
Rabbit Hole (20)	[1] 2.2	8.0	.1	[1] 2.2	6.0	.1	([2])
Tunnel (19)	3.0	8.0	.3	2.5	7.0	.1	16
Peach Mountain (18)	8.0	19.0	.4	6.0	17.5	.3	25
Little Tracy (17)	6.4	14.0	.5	4.7	14.0	.1	26
Upper Four Foot (16½)	3.8	12.0	1.0	2.9	5.3	.2	23
Tracy (16)	7.8	17.8	1.9	6.3	16.0	.1	19
Little Clinton (15½)	[1] 2.0	4.1	.1	[1] 2.0	4.1	.1	([2])
Clinton (15¼)	[1] 2.7	7.1	.1	[1] 2.7	7.1	.1	([2])
Little Diamond (15)	4.6	8.5	1.8	2.8	4.5	.1	39
14½	[1] 1.8	4.0	.1	[1] 1.8	4.0	.1	([2])
Diamond (14)	6.9	13.4	1.0	5.6	10.0	.5	19
Diamond Leader (14L)	[1] 3.8	7.8	.1	[1] 3.8	7.8	.1	([2])
13½	[1] 1.2	1.5	.1	[1] 1.2	1.5	.1	([2])
Little Orchard (13)	4.5	6.0	.6	[1] 4.5	6.0	.6	
Orchard (12)	6.8	14.0	.8	5.0	12.0	.3	26
Primrose (11)	10.0	23.3	1.5	8.4	21.0	.1	16
Holmes (10)	8.9	[3] 84.7	.3	8.2	[3] 80.0	.1	8
Holmes Leader (10L)	[1] 0.3	1.2	.1	[1] .3	1.2	.1	([2])
Lower Four Foot (9½)	4.8	10.0	.4	3.5	5.2	.2	27
Top Split (9)	8.3	30.0	.5	7.4	30.0	.1	11
Middle Split (8½)	8.7	41.4	.9	6.1	14.1	.1	30
8¼	6.2	9.3	.1	4.3	6.7	.1	31
Bottom Split (8)	8.2	[3] 80.0	1.0	6.1	[3] 56.2	.1	25
Skidmore (7)	7.0	20.0	.4	5.4	11.5	.1	23
Skidmore Leader (7L)	3.0	6.5	.1	1.4	2.5	.1	33
Seven Foot (6)	5.2	19.6	.2	3.7	11.5	.1	29
Seven Foot Leader (6L)	1.3	2.8	.2	1.3	2.8	.1	([2])
Buck Mountain (5)	6.9	16.4	.5	4.7	12.5	.2	22
Pottsville Formation							
Sharp Mountain Member							
Little Buck Mountain (4)	7.0	16.0	0.1	4.1	14.0	0.1	41
Scotty Steel 3	10.9	18.7	.8	8.1	12.6	.2	26
Scotty Steel 2	[1] 1.8	4.9	.1	[1] 1.8	4.9	.1	([2])
Schuylkill Member							
Lykens Valley 1	1.9	10.0	.1	1.1	10.0	.1	42
Lykens Valley 1½	1.6	4.0	.5	1.0	2.3	.1	37
Lykens Valley 2	9.2	14.5	.3	8.9	14.5	.3	3
Lykens Valley 3	4.2	10.4	.1	1.3	10.4	.1	69
Lykens Valley 3½	[1] 1.4	4.3	.2	[1] 1.4	4.3	.2	([2])
Tumbling Run Member							
Lykens Valley 3¾	[1] 1.2	2.1	.2	[1] 1.2	2.1	.2	([2])
Lykens Valley 3⅞	2.1	4.0	.2	2.1	4.0	.2	([2])
Lykens Valley 4	4.4	12.0	.8	3.5	5.3	.1	20
Lykens Valley 5	9.3	23.0	.9	8.1	23.0	.1	15
Lykens Valley 6	4.4	12.0	.1	4.2	12.0	.1	4.5
Lykens Valley 7	5.9	12.8	.1	3.3	11.1	.1	34

[1] Thickness data concentrated in a small area and not widespread enough to determine adequately average thickness.

[2] Data insufficient to calculate average percentage of refuse.

[3] Thickened by structural adjustments.

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LYKENS VALLEY NO. 4 COAL BED

The Lykens Valley No. 4, or "Whites," coal bed was mined intensively west of Goodspring in the Donaldson syncline (Wood and Trexler, 1968b, sheet 3; 1968c, sheets 1, 3) and west of Rausch Creek in the Dauphin and Lorberry synclines and the Joliet and Georges Head anticlines (Wood and Trexler, 1968b, sheet 4; 1968c, sheet 2). In these areas this bed averages about 4.4 feet thick and contains an average thickness of 3.5 feet coal (table 3).

Many small mines, trenches, strip pits and prospect pits have been opened on the Lykens Valley No. 4 coal bed east of the area of intensive mining. The bed decreases rapidly in thickness eastward and is not commercially minable east of the west border of the Minersville quadrangle (Wood and Trexler, 1968b, sheet 1).

The Lykens Valley No. 4 coal bed is commonly overlain and underlain by conglomerate and sandstone, but locally, the enclosing rocks are shale and siltstone. The coal generally consists of bands of vitrain, bright attrital, impure, and dull attrital coal, which range in thickness from 1/8 to 1 inch. Thin partings of claystone and shale are common.

Much of the coal produced in the Lykens and Tower City quadrangles was extracted from the Lykens Valley No. 4 coal bed at the Short Mountain, Big Lick, Williamstown, Brookside, Tower City, Valley View, and Porter mines. Workings on this bed extended to nearly 2,200 feet below sea level in the Williamstown mine.

COAL BEDS OF THE SCHUYLKILL MEMBER OF THE POTTSVILLE FORMATION

Four rather persistent coal beds have been recognized in the Schuylkill Member, and a fifth nonpersistent bed has been mapped locally. From the base up, these are the Lykens Valley Nos. 3½, 3, 2, 1½, and 1 coal beds. Other nonpersistent coal beds are shown on the graphic sections of the member (pls. 1–3), but are not shown on the coal maps (Wood and Trexler, 1968b, c) because they are less than 1.2 feet thick. The stratigraphic position of these four persistent beds and the local coal bed are briefly described in the section on the Schuylkill Member. The nonpersistent Lykens Valley No. 3½ coal bed is not described hereafter.

LYKENS VALLEY NO. 3 COAL BED

The Lykens Valley No. 3 coal bed has been mined and prospected at many localities in the area, but it has not been identified in the eastern part of the Minersville quadrangle on Broad and Sharp Mountains (Wood and Trexler, 1968b, sheets 1, 2). It is commonly overlain and underlain by conglomerate and sandstone and averages 4.2 feet thick, of which 1.3 feet is coal (table 3). This bed has been heavily mined in the Tower City and Brookside mines of the Tower City quadrangle (Wood and Trexler, 1968b, sheet 1). Large-scale mining apparently has been unprofitable elsewhere.

LYKENS VALLEY NO. 2 COAL BED

The Lykens Valley No. 2 coal bed has been recognized throughout the area, except on Sharp Mountain in the eastern part of the Minersville quadrangle and on Big Lick Mountain west of Wiconisco in the Lykens quadrangle (Wood and Trexler, 1968b, sheet 2; 1968c, sheet 3).

The Lykens Valley No. 2 is the thickest coal bed of the Schuylkill Member and averages 9.2 feet thick in the western part of the area; 8.9 feet is coal (table 3). It reaches a maximum thickness of about 14.5 feet in the Brookside mine (Wood and Trexler, 1968c, sheet 1). The bed thins east and northeast from the Penag mine in the Tremont quadrangle and is too thin to mine profitably in the northern part of the Minersville quadrangle (Wood and Trexler, 1968b, sheets 1, 3).

In its few exposures the Lykens Valley No. 2 coal bed consists of alternating 1/8 to 1 inch bands of vitrain, bright attrital coal, impure coal, and dull attrital coal. The refuse in the bed averages about 3 percent, which is much lower than the refuse percentages determined for any other bed (table 3).

At about half the localities where the Lykens Valley No. 2 coal bed has been mined, the floor is shale and siltstone. Conglomerate and sandstone forms the floor at all other places. The roof is predominantly conglomerate and sandstone, but in the Brookside and Tower City mines it is shale.

LYKENS VALLEY NO. 1½ COAL BED

The Lykens Valley No. 1½ coal bed has been recognized throughout the area, except in the northern parts of the Swatara Hill and Tremont quadrangles and the Minersville quadrangle (Wood and Trexler, 1968b, sheets 1–3). This bed averages 1.6 feet thick and reaches a maximum thickness of 4.0 feet (table 3) in the Tower City mine (Wood and Trexler, 1968c, sheet 1). It has been exploited by numerous small mines and exposed in many trenches and prospect pits, but it is not considered to be minable under present economic conditions.

LYKENS VALLEY NO. 1 COAL BED

The Lykens Valley No. 1 coal bed lies 5 to 50 feet below the top of the Schuylkill Member and has been exposed at many small mines, prospect pits, and trenches throughout most of the area. The average thickness is 1.9 feet (table 3), and the coal has been worked extensively only in the Lincoln mine (Wood and

Trexler, 1968b, sheet 4), where the bed reaches a maximum thickness of 10.0 feet, of which 4 to 6 feet is good coal.

Where exposed, the Lykens Valley No. 1 coal bed is almost invariably intensively fractured and consists of a mass of "shelly" coal in which all evidence of original bedding has been obliterated. This fracturing suggests that the bed was the site of much differential bedding-plane movement that resulted from structural adjustments in incompetent beds between the competent sandstones and conglomerates of the Schuylkill and Sharp Mountain Members.

At most places the Lykens Valley No. 1 coal bed is too thin and fractured to mine under present economic conditions.

COAL BEDS OF THE SHARP MOUNTAIN MEMBER OF THE POTTSVILLE FORMATION

The Scotty Steel Nos. 2 and 3 coal beds of the Sharp Mountain Member are recognized over much of the area, and the overlying Little Buck Mountain (No. 4) coal bed is known in the northern part of the area in the Minersville and Tremont quadrangles (Wood and Trexler, 1968b, sheets 1–3). Several other nonpersistent coal beds are shown on the graphic sections (pls. 1–3), but are not shown on the coal maps because of their thinness.

SCOTTY STEEL NO. 2 COAL BED

The Scotty Steel No. 2 coal bed is present throughout the area, except east of the Penag mine in the south-central part of the Tremont quadrangle and on Sharp Mountain in the Tower City, Pine Grove, Swatara Hill, and Minersville quadrangles (Wood and Trexler, 1968b, sheets 2–4; 1968c, sheet 2). It averages 1.8 feet thick (table 3) and reaches a maximum of 4.9 feet in Penag mine, but it is too thin to be commercially exploited at most places in the foreseeable future.

The floor and roof rocks of the Scotty Steel No. 2 coal bed vary from place to place, but the floor is predominantly sandstone and conglomerate, and the roof is most commonly shale. Coarse conglomerate, which composes the bulk of the middle and lower parts of the Sharp Mountain Member, generally lies a few inches to a few feet below the coal bed. This stratigraphic relationship was a great aid in tracing the coal beds of the upper part of the member and the lower part of the Llewellyn Formation.

SCOTTY STEEL NO. 3 COAL BED

The Scotty Steel No. 3 coal bed, which has been mined at many places in the area, lies stratigraphically 8 to 100 feet below the Buck Mountain (No. 5) coal bed. It thins out eastward on Sharp Mountain in the Minersville quadrangle (Wood and Trexler, 1968b, sheet 2), is absent or thin in the Peaked Mountain anticline and Little Mountain syncline in the Tremont quadrangle (Wood and Trexler, 1968b, sheet 3), is not known on the limbs of the Joliet anticline and Tremont syncline in the Pine Grove and Tower City quadrangles (Wood and Trexler, 1968b, sheet 4; 1968c, sheets 1, 2), and thins out westward in the central and western parts of the Lykens quadrangle (Wood and Trexler, 1968c, sheet 3). Its average thickness is 10.9 feet, of which 8.1 feet is coal (table 3). A maximum thickness of 18.7 feet is recorded in the Goodspring mine (Wood and Trexler, 1968b, sheet 3).

The No. 3 coal bed is a banded anthracite composed of alternating layers of vitrain, bright attrital coal, impure coal, dull attrital coal, and partings of siltstone and claystone. Individual layers are 1/8 inch to 1 inch thick and average about 1/2 inch.

The Scotty Steel No. 3 coal bed was misidentified and mined as the Buck Mountain coal bed of the Llewellyn Formation at several localities south of the Mine Hill fault complex. At these localities the Buck Mountain is thin and impure, and the Scotty Steel is thick and relatively clean. Misidentification was most common in mines near Donaldson, in mines in the Dam syncline, and at places between Donaldson and the syncline in Tremont quadrangle (Wood and Trexler, 1968b, sheet 3). The Buck Mountain is thick and has been intensively mined north of the complex, whereas the Scotty Steel is thin or absent. South of the complex, the Buck Mountain is thin, and the Scotty Steel is thick and has been intensively mined. Before this survey, the thick Buck Mountain north of the complex had been miscorrelated southward with the thick Scotty Steel.

LITTLE BUCK MOUNTAIN (NO. 4) COAL BED

The Little Buck Mountain (No. 4) coal bed is recognized only in the Tremont and Minersville quadrangles (Wood and Trexler, 1968b, sheets 1–3). North of the Mine Hill fault complex it underlies the Heckschersville, Jugular, Little Mountain, and Rohresville synclines, the south limb of the Powder Hill anticline, and the limbs and crests of the Peaked Mountain and Mine Hill anticlines. South of the fault complex it crops out on the limbs of the Dam and Forestville synclines and Crystal Run anticline and thins out southwestward in the northeastern part of the Tremont quadrangle.

The No. 4 is the uppermost coal bed of the Sharp Mountain Member and commonly lies less than 20 feet below the Buck Mountain coal bed of the Llewellyn Formation. Locally, however, it is as much as 50 feet below. Its thickness averages about 7.0 feet, contains an average of 4.1 feet coal (table 3), and reaches a maximum thickness of about 16.0 feet in the Pine Knot mine (Wood and Trexler, 1968b, sheet 1). On the flanks and crest of the Mine Hill anticline, the Little Buck Mountain is divided into two benches called the Bottom Split (4B) and the Top Split (4T) (Wood and Trexler, 1968b, sheet 1).

Conglomerate, sandstone, siltstone, and shale each constitute the roof and floor in about an equal number of places, but the roof is mainly shale and siltstone where the Little Buck Mountain is less than 15 feet below the Buck Mountain coal bed.

The Little Buck Mountain coal bed has been intensively exploited in many of the large and small underground mines in the Jugular, Heckscherville, Little Mountain, and Rohresville synclines and on the Mine Hill, Peaked Mountain, and Crystal Run anticlines. It has also been extensively strip mined on the crest and upper flanks of the Mine Hill anticline.

COAL BEDS OF THE LLEWELLYN FORMATION

Several hundred million tons of anthracite have been extracted from the Llewellyn Formation in the area. Of the 52 coal beds that have been mined, 38 beds are in this formation. The more productive of these beds are the Bottom Split (No. 8), Middle Split (No. 8½), and Top Split (No. 9) coal beds of the Mammoth coal zone. Less productive coal beds are the Buck Mountain (No. 5), Skidmore (No. 7), Holmes (No. 10), Primrose (No. 11), Orchard (No. 12), Little Orchard (No. 13), Diamond (No. 14), Tracy (No. 16), Little Tracy (No. 17), and Peach Mountain (No. 18). The Seven-Foot Leader (No. 6L), Seven Foot (No. 6), Skidmore Leader (No. 7L), No. 8¼, Lower Four Foot (No. 9½), Holmes Leader (No. 10L), Little Diamond (No. 15), Upper Four Foot (No. 16½), Tunnel (No. 19), and Rabbit Hole (No. 20) coal beds have been mined at many places, but have been of much less economic importance. The Diamond Leader (No. 14L), No. 14½, No. 15R, Clinton (No. 15¼), Little Clinton (No. 15½), and Faust (No. 21) coal beds have been prospected at many places, but except for local thick pods of coal, have been of little economic importance. The 22, 22L, 23½, 24, 25, 26, 27, 28, and 29 coal beds, which lie above the Faust bed in the upper part of the Llewellyn, crop out in small parts of the area, generally are variable from place to place in purity and thickness, and have been little prospected or mined. Other coal beds are locally recognized in the Llewellyn, but are too thin and too limited in areal extent to be of economic importance.

Only those coal beds that have been of economic importance because of production or of stratigraphic importance because of widespread occurrence are described below.

BUCK MOUNTAIN (NO. 5) COAL BED

The Buck Mountain (No. 5), also known as the Twin or Umbahauer, coal bed has been of considerable economic importance in the area. Although it is less than 2 feet thick at some places, this bed has been traced throughout the area. Its average thickness is 6.9 feet, of which 4.7 feet is coal (table 3). A maximum thickness of 16.4 feet is reached in the Pine Knot and Thomaston mines (Wood and Trexler, 1968b, sheet 1).

At most places the Buck Mountain bed is composed mainly of bright attrital coal with thin 1/16 to ½ inch layers of vitrain and dull attrital coal. Bony layers of coal are common near the base of the bed and in the upper 1 foot. The bed is dirty and has been unprofitable to work in the vicinities of the Penag, Goodspring, Joliet, Westwood, New Lincoln, Lincoln, and Tower City mines.

The Buck Mountain commonly rests on a fissile to nonfissile coaly shale sequence 0 to about 10 feet thick. Where this shale is absent, the coal rests directly on the uppermost sandstone or conglomerate of the Sharp Mountain Member of the Pottsville Formation. For example, at many places along Sharp Mountain in the Minersville quadrangle (Wood and Trexler, 1968b, sheet 2), the coal of the Buck Mountain bed is in contact with the upper coarse conglomerate of that member. Locally, the shale at the base of the Buck Mountain contains numerous fossil plant impressions, and at other places, *Stigmaria* impressions are common.

The Buck Mountain is the lowest commercially workable coal bed in the Llewellyn Formation. It has been previously miscorrelated with the Scotty Steel No. 3 coal bed in mines near Donaldson, in mines in the Dam syncline, and in the intervening area (Wood and Trexler, 1968b, sheet 3). (See Scotty Steel No. 3 coal bed for a more complete description.) In addition, it was mined as the Skidmore (No. 7) coal bed in the Williamstown, Big Lick, and Short Mountain mines (Wood and Trexler, 1968c, sheet 3).

About 300 thickness measurements, made largely by employees of mining companies, indicate that the Buck Mountain is thinnest in the Brookside, Tower City, and Valley View mines (Wood and Trexler, 1968c, sheet 1). It increases gradually in thickness toward the northeast from an average of about 4.0 feet in these mines to about 7.8 feet in the Heckscherville syncline and toward the west to about 5.8 feet near Wisconisco (Wood and Trexler, 1968b, sheet 1; 1968c, sheet 3). Similarly, the average thickness of the coal in the bed increases from about 3.0 feet in these mines to 5.5 feet in the Heckscherville syncline and 4.1 feet near Wisconisco.

SEVEN FOOT (NO. 6) COAL BED

The Seven Foot (No. 6) coal bed has been recognized throughout the area, except between Swatara Creek and the Colket mine and east of the Blackwood mine on Sharp Mountain (Wood and Trexler, 1968b, sheets 2, 3). Elsewhere, this bed averages 5.2 feet thick, of which 3.7 feet is coal. It reaches its maximum thickness of 19.6 feet in the Tower City mine (table 3). The average and maximum thicknesses were derived from about 150 measurements, largely made by employees of mining companies. The Seven Foot coal bed has not been as intensively mined as have the adjacent coal beds in the Llewellyn.

SKIDMORE (NO. 7) COAL BED

The Skidmore (No. 7) coal bed has been mined intensively and has been of considerable economic importance. It is less than 1.2 feet thick in the vicinity of Middle Creek (Wood and Trexler, 1968b, sheet 3), but elsewhere averages 7.0 feet, of which 5.4 feet is coal (table 3). The Skidmore attains a maximum thickness of about 20.0 feet in Thomaston mine (Wood and Trexler, 1968b, sheet 1). The average and maximum thicknesses were obtained from about 400 measurements made largely by employees of mining companies. The Skidmore bed is relatively dirty and impure at the Colket, Middle Creek, Goodspring, Joliet, Westwood, and Lincoln mines.

The Skidmore is a banded anthracite that commonly consists, from the top down, of a 4- to 12-inch bench of interbedded layers of bright attrital coal and vitrain; a 6- to 15-inch bench of dull attrital coal, impure coal, or bone, a 6- to 24-inch bench of bright attrital coal containing thin lenses of vitrain, an 8- to 30-inch bench of bright attrital coal and vitrain interbedded in 2-to 5-inch layers, an 8- to 30-inch bench of vitrain, a 6- to 18-inch bench of dull attrital coal, and a basal bench of 2 to 10 inches of bright attrital coal.

Locally, the Skidmore and Bottom Split (No. 8) coal bed are close together. In the Buck Run mine on the north limb of the Rohresville syncline (Wood and Trexler, 1968b, sheet 1), these coal beds almost merge at the surface, and underground mining records indicate either that they merge or are so close that they were mined together. However, at most places 40 to 50 feet of rock separates these coal beds, and the maximum stratigraphic separation is slightly more than 120 feet near the John Veith shafts (Wood and Trexler, 1968b, sheet 2).

The Skidmore coal bed was mined as the Top Split (No. 9) coal bed in the Williamstown, Big Lick, and Short Mountain mines (Wood and Trexler, 1968c, sheet 3). It has been locally mined as the White Ash, Dirt, or Barclaugh bed on Sharp Mountain and as the Billy Best bed at places in the Heckscherville syncline (Wood and Trexler, 1968b, sheets 1, 2).

BOTTOM SPLIT (NO. 8) COAL BED

The most widely mined coal bed in the area is the Bottom Split (No. 8) bed of the Mammoth coal zone; however, the tonnage produced may not exceed that extracted from the Middle Split (No. 8½) and Top Split (No. 9) coal beds.

Locally, the Bottom Split is less than 1.2 feet thick in small areas, but the outcrop of the bed is shown as being continuous on the coal and geologic maps and the structure sections (GQ–689, 690, 691, 692, 698, 699, 701; Wood and Trexler, 1968b, c).

The Bottom Split has been mined under a variety of names. Principal among these names are the Daniel or Jugular beds in the Heckscherville syncline, the Blackheath bed south of the Mine Hill anticline, the White Ash, Big Dirt, Heister, or Barclaugh beds on Sharp Mountain, the No. 9 1/2 beds in the Lykens quadrangle (Wood and Trexler, 1968b, sheets 1, 2, 4; 1968c, sheet 3), and the Mammoth bed at many places throughout the area.

At most places the Bottom Split is underlain and overlain by shale or siltstone. The roof shale commonly contains a well-preserved fossil flora, and the floor shale usually contains impressions of *Stigmaria*, plant branches, tree trunks, and numerous ironstone concretions that range in diameter from about 1/2 inch to 3 feet.

The Bottom Split coal bed averages 8.2 feet thick and contains an average of 6.1 feet of coal (table 3). It reaches a maximum structurally deformed thickness of about 80.0 feet in the Pine Knot mine in the upper plate of Jugular fault (Wood and Trexler, 1968b, sheet 1). The stratigraphic thickness near this locality generally is about 20.0 feet, and the 80.0 foot thickness is the result of disharmonic folding, plastic flow, and fracture flow as the bed was dragged against the lower plate of the Jugular fault. The average and maximum thicknesses were determined from about 500 measurements largely obtained from mined records. The local average thickness of the Bottom Split gradually increases eastward from 2.5 feet in the Lykens quadrangle to 11.0 feet in the eastern part of Minersville quadrangle (Wood and Trexler, 1968b, sheet 1; 1968c, sheet 3).

At the few localities where the Bottom Split could be examined, it is a banded anthracite consisting, from the top down, of the following sequences: 10 to 28 inches of bright attrital coal containing vitrain layers 1/16 to 1/2 inch thick, 8 to 18 inches of bright attrital coal containing local partings of bone or impure coal, 15 to 28 inches of bright attrital coal and thin laminae of vitrain, 6 to 14 inches of dull attrital coal, and 20 to 42 inches of bright attrital coal containing many thin laminae of vitrain. Locally, fusain layers as much as one-sixteenth inch thick are present in the lower 20- to 42-inch bench.

MIDDLE SPLIT (NO. 8½) COAL BED

The Middle Split (No. 8½) coal bed of the Mammoth coal zone has been one of the principal sources of anthracite in the area. It has not been so widely mined as the Bottom Split coal bed, but it has been worked much more intensively at a number of large mines. The Middle Split has been traced throughout the area, except where it merges into the Top Split (No. 9) coal bed from a point near the west border of Minersville quadrangle to near Blackwood mine (Wood and Trexler, 1968b, sheet 2), and except where it is combined with the Bottom Split coal bed for a distance of about 500 feet near the east border of the same quadrangle (Wood and Trexler, 1968b, sheet 1).

The Middle Split has been mined under a variety of names in the area. In the Heckscherville syncline it was known as the Lelar bed, and south of the Mine Hill anticline it was mined as the Middle Branch and Seven Foot bed (Wood and Trexler, 1968b, sheet 1). On Sharp Mountain in the Pine Grove and Tower City quadrangles (Wood and Trexler, 1968b, sheet 4; 1968c, sheet 2), it has been called the Big Dirt, White Ash, Four Foot, and Barclaugh bed, and in the Lykens quadrangle (Wood and Trexler, 1968c, sheet 3) it was mined as the Holmes (No. 10) bed. At many places it was known simply as the Mammoth bed.

Shale and siltstone generally are the floor and roof rocks of the Middle Split, but from the vicinity of New Mines to the Colket mine (Wood and Trexler, 1968b, sheets 2, 3), the bed is overlain by a coarse pebble and cobble conglomerate that resembles the basal conglomerate of the Sharp Mountain Member of the Pottsville Formation.

The Middle Split averages 8.7 feet thick and contains an average of 6.1 feet of coal (table 3). It reaches a maximum thickness of 41.4 feet in Colket mine (Wood and Trexler, 1968b, sheet 3). The average and maximum thicknesses are from about 400 measurements that were largely obtained from mine records. The bed thins in all directions from the Colket mine, but thins less rapidly to the northeast than in other directions.

Where observed, the Middle Split coal bed is a banded anthracite consisting of thin to thick layers of bright attrital coal, vitrain, dull attrital coal, and impure coal. Locally, thin layers of fusain are scattered throughout the bed, and benches of bony coal are present in the upper and lower parts.

TOP SPLIT (NO. 9) COAL BED

The Top Split (No. 9) coal bed has been one of the principal sources of anthracite and is recognized throughout the area. Stratigraphically, it is important because it is the uppermost bed of the Llewellyn Formation that has been intensively mined in almost all the large mines in the area. In the Heckscherville syncline it was known as the Crosby bed (Wood and Trexler, 1968b, sheet 1). South of the Mine Hill anticline it was called the White Ash, or Four Foot, bed, and on Sharp Mountain it was known variously as the Pitch, Dan's, Dirt, Barclaugh, and White Ash beds (Wood and Trexler, 1968b, sheets 1, 2). In the Lykens quadrangle it was mined as the Primrose (No. 11) bed (Wood and Trexler, 1968c, sheet 3). In addition, at many places where the Top Split is exceptionally thick, it has been known as the Mammoth bed.

The roof and floor rock of the Top Split is commonly shale and siltstone, but locally, conglomerate and sandstone rest directly on the coal bed (pls. 1–3). Generally, the roof rock contains a well-preserved fossil flora, and the floor rock contains numerous *Stigmaria* impressions.

The Top Split coal bed averages 8.3 feet thick, of which 7.4 feet is coal (table 3). The maximum thickness is about 30.0 feet in Pine Knot mine (Wood and Trexler, 1968b, sheet 1). The average and maximum thicknesses of the Top Split were obtained from about 600 measurements largely made by mining company employees. The bed gradually thins southwestward and southward from an average thickness of about 11.2 feet in the Heckscherville syncline (Wood and Trexler, 1968b, sheet 1) to an average thickness of 4.8 feet in the western and southern parts of the area. Refuse averages about 11 percent of the bed, which is considerably less than the refuse content of the other coal beds in the Mammoth coal zone (table 3). Despite this low refuse content and the high average thickness, the Top Split coal bed has not been as important commercially as either the Middle or Bottom Splits.

At a few localities where the Top Split was exposed, it was a bright banded anthracite, which consisted from the base up of a 1- to 3-foot unit of bright attrital coal containing thin layers of impure coal or bone near the base, a 2- to 4-foot unit of vitrain containing thin laminae of fusain, and a 2- to 4-foot unit of bright attrital coal containing ⅛- to ½-inch layers of vitrain.

HOLMES (NO. 10) COAL BED

The Holmes (No. 10) coal bed has been an important source of anthracite throughout the area. It was mined as the Church bed in the Heckscherville syncline, as the Black Heath bed in the vicinity of Donaldson and Tremont, as the Pat Martin, Little Orchard (No. 12), or Twin bed in the Lykens quadrangle, as the No. 1 or Peacock bed on Sharp Mountain, and as the Black Mine bed near Silverton (Wood and Trexler, 1968b, sheets 1–4; 1968c, sheets 2, 3).

The Holmes bed averages 8.9 feet thick, of which 8.2 feet is coal (table 3). The maximum recorded thickness is about 84.7 feet in the trough of Heckscherville syncline (Wood and Trexler, 1968b, sheet 1). This enormous thickness, the greatest recorded for any coal bed in the area, was caused by structural flowage and disharmonic flexing of coaly material towards the trough during folding. On the limbs of the syncline this bed averages about 12.0 feet thick; thus, the thickness of the bed has been increased about seven times in the trough. The Holmes averages 10.5 feet thick in the northern part of the Minersville quadrangle, 9.0 feet in the southern part, 10.6 feet in the southern part of the Tremont quadrangle, 9.4 feet in the Tower City and Pine Grove quadrangles, and 4.2 feet in the Lykens quadrangle (Wood and Trexler, 1968b, sheets 1–4; 1968c, sheets 1–3). The average and maximum thicknesses for the area and for specific parts were obtained from about 400 measurements from mining company records.

Conglomerate, shale, and siltstone are the principal roof rocks of the Holmes coal bed, with shale predominating slightly. Sandstone is present at many places, but is not so common as are the other three. Shale is the chief floor rock. Abundant fossil leaves are preserved at many places in the roof shale, and well-preserved impressions of *Stigmaria* are common in the floor shale.

The Holmes is a banded anthracite, which consists of thin to thick layers of vitrain interbedded with similar layers of bright attrital coal, dull attrital coal, and bone. Impure coal or bone generally is concentrated in the lower and upper few inches of the bed.

Refuse averages 8 percent in the Holmes coal bed (table 3), which is less than that determined for the other coal beds of the Llewellyn Formation. Considering the great average thickness of the Holmes and the low refuse content, it is difficult to understand why this bed was not the most heavily mined bed in the area; however, Smith (1895, p. 2111, 2122, 2131) reported that the Holmes was a rough and dirty bed that is locally variable in thickness and purity. Despite the low average amount of refuse, the coal itself apparently contains much ash, which has adversely affected the commercial value and mining.

PRIMROSE (NO. 11) COAL BED

The Primrose (No. 11) coal bed has been traced throughout the central part of the area and has been one of the principal sources of anthracite. It averages 10.0 feet thick, of which 8.4 feet is coal (table 3). A maximum thickness of 23.3 feet is recorded in the Thomaston mine (Wood and Trexler, 1968b, sheet 1). The average and maximum thicknesses were obtained from about 600 measurements, made largely by employees of mining companies. Refuse averages 16 percent (table 3) in the Primrose, which is rather low for the coal beds of the Llewellyn Formation. This bed probably would have been the most intensively mined coal if it had underlain a larger part of the area and if it had been closer stratigraphically to the other important sources of anthracite. At most localities the Primrose is rather clean and, has been considered extremely desirable for mining operations, but at the East Franklin and Goodspring mines it is dirty (Wood and Trexler, 1968b, sheets 3, 4).

Shale is the principal floor rock of the Primrose coal bed. It commonly contains impressions of *Stigmaria* or a well-preserved fossil leaf flora. Shale is also the principal roof rock, but locally either conglomerate or sandstone prevail. A pebble conglomerate overlies the Primrose at many places in the Minersville, Tremont, and Pine Grove quadrangles (Wood and Trexler, 1968b, sheets 1–4).

The Primrose coal bed is a banded anthracite that consists of alternating thin to thick layers of bright attrital coal, dull attrital coal, vitrain, and thin laminae of fusain. The upper and lower parts of the bed contain considerable amounts of bony or impure coal at many places.

The Primrose has been of great economic and stratigraphic importance because it is the uppermost coal bed in the Llewellyn Formation that has been widely mined. The Diamond (No. 14), Tracy (No. 16), and Peach Mountain (No. 18) coal beds also have been mined intensively; but as their outcrop areas are considerably smaller, they have not been as valuable for stratigraphic correlations and structural interpretations.

ORCHARD (NO. 12) COAL BED

The Orchard (No. 12) or lower split of the Twin coal bed has been widely recognized and intensively mined locally. At most places, however, this bed is dirty and variable in thickness. It averages 6.8 feet thick, of which 5.0 feet is impure coal (table 3), and reaches a maximum thickness of 14.0 feet in workings of the East Franklin mine (Wood and Trexler, 1968b, sheet 4). Because of the variable thickness, included impurities, and dirt, the Orchard coal bed has been of limited economic importance, and it does not seem likely that its relative importance will be different in the future.

LITTLE ORCHARD (NO. 13) COAL BED

The Little Orchard (No. 13) or upper split of the Twin coal bed has been mined at many places, but generally is too thin and dirty to be mined profitably. This bed and the Orchard merge in many localities into a single bed that is shown on the maps and illustrations of this report as the Orchard. The Little Orchard averages about 4.5 feet in the Tremont and Minersville quadrangles and reaches a maximum thickness of about 6 feet in the Lytle and Oak Hill mines (Wood and Trexler, 1968b, sheets 1–3). It has been, and probably will be, of little economic importance.

DIAMOND (NO. 14) COAL BED

The Diamond (No. 14) coal bed has been mined at many places, but it has not been so important as the underlying more heavily mined coal beds because of its limited areal extent. It averages 6.9 feet thick, of which 5.6 feet is coal (table 3). It reaches a maximum thickness of 13.4 feet in the Phoenix and Lytle shafts (Wood and Trexler, 1968b, sheets 1, 2). The bed thins gradually from these shafts southwestward to an average thickness of about 4.0 feet in the Tremont quadrangle, thickens to an average of 10.0 feet in the Tower City quadrangle, and thins to about 5.0 feet in the Lykens quadrangle (Wood and Trexler, 1968b, sheet 3; 1968c, sheets 1–3). Southward from the shafts the Diamond maintains an average thickness of about 6.5 feet. The average thickness, the thickness in specific parts of the area, and the maximum thickness are based upon about 100 measurements obtained largely from mining records. Generally, the Diamond is variable in quality and purity, but in the vicinity of Forestville (Wood and Trexler, 1968b, sheets 1, 2) it is exceptionally clean and was mined intensively.

At most localities the floor and roof of the Diamond coal bed are shale and siltstone, but locally, the roof is fine pebble conglomerate. Plant fossils generally are absent or are poorly preserved in the floor and roof shales.

The Diamond coal is a banded anthracite that commonly consists of thin layers of vitrain, bright attrital coal, dull attrital coal, bone, and thin, laminae of fusain. Bone is generally concentrated in the lower and upper few inches of the bed.

TRACY (NO. 16) COAL BED

The Tracy (No. 16) coal bed has been worked at many of the mines that had workings in coal beds in the middle part of the Llewellyn Formation. In the central and eastern parts of the area it averages 7.8 feet thick, of which 6.3 feet is coal (table 3). The bed reaches a maximum thickness of about 17.8 feet in workings of the Middle Creek mine (Wood and Trexler, 1968b, sheet 3). Refuse averages 19 percent; but at many places the coal itself is impure, and the bed has not been mined so intensively as have the underlying beds. The average and maximum thicknesses were determined from about 150 measurements largely made by mining company employees. There is little information concerning the bed in the western and southern parts of the area; but locally in these parts, the bed has been squeezed by adjacent more competent strata so that the thickness is highly variable and the coal is commonly shelly.

Shale and sandstone are the dominant roof rocks, and shale is the principal floor rock. A large assemblage of fossil plant leaves commonly is present in the roof rock.

The Tracy is a banded anthracite that consists of thin to thick layers of vitrain, bright attrital coal, dull attrital coal, and impure coal or bone. Thin laminae of fusain are common at many places. At the Salem mine near Silverton, the Tracy coal bed was mined as the Tunnel coal bed, and north of the town of Minersville it was mined as the Cockle bed (Wood and Trexler, 1968b, sheet 2).

LITTLE TRACY (NO. 17) COAL BED

The Little Tracy (No. 17) coal bed has been mined at many places in the area. It averages about 6.4 feet thick, of which 4.7 feet is usually clean coal (table 3). A maximum thickness of about 14 feet was recorded in workings of the John Veith shafts (Wood and Trexler, 1968b, sheet 2). Locally, this bed has been mined intensively south of the Red Mountain fault, which aided greatly in correlating this and other beds between the vicinities of Llewellyn and Tremont (Wood and Trexler, 1968b, sheets 2–4). The Little Tracy has been of economic value in limited areas and probably will continue to be of value in places where it has not been intensively mined.

PEACH MOUNTAIN (NO. 18) COAL BED

The Peach Mountain (No. 18) is the best known coal bed in the middle part of the Llewellyn Formation and is the uppermost intensively mined bed in the formation. Its thickness is relatively uniform, averaging 8 feet, of which 6 feet is usually clean coal (table 3). A maximum thickness of about 19 feet is recorded in workings from the John Veith shafts (Wood and Trexler, 1968b, sheet 2). The average and maximum thicknesses are from about 150 measurements obtained largely from mining records.

The Peach Mountain has been mined under numerous names, principally as the Spohn, Lewis, Black Mine, and Gate coal beds. The outcrop pattern of this multinamed bed is intricate, and the strip pits on it in the central part of the coal field clearly outline many anticlines and synclines. Workings on this bed and on the Tracy (No. 16) and Little Tracy (No. 17) coal beds south of Red Mountain aided in correlating the strata of the middle part of the Llewellyn from near Tremont to Silverton (Wood and Trexler, 1968b, sheets 2–4).

The roof and floor rocks of the Peach Mountain coal bed are principally shale and siltstone. An abundant fossil plant assemblage is commonly preserved in these rocks. Examination of this assemblage by C. B. Read and Sergius Mamay (oral commun., 1956–58) indicates that the Peach Mountain is probably of early Late Pennsylvanian age and that it probably correlates with the upper part of the Allegheny Formation or the lower part of the Conemaugh Formation of western Pennsylvania.

The Peach Mountain is a banded anthracite consisting of thin to thick layers of bright attrital coal and vitrain with thinner layers of dull attrital coal and impure coal or bone.

Because of its thickness and purity the Peach Mountain probably would have rivaled any split of the Mammoth coal zone in productivity if it had underlain a larger part of the area.

CHARACTERISTICS OF THE COAL

PHYSICAL PROPERTIES

The anthracite and semianthracite of the area are in general distinguishable by their physical properties. The properties of each are perhaps best compared in table 4, which has been modified and expanded from a table published by Turner (1934, p. 332).

Semianthracite is most easily distinguished from anthracite in the field by its tendency to break along prismatic and rectangular fractures rather than along conchoidal fractures. It is also generally much dirtier when handled.

CHEMICAL COMPOSITION

The coal of the area is of semianthracite and anthracite rank, according to analyses of the U.S. Bureau of Mines and to specifications of the American Society for Testing Materials (1954). On a dry-mineral-matter-free basis the percentage of fixed carbon ranges from 90.3 to 98.6 and averages about 94.2 (tables 5–7). Volatile matter on a similar basis ranges in percentage from a maximum of 9.7 in the semianthracite of the western part of the area to a minimum of 1.4 in the anthracite of the eastern part and averages about 5.8 for both types of coal.

Ash content on an as-received basis ranges from about 7.0 to 24.0 percent and averages about 12.2. Sulfur ranges from about 0.5 to 2.0 percent and averages slightly less than 0.7. Variations in ash and sulfur content are unrelated to the rank of the coal.

Calorific value ranges from 10,640 to 14,130 Btu on an as-received basis and from 14,730 to 15,500 Btu on a dry-mineral-matter-free basis. On an as-received basis the coal has an average heat value of 12,650 Btu and on a dry-minneral-matter-free basis, 15,170 Btu. The heat value of semianthracite generally exceeds that of anthracite of similar ash content by several hundred Btu.

Moistures averages about 4.2 percent and ranges from a minimum of 0.9 percent in semianthracite beds of the Pottsville Formation in the East Brookside mine to a maximum of 7.9 percent in anthracite beds of the Llewellyn Formation near Zerbe. The reason for the increase in moisture content in anthracite is not known.

On a moisture-and-ash-free basis, hydrogen ranges from a minimum of 2.3 percent in anthracite to a maximum of 3.8 in semianthracite and averages 2.9 for all coal. Similarly, nitrogen ranges from 0.7 percent in anthracite in the eastern part of the area to 1.5 in semianthracite in the western part and averages about 1.1. Oxygen ranges from 1.1 percent in anthracite to 4.1 in semianthracite and averages about 2.3.

TABLE 4.—*Physical properties of semianthracite and antrhacite from the west-central part of the Southern Anthracite field*

Property	Anthracite	Semianthracite
Luster	Vitreous to dull	Vitreous to dull.
Structure	Laminated to amorphous, laminations commonly obscured because of uniformity of luster.	Laminated to amorphous laminations, generally distinct because of contrasts in luster.
Fracture	Predominantly conchoidal, but locally prismatic and rectangular, usually widely spaced.	Predominantly rectangular and prismatic, rarely conchoidal, commonly closely spaced.
Streak	Chiefly black	Chiefly black.
Hardness	2.75–3.00	2.50–2.75.
True specific gravity.	1.50–1.75	1.45–1.55.
Toughness	23–51 percent friable, averages 35 percent friable.	46–53 percent friable, averages 50 percent friable.
Agglomeration	Noncoherent residue, nonagglomerating.	Noncoherent residue, nonagglomerating.
Ash	Red, gray, creamy, white.	Red, gray, creamy, white.
Stain	Will not stain hands	Commonly stains hands; black or dark gray.
Ash-softening temperature.	2,200°–2,960°C	2,150°–2,450°C.

TABLE 5.—*Chemical analyses of bed samples from the west-central part of the Southern Anthracite field*

[Analyses by U.S. Bureau of Mines]

Mine	Bed [1]	Condition [2]	Percent: Proximate: Moisture	Percent: Proximate: Volatile matter	Percent: Proximate: Fixed carbon	Percent: Proximate: Ash	Percent: Ultimate: Sulfur	Percent: Ultimate: Hydrogen	Percent: Ultimate: Carbon	Percent: Ultimate: Nitrogen	Percent: Ultimate: Oxygen	Calorific value (Btu)	Fusibility of ash (°F): Initial deformation temperature	Fusibility of ash (°F): Softening temperature	Fusibility of ash (°F): Fluid temperature	True specific gravity	Dry-mineral-matter-free basis: Fixed carbon percent	Dry-mineral-matter-free basis: Calorific value (Btu)	Laboratory or index No.
Dauphin County																			
Short Mountain mine	LV 5	1	1.1	9.6	79.5	9.8	0.7	3.2	82.7	1.1	2.5	13,590					[3] 90.3		A1859
		2		9.7	80.4	9.9	.7	3.1	83.6	1.1	1.6	13,740							
		3		10.8	89.2		.8	3.5	92.8	1.2	1.7	15,260							
	LV 5	1	1.9	8.1	80.3	9.7	.5	3.2	81.5	1.0	4.1	13,440		2,210			[3] 91.8		A75152
		2		8.3	81.8	9.9	.5	3.0	83.1	1.0	2.5	13,700							
		3		9.2	90.8		.6	3.4	92.2	1.1	2.7	15,210							
	LV 5	1	2.1	7.5	80.3	10.1	.5	3.3	80.9	1.0	4.2	13,480		2,430			[3] 91.8		A75153
		2		7.6	82.1	10.3	.6	3.1	82.6	1.0	2.4	13,760							
		3		8.5	91.5		.6	3.5	92.1	1.2	2.6	15,350							
	LV 5	1	1.1	8.0	80.4	10.5	.6	3.1	82.1	1.0	2.7	13,560		2,260			[3] 92.1		A75154
		2		8.1	81.3	10.6	.6	3.0	83.0	1.0	1.8	13,710							
		3		9.1	90.0		.7	3.4	92.8	1.1	2.0	15,340							
	LV 5	1	1.2	8.3	80.2	10.3	.5	3.1	81.8	1.0	3.3	13,540		2,230			[3] 91.7		A75155
		2		8.4	81.2	10.4	.5	3.0	82.7	1.1	2.3	13,700							
		3		9.3	90.7		.6	3.4	92.4	1.2	2.4	15,300							
	LV 5	1	3.6	7.8	81.5	7.1	.6	3.4	82.6	1.0	5.3	13,700		2,180			[3] 92.2		A75156
		2		8.1	84.5	7.4	.6	3.1	85.6	1.1	2.2	14,200							
		3		8.8	91.2		.6	3.3	92.4	1.2	2.5	15,330							
Big Lick prospect	LV 7	1	1.3	8.2	72.3	18.2	1.3	3.2	72.9	1.2	3.2	12,130					[2] 92.0		A1861
		2		8.3	73.3	18.4	1.3	3.1	73.9	1.2	2.1	12,290							
		3		10.2	89.8		1.6	3.8	90.5	1.5	2.6	15,060							
Lebanon County																			
Lakmia mine	6(?)	1	2.2	6.6	84.2	7.0	0.7	3.3	85.0	1.0	3.0	14,130		2,330			[3] 93.6		A93499
		2		6.8	86.1	7.1	.7	3.2	86.9	1.0	1.1	14,470							
		3		7.3	92.7		.8	3.4	93.6	1.1	1.1	15,580							
Schuylkill County																			
East Brookside mine	LV 5	1	2.6	8.3	76.2	12.9	0.5	2.9	78.1	0.9	4.7	12,750	2,080	2,160	2,340	1,574	91.5	15,300	A75890
		2		8.5	78.2	13.3	.5	2.7	80.2	1.0	2.3	13,100							
		3		9.9	90.1		.6	3.1	92.5	1.1	2.7	15,100							
	LV 4	1	.9	7.9	83.9	7.3	.5	3.1	85.8	.9	2.4	14,070	2,110	2,190	2,470	1,495	92.1	15,440	A75889
		2		7.9	84.7	7.4	.5	3.0	86.5	1.0	1.6	14,200							
		3		8.6	91.4		.6	3.3	93.4	1.0	1.7	15,330							
	No. 7	1	4.7	7.1	75.9	12.3	.8	3.4	76.2	1.1	6.2	12,580	2,370	2,660	2,730	1,556	92.9	15,370	A75669
		2		7.4	79.7	12.9	.8	3.0	80.0	1.2	2.1	13,200							
		3		8.5	91.5		.9	3.4	91.8	1.4	2.5	15,150							
	No. 8	1	4.8	6.4	78.0	10.8	1.0	3.4	77.9	1.0	5.9	12,910	2,620	2,800	2,910	1,514	93.7	15,500	A75664
		2		6.7	82.0	11.3	1.1	3.0	81.8	1.1	1.7	13,570							
		3		7.6	92.4		1.2	3.4	92.3	1.2	1.9	15,310							
	No. 8	1	3.6	6.5	79.2	10.7	.9	3.3	79.0	1.2	4.9	13,070		2,750			[3] 93.7		A75663
		2		6.8	82.1	11.1	.9	3.0	82.0	1.2	1.8	13,660							
		3		7.6	92.4		1.0	3.3	92.2	1.4	2.1	15,250							
	No. 8½	1	3.6	7.7	74.5	14.2	.6	3.2	75.3	1.2	5.5	12,430	2,710	2,800	2,860	1,483	92.2	15,360	A75662
		2		8.0	77.3	14.7	.6	2.9	78.1	1.2	2.5	12,890							
		3		9.4	90.6		.7	3.4	91.6	1.4	2.9	15,110							

See footnotes at end of table.

TABLE 5.—*Chemical analyses of bed samples from the west-central part of the Southern Anthracite field*—Continued

[Analyses by U.S. Bureau of Mines]

Mine	Bed [1]	Condition [2]	Percent: Proximate: Moisture	Volatile matter	Fixed carbon	Ash	Percent: Ultimate: Sulfur	Hydrogen	Carbon	Nitrogen	Oxygen	Calorific value (Btu)	Fusibility of ash (°F): Initial deformation temperature	Softening temperature	Fluid temperature	True specific gravity	Dry-mineral-matter-free basis: Fixed carbon percent	Calorific value (Btu)	Laboratory or index No.
	No. 9	1	4.2	7.2	64.6	24.0	0.6	2.9	64.9	0.9	6.7	10,640	2,550	2,780	2,920	1,669	92.7	15,260	A75666
		2		7.5	67.4	25.1	.6	2.5	67.8	.9	3.1	11,110							
		3		10.0	90.0		.8	3.4	90.5	1.2	4.1	14,820							
	No. 9	1	3.4	6.6	67.1	22.9	.6	2.9	67.3	.9	5.4	11,070	2,730	2,870	2,960+	1,640	93.7	15,430	A75665
		2		6.8	69.5	23.7	.6	2.7	69.7	.9	2.4	11,460							
		3		8.9	91.1		.8	3.5	91.3	1.2	3.2	15,020							
	No. 10	1	5.0	6.9	78.1	10.0	.8	3.4	78.1	1.0	6.7	12,920	2,350	2,740	2,920	1,489	93.1	15,380	A75667
		2		7.3	82.1	10.6	.9	3.0	82.2	1.1	2.2	13,600							
		3		8.1	91.2		1.0	3.3	91.9	1.2	2.6	15,210							
	No. 11	1	6.9	6.4	68.7	18.0	1.3	3.3	68.2	1.0	8.2	11,280	2,360	2,580	2,730	1,635	93.9	15,380	A75668
		2		6.9	73.7	19.4	1.4	2.7	73.2	1.0	2.3	12,120							
		3		8.5	91.5		1.7	3.3	90.8	1.3	2.9	15,030							
Phoenix Park No. 3	No. 14	1	2.8	2.5	82.1	12.7	.5	2.2	79.2	.7	4.6	12,580					[3] 98.6		5956
		2		2.5	84.4	13.1	.6	2.0	81.5	.7	2.3	12,930							
		3		2.9	97.1		.6	2.3	93.7	.8	2.6	14,870							
Pine Hill mine	No. 6	1	1.6	4.4	83.3	10.7	.5	2.2	81.8	.7	4.1	13,130					[3] 96.0		A10897
		2		4.4	84.7	10.9	.5	2.1	83.1	.7	2.7	13,340							
		3		5.0	95.0		.5	2.3	93.2	.8	3.2	14,960							
Pine Knot mine	No. 5	1	5.3	3.8	74.9	16.0	.6	2.5	73.8	.6	6.5	11,700	2,880	2,980	2,990+	1,710	97.0	15,140	A75479
		2		4.1	79.0	16.9	.6	2.0	77.9	.6	2.0	12,360							
		3		4.9	95.1		.7	2.4	93.7	.8	2.4	14,860							
	No. 6	1	4.5	3.9	83.2	8.5	.5	2.5	82.5	1.2	4.9	12,970	2,940	2,960+		1,662	96.5	15,030	A 75488
		2		4.1	87.1	8.8	.5	2.0	86.4	1.2	1.1	13,580							
		3		4.5	95.5		.6	2.2	94.8	1.3	1.1	14,890							
	No. 7	1	5.1	4.6	80.9	9.4	.6	2.5	80.9	.7	5.9	12,680	2,570	2,780	2,870	1,675	95.7	14,980	A75487
		2		4.9	85.2	9.9	.6	2.0	85.2	.8	1.5	13,360							
		3		5.4	94.6		.7	2.3	94.6	.9	1.5	14,830							
	No. 7	1	3.5	4.0	77.2	15.3	2.0	2.3	75.7	.6	4.1	12,030	2,320	2,500	2,670	1,733	97.5	15,110	A75486
		2		4.1	80.1	15.8	2.1	2.0	78.4	.6	1.1	12,460							
		3		4.9	95.1		2.5	2.3	93.1	.7	1.4	14,800							
	No. 8½	1	4.8	4.3	77.3	13.6	.6	2.5	76.5	.8	6.0	12,900	2,420	2,710	2,810	1,700	96.3	15,040	A75482
		2		4.5	81.2	14.3	.6	2.0	80.3	.8	2.0	12,710							
		3		5.2	94.8		.7	2.4	93.7	1.0	2.2	14,820							
	No. 8	1	5.3	4.3	80.9	9.5	.7	2.7	80.1	.8	6.2	12,750	2,630	2,790	2,860	1,644	96.1	15,130	A75483
		2		4.6	85.3	10.1	.8	2.2	84.6	.8	1.5	13,460							
		3		5.1	94.9		.9	2.5	94.0	.9	1.7	14,960							
	No. 9	1	4.8	4.7	81.6	8.9	.6	2.7	81.0	.8	6.0	12,930	2,340	2,810	2,900	1,625	95.6	15,130	A75484
		2		4.9	85.8	9.3	.6	2.3	85.1	.9	1.8	13,590							
		3		5.5	94.5		.7	2.5	93.8	.9	2.1	14,980							
	No. 10	1	6.5	4.0	76.9	12.6	.8	2.8	75.4	.7	7.7	12,070	2,790	2,860	2,940	1,658	96.7	15,140	A75480
		2		4.3	82.3	13.4	.9	2.2	80.7	.8	2.0	12,920							
		3		4.9	95.1		1.0	2.5	93.2	.9	2.4	14,920							
	No. 10	1	4.6	5.4	77.4	12.6	.5	2.6	77.5	.8	6.0	12,340	2,270	2,360	2,550	1,668	94.8	15,110	A4481
		2		5.7	81.1	13.2	.6	2.2	81.3	.8	1.9	12,940							
		3		6.5	93.5		.7	2.5	93.7	.9	2.2	14,920							
	No. 11	1	6.2	4.9	79.1	9.8	.6	2.8	79.3	.7	6.8	12,540	2,760	2,830	2,910	1,639	95.3	15,090	A75485
		2		5.3	84.3	10.4	.7	2.2	84.6	.8	1.3	13,370							
		3		5.9	94.1		.7	2.5	94.5	.9	1.4	14,930							
	No. 8	1	2.7	6.5	76.4	14.4	.6	2.3	76.1	1.1	5.5	12,080					[3] 93.7		A28774
		2		6.6	78.6	14.8	.7	2.1	78.3	1.1	3.0	12,410							
		3		7.8	92.2		.8	2.5	91.9	1.3	3.5	14,570							
Gold mine run [4]	(?)	1	2.2	6.6	84.2	7.0													
		2		6.8	86.1	7.1													
		3		7.3	92.7		.8	3.4	93.6	1.1	1.1	15,580					[3] 93.6±		S
West Brookside mine	LV 5	1	3.3	3.3	84.3	9.1	.6	3.1	81.4	.8	5.1	13,350					[3] 97.3±		5943
		2		3.4	87.2	9.4	.6	2.8	84.2	.8	2.2	13,810							
		3		3.7	96.3		.6	3.1	92.9	.9	2.4	15,248							

[1] Names and numbers of coal beds are those used in this report and not those used by U.S. Bureau of Mines.
[2] 1, sample as received; 2, dried at 105° C; 3, moisture and ash free.
[3] Dry mineral-matter-free fixed carbon percentages calculated by authors
[4] U.S. Bureau of Mines analysis compiled from data in tables 3-5 (Turner, 1934).

TABLE 6.—*Chemical analyses of breaker samples from the west-central part of the Southern Anthracite field*

[Analyses by U.S. Bureau of Mines]

Breaker	Percent					Calorific value, Btu, as received basis	Dry-mineral-matter-free basis	Samples analized and collected at breaker on specified date	Principal source of coal processed in breaker
	Proximate				Ultimate				
		Dry coal							
	Moisture as received	Volatile matter	Fixed carbon	Ash	Sulfur		(Fixed carbon, percent)		
Schuylkill County									
Blackwood	5.9	8.1	77.4	14.4	0.6	12,240	91.9	7 analyses, Mar. 18, 1948	Coal beds in Llewellyn Formation near village of Llewellyn on Sharp Mountain.
Branch	5.0	6.6	79.5	13.9	.6	12,270	93.8	4 analyses, Oct. 7, 1948.	
Colitz	5.0	5.0	81.8	13.2	.5	12,180	95.6	8 analyses, Sept. 13, 1948	Coal beds in upper part of Llewellyn Formation near Marlyn in Pottsville quadrangle.
Joliet	3.7	7.6	82.3	10.1	.8	13,030	93.7	1 analysis of delivered coal in fiscal year 1948.	Coal beds in Llewellyn Formation near Joliet.
D and B	6.0	5.8	78.6	15.6	.7	11,820	94.8	2 analyses of delivered coal in fiscal year 1952.	Coal beds in Llewellyn Formation near village of Llewellyn.
Kemble	5.1	5.8	81.5	12.7	.5	12,540	94.7	1 analysis of delivered coal in fiscal year 1952.	Coal beds in Pottsville Formation about 1½ miles east of Fountain.
Marlyn	5.0	5.3	78.1	16.6	.4	11,790	95.3	8 analyses, Nov. 17, 1949	Coal beds in Llewellyn Formation southeast of Minersville.
Penna	5.9	5.6	82.0	12.3	.7	12,290	94.8	2 analyses, June 28, 1949	Coal beds near Oak Hill mine.
Schrader	4.3	5.2	80.3	14.4	.6	12,220	95.4	7 analyses, Oct. 11, 1948	Coal beds near Lytle shaft.
Dauphin County									
Franklin-Lykens	2.6	9.2	80.3	10.4	0.5	13,240	90.7	7 analyses, Jan. 20, 1949	Coal beds on Coal Mountain near Lykens Gap.

RELATION OF FIXED CARBON TO DEFORMATION

As early as 1873, Hilt showed that the fixed carbon content of coal beds increases with depth of burial in some places and concluded that the depth of burial was the principal reason for coal increasing in rank from lignite to anthracite.

A detailed study of the coal beds of South Wales (Strahan and Pollard, 1908) showed that the more deeply buried coal in a single shaft commonly had higher fixed carbon content than shallow coal. However, when the coal was considered areally, Strahan and Pollard found that deposits with the highest fixed carbon were located where the overburden apparently had been the least. They concluded that depth of burial was a factor in the increase of fixed carbon, but that differences in original composition were also important.

Ashley (1918, p. 17) determined that differences in amount of fixed carbon in benches of cannel coal versus benches of woody bituminous coal or semianthracite at the same locality gradually disappear as the fixed carbon content approaches 75 percent. This determination showed that variations in original composition were gradually nullified as the fixed carbon content increases. It also proved that in addition to depth of burial and variations in original composition, some other geologic phenomena must influence the fixed carbon content of the higher rank coals.

C. D. White (1915; 1925, p. 269–271; 1935) propounded and supported his theory of progressive regional carbonization of coal in the eastern United States and stated that the controlling factor in carbonization was horizontal stress or thrust. He demonstrated that the chemical variations in lesser rank coals are largely due to original composition but that in the higher rank coals these original differences "yield to the effects of dynamochemical transformation and become less conspicuous" (1925, p. 272). White believed that the weight of overburden was responsible for the early stages of fixed carbon increase but that it was of minor importance in the more advanced stages. He also believed that the weight of overburden provided resistance to horizontal thrust in the progressive metamorphism of coal and that the later stages of fixed carbon increase were the result of combining overburden weight, thrust pressure, and increased temperature resulting from compression and friction. White attempted to prove that horizontal thrust was the controlling factor in convert-

TABLE 7.—*Average chemical analyses of breaker samples from the west-central part of the Southern Anthracite field*

[Form of analysis: As received. Analyses by U.S. Bureau of Mines]

Breaker	Percent: Proximate: Moisture	Percent: Proximate: Volatile matter	Percent: Proximate: Fixed carbon	Percent: Ultimate: Ash	Percent: Ultimate: Sulfur	Calorific value Btu	Dry mineral-matter-free basis: Fixed carbon percent	Dry mineral-matter-free basis: Calorific value Btu	Analyses are averages of samples identified by following U.S. Bur. Mines laboratory or index numbers	Principal source of coal processed in breaker
Dauphin County										
Williamstown	3. 0	8. 7	75. 7	12. 7	0. 7	12, 790	91. 2		83, 84, 86, 91, 94, 95	Coal beds north of Williamstown.
Schuylkill County										
Buck Run	5. 0	4. 6	78. 7	11. 5	0. 7	12, 430	95. 6	15, 080	1380, 1382 1384–1386, 1389, 1390.	Coal beds near Buck Run.
Correale	5. 9	6. 0	75. 1	12. 9	. 6	12, 140	94. 0		1448–1454	Coal beds near Donaldson.
Di Renzo	3. 7	6. 0	78. 7	11. 6	. 5	12, 750	94. 2		H10119–H10123	Coal beds close to breaker, which is 2 miles west of Minersville.
K and P	6. 7	5. 1	77. 4	10. 7	. 5	12, 320	94. 9		G67055–G67058	Coal beds near Minersville.
Legal	3. 1	7. 0	76. 5	13. 4	. 9	12, 670	93. 2		F66917–F66920	Various coal beds as much as 5 miles from breaker, which is 3 miles north of Goodspring.
Lohb	4. 5	5. 4	80. 8	9. 3	. 5	13, 090	94. 7		F48297–F48302	Primrose No. 11 coal bed near Tremont.
Lytle	3. 8	4. 1	79. 4	12. 8	. 6	12, 420	96. 7	15, 080	1624, 1626, 1627, 1632, 1637, 1638.	Various coal beds in Lytle mine.
Oak Hill	3. 6	4. 2	80. 7	11. 6	. 6	12, 650	96. 5	15, 100	1639, 1641, 1645, 1648, 1649.	Various coal beds in Oak Hill mine.
Otto	6. 9	6. 2	74. 7	12. 1	. 5	11, 770	93. 6	14, 730	1373, 1374, 1376, 1377, 1379.	Various coal beds in Otto mine.
Penag	3. 5	6. 1	79. 4	11. 0	. 5	12, 980	93. 9		F80974–F80981	Lykens Valley No. 2 coal bed.
Sherman	4. 8	5. 7	76. 6	13. 0	. 6	12, 350	94. 5		1813, 1815, 1816, 1818–1820, 1822.	
Swatara	7. 9	4. 5	78. 7	8. 8	. 8	12, 450	95. 6		G70356–G70363	Coal beds near Zerbe.
Primrose	5. 3	4. 4	78. 1	12. 2	. 5	12, 300	96. 0	15, 110	1653, 1655, 1657, 1659, 1662, 1665, 1668, 1670.	
Indian Head	4. 6	4. 9	76. 1	14. 3	. 5	12, 160	95. 4	15, 240	1928, 1931, 1935, 1938, 1942, 1945, 1949, 1953.	Coal beds near Indian Head breaker, which is between Newton and Tremont.
West-West	6. 6	5. 9	76. 9	10. 6	. 5	12, 150	94. 0	14, 840	1671 and 1673	Coal beds south of Minersville.
New Castle	7. 4	5. 0	78. 3	9. 3	. 6	12, 350	95. 2		1828 and 1829	Coal beds near New Castle, delivered coal, 2 miles east of Pine Knot mine.
Westwood	2. 9	7. 6	76. 5	12. 0	. 7	12, 920	92. 3		1956, 1958, 1960, 1964, 1968, 1970.	Various coal beds near Joliet.

ing bituminous coal to anthracite by citing the eastward progressive regional carbonization across the Appalachian basin into the area of this report (1925, p. 257–258, 265–266).

In 1934, Turner (p. 339–342) strongly endorsed White's theory of progressive regional carbonization. He also presented a map (Turner, fig. 3) on which lines connecting points of equal volatile matter cross the modern anticlinal and synclinal axes at nearly 90° in the western part of the Anthracite region. Turner (1934, p. 339–340) stated that this divergence of nearly 90° between lines of equal volatile matter and structural axes "was brought about by thrust pressures applied prior to and in a different direction from the immediate pressures that caused the present anticlines and synclines."

A year later, Hendricks (1935, fig. 1), in a report on the Arkansas-Oklahoma coal field, showed lines of equal fixed carbon that trend at a large angle to the known structural trends. He attributed this to a regional increase in unrelieved thrust pressure, during the structural deformation of the coal basin and Ouachita Mountains to the south, that coincided with the regional increase in fixed carbon content. However, in two areas where the number of analyses was sufficient to provide detailed control (Hendricks, 1935, fig. 1 and p. 946), the maximum fixed carbon content of the coals was found "to coincide closely with the location of the most tightly folded anticline along lines drawn at right angles to the Ouachita Mountains and extending through these two areas," Hendricks' concluded (p. 947), "The major variations in the fixed carbon content of the coals are directly related to structural deformation produced by pressure exerted from the south."

Barnes (1962) found that coal in flat-lying undeformed Tertiary beds in the Cook Inlet area, Alaska, increases in rank vertically from subbituminous C at the surface to high-volatile B bituminous coal at a depth of about 11,000 feet. He also found that at the surface where these beds are deformed structurally, the coal increases in rank from subbituminous to anthracite in a distance of about 55 miles in the direction of increasing structural deformation.

Hendricks (1935) and Turner (1934) agreed that deformation was the major factor in determining the fixed carbon content of a coal at a specific locality. They arrived at diametrically opposed views, however, as to the timing of this deformation. Hendricks maintained that variations in fixed carbon in the Arkansas-Oklahoma coal field were largely due to the forces that deformed the Ouachita Mountains whereas Turner believed that such variations in the Anthracite region were caused by forces preceding those that formed the modern Appalachians.

All available coal analyses of bed, breaker, and tipple samples in the area were assembled to determine whether variations in fixed carbon are due to an early episode of deformation (Turner's theory), to deformation that resulted in the modern structural features (Hendricks' theory), or to depth of burial. Most of these samples were collected and analyzed by the U.S. Bureau of Mines, and some are listed in tables 5–7. After assembly, the fixed carbon content of each analyzed coal sample was calculated according to specifications of the American Society for Testing Materials (1954). The calculated fixed carbon of each sample or group of samples was then plotted on a base map at the correct geographic position. Finally, a contour map showing lines of equal fixed carbon content was constructed (fig. 42).

Normally, only bed or face analyses are used in determining the rank of coal. However, in the Anthracite region bed analyses are lacking, and present conditions

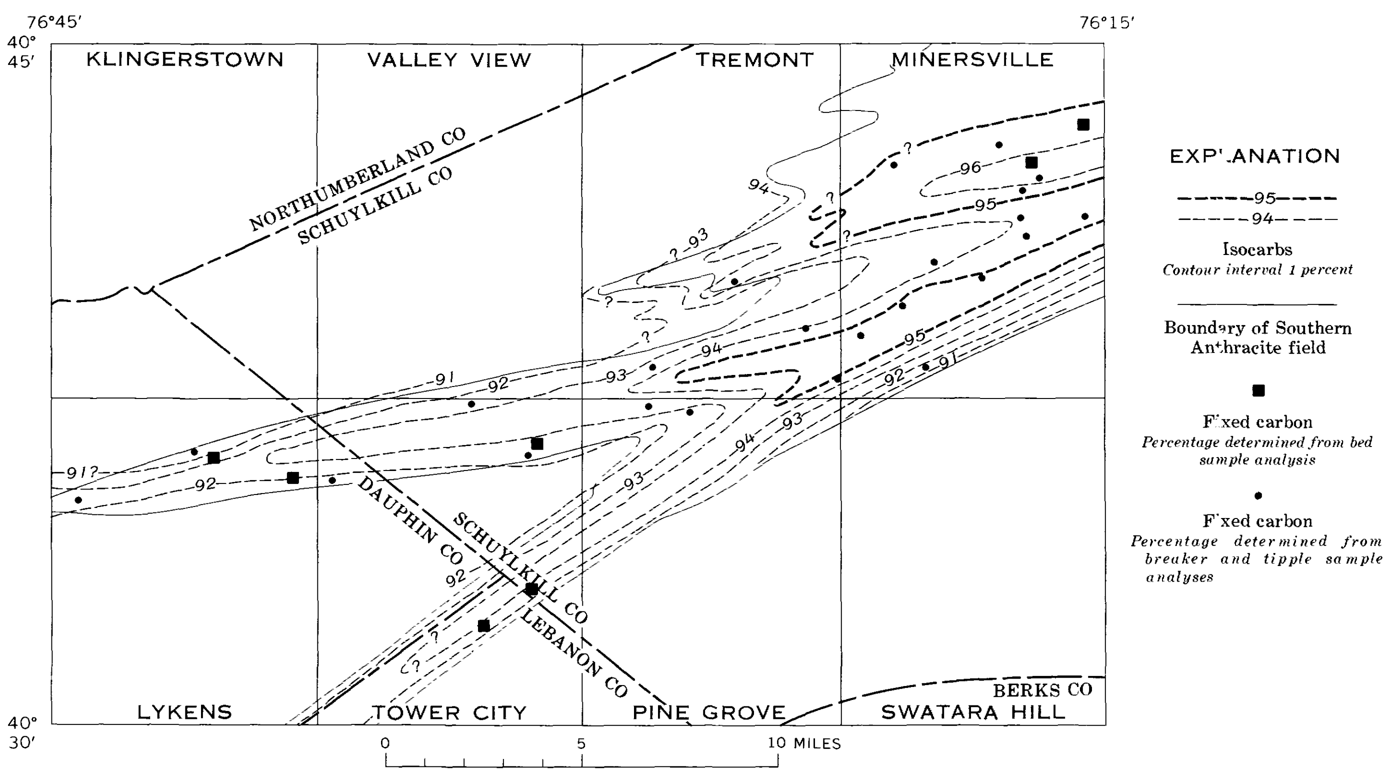

FIGURE 42.—Isocarb map of the west-central part of the Southern Anthracite field, Pennsylvania. Samples for bed, breaker, and tipple analyses collected and analyzed by the U.S. Bureau of Mines.

are not favorable for a program of sample collection and analysis. Because of the lack of proper analyses, it was decided to incorporate pertinent data from breaker and tipple analyses in constructing figure 42. This decision was made with the full understanding that such data are not customarily considered to be satisfactory for rank determination because the source of the analyzed coal generally is not known with sufficient accuracy. This possible objection is not pertinent with reference to figure 42 because the source of the coal for each breaker and tipple analysis is known with considerable accuracy.

The part of Turner's map (1934, fig. 3) that covers the area was based on 31 bed analyses of coal collected in seven localities. When these analyses are plotted at their correct locations on a map, it is possible to draw lines, as he did, that connect points of equal fixed carbon at nearly 90° to the structural axes. However, when the breaker and tipple analyses are located properly, it is obvious that such connections are not correct.

Figure 42 shows that lines of equal fixed carbon in the west-central part of the Southern Anthracite field regionally parallel the trend of the larger structural features and locally parallel the smaller structural features. It also shows that the fixed carbon increases eastward and toward the cores of the more highly deformed large synclines. The regional eastward increase in fixed carbon is in the direction of the regional plunge of the major structural features and also in the direction of greater depth of burial. Turner (1934, fig. 3) and White (1925, fig. 1) recognized the eastward increase in fixed carbon, but they did not identify the regional or local parallelism to structural features. The parallelism between the lines of fixed carbon and structural features shown in figure 42 points inevitably to the fact that the variations in fixed carbon were largely determined at the time the structural features were deformed. Secondarily, the regional eastward increase in fixed carbon must have been caused by a greater depth of burial, which substantiates Hilt's law (1873) and corroborates the evidence from the South Wales coal field (Strahan and Pollard, 1908) and the Cook Inlet area, Alaska (Barnes, 1962). White's (1925, fig. 1), Hendricks' (1935, p. 947), and Barnes' (1962) conclusion that fixed carbon increases toward areas of greater deformation are validated by the area. Finally, the conclusion reached by White and Turner (1934, fig. 3) that the lines of equal fixed carbon in the area trend across the structural axes at nearly right angles is not validated.

The analyses from East Brookside and the Pine Knot mines (table 5) seem to contradict one another, in proving that higher rank coal is concentrated in the more highly deformed areas. According to stratigraphic position, the fixed carbon content of beds in the East Brookside mine seemingly increases as the overburden decreases; but when the analyses are properly located three dimensionally, the fixed carbon content actually increases toward the core of the tightly folded Tremont syncline. In contrast, in the Pine Knot mine, when the analyses are located correctly, the fixed carbon content not only increases as the overburden increases, but also increases toward the complexly deformed Jugular thrust fault. Thus, despite the apparent reversal with reference to stratigraphy, the fixed carbon content in both mines increases toward areas of greater deformation.

In conclusion, the regional eastward increase of fixed carbon in the area with greater depth of burial substantiates Hilt's law. The parallelism of equal fixed carbon lines with structural features also substantiates White's, Hendricks', Turner's, and Barnes' conclusion that structural deformation is exceedingly important in determining the rank of anthracite and semianthracite.

AMOUNT OF COAL IN THE REPORT AREA

During the last seven decades, three estimates of original tonnage of coal preserved in the Anthracite region have been made. The earliest of these was by Smith (1895, p. 2147–2152). Ashmead (1926, table 20), after an examination of mine maps, revised Smith's estimate. Subsequently, in 1945, Ashley revised Ashmead's estimate slightly to include county and culm bank reserves and to show coal produced by river dredges.

Smith (1895, p. 2150) determined that the Southern Anthracite field originally contained 9,198,435,263 gross tons, or 10,302,247,474 net tons, of coal in beds more than 2.0 feet thick. He used 2,040 net tons of coal per acre foot over most of the field, but locally used 2,190 net tons per acre foot. He did not divide the coal tonnage into categories based on coal thickness, amount of overburden, and reliability of data. The first step in Smith's method was to determine the probable average cumulative thickness of all coal beds along 21 cross sections. He then computed the probable average cumulative thickness of coal in the 22 areas that lay between the cross sections by averaging the probable cumulative thickness of coal in the adjacent cross sections. The tonnage of coal was then determined by multiplying the average thickness of coal for each of the 22 areas by the acreage of each of the areas. The combined areas lying between his cross sections 20 through 29 is approximately the same as the report area. Between and adjacent to his cross sections 20 through 29, he estimated 4,451,176,000 gross tons, or 4,985,317,000 net tons of original coal, which is 48.39 percent of the total original coal (10,302,247,474 net tons) that he estimated for the Southern Anthracite field.

Ashmead in 1926 (table 20) estimated that the original content of coal in the Southern Anthracite field was 10,114,000,000 gross tons, or 11,327,736,000 net tons. He apparently followed Smith in not calculating reserves in beds less than 2.0 feet thick and by using 2,040 and 2,190 net tons per acre. Assuming that 48.39 percent of the total original coal in the Southern Anthracite field is in the report area as determined from Smith's estimate, 5,481,490,000 tons of coal originally underlay the area according to Ashmead's estimate.

In 1945, Ashley (table 2) revised Ashmead's estimate of the original coal in the Southern Anthracite field to 10,880,000,000 net tons. Assuming that 48.39 percent of the total original coal in the field underlay the area, the original coal in the area was 5,264,832,000 net tons according to Ashley's estimate.

In this report the original coal content of the area was calculated as follows:

Using the coal cross sections (Wood and Trexler, 1968b, sheet 4; 1968c, sheets 5, 6) the underground length of each coal bed was measured to 0.04 mile (about 200 ft.). The average thickness of coal in each coal bed on each cross section was then determined, taking into account the thousands of surface and subsurface measurements. Subsequently, the average thickness of coal in each coal bed was averaged between adjacent cross sections. In addition, the length of each coal bed as determined from adjacent cross sections was also averaged. After this, the average distance between adjacent cross sections was determined to 0.1 mile (about 500 ft.). Then, the average distance between cross sections was multiplied by the averaged length of each coal bed and then by the average weight of the coal in tons per square mile (measured to 0.10 of a foot and 2,000 tons per acre-ft., or 1,280,000 tons per sq. mi. ft.).

The calculations made for this report indicate that about 7.6 billion net tons of anthracite originally underlay the area with about 0.44 billion tons being in Dauphin County, 0.66 billion tons in Lebanon County, and 6.50 billion tons in Schuylkill County. About 60 percent of the original coal lay within 2,000 feet of the surface, 15 percent lay between 2,000 and 3,000 feet, 20 percent lay between 3,000 and 6,000 feet, and 5 percent lay below 6,000 feet.

About 6.3 billion tons, or 83 percent, of the 7.6 billion net tons of original coal was in beds more than 42 inches thick; about 0.6 billion tons, or about 8 percent, was in beds from 28 to 42 inches thick; and 0.7 billion tons, or about 9 percent, was in beds from 14 to 28 inches thick. In descending stratigraphic order, the original tonnage of coal was distributed as follows: Peach Mountain (No. 18) coal bed, 180 million tons; Tracy (No. 16) coal bed, 260 million tons; Diamond (No. 14) coal bed, 320 million tons; Orchard (No. 12) coal bed, 300 million tons; Primrose (No. 11) coal bed, 570 million tons; Holmes (No. 10) coal bed, 490 million tons; Top Split (No. 9), Middle Split (No. 8½), and Bottom Split (No. 8) coal beds of the Mammoth coal zone, 1.75 billion tons; Skidmore (No. 7) coal bed, 410 million tons; Buck Mountain (No. 5) coal bed, 470 million tons; Lykens Valley No. 4 coal bed, 390 million tons; Lykens Valley No. 5 coal bed, 510 million tons; all other coal beds, 1.95 billion tons.

The present estimate of the original tonnage of anthracite which underlay the area is about 2,615 million tons greater than that of Smith (1895), about 2,118 million tons greater than that calculated by the authors from Ashmead's 1926 estimate, and about 2,336 million tons greater than that calculated by the authors from Ashley's 1945 estimate.

The great discrepancy between the present estimate and those of Smith, Ashmead, and Ashley is due to the following:

1. The coal beds were found to descend to much greater depths than thought previously,
2. Much previously unknown thrust faulting has repeated coal beds at many places so that a greater volume of coal is preserved than thought heretofore,
3. Stratigraphic miscorrelations and previously unrecognized effects of complex folding incorrectly reduced the earlier estimates,
4. Tonnages of coal in beds from 14 to 24 inches thick were not included in the early estimates,
5. Much more complete data for determining the average thickness of coal beds and for determining the three-dimensional configuration of coal beds were available for this estimate than for the previous estimates.

About 318 million tons of anthracite is estimated to have been extracted from the area as of January 1, 1961. Thus, of the original content of anthracite, about 7.282 billion tons are estimated to remain in ground as of that date. Combined underground and strip mining in the area probably has extracted an average of about 65 percent of the coal available in the mined areas. In addition to the 318 million tons estimated to have been mined, about 171 million tons of the original content of coal was disturbed in mining operations. Therefore, about 489 million tons of the coal originally present had been extracted and disturbed in mining as of January 1, 1961, leaving about 7.1 billion tons as virgin, or undisturbed coal.

About 75 percent of the original coal lay within 3,000 feet of the surface. Therefore, about 5.7 billion tons of the 7.6 billion tons of original content of coal was within 3,000 feet of the surface. The mined and disturbed-in-mining tonnage of 489 million tons as of January 1, 1961, was almost all extracted from above 3,000 feet. Thus, about 5.2 billion tons of virgin undisturbed coal remained above 3,000 feet as of January 1, 1961, and of this amount probably about half, or 2.6 billion tons, is recoverable in the future. About 1.9 billion tons of this coal is estimated to be in beds 42 inches or more thick. This 1.9 billion tons of thick recoverable coal is distributed approximately as follows: Dauphin County, 110 million tons; Lebanon County, 170 million tons; and Schuylkill County, 1,620 million tons. In addition to the 5.7 billion tons above 3,000 feet, approximately 1.9 billion tons of original coal lies below 3,000 feet. It is not possible to say how much, if any, of this deeply buried coal will be recoverable in the future.

SAND AND GRAVEL

Sand has been excavated at several points in the area where weathering has caused the Montebello Sandstone Member of the Mahantango Formation to disintegrate. Principal among these localities are the crests of Fisher Ridge and Hooflander Mountain in the Klingerstown quadrangle (GQ–700) and the crest of Swope Mountain in the Swatara Hill quadrangle (GQ–689).

Gravel has been quarried from the Sharp Mountain Member of the Pottsville Formation on Broad Mountain north of the villages of Heckscherville and Mount Pleasant (Wood and Trexler, 1968b, pt. 1, sheet 1). This gravel was used as base material in highway construction. The conglomerate units of the Sharp Mountain are suitable for quarrying at many other places where they have weathered into gravel.

ROAD METAL

The upper beds of the upper shale member of the Mahantango Formation and the lower beds of the Trimmers Rock Sandstone have been quarried for road metal in numerous shallow pits in the Pine Grove and Swatara Hill quadrangles. Some of these pits are large and have been excavated extensively during recent years. As road building continues, these rocks will probably be quarried increasingly for highway construction materials.

BUILDING STONE

Conglomerate and sandstone of the Tuscarora Sandstone, Pocono Formation, and Pottsville Formation have been widely used in the past for building materials in the Anthracite region. However, because of the ease with which most rocks of these formations disintegrate during weathering, it does not seem likely that they will be used as widely in the future. This usage probably will decrease because manmade materials are more uniform in size, more durable, and easy to manufacture.

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