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LATE WISCONSINAN TO PRE-ILLINOIAN(G?) GLACIAL AND PERIGLACIAL EVENTS IN EASTERN PENNSYLVANIA

(GUIDEBOOK FOR THE 57TH FIELD CONFERENCE FRIENDS OF THE PLEISTOCENE NORTHEASTERN SECTION MAY 20-22, 1994 , HAZLETON, PENNSYLVANIA)

by
Duane D. Braun[1]
Edward J. Ciolkosz[2]
Jon D. Inners[3]
Jack B. Epstein[4]

With Contributions by
[5]G. Michael Clark
[6]Ira D. Sasowsky
[7]Robin Koberle

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[1]Bloomsburg University, Bloomsburg, PA
[2]Penn State University, State College, PA
[3]Pennsylvania Geological Survey, Harrisburg, PA
[4]U.S. Geological Survey, Reston, VA
[5]University of Tennessee, Knoxville, TN
[6]Nittany Geoscience
[7]Pagnotti Enterprises

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57TH FIELD CONFERENCE
FRIENDS OF THE PLEISTOCENE
NORTHEASTERN SECTION

MAY 20-22, 1994
HAZLETON, PA.

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THIS GUIDEBOOK IS DEDICATED TO THOSE WHO HAVE TRAVERSED THIS AREA BEFORE:

	*
Pre-Friend	
Edward H. Williams, Jr	(1895, 1917)
Frank Leverett	(1934)
Past Friend	
George H. Crowl	(1975, 1980)
Present Friends	
Louis C. Peltier	(1949)
William D. Sevon	(1975, 1980)

* Date of most significant publication (cited in References).

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LATE WISCONSINAN TO PRE-ILLINOIAN (G ?) GLACIAL EVENTS IN EASTERN PENNSYLVANIA

by: Duane D. Braun, Geography & Earth Science
Bloomsburg University, Bloomsburg, PA 17815

Introduction

The field trip will make a transect from the late Wisconsinan limit to the pre-Illinoian (G ?) limit, the maximum southern extent of any evidence of glaciation. The late Wisconsinan terminus shows how the Laurentide Glacier interacts with the moderate relief landscape of the Appalachian Plateau and Valley and Ridge in eastern Pennsylvania. The trend, landform, and depositional style of the late Wisconsinan limit is used as a model for older glacial limits in eastern Pennsylvania. The trip will start at the Late Wisconsinan border and travel southward to examine progressively older glacial deposits. It will be emphasized, that due to the increasingly greater erosional truncation of the incrementally older glacial deposits, there is considerable uncertainty as to the difference in the degree of weathering between Illinoian and pre-Illinoian aged material. Another peculiarity of eastern Pennsylvania glacial geology, is that there is almost a complete lack of absolute dates and stratigraphic superposition of glacial deposits of different glacial stages is almost never observed. Each glacial advance into eastern Pennsylvania effectively removes nearly all traces of earlier glacial advances and demonstrates the erosive nature of the ice as it crosses this moderate relief landscape.

Number and age of glacial events

During the last decade the full multiplicity of cold-warm alternations and associated glacial advances has been recognized from marine oxygen isotope records and radiometric dating of terrestrial deposits of interbedded volcanic and glacial material. The oxygen isotope record has been used as a time-space diagram of glaciation (Fig. 1) (Sibrava, 1986; Shackleton, 1987; Mix, 1987; Braun, 1989c). About ten isotope maxima approach or exceed the amplitude of the late Wisconsinan oxygen isotope maximum and would be expected to have ice extent similar to or greater than the late Wisconsinan limit (LW line on Fig. 1). Four early to middle Pleistocene isotope maxima, stages 6,12,16, and 22 (>W line on Fig. 1) exceed the late Wisconsinan maximum. The terrestrial record suggests that there may be as many as five or six middle and early Pleistocene ice advances more extensive than the late Wisconsinan advance of the North American Laurentide ice sheet (Bowen and others, 1986) (Table 1). Nearly all Pleistocene isotope maxima approach the amplitude of the early Wisconsinan isotope maxima (Fig. 1). Early Wisconsinan ice is now thought to have advanced to the Great Lakes (Eyles and Westgate, 1987; Eyles and Schawrcz, 1991), or into the Great Lake basins (Hicock and Dreimanis, 1989),or even to the north side of the Appalachian Plateau (Muller and Calkin, 1993; Young and others, 1993) but not into Pennsylvania (Braun, 1988; Ridge and others, 1990).

Thus during the Pleistocene as many as ten glaciations may have approached the late Wisconsinan terminus and four probably reached beyond that limit in eastern Pennsylvania (Fig. 1). The late Wisconsinan advance has destroyed nearly all traces of previous advances right up to the late Wisconsinan terminal margin (LW on Fig. 2 & 3). The four or five early to middle Pleistocene advances more extensive than late Wisconsinan have left a fragmentary record as much as 45 miles (70 km) to the southwest of the late Wisconsinan margin (I & PI on Fig. 2 & 3). The only area of older glacial deposits that has a significant thickness and retains limited relict constructional topography is in a belt 3 to 15 (5-25 km) wide in front of and sub-parallel to late Wisconsinan boundary (I on Fig. 2 & 3). This margin is currently interpreted as the late Illinoian limit, the most recent glacial advance to extend beyond the late Wisconsinan limit (Fig.1 & Table 1, Isotope Stage 6). Further to the southwest and sub-parallel to the Illinoian limit is a 12 to 30 mile (20-50 km) wide belt of thin, very discontinuous to almost nonexistent glacial drift materials of pre-Illinoian age (PI on Fig. 2 & 3). These materials are most likely of pre-Illinoian B, D, and/or G age (Table 1), the most extreme isotope stages 12, 16, and 22 (Fig.1).

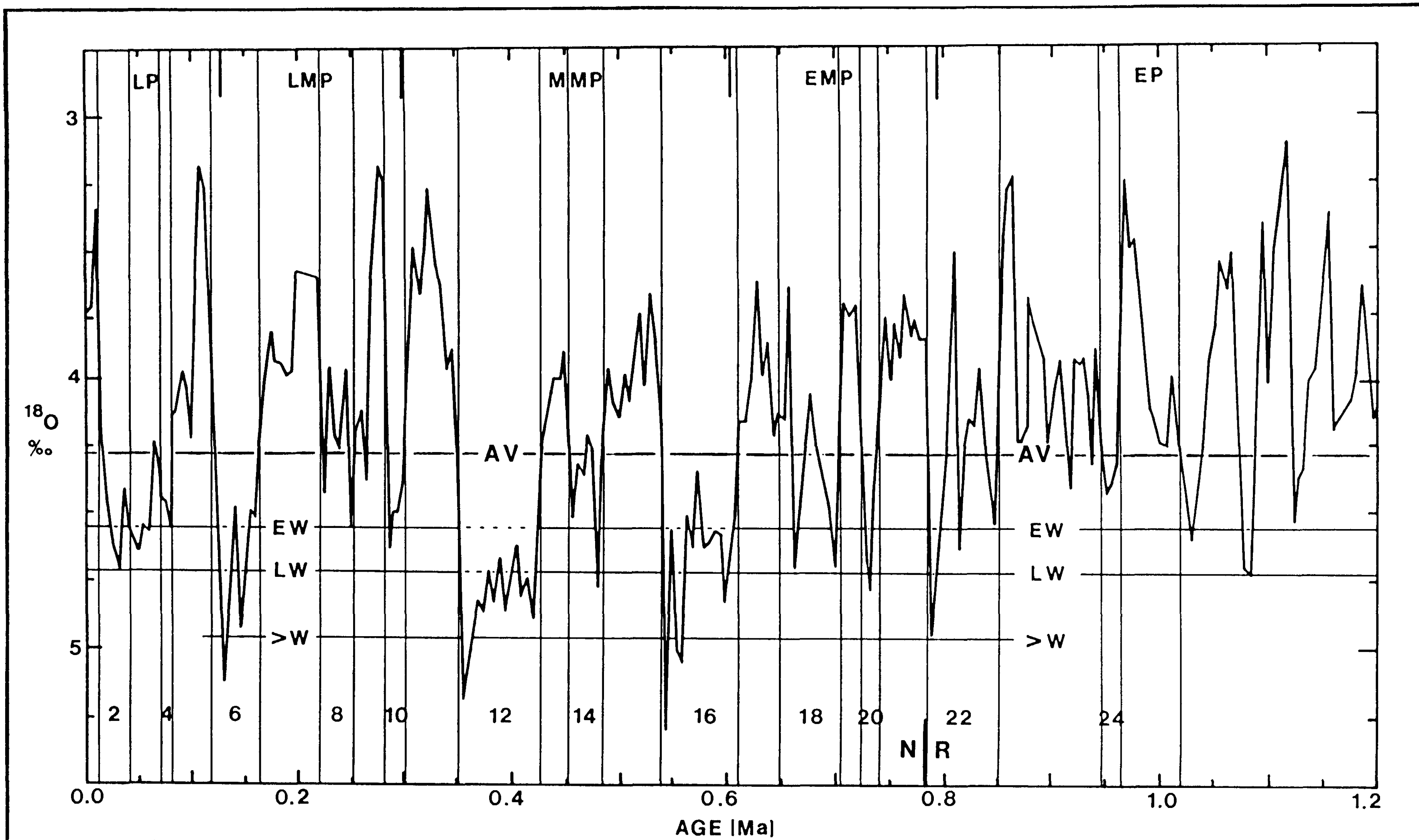
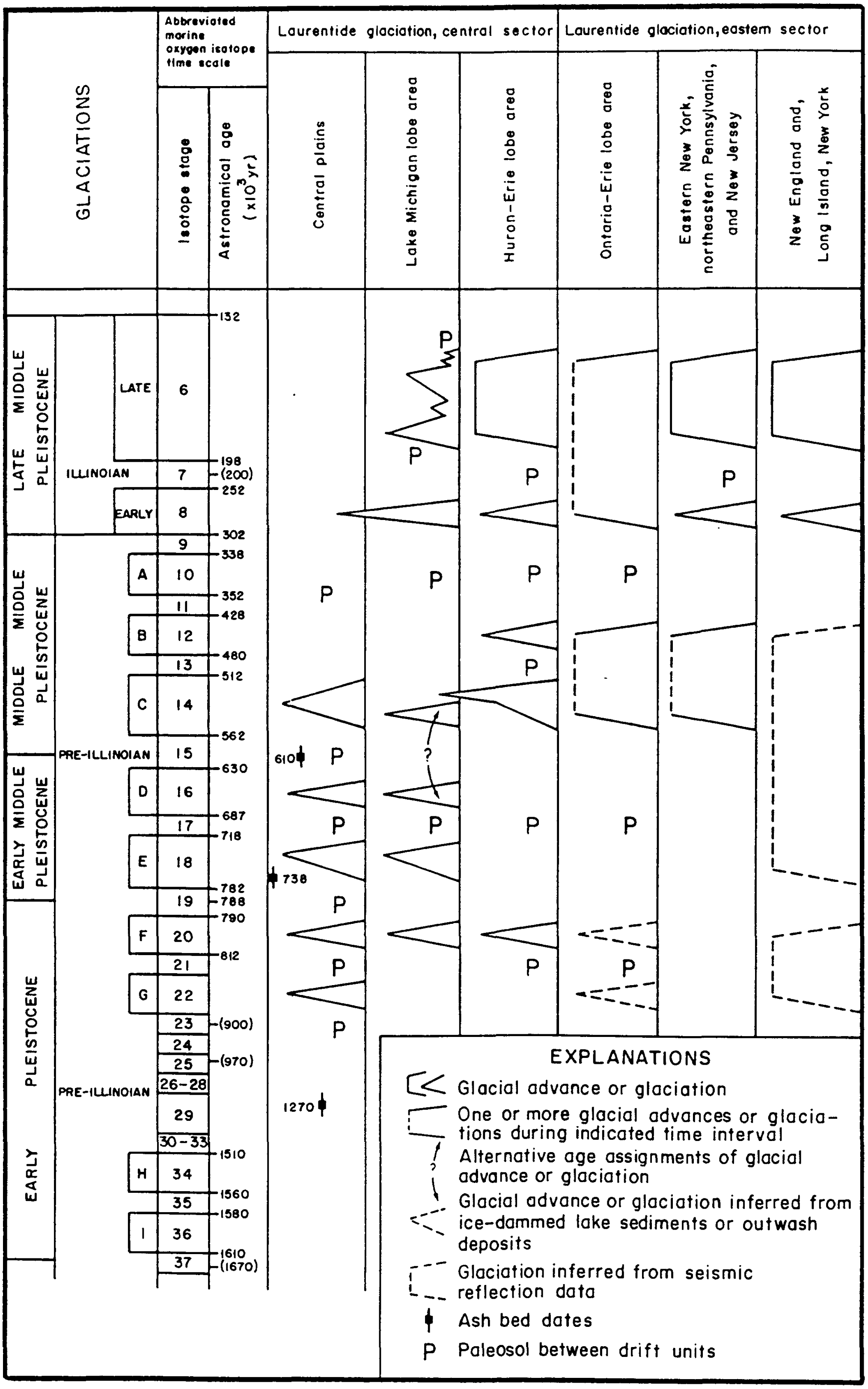


Figure 1. Oxygen isotope record showing the number and amplitude of glacial events during the Pleistocene (Braun, 1989c, Fig. 2A). Ten or more events have a amplitude similar to that of the late Wisconsinan (LW) and should have brought ice near or into eastern Pennsylvania. Four events have a markedly greater amplitude (>W) and should have permitted ice to advance beyond the late Wisconsinan limit. Periglacial conditions should have reached eastern Pennsylvania under early Wisconsinan (EW) or even average conditions (AV)(Porter, 1989). LP, late Pleistocene; LMP, late middle Pleistocene; MMP, middle middle Pleistocene; EMP, early middle Pleistocene; EP, early Pleistocene; N/R, Normal - Reversed magnetic polarity boundary.

Table 1. Illinoian and older Pleistocene time divisions, isotope stages, and Central Plains to New England glacial advance record. (Modified from Richmond and Fullerton, 1986)



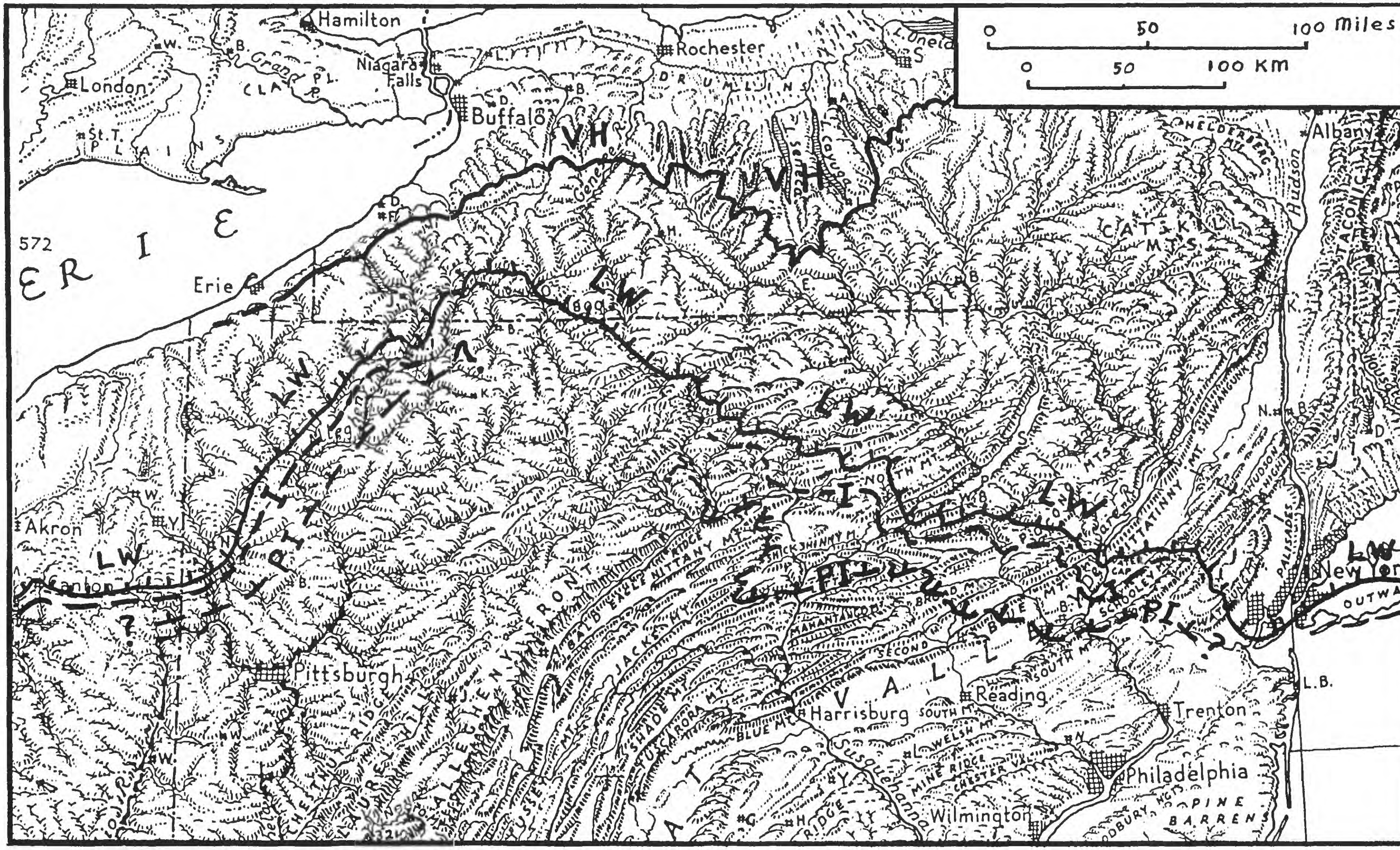


Figure 2. Ice margins across Pennsylvania and adjacent states. VH, Valley Heads; LW, late Wisconsinan; I, late Illionian; PI, pre-Illinoian (base map, Raisz, Physiographic map of U.S.A.)

The late Wisconsinan terminus, a model for older ice margins

The well-defined late Wisconsinan or Olean terminal margin (Lewis, 1884; Crowl and Sevon, 1980) shows how ice interacts with the landscape of northeastern Pennsylvania and can be used as a model for the form of pre-Wisconsinan ice margins. The overall trend of this margin across northeastern Pennsylvania is N60W (Fig. 2). Hilltop striations on the Appalachian and Pocono plateaus within 30 miles (50 km) of the margin indicate a regional ice flow direction of S10W to S30W. Where ice entered the Ridge and Valley province, especially within the North Branch Susquehanna area, striations within 6 miles (10 km) of the margin show ice flow to be sub-parallel to bedrock strike.

The local relief of northeastern Pennsylvania Valley and Ridge is typically about 600 to 1000 feet (200-300 m) from valley floor to adjacent ridgetop. This local relief caused lobation of the ice front along the general trend of the ice margin (Fig. 2 & 3). In the strike valleys, the ice front projected as a lobe 2 to 5 miles (3-8 km) farther than where the ice front crossed adjacent strike ridges (Braun, 1988). Even in the narrow, deep valleys of the Appalachian Plateau, valley glacier-like ice lobes projected only 3 to 5 miles (4-8 km) down valley from where the ice front crossed the plateau top.

The types of deposits and landforms along the late Wisconsinan margin can also be used as a model for older glaciations. The margin generally does not display true knob and kettle (hummock and swale) morainal topography (distinct end moraine with greater than 10 feet of local relief of Crowl and Sevon, 1980). Distinctive knob and kettle topography occurs only across the Pocono plateau and on the flank of some major strike ridges in the Valley and Ridge. Along most of the margin there is either no morainal topography at all (ground moraine of Crowl and Sevon, 1980) or less than 10 feet local relief morainal topography (indistinct end moraine of Crowl and Sevon, 1980). Strike valleys draining away from the late Wisconsinan margin contain prominent heads of outwash (kame fans or kame deltas)(Stop 1, Day 1). Outwash deposits at the margin approach 200 feet (60 m) in thickness and bury the pre-glacial landscape thereby causing local stream derangement. Outwash valley trains slope gently away from the margin. Recessional positions are marked by a series of heads-of-outwash, the New England stagnation-zone retreat model (Koteff, 1974). Where valleys draining away from the ice are relatively narrow, as on the Appalachian and Pocono Plateaus, outwash deposits have been mostly removed by paraglacial and post-glacial erosion.

The degree of glacial modification progressively lessens as the late Wisconsinan (Olean) terminus is approached. This same region has some of the highest and steepest local relief (1600 ft or 500 m) of the northern Pennsylvania Appalachian Plateau. The area is characterized by hillslopes that are devoid or almost so of glacial drift, flat topped ridges with a thin veneer of glacial till, and valleys partly filled by thick kame and outwash deposits. A number of tributaries to the North Branch Susquehanna River are northeast draining and contained proglacial lakes impounded by the ice (Braun, 1987, 1989a). Existing small lakes in the region are generally a result of drift blockages of local drainages, sometimes in conjunction with glacial scour of the bedrock (Cameron, 1970; Coates, 1974).

Along the North Branch Susquehanna River at Wilkes-Barre, Pennsylvania, ice flow was parallel to the Lackawanna synclinal axis. There, over a distance of 10 miles (15 Km), as much as 400 feet (125 m) of sandstone, shale, and anthracite coal was scoured out below present river level (Ash, 1950). Resistant sandstone outcrops in the present river bed at watergaps both immediately upstream and downstream of the site. The remaining question is: How much erosion has occurred generally and how could one tell without the situation just described?!

Early Wisconsinan ice, less extensive than late Wisconsinan ice

DELINEATION OF AN EARLY WISCONSINAN ICE ADVANCE

The existence of an early and/or mid Wisconsinan glacial advance beyond the late Wisconsinan margin in eastern Pennsylvania was first proposed by Connally (unpublished, 1972) and Sevon (1973, 1974). Additional areas of early Wisconsinan were described by Crowl and others (1975), Sevon and others (1975), and Cadwell (unpublished), in the region that had been mapped

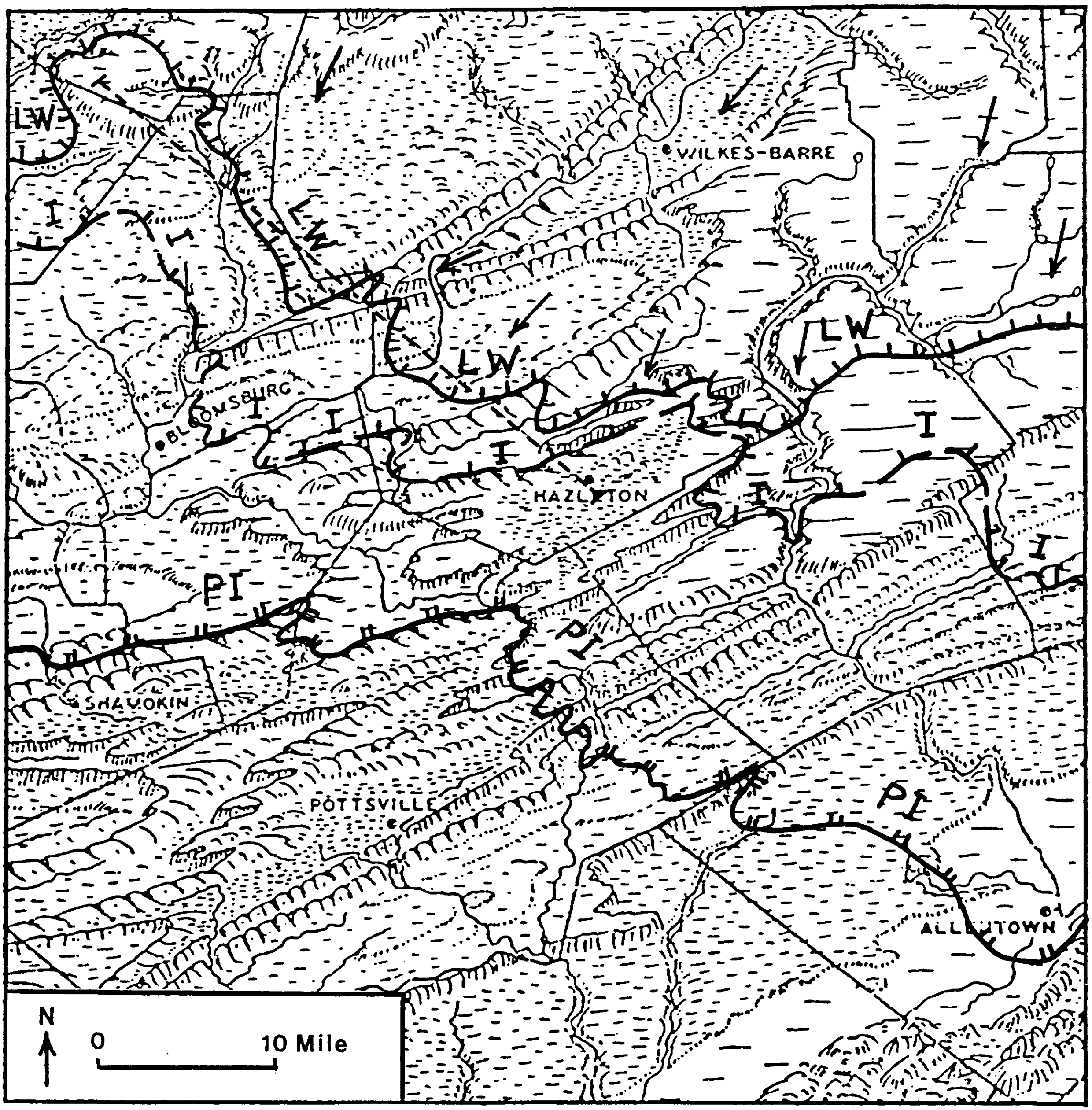


Figure 3. Ice margins across the Valley and Ridge of eastern Pennsylvania. LW, late Wisconsinan (Crowl and Sevon, 1980); I, late Illinoian (Braun, 1988 and ongoing work); PI, pre-Illinoian (Braun, 1988 and ongoing work). East of Hazelton, the late Wisconsinan margin is at the base of a strike ridge while the late Illinoian margin surmouts the ridge, producing an apparent close proximity for the two margins. Short arrows represent bedrock striations within the late Wisconsinan margin. The dashed line is the line of view extending northwest feom Stop 6 at Hazelton to North Mountain on the edge of the Appalachian Plateau. (Physiographic map from Deasy and Griess, 1963)

as Illinoian by Leverett (1934). Sevon and others (1975) assigned an early Wisconsinan age to the deposits and gave three lines of evidence to support that assignment. First, the drift did not possess a Sangamon weathering profile typical of Illinoian drift, either in depth of oxidation or concentration of iron oxide, but was weathered to a greater depth and possessed a thicker soil profile than late Wisconsinan drift. Second, the drift had not been as extensively eroded as Illinoian drift, but had been sufficiently eroded and colluviated such that almost all original constructional topography had been destroyed. Third, the areal distribution of the drift indicated deposition by an ice advance of less magnitude than that of the Illinoian, but locally of greater magnitude than that of the late Wisconsin. Due to uncertainty as to the precise age of the early Wisconsinan material, the time-stratigraphic term Altonian, as defined in Illinois (Willman and Frye, 1970) was used. The term covers a longer time interval of the early Wisconsinan than the more narrowly defined time-stratigraphy developed for areas adjacent to Pennsylvania by Dreimanis and Karrow (1972).

Bucek (1975), in mapping the Pleistocene geology of the Williamsport area, proposed a series of rock-stratigraphic and soil-stratigraphic units for the late Wisconsinan, the early Wisconsinan, and the Illinoian. The Altonian-age rock-stratigraphic unit was named the Warrensville till and was tentatively correlated with the Titusville till in northwestern Pennsylvania, at that time thought to be 35 Ka old (White, 1969).

Marchand suggested that early Wisconsinan ice tongues extended down the North Branch Susquehanna to Sunbury and down the West Branch Susquehanna to New Columbia (Marchand and others,1978) . He noted that the early Wisconsinan ice carried a high percentage of red Catskill sandstone and shale fragments compared to other glaciations. Supposedly some early Wisconsinan ice-contact-stratified drift existed, but no specific sites were mentioned. Early Wisconsinan outwash, loess, and colluvium have been severely eroded and altered by Woodfordian glacial and periglacial activity. No evidence of recessional or terminal moraines exists, suggesting no stillstand of the early Wisconsinan ice when it was outside of the Woodfordian margin.

Crowl and Sevon (1980), in retracing the late Wisconsinan margin, noted a number of small patches of early Wisconsinan drift outside of the Woodfordian border. Berg and others (1980) mapped long narrow tongues of early Wisconsinan ice extending down strike valleys away from the late Wisconsinan margin. Detailed study of soils in till derived from redbeds at two sites previously identified as early Wisconsinan (Levine, 1981; Levine and Ciolkosz, 1983) showed that several different parameters mutually suggested a stage of soil development between that of the late Wisconsinan and the Illinoian.

PROBLEMS WITH THE EARLY WISCONSINAN ICE ADVANCE

The first problem with the delineation of the early Wisconsinan is with the map pattern of long narrow lobes extending down strike valleys (Fig. 4), a style of ice margin first used by Leverett (1934) for his Illinoian border (Fig.8). From the distribution of supposedly early Wisconsinan drift, lobes were mapped extending as far as 30 km down strike valleys to either side of the Pocono plateau (Crowl and others, 1975; Sevon and others, 1975). A series of narrow lobes extending down some but not all strike valleys were mapped across the remainder of the Valley and Ridge (Berg and others, 1980). This pattern of long narrow lobes differs markedly from the 3-8 km or so lobation of the late Wisconsinan (Woodfordian) ice described earlier. For the early Wisconsinan ice not to overtop adjacent strike ridges, the ice could only thicken about 200 m in the 30 km up-ice from the margin. This is a far smaller thickening up-ice than suggested by ice surface gradients measured from the late Wisconsinan lobes (Crowl and Sevon, 1980). It is also an ice surface gradient far smaller than is observed at the terminus of present day ice sheets (Nye, 1952; Weertman, 1973; Paterson, 1969) and would require the glacier's bed be as fluid as that suggested in the Great Lakes basins (Boulton and others, 1985).

An additional problem with the mapped distribution of early Wisconsinan deposits occurs in the strike valley to the south of the Pocono plateau and along the North Branch Susquehanna River.

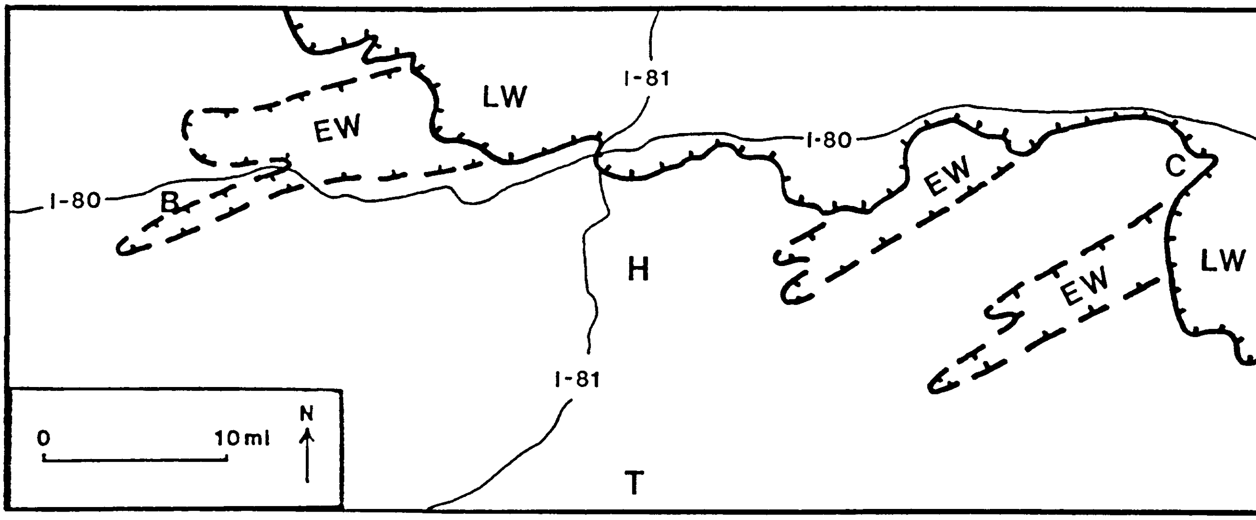


Figure 4. Map showing long narrow lobes of early Wisconsinan ice (EW) projecting down some but not all strike valleys in front of the late Wisconsinan ice margin (LW)(Berg and others, 1980). H, Hazelton

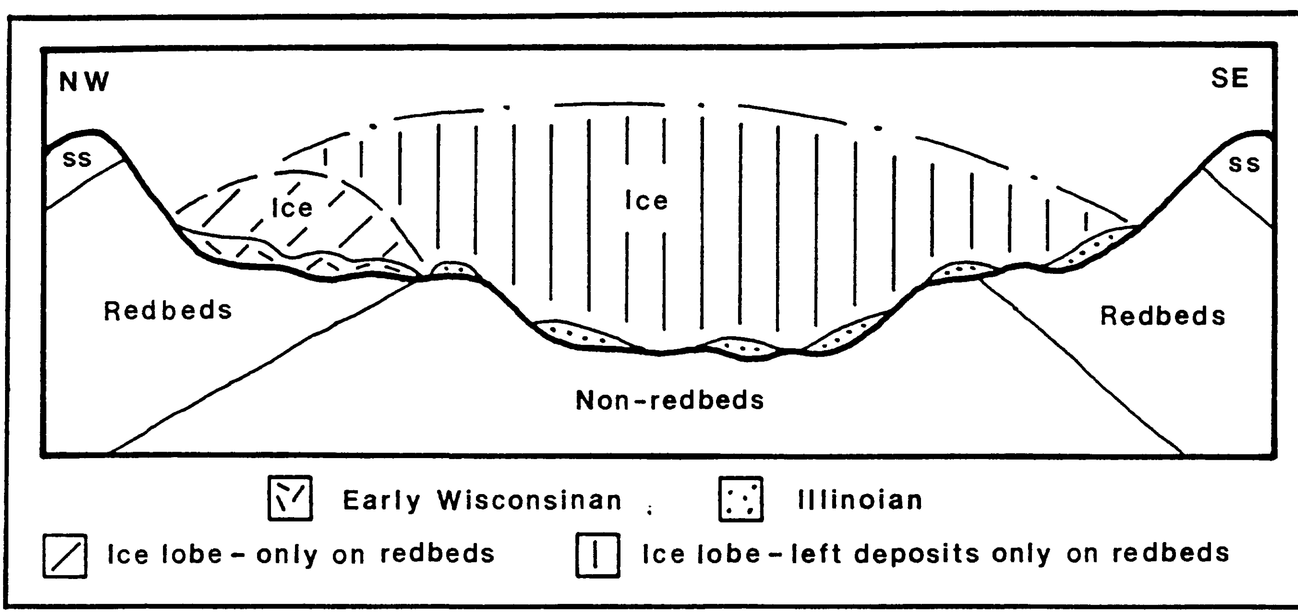


Figure 5. Cross-section sketch showing the distribution of different aged glacial deposits within a strike valley. Ice flow is from the northeast ("out of the paper")(Braun, 1988). The early Wisconsinan deposits are primarily on redbeds, while the Illinoian deposits are primarily on non-redbeds. — — — Surface of early Wisconsinan ice lobe just covering deposits of that age. — · — Surface of early Wisconsinan ice lobe covering the whole strike valley but leaving deposits only on the redbeds.

There, where ice flow is parallel to strike, the early Wisconsinan material only occupies the north side of each valley (Fig. 5). That side is underlain by the Catskill redbeds and is a rolling upland that is higher than the adjacent shale and carbonate lowland. From the mapped distribution of deposits, the ice only flowed in a narrow tongue along the higher Catskill belt and not along the adjacent lower shale and carbonate belt, a physically impossible situation (Fig. 5). Or, ice occupied the whole valley, only leaving deposits on the Catskill Formation while not eroding the Illinoian deposits that are common on the the lower shale and carbonate area, a highly improbable situation (Braun, 1988).

Another problem with the mapped pattern of the early Wisconsinan deposits is that there are often 100 meters or more across areas of Illinoian deposits scattered among the areas of early Wisconsinan deposits (Fig. 6). In redbed areas, Illinoian patches are scarce within the discontinuous early Wisconsinan veneer while in non-redbed areas, Illinoian patches are common and the early Wisconsinan patches are scarce. This map pattern implies that early Wisconsinan ice did not effectively scour the pre-existing Illinoian deposits and often did not deposit any material on top of the Illinoian deposits. There is no published description of a site where early Wisconsinan material overlies Illinoian material, even though Illinoian material often lies beside early Wisconsinan material at the ground surface.

This patchwork quilt of younger and older deposits is not what is observed within the late Wisconsinan margin. There are no known patches of early Wisconsinan or Illinoian deposits at the ground surface within the area of late Wisconsinan ice cover in eastern Pennsylvania. The Late Wisconsinan ice effectively removed all older material right up to the very edge of its advance. This suggests that either the early Wisconsinan ice was uniquely non-erosive and selective in its deposition or that the material mapped as early Wisconsinan is really Illinoian. If there were no early Wisconsinan patches, what would be left is a patchwork quilt of different glacial drift lithofacies of Illinoian age.

A further problem with the way the early Wisconsinan has been delineated is that nearly all areas mapped as early Wisconsinan-aged drift are on redbeds or in material dominated by redbed fragments from the Bloomsburg, Catskill or Mauch Chunk Formations. In the strike valleys to either side of the Poconos and in the North Branch Susquehanna lowland, over 95 % of the mapped early Wisconsinan drift patches are directly on redbeds (Crowl and others, 1975; Sevon and others, 1975; Berg, 1975; Inners, 1978; Crowl and Sevon, 1980; Inners, 1981). In the West Branch Susquehanna lowland between Bald Eagle Mtn. and the Appalachian Plateau, over 90% of the mapped early Wisconsinan drift patches are on redbeds east of Williamsport (Bucek, 1975; Wells and Bucek, 1980; Crowl and Sevon, 1980). West of Williamsport more than 80 % of the early Wisconsinan drift is on redbeds (Faill and others, 1977). The material mapped as early Wisconsinan outside of the redbed areas is still dominated by redbed material. Overall, the map pattern shows that material identified as early Wisconsinan is nearly always redbed material and suggests that what is being mapped is a lithofacies variation rather than an age variation.

Furthermore, the soils considered thus far as characteristic of early Wisconsinan weathering are derived from redbed material (Crowl and others,1975; Sevon and others, 1975; Bucek, 1975; Marchand and others, 1978; Levine, 1981; Levine and Ciolkosz, 1983). The early Wisconsinan-aged Warrensville till type locality is on the Catskill redbeds (Bucek, 1975). The proposed soil-stratigraphic units developed in the till are the Leck Kill, Meckesville, and Albrights series, all dominated by redbed parent material (Bucek, 1975). Marchand suggested that early Wisconsinan ice carried a high percentage of red Catskill fragments and that was reflected in the soils developed from early Wisconsinan till (Marchand and others, 1978). In a comparison of soil development in till of various ages, the two early Wisconsinan-aged soils were redbed derived (Leck Kill series). The five Illinoian-aged soils had no specified parent material (Levine, 1981; Levine and Ciolkosz, 1983), but were located on non-red bedrock. Using clay accumulation data, Levine and Ciolkosz (1983) developed a regression equation to predict the absolute age of early Wisconsinan and Illinoian soils. While the 41,000 BP predicted age for the early Wisconsinan soils was reasonable, again the four early Wisconsinan soils used were redbed derived. Thus far, there has been no published characterization of a non-redbed, non-truncated, early Wisconsinan-aged soil in eastern Pennsylvania.

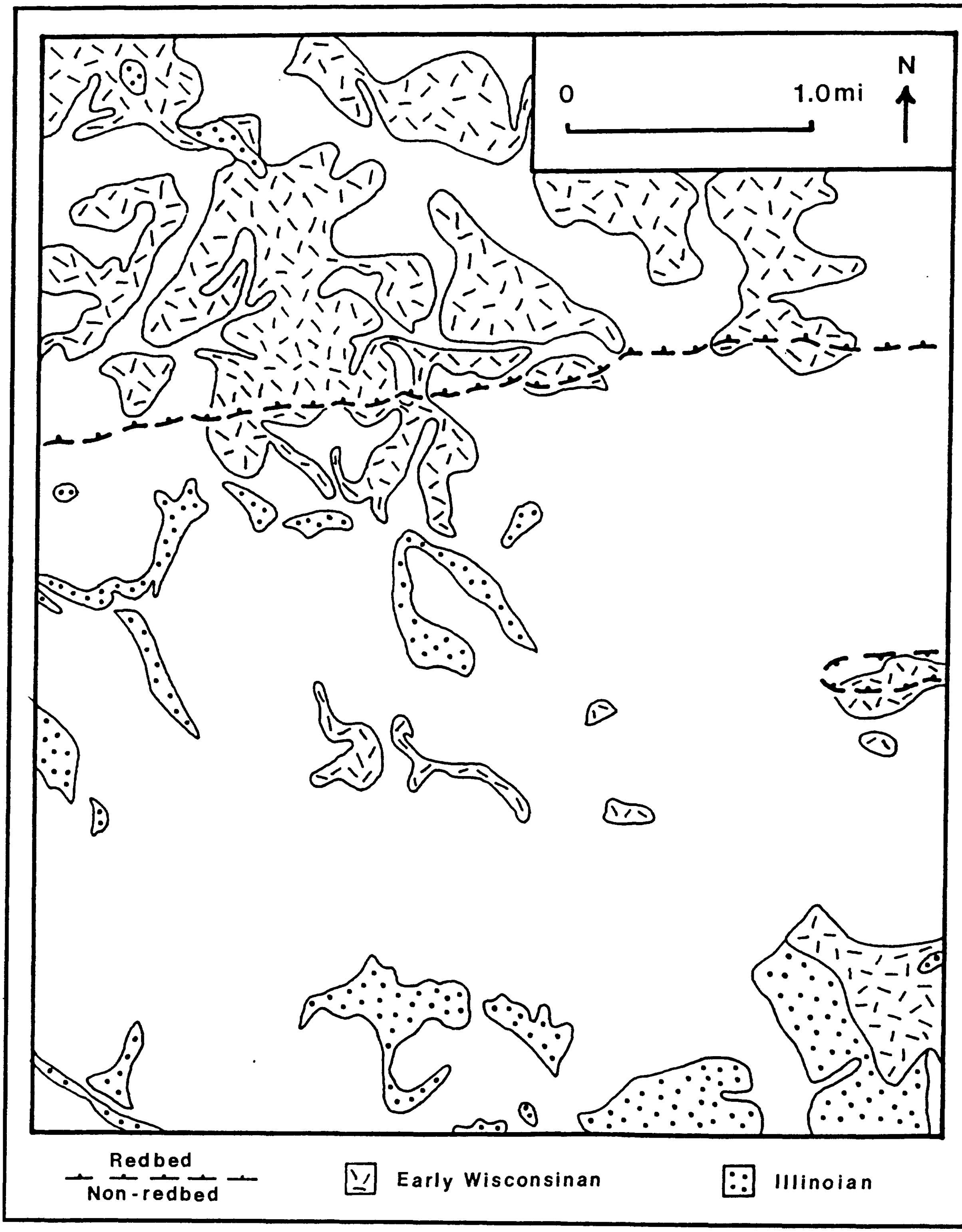


Figure 6. Map of "patch work quilt" of early Wisconsinan and Illinoian deposits. Early Wisconsinan deposits predominate on redbed areas, while less continuous Illinoian deposits predominate on non-redbed areas. (Simplified from Wells and Bucek, 1980).

This consistent association of supposed early Wisconsinan age with redbed material suggests an altemative hypothesis: what is being differentiated is parent material rather than time. The intermediate degree of weathering of the "early Wisconsinan" material could be caused by a slower production or illuviation of very fine oxide particles in the hematitic redbed parent material as compared to the non-redbed, non-hematitic parent material. The bright reddish-yellow color (2.5 to 7.5 YR) characteristic of Illinoian-aged material may be a product of very fine iron oxide particles (Ciolkosz, personal communication). If such particles are produced or transported more slowly in the redbed tills, then the material would look less "bright" and therefore younger. That this is the case, is suggested by the bright reddish-yellow weathering of non-redbed clasts and lenses of high conductivity sandy gravel within lower conductivity redbed dominated till. This is shown in strip mine exposures in the anthracite region (Stop 2, 4, & 5).

The contention that redbed material retards the development of red-yellow weathering coloration is further supported by re-examination of a large kame delta near Weatherly. The deposit was originally designated as Illinoian in age (Leverett, 1934) and more recently designated as early Wisconsinan in age (Crowl and others, 1975; Sevon and others, 1975). Exposures in the eastern end of the delta, proximal to the ice, show sand and gravel capped by redbed diamict (lodgement and/or re-sedimented till) that exhibits weathering intermediate between that of early Wisconsinan and Illinoian tills. Soils mapping , verified by field checking, shows the central and least eroded part of the kame delta has a deeply weathered Allenwood soil surrounded by a periphery of less weathered Tunkhannock soil. On the south and west sides of the deposit, erosion appears to have truncated the weathering profile, leaving the less weathered material at the surface. On the east side of the deposit, it is the redbed diamicton cover that has produced a less weathered appearance than in the center of the deposit. That the same sand and gravel deposit exhibits deep weathering where exposed at the ground surface and markedly less deep weathering where covered by a veneer of redbed till, implies that the till is retarding the relative degree of weathering development.

To summarize, there are three basic problems with the early Wisconsinan as it has been delineated in eastem Pennsylvania. The first problem is that the map pattem of long narrow extremely low gradient ice tongues in the strike valleys is physically impossible. The second problem is that the patchwork quilt map pattem of early Wisconsinan and Illinoian deposits requires the early Wisconsinan ice to be uniquely selective in its erosion and deposition. The third problem is that the supposed early Wisconsinan intermediate degree of weathering is consistently observed on redbed material and not on non-redbed material. This suggests that what is being differentiated is parent material rather than time. A simple solution to these problems is that the supposed early Wisconsinan material actually represents a redbed-derived lithofacies of Illinoian age.

Illinoian glaciation

Leverett (1934) correlated some of the older glacial deposits closest to the Wisconsinan terminus with the Illinoian of Illinois on the basis that they were distinctly more weathered and eroded than the Wisconsinan material but less weathered and eroded than glacial material farther to the south. Leverett mapped several long tongues of Illinoian deposits down strike valleys (Fig. 8). Some tongues were so long and narrow that had they actually represented ice lobes, the ice would have had impossibly low ice gradient and would be today considered physically impossible by glaciologists (Paterson, 1969; Weertman, 1973).

Sevon and others (1975) emphasized that while the Illinoian of Illinois could not be directly traced into Pennsylvania, the Illinoian age assignment was reasonable from the degree of weathering and erosion of the deposits. A series of tongues of Illinoian deposits, even longer than Leverett's, were mapped down the strike valleys of the region (Fig.7) (Berg and others, 1980). Again, many of the tongues required impossibly low ice gradients and could not mark an actual ice lobe margin. What the tongues actually represented were lines drawn around the known distribution, as of late 1970's, of weathered diamicts thought to have a glacial origin. The longest tongues

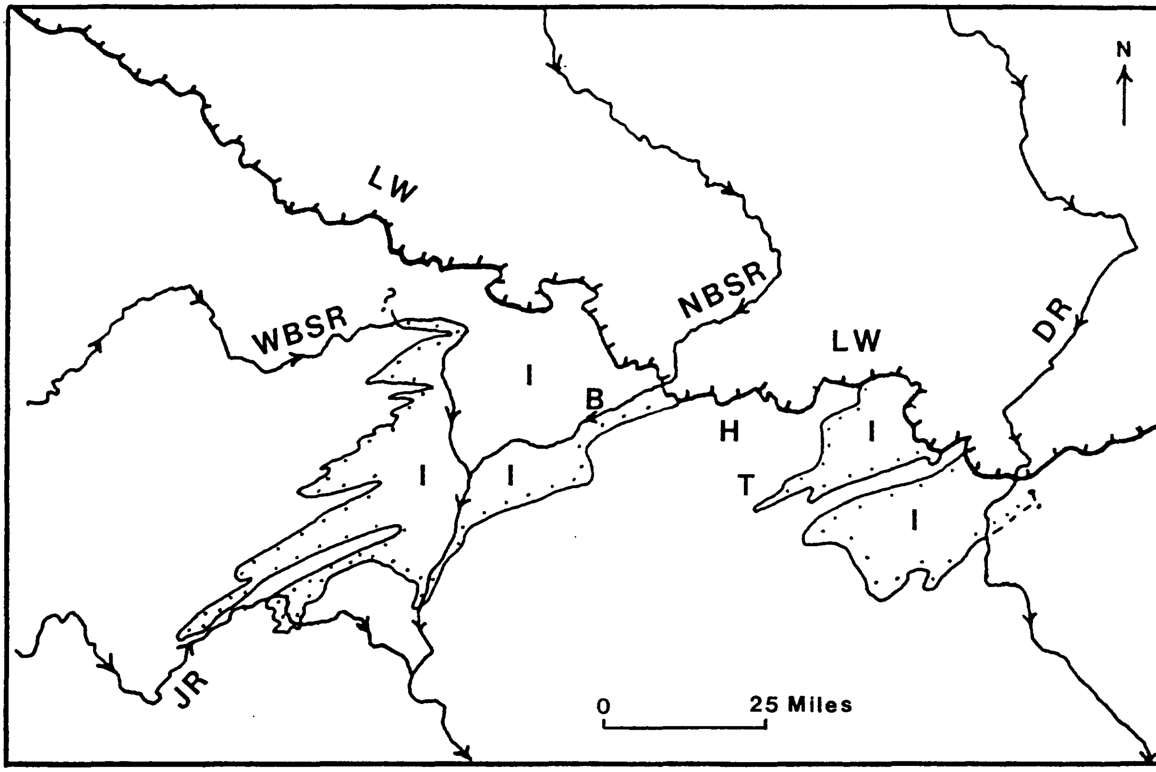


Figure 7. Map showing that, in contrast to the late Wisconsinan margin, the late Illinoian margin has been portrayed as having extremely long lobes projecting down some but not all lowlands (Berg and others, 1980). LW, late Wisconsinan; I, Illinoian; B, Bloomsburg; H, Hazelton; T, Tamaqua; DR, Delaware River; JR, Juniata River; NBSR, North Branch Susquehanna River; WSBR, West Branch Susquehanna River.

projected into the Juniata River valley and were placed on the map by Sevon (personal communication) to accommodate diamict deposits at the ends of the tongues thought at the time to be glacial till (Higbee, 1967). Kaktins (1986) demonstrated that the diamict in question was not glacial in origin.

For the Illinoian-aged deposits in the Williamsport, Pennsylvania area, a formal rock-stratigraphic name was proposed, the Muncy drift (Bucek, 1975; Wells and Bucek, 1980). Marchand proposed that there are two different Illinoian aged drifts in the region around the confluence of the North and West Branches of the Susquehanna (Marchand and others, 1978). The drifts were differentiated on the basis of the relative degree of weathering and soil development. The older Laurelton drift was correlated with the 260 Ka oxygen isotope stage 8 (Harden and Taylor, 1983), placing it in the early Illinoian of Richmond and Fullerton (1986). The younger White Deer drift would then be placed in the 150 Ka late Illinoian of Richmond and Fullerton (1986). At present, both materials are considered to be pre-Illinoian in age because of the degree of dissection of the deposits and because a reasonable glacial limit cannot be drawn using the presently mapped distribution of deposits (Braun, 1988, 1989b) .

Ongoing work has delineated a late Illinoian ice terminus that runs 2 to 11 miles (3-18 km) southwest of and sub-parallel to the late Wisconsinan ice margin (Braun, 1985, 1988, 1989b)(Fig.3). The margin, named the Bloomsburg Margin due to its prominent expression there, has been traced from the Great Valley northwest across the Valley and Ridge to Williamsport. The Bloomsburg Margin is essentially identical to the late Wisconsinan margin in regard to overall trend, degree of lobation, thick frontal kame deposits in valleys, derangement of local preglacial stream patterns, and relative lack of expression in upland areas. The late Illinoian terminal kame deposits often are 100 to 200 feet (30-60 m) thick and have as much or more volume than the late Wisconsinan terminal kames. The Bloomsburg Margin is the only place that the North Branch Susquehanna River has been forced from its pre-glacial channel by blockage from glacial deposits (Fig. 13)(Optional trip, Stop 1). This margin differs from the late Wisconsinan margin in that it is deeply weathered and eroded though in places, such as at Bloomsburg, constructional topography remains.

The maximum expectable degree of weathering of the Bloomsburg margin is observed at the Bloomsburg Interchange 35 of I-80. The site is a flat topped hill that projects northward from the side of a higher anticlinal ridge. The near horizontal hilltop surface extends eastward into a saddle that marks the preglacial water-gap of Fishing Creek through the anticlinal ridge (Fig. 13). That water-gap is now buried by at least 50 meters of sand and gravel that is part of the eastern end of the frontal kame. The stratigraphic sequence at the site starts with 0.5 to 1.0 meters of brown loess of late Wisconsinan age. The loess is underlain by 1 to 3 meters of oxidized yellowish red (5YR 5-6/6-8) roundstone diamict that is in turn underlain by 5 to 10 meters of yellowish red stratified sand and gravel. There is often a 1 meter or so thick transition zone from the diamict to the well-stratified sand and gravel where stratification is partly preserved. Beneath the strongly oxidized material is 10 to 20 meters of brown to gray slightly oxidized to unoxidized sand and gravel.

The roundstone diamict is not a glacial till but rather a result of post-depositional weathering of the kame sand and gravel (Braun and Kaktins, 1986). Stratification has been destroyed and fine matrix has been produced by a combination of clast dissolution and disintegration, clay illuviation, bioturbation, and cryoturbation. Other such roundstone diamicts also cap the higher fluvial terrace surfaces in the Valley and Ridge for 100's of kilometers south of the glacial limit.

The Bloomsburg Margin is considered late Illinoian in age primarily because at Bloomsburg and several other sites the gently sloping top surface of the frontal kame (head-of-outwash) retains its original constructional form. That form would have been lost and other younger glacial drift would have been deposited on top of the surface had it been overridden by another glaciation. The top surface of the kames is in places capped by late Wisconsinan loess or colluvium showing that there has been little erosion during the late Wisconsinan. That the Bloomsburg Margin is unaffected by later glacial activity indicates that it was formed by the last glaciation to reach beyond the late Wisconsinan border. This does not necessarily make the margin late Illinoian in age but the older a deposit is, the less likely it is for any original constructional form to be retained as repeated periglacial and interglacial erosional episodes dissect the landscape.

The difference in the degree of erosion and the continuity of glacial deposits to either side of the Bloomsburg margin also argue for a younger (late Illinoian) rather than older (pre-Illinoian) age. In the rolling uplands to the northeast of the Bloomsburg margin glacial deposits are observed on many broad hilltops and colluvium derived from such deposits is observed in many first order hollows. To the southwest of the Bloomsburg margin, the rolling uplands have very few remnants of glacial material either on hilltops or in hollows. Also the thickness and continuity of the kame deposits along the Bloomsburg margin is markedly greater than anywhere to the southwest of that margin.

Indirect supporting evidence for the late Illinoian age of the Bloomsburg margin is regional and global correlations of an ice advance of that age that is slightly more extensive than the late Wisconsinan. The regional correlation is that in western Pennsylvania and eastern Ohio a narrow late Illinoian fringe lies just outside the late Wisconsinan limit (Fullerton, 1986) at a distance similar to that of the Bloomsburg margin (Fig. 2). The global correlation is based on the oxygen isotope record (Fig. 1). Currently the Illinoian is subdivided into the late Illinoian, with a glacial maximum at about 150 Ka, and the early Illinoian, with a glacial maximum between 250-270 Ka (Richmond and Fullerton, 1986; Martinson and others, 1987) (Table 1). These ages are from correlation of United States glacial advances with the oxygen isotope stages and are not from actual dating of glacial deposits though a growing number of dated glacial deposits do fall near the appropriate glacial maximums (Richmond and Fullerton, 1986). Calculations of relative ice volume from the isotope data suggest that late Illinoian ice volume was 5% greater than the late Wisconsinan and that the early Illinoian ice volume was 20% less than the late Wisconsinan (Shackleton, 1987).

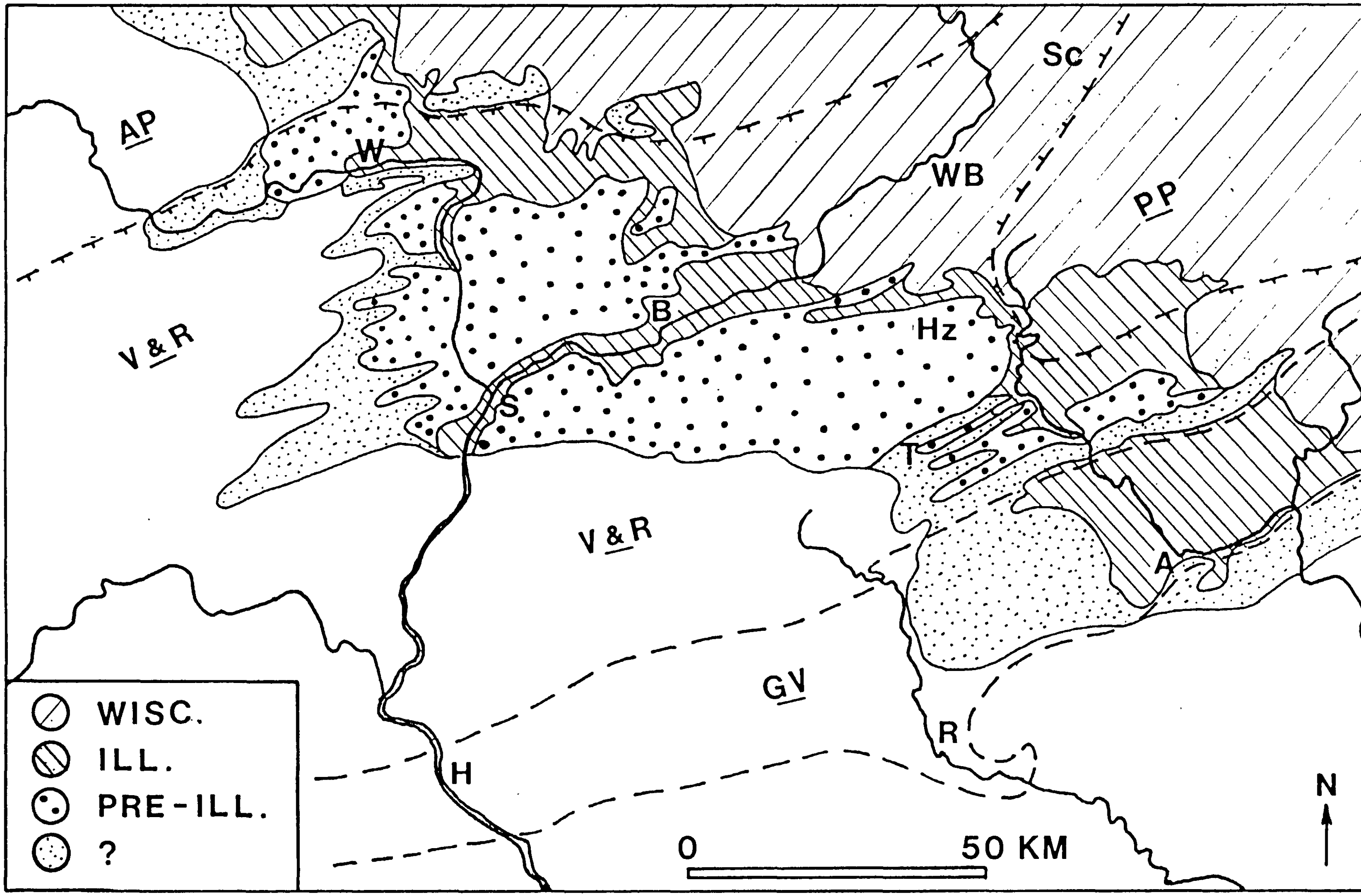


Figure 8. Leverett's (1934) map of glacial margins in eastern Pennsylvania. The lobation of all older margins is markedly dissimilar from that of the Wisconsinan margin. AP, Appalachian Plateau; GV, Great Valley; PP, Pocono Plateau; V&R, Valley and Ridge; A, Allentown; B, Bloomsburg; H, Harrisburg; Hz, Hazleton; S, Sunbury; Sc, ScrantonT, Tamaqua; W, Williamsport; WB, Wilkes Barre.

Of the two Illinoian advances, only the 150 Ka late Illinoian would have been of sufficient volume to extend a relatively short distance beyond the late Wisconsinan border. Some work in the mid-western United States though suggests that there, early Illinoian ice was more extensive than late Illinoian ice (Hallberg, 1986; Richmond and Fullerton, 1986) (Table 1). Yet another option is that the Bloomsburg margin is middle Pleistocene stage 12 (Fig. 1) or pre-Illinoian-B, centered around 450 Ka (Table 1). The projected ice volume for stages 12 is 15% greater than late Wisconsinan and 10% greater than late Illinoian (Shackleton, 1987).

The great depth of weathering of the Bloomsburg Margin would tend to suggest an older than late Illinoian age. Marchand thought that the weathering in the sand and gravel at Bloomsburg was equivalent to either his pre-Illinoian Penny Hill or early Illinoian Laurelton drift (Marchand and others, 1978). The exceptional depth of weathering though may be related to the unique character of the deposit rather than an exceptional age of weathering (Braun, 1988). The kames are exceptionally well drained sand and gravel deposits, often occupying hilltop or drainage divide landscape positions, and especially at Bloomsburg, contain an abundance of readily weatherable clasts from the strike valley. This would tend to produce deeper weathering more rapidly than on nearby till areas that are less well drained.

At present , the weight of "circumstantial " evidence is considered to favor a late Illinoian age rather than an early Illinoian or pre-Illinoian B age for the Bloomsburg ice margin (Braun, 1988, 1989b). As of yet there are no absolute age dates on the Bloomsburg margin material. Hopefully in the future some dating technique, such as cosmogenic isotopes, will provide absolute age constraints on the relative ages estimated from the degree of weathering and erosion.

Pre-Illinoian glaciation

In the 1880's several workers identified glacial material southwest of the Wisconsinan margin (extramorainic drift) but Williams (1895,1917) was the first to map an ice margin across the region. Williams (1895,1917) used the trend of the Wisconsinan limit as a model for the trend of what he thought was just the earliest advance of the Wisconsinan ice even though he called it the Kansan limit (Williams, 1894, 1898). Leverett (1934) revised Williams work, recognizing older, more weathered and eroded glacial material southwest of the Illinoian border (Fig. 8). He thought the scanty deposits of older drift were from a single pre-Illinoian glaciation. Since there was no superposition of two older drifts as there was in the Central Plains, he could find no way to determine whether the pre-Illinoian drift was Kansan or Nebraskan. He noted that the older drift had been named Jerseyan in New Jersey and used that name as an alternate term for the pre-Illinoian drift in eastern Pennsylvania. Leverett also recognized "areas of questionable location" of pre-Illinoian deposits farther to the south and west of his pre-Illinoian boundary, essentially areas that he could not verify glacial materials within Williams "Kansan" limit (Fig. 8).

Sevon and others (1975) observed no evidence for pre-Illinoian glaciation in northeastern Pennsylvania, but also did not re-examine Leverett's pre-Illinoian sites. They thought that strip mining had either destroyed possible pre-Illinoian deposits in the anthracite region or made them difficult to interpret and did not examine that area. In the central Susquehanna lowlands, the Penny Hill drift, first described by Peltier (1949), was thought to be pre-Illinoian in age by Marchand (Marchand and others, 1978). The Penny Hill drift has been correlated with the 330 Ka isotope stage 10 (Harden and Taylor, 1983), placing it in the pre-Illinoian A of Richmond and Fullerton (1986). As noted before in discussing the isotope record, stage 12, 16 and 22 indicate maximum pre-Illinoian ice extent and are better candidates than stage 10 as an age for these highly weathered deposits.

The surficial geology map of part of the central Susquehanna lowlands (Fig. 9) has the same problematic map patterns as discussed before with regard to the supposedly early Wisconsinan deposits (Marchand and Crowl, 1991; from field work primarily done in the 1970's). There is a "patch work quilt" pattern of supposedly younger glacial materials (early Wisconsinan till, dt and late Illinoian White Deer till, wt) surrounded by older material (early Illinoian Laurelton drift, ld). The two small early Wisconsinan till patches are on the Bloomsburg redbeds while the White Deer till is noted as having a limestone component, unlike that of other pre-Wisconsinan drifts. This suggests that again there are problems separating parent material effects from time effects in these weathering profiles. Also, the younger material is far less continuous than the older material, the opposite of what would be expected. The undifferentiated pre-Wisconsinan deposit map unit covers the most area and indicates that differentiating the units by weathering profile characteristics is problematic more often than not. Then, as noted before regarding past mapping of older then late Wisconsinan glacial boundaries, exceptionally long and low to reverse gradient pre-Wisconsinan ice lobes are mapped down strike valleys (Fig. 10). The boundary actually represents a line drawn at the base of the ridges separating where the glacial deposits have been completely eroded away from where there are remnant glacial deposits.

This surficial geology map shows that depending primarily on the degree of weathering profile development leads to glacial deposit and glacial boundary map patterns that are totally unlike that of the late Wisconsinan and in some regards are irrational. Probably what these map patterns are showing is the complex interplay of different depths of truncation of weathering profiles on different parent materials on different landscape positions. It is highly unlikely that there are any non-truncated weathering profiles on the oldest drift, especially the Penny Hill drift (Stop 11). For instance, how does one reliably differentiate between a deeply truncated Penny Hill soil and a less deeply truncated Laurelton soil ? What this work may best demonstrate is that in the eroding moderate relief Valley And Ridge landscape, one may accurately differentiate the remnant amount of weathering profile development but that differentiation does not yield a reasonable distribution or age for pre-Wisconsinan-aged glaciers. These various drift deposits can be explained more simply and rationally as a complex of different degrees of truncation of a pre-Illinoian weathering profile coupled with different lithofacies of pre-Illinoian drift.

Ongoing work is retracing the pre-Illinoian-aged maximum Laurentide ice limit across eastern Pennsylvania (Fig. 2 & 3)(Braun, 1988, 1989b). The margin is considered to be pre-Illinoian in age because post-glacial erosion has almost totally erased surface evidence of the glaciation. In the anthracite region, glacial deposits are only exposed in strike-valley floor strip mines buried under as much as 65 feet (20 m) of colluvium. At three strip mine sites, well-developed striated pavement shows that ice flow was S 10^0 W to S 40^0 W, the same as for the late Wisconsinan. The glacial materials contain abundant Mauch Chunk redbed erratics and are thus readily separable from other non-glacial surficial deposits (Stop 2, 4, & 5). Colluvium in the anthracite synclines is dominated by Pottsville conglomerate fragments from the adjacent strike-ridges and sandstone and shale fragments from the Llewellyn Formation (Stop 5). Local proglacial lakes were impounded within the pre-Illinoian margin in strike-valleys within the anthracite synclines. In those areas till, sand and gravel, and clay deposits as thick as 30 m overlie the coal seams (Stop 5). Intervening strike-ridges show extensive areas of weathered bedrock tors and no evidence of glaciation (Stop 6).

Both northwest and southeast of the anthracite region are broad strike-valleys with "rolling hill" uplands in the central Susquehanna , the Little Schuylkill, and the Lehigh basins. In those areas, fragmentary remnants of pre-Illinoian deposits are preserved primarily on broad interfluves (Stop 7, 8, & 11) , heads of first order hollows, or along strike ridge toeslopes where younger materials eroded from higher on the ridge bury the older glacial material (Stop 5). At these sites, kame sand and gravel as thick as 65 feet (20 m) are most frequently observed and occasionally in-situ till and varves are exposed. In first order stream hollows much of the pre-Illinoian diamict is in transport downslope from its original depositional position, material that some have called "colluviated till". The most common evidence for pre-Illinoian glaciation in the gently rolling strike valleys are lags of occasional erratic cobbles and boulders imbedded in much younger residual or colluvial material (Stop 7).

Outside of the late Wisconsinan limit, from the northern part of the Great Valley to the West Branch Susquehanna River, all sites mentioned by Williams (1917) and Leverett (1934) have been re-examined. Study of Williams' work showed that the effect of periglacial activity in eroding and transporting material down slope was unknown. Re-examination of many of his sites showed that what he had called "Pottsville conglomerate drift" was actually late Wisconsinan colluvium.

Figure 9. Surficial geology map of part of central Pennsylvania (Marchand and Crowl, 1991) showing a "patch work quilt" pattern of younger deposits (dt, wt) surrounded by more extensive older deposits (ld, ud). The widest spread deposit (ud) is undifferentiated with respect to age or material.

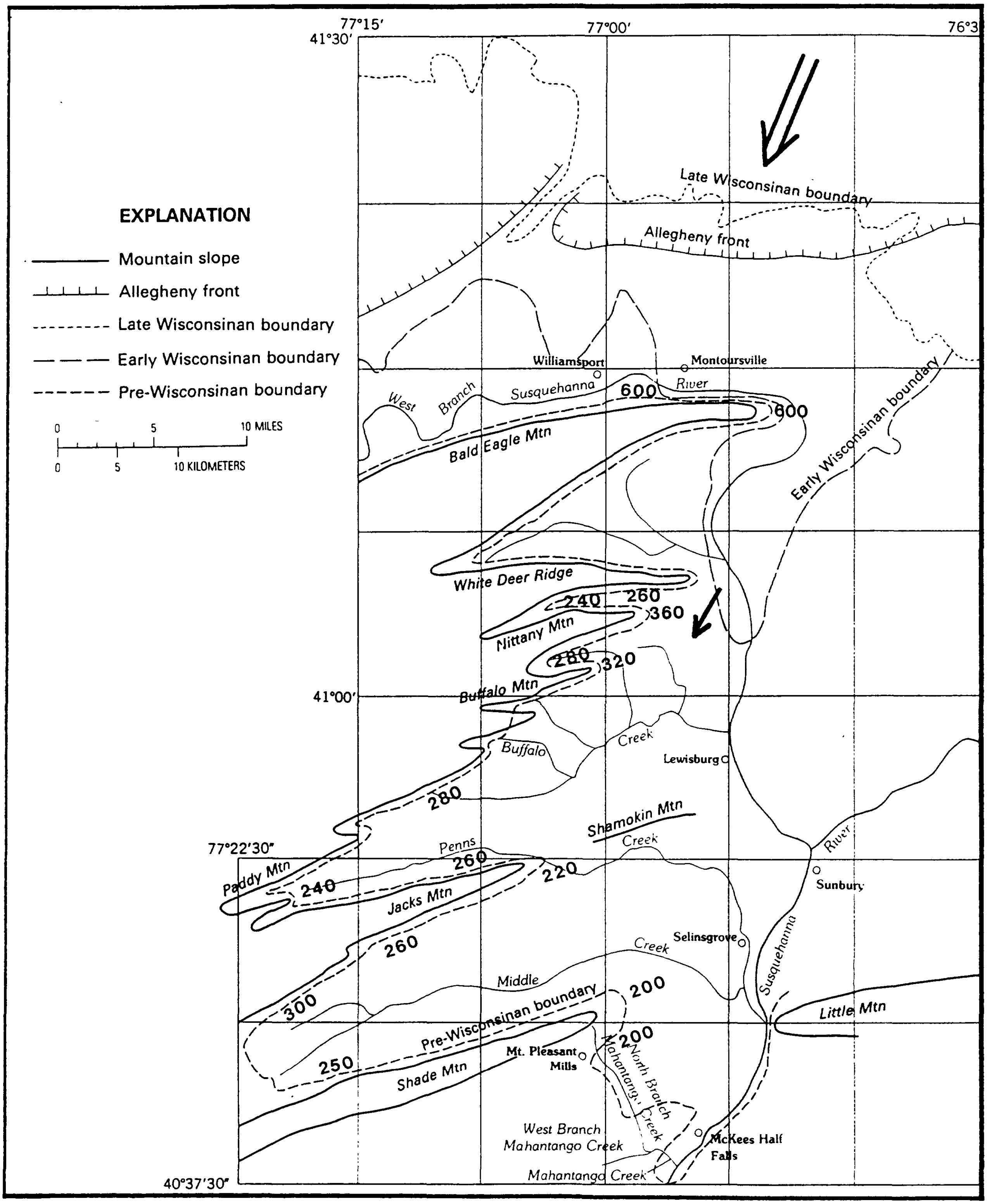


Figure 10. Pre-Wisconsinan glacial boundary with extremely low to reverse ice surface gradients. The numbers are the elevation, in feet, at various sites along of the boundary. Single arrow is hilltop striation at New Columbus, Pennsylvania (Braun, unpublished data at same site as Stop 4, Marchand and others, 1978). Double sided arrow is regional ice flow direction for the late Wisconsinan glacier.

Many sites where heobserved anthracite beds being overturned and moved southward are actually classic examples of creep. The freshness of the coal under the "Pottsville drift" was the primary reason he thought the glaciation of the area was similar in age to the Wisconsinan glaciation to the north. The truncation of the coal was Wisconsinan in age but from periglacial rather than glacial activity. While Leverett (1934) better recognized the influence of slope processes, some of his sites were also colluvium where a mixture of both Pottsville and Llewellyn fragments gives the colluvium a till-like diamict texture.

Pre-Illinoian terrace remnants have been identified along the main-stem Susquehanna River (Leverett, 1934; Peltier, 1949; Pazzaglia and Gardner, 1992). The highest glacial gravels containing crystalline erratics from outside the basin occur at Paxton at 125 feet (38 m) above the river (Peltier, 1949) and at Dauphin at 195 feet (59 m) above the river (Leverett, 1934). The Dauphin site is a gravel lag on a bedrock bench. This implies that the Susquehanna River has incised into bedrock at least 195 feet (59 m) since the first pre-Illinoian glacial advance entered the Susquehanna River basin. These very fragmentary, high level outwash remnants may represent a glacial advance older than at the presently identified maximum pre-Illinoian glacial limit.

The age of the pre-Illinoian limit and whether or not multiple advances are present within that limit is an open debate at present. As noted previously, the isotope record suggests that four times in the middle to early Pleistocene ice was extensive enough to reach beyond the late Wisconsinan limit (Fig. 1). Middle Pleistocene stage 12 or pre-Illinoian-B, centered around 450 Ka and stage 16 or pre-Illinoian-D, at about 650 Ka (Table 1), should have reached the farthest (Braun, 1988, 1989c). The projected ice volumes for stages 12 and 16 are 15% greater than late Wisconsinan and 10% greater than late Illinoian (Shackleton, 1987). That would place the pre-Illinoian margin about twice as far from the late Illinoian margin as the late Illinoian is from the late Wisconsinan and that is what is observed (Fig. 3). Also these pre-Illinoian B and D glaciations are relatively young pre-Illinoian advances and would be more likely to be preserved in the eroding moderate relief Valley and Ridge than yet older advances (Braun, 1988, 1989b).

Recent paleomagnetic information (Gardner and others, in prep.; Sasowsky, this guidebook) suggests that the pre-Illinoian limit is more likely to be early Pleistocene or even Pliocene in age. Samples of pre-Illinoian-aged varves from the West Branch Susquehanna valley at Antes Fort and clay drapes from the Eastern Middle Anthracite field at Jeansville have a strong reversed polarity magnetism. Both sites are 13 to 15 kilometers inside the pre-Illinoian limit (as it is presently mapped). The reversed polarity suggests an early Pleistocene age, older than 788 Ka (Table 1). The most extensive early Pleistocene event is Stage 22 (Fig. 1) or pre-Illinoian-G at about 850 Ka (Table 1). A Pliocene event either at 2.1 Ma or 2.3-2.4 Ma is a less likely choice considering the preservation of such old material on the eroding Appalachian landscape (Braun, 1989b). It remains possible that the reversed polarity material was deposited during the Emperor reversed polarity subchron at about 450 Ka. That would permit the material to be from the pre-Illinoian-B event, in agreement with the isotope record and requiring a shorter period of preservation. But in the mid-western United States, material correlated with pre-Illinoian-B has a normal polarity though the actual age of the material is not very well constrained (Table 1). Given the present evidence, the most likely age for the maximum glacial limit in eastern Pennsylvania is 850 Ka, the early Pleistocene pre-Illinoian-G event. A second less likely age would be 450 Ka, the middle Pleistocene pre-Illinoian-B event.

Material with a strong normal polarity magnetism is located 2.5 kilometers east of the reversed site in the Eastern Middle Anthracite field near Beaver Meadows. A site with mostly normal polarity samples mixed with a few reversed samples is located a few kilometers from the pre-Illinoian limit near Tamaqua and is not considered reliable without additional sampling. The Beaver Meadows site is less than 788 Ka and may be either pre-Illinoian B, D, or some other event if the isotope record does not accurately reflect ice volume. Hopefully in the future at least a normal versus reversed polarity material "age" boundary can be drawn through this region.

Conclusions

The late Wisconsinan ice margin should be used as a model for how the Laurentide glacier, each time it advances, interacts with the moderate relief landscape of the Valley and Ridge of eastern Pennsylvania. Previous work had consistantly produced pre-late Wisconsinan ice margin and deposit patterns that were markedly dissimilar from the late Wisconsinan patterns and that were sometimes physically impossible. On going work has developed a simple pattern of progressively older glacial limits subparallel to and incrementally to the southwest of the late Wisconsinan border. The first glacial margin outside the late Wisconsinan limit, the late Illinoian Bloomsburg margin, is marked by thick head-of-outwash deposits that are now deeply weathered and eroded. The depth of weathering suggests the possibility that the Bloomsburg margin could be pre-Illinoian B in age. Farther to the southwest exist fragmentary remnants of reworked or deeply truncated deposits of pre-Illinoian age. Those deposits near the limit of evidence for glaciation have a reversed polarity magnetism, suggesting an older than 788 Ka maximal advance of the Laurentide glacier during the pre-Illinoian G event at about 850 Ka.

PALEOMAGNETISM OF GLACIAL SEDIMENTS FROM THREE LOCATIONS IN EASTERN PENNSYLVANIA

by: Ira D. Sasowsky, Nittany Geoscience, Inc.
120 Radnor Road, State College, PA 16801
(814) 231-2170

Paleomagnetic samples were collected from glacial sediment exposures at the Beaver Meadow (Fig. 34), Jeansville (Fig. 34), and Tamaqua (Fig. 45) locales. A total of 16 samples were collected. Pairs of samples were taken from each of three intervals at the Beaver Meadow locale and each of two intervals from both the Jeansville and Tamaqua locales.

Collection methods used were those suggested by Schmidt (Undated), and involved clearing a fresh face in the outcrop, and then inserting the plastic sampling cube into the face of the material. If the material was well-indurated, the face was first prepared by cutting away material until the cube could be pressed on. Once in place, the sample number was affixed to the cube, the orientation was measured, and the sample was excavated and capped. If the bedding in the outcrop was significantly different than horizontal, bedding strike and dip were measured to allow for correction of the sample orientation.

Samples were analyzed at the University of Pittsburgh Paleomagnetism Laboratory. Measurements were made using a cryogenic magnetometer in a low-field (shielded) room. The Natural Remanent Magnetism of each sample was first measured. The samples were then subjected to increasingly stronger alternating-field (AF) demagnetization, followed by remeasurement of their magnetization. Peak intensity steps used for the demagnetization were typically 0, 10, 20, 30, 50, 80, and 120 millitesla (mT). AF demagnetization was used to assure that a true paleomagnetic direction was being measured, as opposed to a more recently acquired viscous magnetic component.

Demagnetization paths were plotted on vector end-point (Zijderveld) diagrams. A wide variety of behaviors were observed, but the majority of samples clearly indicated either a normal or reverse polarity. Typical Zijderveld plots are shown in Figure 11. Some of the samples had non-linear demagnetization paths, indicating the preservation of more than one magnetic signal. Non-convergence on the origin for several of the samples shows that a high-coercivity component remained even after application of the 120 mT peak field.

Paleomagnetic field direction (vectors) were extracted from the demagnetization data using the methods of Kirschvink (1980), and are presented on Figure 12. Also illustrated on this Figure are the present day-field direction, and the geocentric, axially-centered dipole directions (normal and reversed polarity) for eastern Pennsylvania. Tamaqua samples T4, T5, and T6 give directions which are unclear as to their polarity. Jeansville samples JE1, JE2, JW1, and JW2 yield field directions which indicate reversed polarity, but with substantial shallowing of the inclination. This shallowing suggests that the recorded magnetization is a depositional remanent magnetism (DRM) which was acquired when the sediments were deposited. The remainder of the samples cluster near the present-day field direction and the normal, axially-centered dipole, indicating that they were deposited during a period of normal polarity. Inclination shallowing was not seen for the normal polarity samples. Fisher (1953) statistics were used to generate mean site directions for the samples, but the relatively small number of samples taken precluded any accurate determination, and such results are not presented here.

The reversed polarity sediments present at the Jeansville site indicate that these sediments were deposited at least 788,000 years ago, the age of the transition between the Matayuma reversed-polarity chron and the Brunhes normal-polarity chron (Spell and McDougall, 1992). The presence of these reversed sediments provides a useful constraint on the age of glaciations in this part of Pennsylvania. Gardner and others (1994) report on similar age constraints on glacial deposits further to the West. The normal polarity results from the other sites in the present study do not provide a minimum age for deposition of the sediments, but can help to separate deposits from different glaciations.

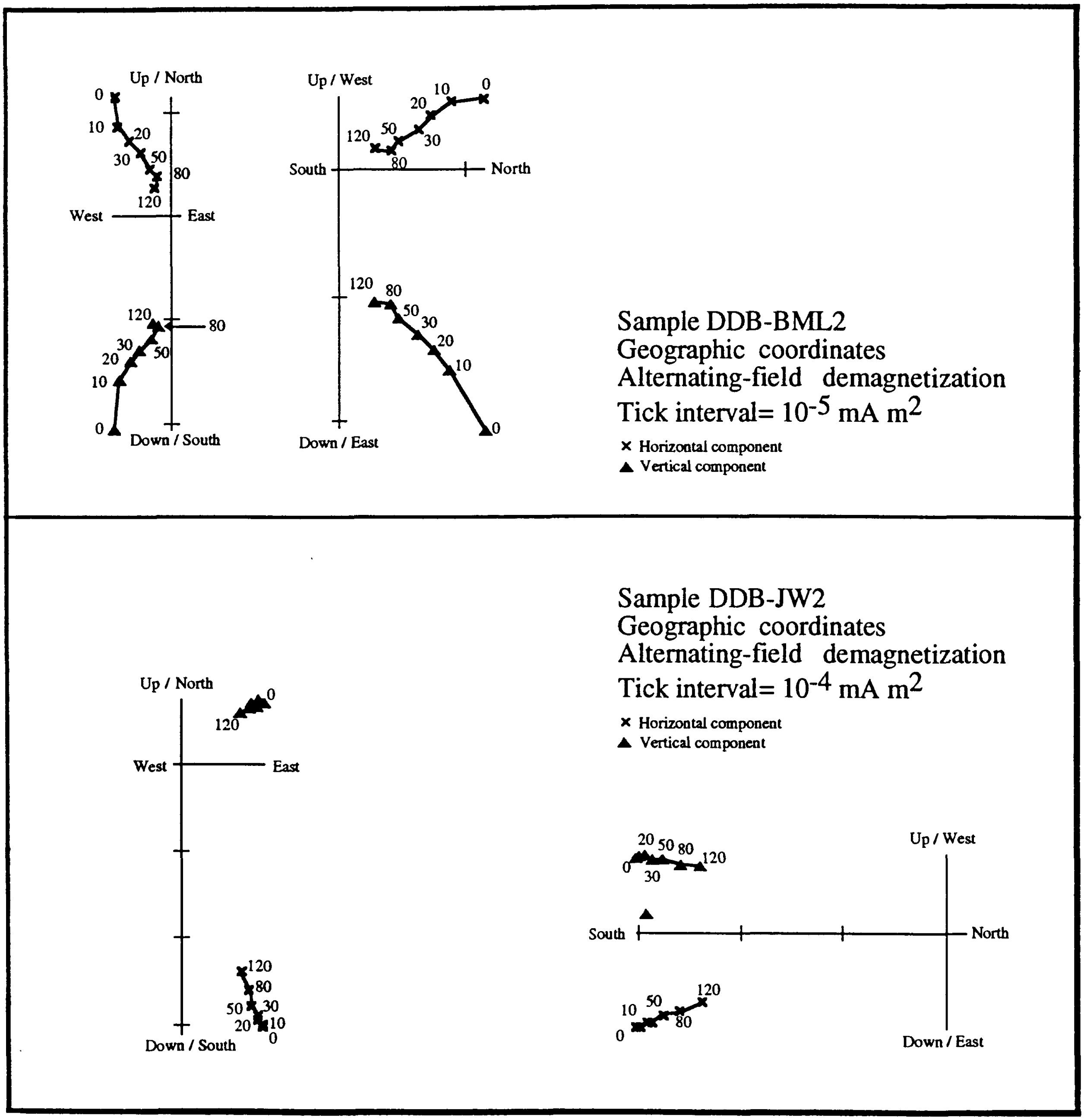


Figure 11. Representative Zijderveld-plot pairs for samples from glacial deposits, eastern Pennsylvania. Sample DDB-BML2 is from the Beaver Meadow location, and exhibits normal polarity, as did the other samples from that location. Sample DDB-JW2 is from the Jeansville (West) site, and exhibits reversed polarity, as did the other samples from that site. Numbers along demagnetization paths are peak alternating-field strengths in millitesla.

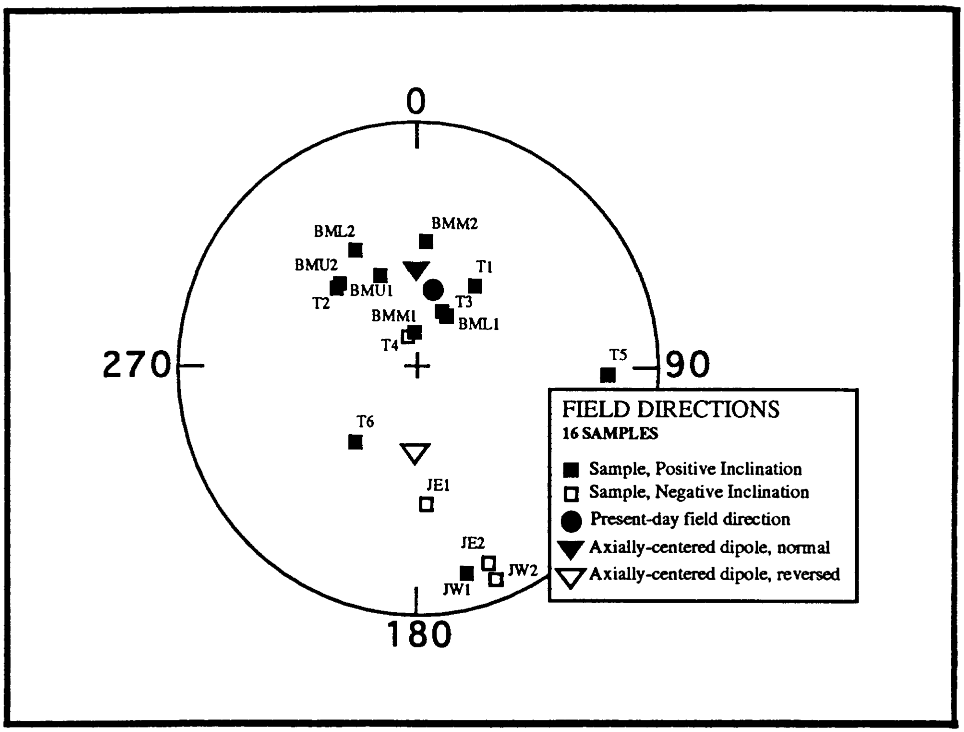


Figure 12. Polar Lambert equal-area plot of paleomagnetic field directions for 16 samples from eastern Pennsylvania. BM, Beaver Meadow site, located at N on figure 34 ; JE, Jeansville site, located at R on figure 34 ; T, Tamaqua site, located at N ? on figure 45 .

SOIL DEVELOPMENT IN TILL IN NORTHEASTERN PENNSYLVANIA

by: Edward J. Ciolkosz and Nelson C. Thurman
Agronomy Department, Penn State University

Introduction

Soil development in glacial till in Northeastern Pennsylvania is an interesting topic. Of particular interest is the relation ship of soil development to the age of the till material. The youngest till is of late Wisconsinan age (Woodfordian - 18,000 yrs BP). The age of the older tills is not known with any great certainty. Regardless of this uncertainty there is a distinctive difference in the degree of soil development between the younger (Woodfordian) and the older till soils. Previous studies on this topic are presented in Marchand and others (1978), Levine and Ciolkosz (1983), and Ciolkosz and others (1985). The following presentation is a brief update of those publications.

Late Wisconsinan Till Soils

Based on color, within the late Wisconsinan (Woodfordian) materials there are two major kinds of till. These are brown-tills (10YR 4/4) and red-tills (2.5YR 4/4). These tills derive their color from the bedrock that was eroded to form the till. The red material comes primarily from the Catskill formation and the brown from other Paleozoic rocks. Regardless of color the soils in these two tills generally show the same degree of pedological development. They have ochric epipedons (A horizon), cambic (Bw), and fragipan (Bx) subsurface horizons and are classified as Fragiochrepts or Fragiaquepts depending on their internal drainage (Ciolkosz and others, 1989). Thus in the last 18,000 years there has not been a significant parent material impact on the pathway of genesis of the soils developed in these two different color tills. Although not major, there appears to be a slight effect above the fragipan in the cambic horizon of the brown-till soils. This horizon tends to show a slightly brighter color (higher chroma) than the fragipan whereas in the red-till soil there generally is no difference in color between the cambic horizon and the underlying fragipan. There also is no major difference in either the iron oxide content of the red and brown till soils (Ciolkosz and others, 1993; MacFie, 1991) or their total iron content (Ciolkosz and others, 1993). Generally these soils have about 1% (as Fe) iron oxides and 3 to 5% (as Fe) total iron. In addition the clay content is also similar between these soils (10 to 20%) (Ciolkosz and Thurman, 1994).

Soil Color

The color of well-drained soil subsurface horizons is determined by the iron oxide mineral type (Ciolkosz and Dobos, 1990; Schwertmann, 1993). Red soils are dominated by hematite (Fe_2O_3) and brown soils by goethite (FeOOH). Once formed, both hematite and goethite are stable in soil oxidizing environments (well drained soils) and , according to Schwertmann (1993), there is no indication of a solid-state transformation of goethite to hematite by simple dehydration in soils. Although this appears to be the case, brown soil samples can be turned red rapidly in the laboratory by hearing them in a muffle furnace for a few hours at 600° C (LaFleur, 1970; Steele and Ciolkosz, 1980). As judged by the DTA endothermic temperatures of 280 to 400° C (Schwertmann and Taylor, 1989), the temperature of dehydration of goethite is lower than 600° C, although still much higher than temperatures that exist in soils. This indicates that in the soil there is not enough energy to directly convert goethite to hematite.

According to Schwertmann (1993) the soil factors that promote the formation of hematite over goethite as Fe is released from primary minerals by weathering are (1) high temperatures, (2) low water activity, (3) neutral pH, (4) high Fe content in the parent rock and rapid Fe release via weathering, and (5) a rapid turnover of biomass (rapid oxidation of the organic material added to the soil). It appears that in a udic (humid), mesic (temperate) soil environment typical of Pennsylvania today that goethite is the most likely pedogenic iron oxide mineral being formed. Two exceptions may be the soils forming from ultrabasic rocks (e.g., diabase in southeastern PA) and limestones. Thus the only significant difference between the late Wisconsinan brown and red till soils is their iron oxide mineralogy which is primarily inherited from the till parent material.

Pedogenic reddening (rubification) produces high chroma (6 to 8) colors with hues of 5YR to 2.5YR and apparently requires warmer temperatures than exist in central Pennsylvania today. If we observe the oxygen-isotope record as presented by Denton and Hughes (1983) (Figure 13) and assume that it is a surrogate for mean annual temperature, there doesn't appear to be a great amount of temperature forcing of the pedogenic process between the Illinoian (132,000 yrs BP) and the Holocene (10,000 yrs BP) except in isotope stage 5e. This is somewhat perplexing because the

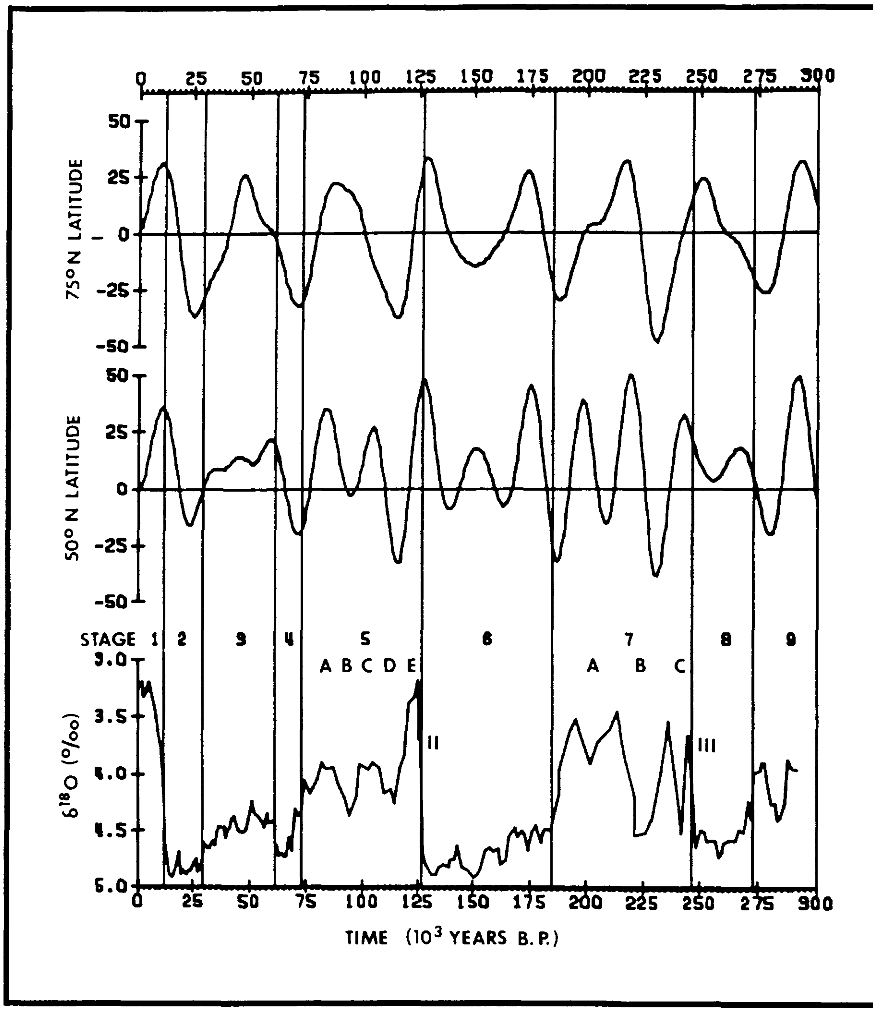


Figure 13. Comparison of Northern Hemisphere summer half-year insolation at 75°N and 50°N and the oxygen-isotope record. Reprinted from Denton and Hughes (1983) with permission of the publisher.

Sangamonian (132,000 - 75,000 BP) has always been projected as a time warmer than the present. The key to this perplexing problem may be the half year insolation (the amount of incoming solar radiation per area of the earth's surface). The data presented by Denton and Hughes (1983) (Figure 13) shows that for a significant period of time at mid latitudes (50°N) the summer half year insolation during the Sangamonian was greater than today. This would result in warmer summers in Pennsylvania, although the winters would be colder. The overall effect would be to accelerate pedogenesis during the Sangamonian because very little pedogenesis occurs today during the winter in mid latitudes, particularly with regards to oxidation and rubification. This type of evaluation of soils needs to be explored in greater detail in the future, and may provide some answers to the many questions we have about soil genesis in Pennsylvania.

Early Wisconsinan (?) Till Soils

Levine and Ciolkosz (1983), as a part of a chronosequence study of soil development in till published data on two soil pedons (Leck Kill) that were believed to be early Wisconsinan age (Altonian - 40,000 yrs BP). The Leck Kill pedons were developed in red till material, and differed from Woodfordian red till soils in that they had an eluvial (E) horizon, an argillic (Bt) horizon, slightly redder (higher chroma) colors in the B horizon, and no fragipan. These soils showed a distinctive increase in pedological development beyond the Woodfordian soils (see the data in Table 2 and Figure 14). The Leck Kill soils also indicate that the red till material can be rubified. This is logical because the Woodfordian age till soils have about one-third of their total iron in the oxide form (Table 2). Thus two-thirds of the iron is still in minerals that can be weathered and the iron can be released and form high chroma pedogenic hematite.

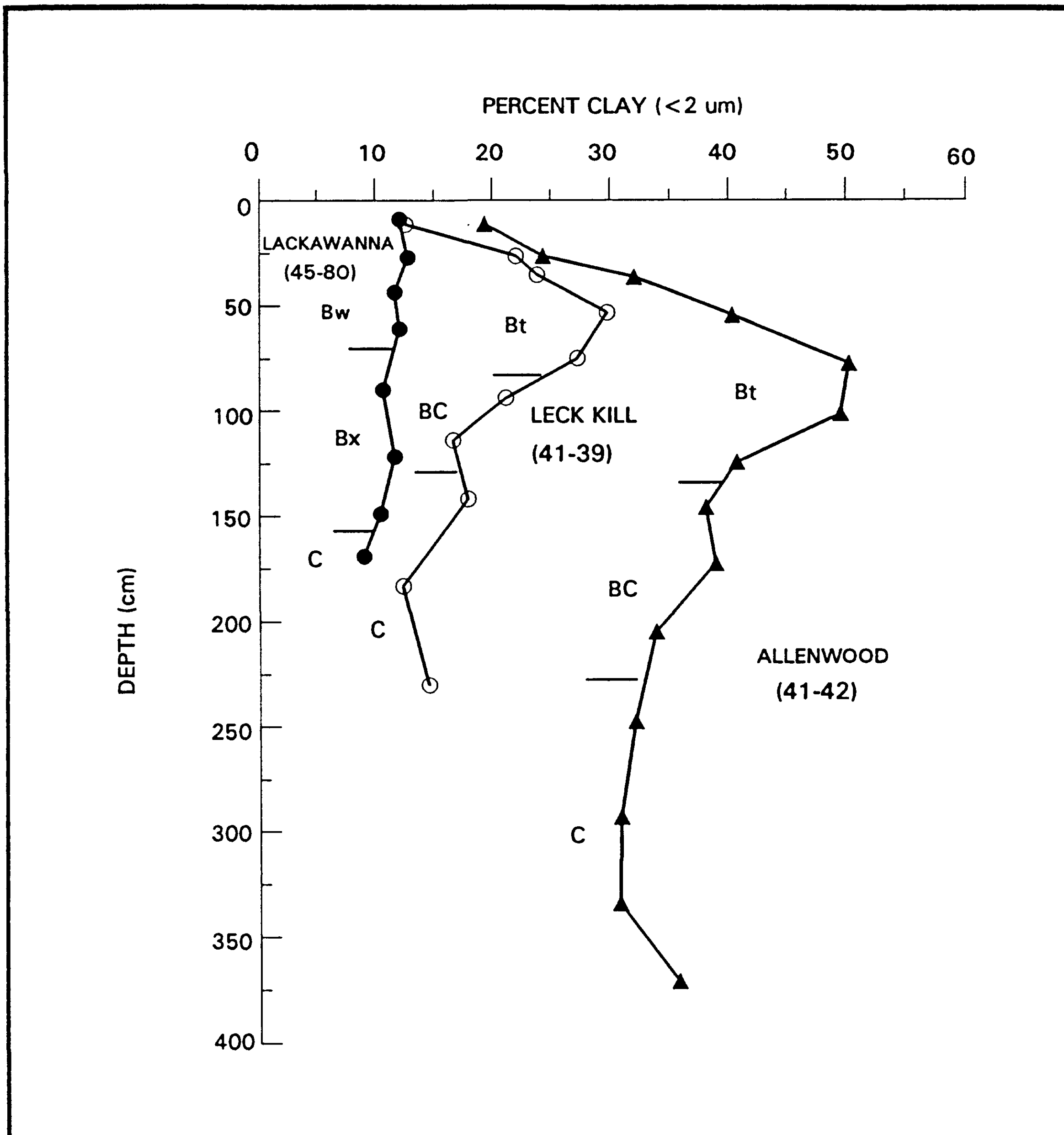


Figure 14. Clay content with depth of the Lackawanna (late Wisconsinan Age), Leck Kill (early Wisconsinan ? Age), and Allenwood (Pre-Wisconsinan Age) soils in Northeastern Pennsylvania. These soils have developed from similar glacial till parent materials (from Ciolkosz and others, 1994).

Table 2. Clay, color, iron oxide (as DBD-Fe) and total iron content (as Fe) in three glacial till soils from northeastern Pennsylvania. The clay content of these three soils is plotted with depth in Figure 15. Data taken from Ciolkosz and others (1993) and Levine and Ciolkosz (1983).

Soil Name and Drainage	Soil and Horizon Number	Depth in cm	Horizon	pH	Color	Percent Total Clay+	Percent Fine Clay+	Percent CBD Fe	Percent Total Fe	Ratios $\frac{Fe_d}{Fe_t}$	Ratios $\frac{(Fe_d-Fe_o)}{Fe_t}$*
					Late Wisconsinan Till Soil						
Lackawanna	45-80-01	0-18	Ap	5.3	7.5YR 5/4	12.1	1.9	0.98	2.96	33.1	29.4
Well	45-80-02	18-36	Bw1	5.5	5YR 4/3	12.8	2.2	0.77	3.35	23.0	20.6
drained	45-80-03	36-51	Bw2	5.7	5YR 4/3	11.7	1.9	0.77	3.25	23.7	20.3
	45-80-04	51-71	Bw3	5.8	5YR 4/3	12.1	1.9	0.70	3.20	21.9	18.7
	45-80-05	71-109	Bx1	5.6	2.5YR 4/4	10.7	1.7	0.91	3.36	27.1	25.0
	45-80-06	109-135	Bx2	5.3	2.5YR 4/4	11.7	1.9	0.84	3.63	23.1	19.0
	45-80-07	135-163	Bx3	5.1	2.5YR 4/4	10.5	1.9	0.91	3.36	27.1	24.4
	45-80-08	163-175	C	5.1	5YR 4/3	9.1	0.2	0.84	3.56	23.6	21.6
	45-80-09	175-201	R	.	10R 4/2	.	.	.	.	.	.
					Early Wisconsinan (?) Till Soil						
Leck Kill	41-39-01	0-23	Ap	6.1	10YR 3/3	12.6	2.0	1.12	2.48	45.1	37.1
Well	41-39-02	23-30	E	6.0	7.5YR 5/6	22.1	2.9	1.12	3.48	32.2	28.7
drained	41-39-03	30-41	BE	6.0	5YR 5/6	23.9	5.0	1.26	3.90	32.3	29.2
	41-39-04	41-66	Bt1	5.2	5YR 4/5	29.8	6.7	2.24	4.25	52.7	50.1
	41-39-05	66-84	Bt2	4.7	5YR 4/5	27.3	5.8	1.96	4.40	44.5	40.9
	41-39-06	84-104	BC1	4.7	5YR 4/4	21.2	4.8	1.19	3.87	30.7	27.9
	41-39-07	104-124	BC2	4.6	5YR 4/4	16.7	3.3	1.12	3.19	35.1	31.3
	41-39-08	124-160	C1	4.6	5YR 4/4	18.0	2.8	1.26	3.56	35.4	32.3
	41-39-09	160-206	C2	4.5	5YR 4/4	12.4	2.3	1.47	3.27	44.9	41.5
	41-39-10	206-254	C3	4.6	5YR 4/4	14.6	2.5	1.47	3.34	44.0	40.7
					Pre-Wisconsinan Till Soil						
Allenwood	41-42-01	0-23	Ap	5.4	10YR 4/3	19.4	8.3	1.75	2.98	58.7	52.0
Well	41-42-02	23-30	E	5.9	7.5YR 5/4	24.4	11.4	2.38	3.88	61.3	58.4
drained	41-42-03	30-43	BE	6.9	7.5YR 5/6	32.1	15.1	2.73	4.81	56.7	54.8
	41-42-04	43-66	Bt1	6.3	5YR 5/8	40.4	19.8	2.94	5.78	50.8	48.9
	41-42-05	66-89	Bt2	5.6	5YR 5/8	50.3	24.6	2.94	5.94	49.4	46.9
	41-42-06	89-114	Bt3	5.2	5YR 5/8	49.6	24.3	2.94	6.08	48.3	46.0
	41-42-07	114-135	Bt4	5.1	5YR 5/6	40.8	19.6	2.87	5.94	48.3	46.2
	41-42-08	135-157	BC1	5.1	5YR 5/4	38.1	18.3	2.80	5.72	48.9	46.5
	41-42-09	157-188	BC2	5.2	5YR 5/4	39.0	18.7	2.80	5.81	48.1	45.6
	41-42-10	188-221	BC3	5.2	5YR 5/4	33.9	15.9	2.66	5.18	51.3	48.0
	41-42-11	221-274	C1	5.1	7.5YR 5/4	32.2	15.1	2.24	5.16	43.4	39.7
	41-42-12	274-312	C2	5.1	7.5YR 5/4	31.0	14.6	2.73	5.50	49.6	45.8
	41-42-13	312-356	C3	5.2	7.5YR 5/4	30.9	14.3	2.45	5.49	44.6	39.5

*d = Citrate-Dithionite-Bicarbonate (CBD); t = total, o = oxalate

+Total clay = $< 2\ \mu m$, fine clay = $< 0.2\ \mu m$

Subsequent to the publication of Levine and Ciolkosz's paper, Eyles and Westgage (1987) proposed that no early Wisconsinan ice has crossed the Lake Ontario basin, thus an early Wisconsinan till in Pennsylvania was unlikely. Braun (1988) and Ridge and others (1990) have supported this position in recent publications. Thus the problem is: are the supporters of no early Wisconsinan till in Pennsylvania wrong, or are the soils studied by Levine and Ciolkosz pre-Wisconsinan in age. If these soils are pre-Wisconsinan, they most likely would be Sangamonian in age. When these soils are compared to Sangamonian soils of Illinois (Follmer and others, 1979) similarities are noted particularly in argillic horizon development. Long range soil development correlation from Illinois to Pennsylvania, particularly in different parent materials, is fraught with danger. Although this is the case, Follmer (1994) indicates that Sangamonian soils of Illinois show slightly better developed argillic horizons than Woodfordian age soils and that redness development in the B horizon is found but it is not common. Thus the Altonian age soils of Levine and Ciolkosz may be pre-Wisconsinan in age. If that is the case, the Allenwood soils that have been projected as pre-Wisconsinan age should logically be Pre-Illinoian.

This scenario creates some problems with regards to soil development in colluvium in the Ridge and Valley region. In this area, Ciolkosz and others (1990) have interpreted the brown colluvium found at the surface as Woodfordian age and the underlying red colluvium as a pre-Wisconsinan (Illinoian) buried paleosol that has been extensively rubified. Ciolkosz and others (1990) have not recognized any colluvial soil that would be equivalent in age to the Altonian soil of Levine and Ciolkosz (1983). Although this is the case, more extensive study in the redrock colluvial areas may show some buried Leck Kill type soils under Woodfordian red colluvial material. A soil characterization site in Schuylkill County may hold promise for this study. This soil (Meckesville, 54-8; Ciolkosz and others, 1979) shows a buried paleosol at a depth of 105 cm in which the Illite Weathering Products / Illite ratio decreases progressively with depth and indicates a significant amount of pedogenesis. Another problem with projecting the Altonian soil as Illinoian age is Braun (1988) indicates that a kame at the Lightstreet exit of I-80 as late Illinoian in age. The soil formed in the kame was studied by Ciolkosz and Crowl (1972) and it appears much older than the Altonian soil of Levine and Ciolkosz (1983) or a kame at Hughesville (about 20 miles northwest) that is about the same distance outside of the Wisconsinan border as the Lightstreet kame. The Lightstreet kame appears to show soil development more similar to a kame in Buffalo Valley (10 miles west of Lewisburg) which, according to the map of Marchand and Crowl (1991), is Laurelton which they indicate as early Illinoian in age.

Pre-Wisconsinan Till Soils

Marchand (1978) proposed three pre-Wisconsinan tills in the central Susquehanna River Valley (White Deer, Laurelton, and Penny Hill). Hardin and Taylor (1983) projected ages of 330,000 yrs BP (isotope stage 10) for the oldest (Penny Hill) and 260,000 yrs BP (isotope stage 8) for the next oldest (Laurelton). Hardin and Talyor (1983) also projected an age of 60,000 yrs BP for the Altonian soils studied by Levine and Ciolkosz (1983). Levine and Ciolkosz (1983) attempted to project the age of pre-Wisconsinan soils but were unsuccessful because their method required a comparison of the argillic horizon with unaltered till, and no unaltered till was available for the soils studied.

All the methods used to project pre-Wisconsinan age soils have limitations and to date none has provided irrefutable ages for these deposits. More intensive soils work is needed to sort out the influence of the red-rock (redbed) parent material on soil development. In particular the claim of Braun (1988) that redbed material resists pedogenesis, particularly rubification, and the counter claim by Ciolkosz and others (1984) that soils developed from redbeds show better development in the form of argillic horizons, soil depth and reduced rock fragment content than soils developed from brown or gray rocks.

Another interesting but anomalous feature of the pre-Wisconsinan till soils is their clay mineralogy. From visual observations of very deep oxidation (rubification) and extensive argillic horizon devlopment it would be expected that a mature mineralogical suite of low amounts of illite and high amounts of kaolinite would be present. This is not the case: what is observed is a deep (well into the lower B and C horizons) conversion of illite to expandable minerals such as vermiculite with very little accumulation of kaolinite. The illite to expandable type conversion is also common in Woodfordian till soils but it is confined to the upper parts (A and upper B horizons) of the soil. It is unknown why this anomaly exists and like many of the problems associated with the till soils of northeastern Pennsylvania, it needs further study.

LACK OF EVIDENCE IN THE SUSQUEHANNA VALLEY FOR HYPOTHESIZED LATE WISCONSINAN CATASTROPHIC DISCHARGES

by: Duane D. Braun, Geography & Earth Science,
Bloomsburg University, Bloomsburg, PA 17815

Introduction

Shaw (1989) has proposed that the Susquehanna valley be examined for evidence of catastrophic discharges on the order of 10^6 m^3/s. Such discharges are hypothesized to come from subglacial floods that formed drumlins and other features in New York State and Ontario (Shaw, 1989; Shaw and Gilbert, 1990). Drumlin-forming discharges are estimated using a minimum velocity of 2 m/s, drumlin field width, and mean drumlin height (technique of Shaw, 1989). The part of the New York State drumlin field oriented directly towards the Finger Lakes and the Susquehanna River basin is about 120 km in width (Figure 15). Average drumlin height is about 20 m (Mills, 1980; Muller, personal communication). These quantities yield a peak drumlin-forming discharge of 4.8 x 10^6 m^3/s. These catastrophic events are related to hypothesized glacial meltwater floods into the Gulf of Mexico (Emiliani and others, 1978) and the north Atlantic (Berger and others, 1977) starting at about 16 Ka and culminating in a series of three peaks between 14 and 12.5 Ka (Shaw, 1989; Fig. 5). At 16 Ka, the terminus of the Wisconsinan ice sheet was between the Pennsylvania - New York State boundary and the Valley Heads ice margin (Fig. 15) (Fullerton, 1986) . From 14 Ka to 12.5 Ka, the Wisconsinan terminus was withdrawing from the Valley Heads margin (Fullerton, 1986).

The Susquehanna valley is an ideal site to look for evidence to test the hypothesis of catastrophic subglacial floods for three reasons. First, where the Susquehanna River crosses the Appalachian Plateau, the valley has a relatively narrow cross-section with incised meanders that should display an abundance of bedrock scour features like those in the Channeled Scabland of the Columbia Plateau. Second, where the Susquehanna River crosses the Valley and Ridge Province, there is a series of narrow water gaps and broad lowlands that would be ideal sites for catastrophic flood deposits such as those on the Channeled Scabland. Third, unlike the Mississippi tributaries where large proglacial lakes cause problems in differentiating between catastrophic glacial lake outbursts and catastrophic subglacial outbursts (Kehew and Lord, 1990; Shaw, 1990), the Susquehanna basin contains no proglacial lakes of the size that could produce 10^6 m^3/s scale discharges.

The channeled scabland - a model for catastrophic flood flows

The subglacial floods would become surface floods south of the glacial margin and be like those on the Channeled Scabland. While the total Channeled Scabland discharge (21.3 x 10^3 m^3/s) (Baker, 1973) was 4.5 times the proposed subglacial flood, the discharge of individual coulees (2-8 x 10^3 m^3/s) is similar to the hypothesized subglacial derived flood. The North Branch Susquehanna valley is also of the same cross-section scale as Grand Coulee, 2-10 km wide and 200 m or so deep. In its 320-450 km passage across the Columbia Plateau, the Scablands flood discharge was reduced 57% (to 9.15 x 10^6 m^3/s) by channel anastomosis and 1000 km^3 of storage in broad shallow (100 m deep) basins (Baker, 1973). The North Branch Susquehanna valley, with a 225-290 km travel distance, several channels that become a single one in Pennsylvania, and only two small basin like lowlands, would be expected to have a smaller peak flow reduction than the Channeled Scabland flood. Due to lack of evidence of high water marks to calculate volumes in the Susquehanna system, an arbitrary reduction of the hypothesized Susquehanna flood of 25% (3.6 x 10^6 m^3/s) has been used for the Appalachian Plateau and (3.2 x 10^6 m^3/s) has been used for the Valley and Ridge. This reduction in peak discharge is probably too large and is used to provide a conservative estimate of the expected height of flood features in the North Branch Susquehanna valley.

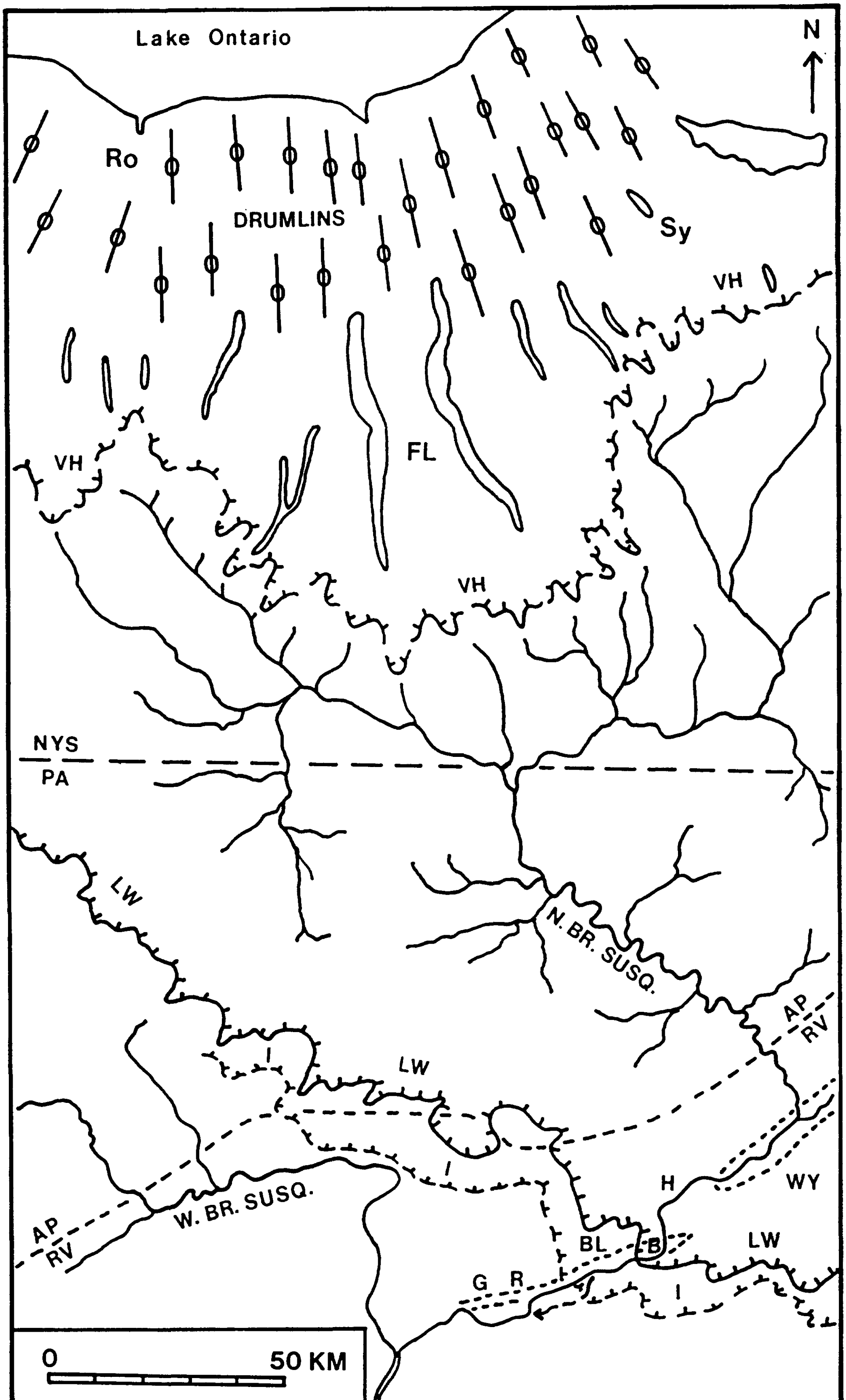


Figure 15. Location map showing relationship of the drumlin field to the North Branch Susquehanna River. FL, Finger lakes; VH, Valley Heads ice margin; LW, late Wisconsinan terminus; AP, Applachian Plateau; RV, Ridge and Valley; B, Berwick; BL, Bloomsburg; G, Grovania; R, Rupert water gap; WY, Wyoming Valley.

Since individual Scabland coulees had discharges in the range of the proposed Susquehanna flood, they could be used as hydraulic models of Susquehanna floods. For Scabland coulees without abrupt width constrictions or abrupt steeping, flow velocities were 10-20 m/s, water depths are 50-180 m and water surface slopes of 0.76 - 1.86 m/km were parallel to bed slopes (Baker, 1973, 1978a). At abrupt constrictions or steepenings, flow velocities approach 30 m/s, depths were 50 - 170 m, and water surface slopes were 1.14 - 18.29 m/km (Baker, 1973, 1978a).

The Rupert water gap - the most critical flow constriction

The smallest cross-section along the North Branch Susquehanna, less than one-half of any other, is the Rupert water gap. At a 10 m/s velocity, only a 0.74 x 10^6 m^3/s discharge would fit through the Rupert gap entrance. A higher velocity is unlikely in that the bed slope in the gap is only 0.55 m/km and the 10 m deep largest historic flood had only a 0.4 m/km water surface slope, values that are less than the lowest scabland slopes. The hypothesized drumlin-forming flood would spill about a 2.46 x 10^6 m^3/s discharge down an adjacent strike valley and across the hilltops surrounding the Rupert gap.

The Rupert water gap is the smallest one along the North Branch Susquehanna because the river has occupied that gap only since about 150 Ka, the late Illinoian glaciation (Braun 1985; 1988; 1989b). The Rupert gap had been originally cut by Fishing Creek, a tributary of the Susquehanna. Deposits at the late Illinoian terminus (Braun and others, 1984; 1988) diverted the North Branch Susquehanna down a nearby strike valley and into the Rupert gap (Fig. 23). Immediately to the west of where the Susquehanna makes a 90 degree turn to enter the Rupert gap, the strike valley continues westward to provide an alternate low level route for the river to return to its original course (Fig. 23). A low divide in the strike valley at Grovania is only 50 m above the present bed of the river in the Rupert gap and has a cross-section larger than the Rupert gap.

It is this unique combination of a extremely small constriction and an adjacent low level alternate flood route, that makes the Rupert-Grovania site the most critical test site for the catastrophic subglacial flood hypothesis. Any flood greater than 50 m deep and discharge greater than 0.32 x 10^6 m^3/s, would have spilled over the Grovania divide. The 3.2 x 10^6 m^3/s flood would have over-topped the Grovania divide by 100 m, completely filled the cross-section, and spilled across the adjacent hilltops. The steep slope of the strike valley floor to the west of the divide (9.14 x 10^{-3}), more than ten times that of the present river, should have had permitted spectacular scour features to have been cut in the valley. Instead of scour features, there are undisturbed pre-Wisconsinan glacial deposits and residuum exhibiting pre-Wisconsin-aged ultisol soil development and Wisconsinan periglacial deposits showing inceptisol development (Fig. 24). Neither Wisconsinan nor pre-Wisconsinan catastrophic floods of greater than 0.32 x 10^6 m^3/s discharge could have passed through the region without disturbing these deposits.

Catastrophic floods of less than 50 m depth and discharges of less than 0.32 x 10^6 m^3/s would have been ponded in the strike valley between the Grovania divide and the Rupert gap. This would have been an ideal site for eddy bar deposits and slack water deposits, but no such deposits have been observed. What is observed are undisturbed pre-Wisconsinan glacial deposits having ultisol development and Wisconsinan periglacial deposits and loess displaying inceptisol development (Fig. 24). Non-flood deposits extend down to Wisconsinan outwash terraces that lie 15 m above the bed of the river at the entrance to the Rupert water-gap. Coincidentally, this 15 m height is the elevation of the calculated probable maximum flood for the present river, a flood having a 2 m/s average velocity and a 1.32 x 10^4 m^3/s peak discharge (U.S. Engineer Corps, 1974). The largest historic peak discharge, a 10 m deep "350 year" event, is 1.05 x 10^4 m^3/s (U.S. Engineer Corps, 1974). These 10^4 scale discharges are the only ones that can be supported by the evidence at the Rupert-Grovania site. Such discharges are 100's of times smaller than the "drumlin-forming" discharges proposed by Shaw (1989).

Undisturbed late Wisconsinan landforms in the Susquehanna valley

A site where erosion of late Wisconsinan glacial landforms would have been expected is on the outside of Bell Bend where the river makes a 90 degree turn after it has entered the Berwick lowland (Fig. 15). This site is the late Wisconsinan terminus and is marked by a 50 m thick head-of-outwash complex (Inners, 1978; Crowl and Sevon, 1980)(Stop 1). It is capped by glacial diamict and ice-contact-stratified-drift that exhibits a well developed knob and kettle moraine topography (Braun and Inners, 1988). A 3.2 x 10^6 m^3/s scale flood should have covered the site by at least 10's m and should have reworked the area into a series of streamlined or giant current rippled surfaces like those in the Channeled Scabland (Bretz and others, 1956; Baker, 1973, 1978b).

On the north or inside of Bell Bend, an extensive near planar gravel-capped surface exists (Inners, 1978). This is just downstream of a water gap where the river enters the Berwick lowland (Fig. 15). It is a site where a catastrophic flood would have been expected to deposit a large scale, foreset bedded expansion bar covered with giant current ripples (Baker, 1973; 1978b). What is observed is normal outwash, a series of faintly expressed diamond shaped bars with near horizontal stratification.

Upstream of the Berwick lowland is the Wyoming Valley lowland (Fig. 15). Where the river enters the lowland through a water gap in a strike ridge, it undergoes the most abrupt expansion of any site along its course. There is an expansion bar, Scovell Island, of a scale appropriate to the present river at the site today. This would have been another ideal site for catastrophic floods to deposit large scale foreset bedded, expansion bars with giant current ripples. What is observed are normal outwash features with near planar surfaces and near planar stratification on the valley floor and till-mantled slopes rising above the valley floor (Hollowell, 1971).

Conclusions

The drumlin forming catastrophic subglacial flood hypothesis suggests that a 4.8 x 10^6 m^3/s peak discharge formed the New York State drumlin field (Shaw, 1989; Shaw and Gilbert, 1990). That flood can only discharge through the Finger Lakes - North Branch Susquehanna River system. The Channeled Scabland was used as a hydraulic model for the catastrophic flood to estimate peak flow reductions from channel storage and flow velocities. Resultant discharges yield flow depths on the order of 200 m that, particularly in the Valley and Ridge constrictions, overfill the river valley and spill across the surrounding hilltops.

The smallest valley cross-section, from a late Illinoian stream derangement of the river course, can only pass 0.74 x 10^6 m^3/s (23%) of the hypothesized flood flow. The remainder of the flood would have spilled down an adjacent strike valley with a depth of more than 100 m and overtopped the surrounding hilltops. Instead of flood features, undisturbed pre-Wisconsinan and Wisconsinan deposits mantle the valley floor. Outwash deposits exist up to 15 m above the present river bed. Floods of that height would have had discharges on the order of 1.32 x 10^4 m^3/s, flows 100's of times smaller than the hypothesized drumlin forming discharges.

Within the late Wisconsin margin, glacial landforms in the North Branch Susquehanna valley are undisturbed by catastrophic flooding. In the Berwick lowland where large scale expansion bars with giant current ripples would have been expected, undisturbed knob and kettle moraine and planar outwash surfaces are present. In the Wyoming Valley lowland there is also a lack of catastrophic flood deposits with only normal outwash surfaces on the valley floor and till-mantled valley side slopes being present.

The evidence in the North Branch Susquehanna valley negates the possibility of catastrophic floods having flowed out of the Finger Lakes region. This in turn negates the catastrophic subglacial flood origin for the New York State drumlin field. If this drumlin field, one of the best developed in North America, is not a result of catastrophic floods, then other such features should not be expected to be the result of such floods. The North Branch Susquehanna valley evidence can be considered a crucial test that falsifies the subglacial flood hypothesis for drumlin formation and calls into doubt the overall hypothesis of catastrophic subglacial floods.

THE UBIQUITOUS BOULDER COLLUVIUM OF EASTERN PENNSYLVANIA, A RELICT OF PERIGLACIAL MOBILIZATION OF THE ENTIRE LAND SURFACE

by: Duane D. Braun, Geography & Earth Science
Bloomsburg University, Bloomsburg PA 17815

The oxygen isotope record (Fig. 1) shows that during the late and middle Pleistocene, the last 800,000 years, there have been at least 10 interglacial to glacial climate oscillations. The landforms of eastern Pennsylvania have developed under the alternation of two climatic extremes, the present warm temperate and the cold glacial-periglacial. Under temperate conditions, fluvial and debris flow erosion processes tend to incise the drainage ways more than adjacent slopes. The intensity and frequency of temperate climate rain events that generate the most dramatic erosional effects increases southward in the Appalachians. Under periglacial conditions erosion processes such as gelifluction generally affect the entire slope and tend to fill headward drainage ways. The intensity of periglacial effects should decrease southward in the Applachians, opposite to that of the temperate rain event effects.

It should be noted that while not all cold episodes brought Laurentide ice directly into eastern Pennsylvania, each cold event should have brought periglacial climate conditions to Pennsylvania. Also, during the late and middle Pleistocene the average climate has been much colder than at present (Fig. 1), equivalent to the 12 Ka Stage 2 to 1 transition (Porter, 1989). On the average then, while Laurentide ice occupied much of northern Canada, the climate in the highlands of eastern Pennsylvania should have been near to or at periglacial cold conditions with tundra vegetation (Jacobson and others, 1987). The area near the glacial limit in Pennsylvania should show the maximum degree of periglacial landscape modification of any place in the Applachians. It is an area of moderate altitude that was the closest to the Laurentide glacier for the longest period of time without actually having been overriden by the ice. Paleoperiglacial features, both their expression and areal extent, should gradually decrease southward from here.

Colluvial materials in eastern Pennsylvania thought to be paleoperiglacial in origin were first identified by Louis Peltier (1949) and called congeliturbate (Fig. 16). At present, paleoperiglacial features have been identified in the Appalachian Highlands for a considerable distance south of the late Wisconsinan glacial limit (Clark and Ciolkosz, 1988; Gardner and others, 1991; Clark and others, 1992; Clark, this guidebook). Some think that most if not all of the ubiquitous colluvial deposits in the Appalachians are the result of periglacial activity (Pewe, 1983; Ciolkosz and others, 1986; Clark and Ciolkosz, 1988; Braun, 1989c, Jacobson and others, 1989; Clark and others, 1992; Ridge and others, 1992). The area covered by colluvial soils decreases southward in the Appalachians (Ciolkosz and others, 1986), the expectable pattern if such deposits are related to periglacial activity.

Such features are also observable within the late Wisconsinan limit in Pennsylvania, as first noted by Peltier (1949), and indicate significant movement of boulder material across the top of glacial deposits during the recession of the last Laurentide ice sheet (Fig. 17). In the mapping of the late Wisconsinan glacial deposits of sixty-four 7-1/2' quadrangles in north central Pennsylvania (Sevon and Braun, in prep.), boulder colluvium from bedrock ledges on the upper ridge slopes was often observed mantling glacial deposits on the toeslopes and extending to the valley floors. On gentle toeslopes the upper 1 to 3 meters of till typically shows a strong downslope directed fabric, so called "colluviated till". At a hanging delta site, boulder colluvium from upslope bedrock ledges has been transported 150 meters across the near horizontal delta top and down the delta foreset face (Gardner and others, 1993).

As noted above, the most severe periglacial effects would be expected near the glacial limit in Pennsylvania. At the plunging nose of the Southern anthracite coal field near Nesquehoning, Pennsylvania, pre-Illinoian till is overlain by as much as 65 feet (20 m) of boulder-mantled colluvium. The volume of colluvium at the site requires at least 29 feet (9 m) of ridge-top erosion since glaciation (Braun, 1989b). This is a minimum erosion depth that assumes a 100 percent sediment trap efficiency for the mostly sand- and pebble- sized colluvial material at the site.

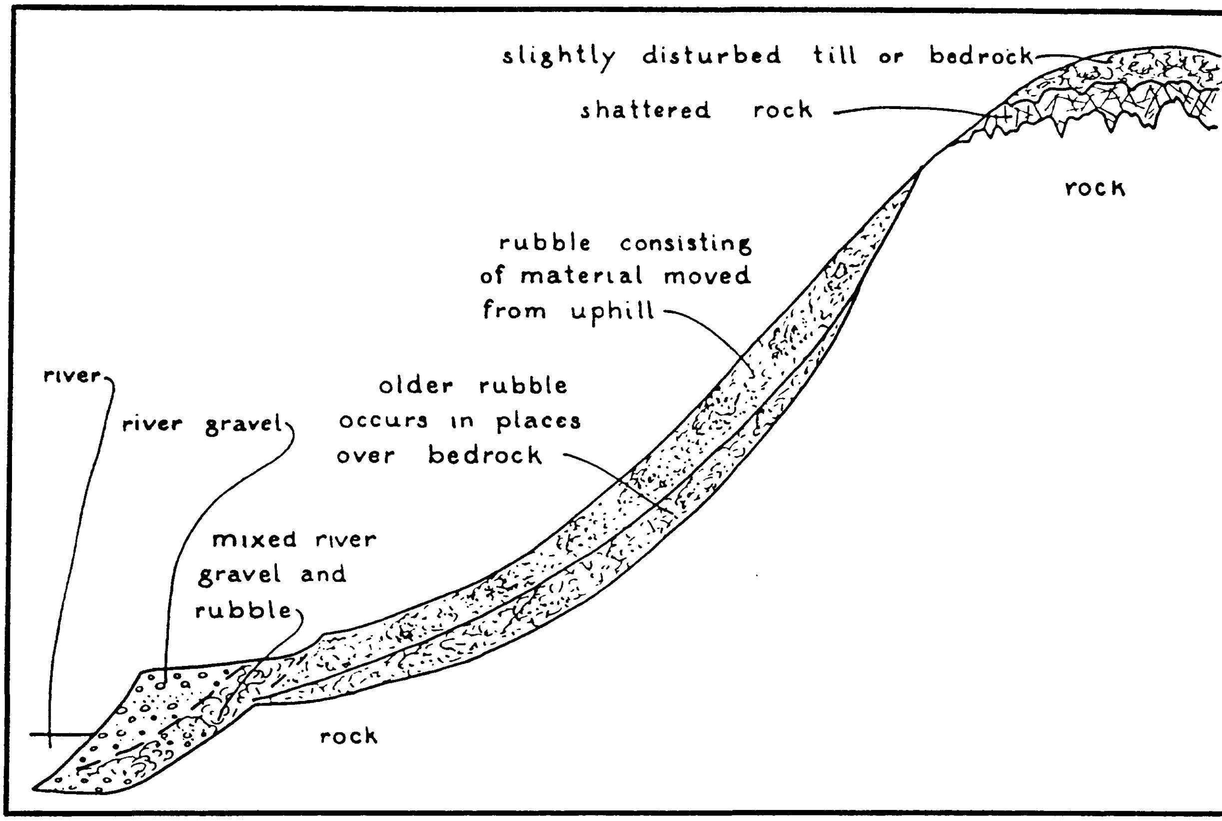


Figure 16. Schematic diagram of the distribution of colluvium (rubble) on a hillslope in eastern Pennsylvania. (From Peltier, 1949; Fig. 2)

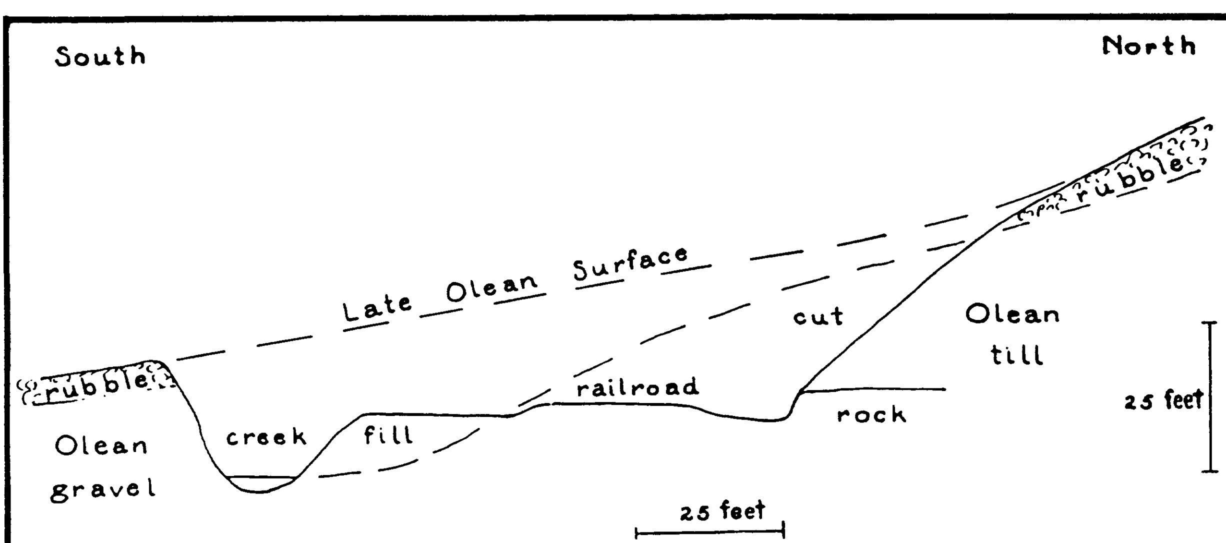


Figure 17. Cross section showing colluvium (rubble) overlying late Wisconsinan (Olean) gravel and till in northeastern Pennsylvania. (From Peltier, 1949, Fig. 17)

Depending on the actual age of the pre-Illinoian glacier that crossed the site, from 3 to 10 periglacial episodes have affected the site. The region between the late Wisconsinan limit and the pre-Illinoian limit is the ideal place to study the colluvium because there are some age constraints on the time of formation of the material in a place where the periglacial influence is at its maximum.

The surficial material that most needs explanation as to its genesis in eastern Pennsylvania is the sheet-like deposit of boulder-mantled diamict that uniformly covers the slopes of the strike ridges (Fig. 34 & 45). The map pattern itself indicates that the boulder colluvium has been generated by a process that affects the entire slope, both vertically and laterally, more or less evenly. This evenness of erosion is likely to be the source of the smooth concave slope of the ridges that runs along strike without interruption by first order hollows. The transport of the boulder-sized material across very gentle toeslopes suggests a process that "fluidizes" the material (greatly reduces the shear strength). Temperate climate processes that can do this are associated with rare rainfall events that are large enough to produce debris flow activity on the slopes. But this activity tends to generate a series of long narrow flows that primarily affect zero and first order hollows and only affects a small percentage of the overall mountain side (Kite and others, 1987; Jacobson and others, 1989). To produce the boulder colluvium mantle in eastern Pennsylvania by this process would necessitate a large number of debris flows "side by side" from a large number of rare rainfall events. It is doubtful that there has been enough time in the Holocene to do this and largest historic rain events in eastern Pennsylvania have not produced debris flows in the boulder colluvium areas. Throughout the late and middle Pleistocene, warm temperate climate conditions occur for less than 10 % of of entire interval (Fig. 1); would all the interglacials collectively provide enough time for the rare rain events to do their work ? Of course this also assumes that the intervening colder intervals would have had no effect on the landscape.

The map pattern and the boulder-mantled surface of the boulder colluvium is most simply explained by a process that "fluidizes" the material, tends to heave large clasts to the surface, and affects the entire slope - periglacial gelifluction. Also, colder than present climates have been the more prevelant climate in eastern Pennsylvania during the last 850,000 years and it would be expectable that processes related to those cold climates have dominated the landscape. In addition, the degree of soil development in the boulder colluvium indicates landscape stability throughout much if not all of the Holocene (Ciolkosz and others, 1986). Some specialists in present day periglacial processes (Discussions on previous field trips in this region) are not convinced by any of the above because they know of no present day analogies of such boulder colluvium deposits. Are there really no such boulder colluvium deposits in the Artic today, such as along the strike ridges in the Brooks Range of Alaska ?

The critical future research need is to obtain the three dimensional geometry, stratigraphy, and age of the ubiquitous colluvium deposits in eastern Pennsylvania. Another worthwhile project would be to compare the slope morhology and associated processes on a single strike ridge starting from here and going to the southern edge of that ridge, for instance Blue Mountain in eastern Pennsylvania to Clinch Mountain in eastern Tennessee. A new generation of dating techniques such as cosmogenic istopes will be necessary to date the colluvium. Another key research area is to better determine the mode of deposition of the colluvium by finding those apparently elusive modern day analogies.

SORTED PATTERNED GROUND IN THE MIDDLE AND SOUTHERN ANTHRACITE FIELDS AREAS, RIDGE AND VALLEY PROVINCE, PENNSYLVANIA

by: G. Michael Clark, Geological Sciences
University of Tennessee, Knoxville, TN 37996-1410

Introduction

A number of different types of periglacial features have been reported in the Central Appalachians (Ciolkosz and others, 1986; Clark and Ciolkosz, 1988). One of the most promising groups of features for climatic and slope stability studies is sorted patterned ground. The nongenetic term 'sorted patterned ground' is a group term for more or less symmetrical forms in soils that are characterized by the separation of stones (pebbles, cobbles, blocks, boulders) from fines (sands, silts, clays) to produce specific geometric forms named: circles, nets, polygons, steps, and stripes (Washburn, 1956, 1980). In the Central Appalachians, these features exist at several size scales. A small-scale (miniature) forms (up to about 40 cm in diameter and involving size classes no larger than pebbles and cobbles) can be found in wet areas on previously-disturbed land that lacks vegetative cover. These diminutive forms can be active in the present climate. Intermediate-scale forms (about 50 cm to 2 m in diameter) appear to be inactive today and may be features that formed during prior major disturbances in historic times (logging, mining, fires) and/or during colder times, such as the Neoglacial geologic-climatic unit of Porter and Denton (1967) and its severe and terminal(?) event, The Little Ice Age (Grove, 1991). Large-scale forms (greater than 2 m in diameter) are completely inactive or truly fossil, as indicated by both weathering features and soil-geomorphic criteria (Rapp, 1967; Clark, 1992/1993; Clark and Ciolkosz, 1988; Clark and Ciolkosz, unpublished data) and are the forms that will be discussed below. Most, but not all, large-scale (> 2 m mesh diameter or stone border width) sorted patterned ground is of cold-climate, or 'periglacial' origin (Washburn, 1980; Williams and Smith, 1989), although genesis under certain other environmental conditions is known.

General Site Characteristics

In the Middle and Southern Anthracite fields, medium to thick-bedded, quartz-rich sandstone and conglomerate lithologies that break up along bedding and joints to produce large blocks (about 0.5 to several meters in a-axis dimensions) provide ideal source rocks for the development of sorted patterned ground. Large scale sorted nets and stripes commonly occur in areas underlain by, or downslope from, such lithologies. Tables 3 and 4 provide location and descriptive data for several selected sorted patterned ground finds in the anthracite fields.

Almost all of the finds of sorted stone nets to date (1994) have been on nearly horizontal to very gently-sloping upland areas. Many more sorted net localities may occur on the isolated high flat ridge-crest areas in the Middle and Southern Anthracite fields that lack road or trail access. Sorted nets occur on level to gently-sloping land with gradients up to 3°. Mesh (stone-free center) diameters range from 3 to 10 m; nets on slope land tend to e elongate. Stone border widths rage from 1 to about 5 m. At least some of the sorted nets once may have been well-shaped sorted polygons, as suggested by slightly angular corners in the stone mesh intersections.

Finds of sorted stone stripes are more common, having been reported by a number of authors (Crowl and Sevon, 1980), and tend to occur on land with higher slope gradients; slope angles range from 1.5 to 10°. Stone stripe widths range from about 3 to 11 m. Sorted stripes occur as solitary forms, in complexes, and in conjunction downslope from sorted stone nets that exists on gently-sloping upland surfaces, and with block fields, block slopes, and block streams. It is common to find sorted stripes either merging downslope into block streams or having an apparent surface source upslope from block slopes or block streams. Some to much of the block and boulder material, though, may have come from subsurface bedrock sources.

Discussion

In the Central Appalachians of Pennsylvania, Maryland, Virginia and West Virginia, Clark and Ciolkosz (unpublished data) have located more than 50 sites where large-scale sorted nets occur, and more than 150 locations where large-scale sorted stripes are present (Clark, 1992/1993, Figure 8, p. 48). These finds, however, should not be regarded as providing a true picture of sorted patterned ground distribution. For example, there are as yet few finds of sorted patterned ground in the Appalachian Plateau province. In part, this may be because many areas on the Plateau lack exposures of suitable resistant lithologies. On the other hand, there has been no systematic reconnaissance effort made to locate patterns in this geomorphic province, although there are several finds. The report of sorted stripes 10 to 15 feet wide and many tens of feet long on the slopes of Montogomery Creek in the southern half of the Penfield quadrangle (Edmunds and Berg,

Table 3. Representative large-scale sorted net ground finds in the Central and Southern Anthracite Fields. (Clark, Ed., 1992/1993, Figure 8, p. 48, and Clark and Ciolkosz, unpublished data.)

Quadrangle (U.S.G.S. 7.5') Location	Latitude Longitude deg min sec	Elevation (m)	Remarks
Ashland Wilburton area	40 49 04 76 22 10	539	With local slope angles of 3°, sorted nets are elongated up-down slope
Conyngham L. Sugarloaf Mtn.	40 56 03 76 05 15	605	Excellently-developed centers; borders w/large blocks incomplete in places
Delano Bears Head	40 51 02 76 04 45	634	Crestal area bordering fire tower with nearly equant sorted nets
Tower City Stony Mtn.	40 33 15 76 32 15	472	Sorted nets are transitional with sorted stripes on slopes of 3° and greater

Table 4. Representative large-scale sorted stripe finds in the Central and Southern Anthracite Fields areas. (Clark, Ed., 1992/1993, Figure 8, p. 48; Clark and Ciolkosz, unpublished data; Crowl and Sevon, 1980.)

Quadrangle (U.S.G.S. 7.5') Location	Latitude Longitude deg min sec	Elevation (m)	Remarks
Catawissa (site 1) Catawissa Mtn.	40 56 24 76 25 28	553	Large sorted-stripe complex with tabular blocks that display excellent orientations of a-axes and ab-planes in both headward source area and in individual stripes downslope
Lykens (site 1) Broad Mtn.	40 30 52 76 44 49	472	Excellent example of sorted stripe transitional to a block stream
Lykens (site 2a) Rattlesnake Den	40 31 00 76 44 51	454	Excellently-defined single sorted stripe
Pine Grove US 209 & I-81	40 37 08 76 26 33	372-384	Excellent example of sorted stripe complex

1971, p. 61) is a case in point. In the Middle and Southern Anthracite Fields, there are finds of sorted patterned ground in many areas where source rocks occur and where slopes are not overly steep. Lack of road or trail access would make the production of a more complete inventory a time-consuming and inefficient process. Also, mining activity has no doubt destroyed some features, so that a complete inventory will never be possible in disrupted land areas.

Some sorted stripes are physically continuous with block streams; stripes enter some block streams from upslope directions, apparently as feeders, and others emerge at the distal termini of some block streams, apparently as disseminators. Sorted circles, defined as nests of smaller and more rounded cobbles and boulders , are present in both block streams and block fields. Thus there is a physical, if not necessarily genetic, linkage between the origins of at least some sorted patterned ground and some parts of some block streams and block fields.

The results of experimental studies provide information about conditions under which sorted patterned ground may originate. For example, Coutard and others (1988) conducted an experimental study of frost heave and frost creep in an insulated tank. The materials consisted of loam, sandy loam, and a poorly sorted sediment mixture that included gravel. The experimental slope had a 12° inclination and was designed to include subsurface water flow. They found that processes changed downslope and intensified in loam where water supply was greater. The largest movements occurred just after thawing fronts passed through the materials. They demonstrated that frost creep was able to deform sorted polygons into sorted stripes and that such activity took place with the progressive disappearance of gravel-rich wedges transverse to the slope direction.

Most, but not all, large-scale (> 2 m mesh or stone border diameter) sorted patterned ground is of cold-climate, or periglacial origin (Washburn, 1980; Williams and Smith, 1989). The requisite conditions for other types of origins for these large features are lacking today in the Central Appalachians. Neither are there soils with large percentages of expanding clay minerals (Vertisols), nor are there soils that contain large amounts of salt (certain Aridisols). Some authors, for instance Goldthwait (1976), interpret sorted polygons and nets over 2 m in diameter as features requiring permafrost for development. Regardless of the ground thermal state(s) that accompanied the development of large-scale patterns, the above authors agree that the formation of large-scale patterns over broad areas occurs only above treeline. Of course, active development of sorted patterned ground features on intermediate or small scales often occurs in local azonal conditions with frost-susceptible (silt-rich) soils, as for example small sites with high water tables and in areas where forest and ground cover have been removed by human activity.

If large-scale sorted patterned ground features date from one or more pre-Holocene cold-phase events and are thus so old, how have they survived the effects of long-prolonged tree bole and root growth and tree throw? The answer may lie in the subsurface, as these features that can be seen in excavations are of the "deep rooted" variety, with tabular blocks tending to be oriented with their ab planes (maximum projection area) vertical and anchored firmly in the subsoil. There are few places, such as road cuts, where the maximum depth of stone concentrations can be seen without backhoe excavation, but for the larger stripes and nets observed to date (1994), the approximate depth to base of contiguous stones ranges from about 1 m to a maximum of about 1.7 to 1.8 meters. If permafrost existed during sorted stripe development, this stone depth might have some relationship with the thickness of the active layer. On sloping land on the other hand, depth to base of contiguous blocks could be a measure of the effective depth of highly-disruptive seasonal frost activity that was capable of segregating, orienting, and transporting large blocks of rock.

TORS IN THE MIDDLE AND SOUTHERN ANTHRACITE FIELDS , RIDGE AND VALLEY PROVINCE, PENNSYLVANIA

by: G. Michael Clark, Geological Sciences
University of Tennessee, Knoxville, TN 37996-1410

Introduction

Tors are large, free-standing, essentially-in-place, tower-like masses or piles of residual rock that remain unconsumed by erosional episodes. Tors have been extensively studied in Australasia, Europe, and Africa, but have received little detailed attention in North America with respect to their origins and significances. Exceptions include Cunningham (1969), Bailey (1983), Inners (1988), and Braun and Inners (1990); the last two authors studied prominent tors in the Anthracite region. Also, a number of visually spectacular tors near roads and trails in Pennsylvania have been identified and described as scenic geological features (Geyer and Bolles, 1979, 1987). Examples in the Appalachian Mountain geomorphic subsection are Wolf Rocks, Monroe County, and Devils Pulpit, Carbon County. Table 5 contains basic information on selected tors in the anthracite fields.

Table 5. Tor characteristics at two selected sites in the anthracite fields (data from D.D. Braun, p. 108, IN: Clark, ed., 1992/1993).

ROCK CHARACTERISTIC	GREEN MT. TORS (Freeland 7.5' quadrangle)	HAZLETON TORS (Hazleton 7.5' quadrangle)
Bedrock lithology (Pottsville Formation)	Quartz-pebble conglomerate and coarse-grained qtz ss	Quartz-pebble conglomerate
Bedding Characteristics	Planar-bedded to x-bedded, beds 0.25 to 1.0 m thick	Planar to x-bedded, beds 0.5 to 2.0 m thick
Bedrock strike	N 75° to 80° E	N 75° E
Bedrock dip	10° to 20° SE	Horizontal to 5° NW
Structure setting	Crest of homoclinal ridge	Crest of anticlinal ridge
Dominant joint orientation	N 40° - 50° W (dip joints) N 50° - 60° E (strike joints)	N 35° - 40° W (dip joints) N 55° - 65° E (strike joints)
Joint spacing	1.0 to 7.0 m (dip joints) 2.0 to 8.0 m (strike joints)	2.0 to 5.5 m (dip joints) 2.0 to 6.5 m (strike joints)
Tor dimensions: Height Length Width	 2.0 to 5.0 m 3.0 to 17.0 m (along strike) 2.0 to 8.0 m (along strike)	 2.0 to 6.5 m 3.0 to 20.0 m (along strike) 2.0 to 6.0 m (across strike)

Some hypotheses of origin

Because tors do not fit convenietly into models of uniform landscape reduction under constant environmental conditions, a number of hypotheses have been proposed to explain the origin of these topographic eminences. When discussing the following hypotheses of origin, several considerations may be of intellectual assistance. One of these conditions is that tors in some regions may have evolved under conditions of significant climatic, tectonic, or other environmental change and thus would be polygenetic in origin. These features could be called tors of compound, or of multigenetic, origin. A second situation is that, regardless of the environment(s) of formation, tors can be produced under conditions of either downwasting or backwasting, or some combination of the two. For instance, it would be well to remember that authors who subscribe to backwasting as a primary process group generally agree that the production of deep chemical alteration need not be involved in tor production in summit areas, but might well occur on gentle upper valley sides. In areas where such tor evolution is active, the lower surfaces might eventually encroach upon and consume the summit areas (King, 1966), producing one or more new geomorphic surfaces. Several hypotheses of tor origin are given below.

First, within a cyclical framework of evolution, tors may be the products of a two-step development involving an initial climatic situation and a subsequent environmental change. The first stage requires hot, humid tropical conditions, with production of a deep, differentially-weathered regolith with bedrock pinnacles at the base. The second stage mandates the removal of surrounding weathered fine clastic sediment to exhume the remaining firm rock towers. Driving processes may include climatic change (Linton, 1955), tectonic or epeirogenic rejuvenation (Falconer, 1912), or internally-driven subsurface weathering and removal processes that do not require climatic or tectonic input (Thomas, 1965). Under the two-cycle hypothesis described above, angularly-shaped tors in areas without present-day tropical climates would be interpreted as palimpsests that may have had their originally-rounded morphology altered or destroyed by subsequent processes.

Second, tors could originate and evolve under humid tropical climatic environments and could exist as subaerial features throughout their existence (Macar, 1957). In this scenario, those tors found in non-tropical environments today would be interpreted as subaerial survivors of subsequent processes that may be acting to modify or destroy them in order to produce forms that would be in harmony with the new climatic environment.

In a third hypothesis, tors are of a periglacial origin, and are the last remnants left by erosional processes involved in the production of cryoplanation terraces in upland areas above the forest limit under extreme periglacial environments (Fitzpatrick, 1958). Cryoplanation terraces (Demek, 1969) may develop in extremely could, relatively dry, continental environments such as interior Alaska (Reger and Pewe, 1976) and in wet maritime areas, such as Iceland, where mean annual air temperatures are near 0^{o} C (Priesnitz, 1988). Tors existing in areas that are now below the forest limit would be interpreted as fossil forms that are being modified by subsequent environmental conditions (Palmer and Radley, 1961).

Research on the origins and ages of tors in the Central Appalachians is just beginning. With judicious use of geochemical and geophysical research tools, it will not only be possible to determine numerical values for their exposure and erosional histories, but also to relate them to the surrounding rock and regolith. By use of geophysical prospecting techniques and soil stratigraphic study controlled with TL and C-14 numerical age control dates, the dates derived from tors can be related to the nature of the regolith and its underlying bedrock at sites immediately bordering the dated tors.

Tors as sites for the determination of exposure histories and erosion rates in upland areas

Tors may surmount topographic crests, occur on upper slope breaks of ridges, and exist on spurs; these are locations where knowledge of their geomorphic history would be useful in glacial, paraglacial, and periglacial studies. Until recently, however, there have been few detailed investigations about the structural and lithologic properties of tors. If one were to propose a world type area for tors, Dartmoor would be a prime candidate. For instance, Gerrard (1990, p. 230) noted that Dartmoor has long been regarded as an important area and as a key to the understanding of Cenozoic landscape evolution in Britain. Ehlen (1990, 1992) showed that the Dartmoor tors are the most resistant, and slowest weathering and eroding, remnants of lithology in an area beyond the Late Quaternary continental glaciation limits, and presented a semi-quantitative classification of tors on Dartmoor. She differentiated summit tors, valleyside tors, and spur tors on the bases of variation in relief; joint spacing; and composition, texture, and structure of the bedrock. Although the Dartmoor tors are of different lithologies (plutonic rocks) than those studied here (orthoquartzites), tors studied to date in the anthracite fields (Inners, 1988; Braun and Inners, 1990) share certain similar overall structural characteristics with the sheet structure and jointing in the Dartmoor tors.

For example, sedimentary bedding is thick to massive, dips are gentle to nearly horizontal, and joints are widely space, thus minimizing any complicating factors of structure or lithology. These two authors concluded that the tors are products of prolonged and continuous exposed bedrock weathering. The diagnostic formative weathering environments are interpreted to have been solely ("single stage" in their terminology) periglacial (Braun and Inners, 1990).

Recent advances in methodologies for dating geomorphic surfaces (Phillips and Dorn, 1991) now provide numerical exposure ages and erosion rates for rocks that can be assumed to have been exposed to cosmogenic bombardment. From several standpoints, tors can be defended as excellent sample points for diamond drill coring to obtain samples for AMS cosmogenic dating. All of the rock units are composed almost exclusively of silica-cemented detrital quartz grains, many of which lack evidence of internal fracturing and are thus ideally suited for dating. Weathering effects on the tops of the tors produce slow in situ grain-from-grain disintegration of orthoquartzite, and none of the field sites display rind- or shell-like exfoliation features. Weathering profile characteristics in the tors are also ideal for AMS dating. Individual (naturally sealed) quartz grains are easily separated from the matrix at various shallow depths in the weathering profiles, or can be disaggregated in deeper firmer rock. Thus, unwanted mineralogy can be excluded from the samples for chemical separation and contaminant "garden variety" Cl and other atoms can be eliminated from small samples required for AMS dating that contain fluid inclusions, and hence Cl of interest.

Long-term, continuous, exposure to cosmogenic bombardment during weathering and erosional development is inferred from all of the field evidence independently collected to date by different workers (Braun and Inners, 1990; Clark and Hedges, 1992). Of especial importance is the presence of in situ, unmodified, Opferkessel (Hedges, 1969) which demonstrates that the sample surfaces have not been vegetated, covered overwise by parent material or soil, nor tilted (if these basins are horizontal) during the time required for these closed depressions in essentially pure quartzites to form, as these features can develop and persist only on exposed rock surfaces (Hedges, 1969). Conversely, very recent surficial mass movements are shown by out-of-place blocks with tilted Opferkessel, as at the Hazleton water tower area tor site. Hedges (1969) conducted a thorough study of Opferkessel, including many Appalachian localities, and concluded that these features have developed very slowly under modern (Holocene) climatic conditions, although quantitative rate(s) of development are not yet known beyond historical records. The low solubility of silica in natural waters poses problems for models of inorganic chemical solution as an agent of solution pan formation. A productive area for research on Opferkessel is biogeochemical activity. For example, observation by Bennett and Siegel (1987) identified a direct link between the presence of dissolved organic carbon and enhanced rates of dissolution of silica. Folk (1993) provided evidence that certain nannobacteria may be active weathering agents as well.

A final potential problem that must be addressed in selecting sites for cores for AMS dating is the danger of loss of thick weathering rinds during range or forest fires. Blackwelder (1927) emphasized the concern about fire as an agent of spallogenic weathering on exposed rocks. The high-standing topographic position of tors for drillings minimizes to every extent possible the danger of rock loss by fire. Studies of Late Quaternary vegetational history on the high-elevation local broad upland sites indicates that neither range fires nor high-temperature crown fires have been likely for many thousands of years because of the lack of vegetation types capable of propagating such high temperature fires (Delcourt and Delcourt, 1987). These sites are interpreted to have had tundra vegetation during the Late Pleistocene, late-glacial maximum between 25-16.5 Ka (Delcourt and Delcourt, 1987). There is, of course, a subsequent forest history and fire record, but in the northern hardwood and spruce stands that have occupied the sample sites during the last 16.5 Ka, extreme fire temperatures were likely only in dried organic soil horizons that are several meters below the tor tops. The same rationale holds for burns of the slash-cut that accompanied forest cutting in the high Appalachians during the peak of logging operations between ca. 1890 and 1920. Finally, evidence of weathering types that are known to be produced by range and forest fires are absent on the tor tops.

Future work will involve additional sampling and laboratory analyses. Cosmogenic signatures of, for example Cl-36, can be detected and measured to depths of more than one meter, a depth that virtually guarantees that long-term numerical age dates can be obtained by use of AMS techniques. The selection of specific tor summits as the best sample drilling points will be fully refined with preliminary Cl-36 dating before the main expenditure of funding for the two additional isotopes (Be-10 and Al-26). This additional safeguard will be a critical test to determine the very best drilling points and depths from which to extract more cores for analysis of the additional isotopes. Recent successes in the use of cosmogenically-produced isotopes (Evenson and others, 1993; Gosse and others, 1993) bodes well for use of these procedures. Laboratory techniques for sample chemical preparation for AMS dating continue to be expanded (Vogt and others, 1993) and AMS procedures continue to be refined to give more precise values (Elmore and others, 1993).

Discussion

In the FOP field trip area, tors occur both within and outside of the Late Cenozoic glacial borders. They commonly surmount and border the low-relief upland surfaces that often break abruptly into the blocky slopes of the ridge sides. Many early workers in the Central Appalachians followed Davis (1889) and interpreted the low-relief, nearly-level accordant upland surfaces in the Ridge and Valley province - where many tors occur - as remnants of one or more peneplains. Incidentally, Monmonier (1967) used trend-surface analysis to demonstrate conclusively that local broad upland ridge crests in the Ridge and Valley province in Pennsylvania actually are highly accordant! Nonetheless, upland surface origin has remained a neglected problem area perhaps because until recently there have been no obvious long term numerical dating techniques available to attack and model environments and geomorphic processes in ridge crest areas, and perhaps because researchers chose not to approach a problem that smacked of denudation chronology. Most modern geomorphologists would probably question the interpretation that ridge crests are remnants of once-continuous peneplains, or of once-extensive surfaces of other origins. The painful fact remains, however, that certain flat topped high-elevation summit levels - named local broad uplands by Monmonier (1967) - actually do truncate structure and lithology in a number of areas (Clark and Hedges, 1992). In order to explain these upland relationships without resort to cyclical, closed-system thinking, Clark and Hedges proposed a hypothesis of paleoperiglaciation in which they viewed the local broad uplands as incipient surfaces of cryoplanation as opposed to remnants of once-continuous landscapes.

Regardless of their origins, the accordant local broad uplands are prime source area candidates for the vast volumes of diamictons that mantle the slopes of ridges, and extend into many valley areas as well. When, and under what conditions, did the production of regolith and the mass transfer of it occur? If some problems related to temporal- and erosional-rate constraints can be solved by dating tors, the difficulty that a number of workers in the Central Appalachians have called 'the dating impasse' may well have been broken.

BEDROCK GEOLOGY FROM THE NORTH BRANCH SUSQUEHANNA VALLEY TO THE EASTERN MIDDLE ANTHRACITE FIELD

by Jon D. Inners, Pennsylvania Geologic Survey,
P.O. Box 2357, Harrisburg, PA 17120

Valley and Ridge Overview

The Day-1 field trip route between Hazleton and Berwick and back again traverses rocks of the Valley and Ridge province that range from middle Devonian to late Pennsylvanian in age (Table 6 & 7 and Fig. 18). The area's distinctive linear topography is the result of differential erosion of folded sedimentary-rock strata that generally strike east-northeast. High ridges (1500-2000 feet in elevation) are mostly held up by upper Devonian to lower Mississippian and lower Pennsylvanian quartzitic sandstones and conglomerates. Secondary ridges and uplands (900 to 1100 feet) in the Berwick area are underlain by upper Devonian sandstones, siltstones, and shales. The North Branch lowland (500-700 feet) at Berwick has been eroded from Silurian to middle Devonian shales and limestones, and the broad, rolling valley just north of Hazleton (Conyngham Valley) (800-1000 feet) is developed on upper Mississippian red shales and sandstones. The high valleys (1300-1500 feet) of the Eastern Middle Anthracite field are underlain by middle to late Pennsylvanian sandstones, conglomerates, shales, and coals. Only in the vicinity of Berwick, where the field trip route parallels the late Wisconsinan "terminal moraine," do conspicuous glacial landforms overprint the bedrock topography.

Deformation in this part of the Valley and Ridge is entirely Alleghanian and dates from about 280 Ma. From the Mahantango Formation to the Pottsville Formation, broad folding is the dominant type of deformation. In the heterolithic, coal-bearing Llewellyn Formation, however, numerous faults cut the intricately folded strata. Summaries of the structural geology of Day-1 route can be found in Inners (1978, 1988); Wood and Bergin (1970); Nickelsen, (1979, 1987) and Wood and others (1986). Detailed 7.5-minute bedrock maps of the Berwick and Conyngham quadrangles have been prepared by Inners (1978) and Schasse and others (unpublished), respectively. Reconnaissance maps of other quadrangles on the Day-1 itinerary are included in Berg and Dodge (1981).

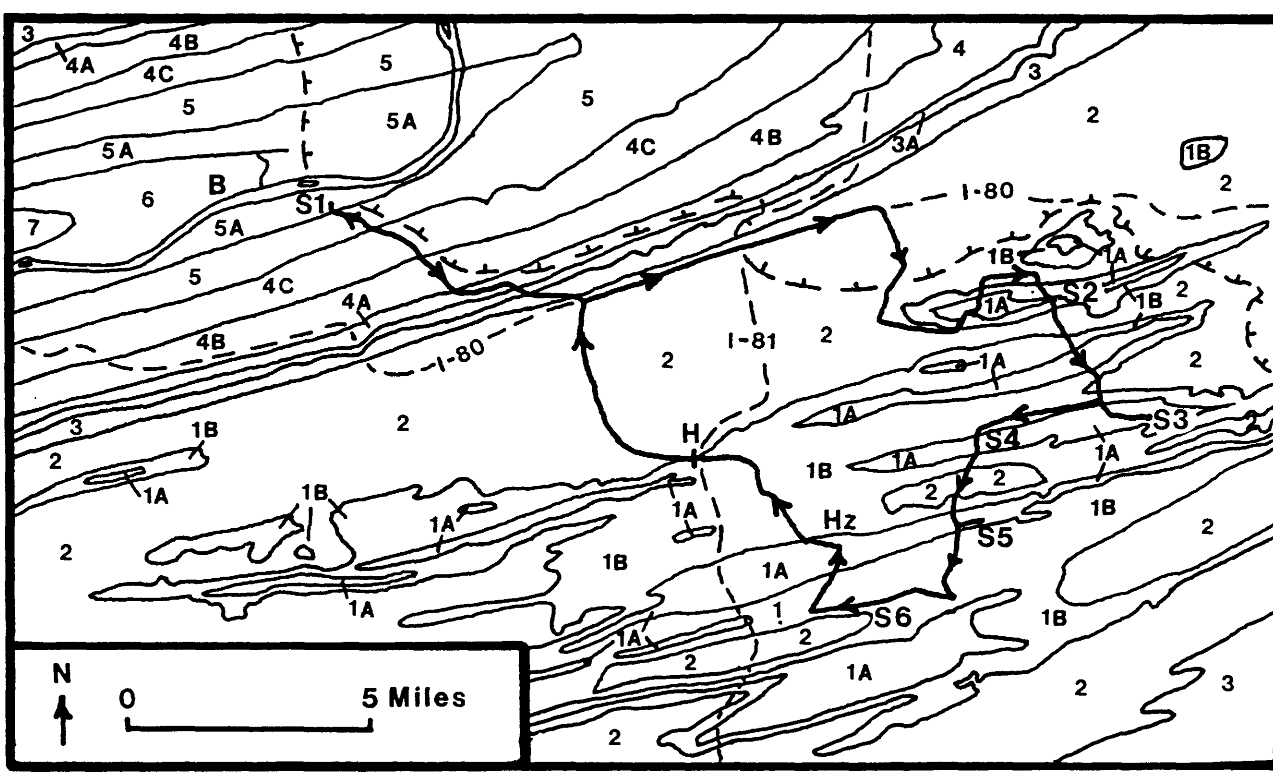


Figure 18. Generalized bedrock geology of the Valley and Ridge in the area of the Day-1 trip (after Berg and others, 1980). Numbers are keyed to Table 6. B; Berwick; H, Hazleton; F, Freeland; W, Weatherly.

Table 6. Bedrock units exposed in Day 1field trip area. Formations are grouped into map units shown in figure 18. A general description of the formations is given in table 5. **R**, major ridge; **r**, minor ridge.

PERIOD	MAP UNIT	RIDGE	FORMATION
PENNSYLVANIAN	1A 1B	 R	Llwewllyn Pottsville
MISSISSIPPIAN	2		Mauch Chunk
	3	R	Pocono
MISS. - DEV.	3A		Spechty Kopf
DEVONIAN	 4A 4B 4C		Catskill Duncannon Sherman Creek Irish Valley
	5	r	Trimmers Rock Harrell Mahantango Marcellus
	6		Onondaga Old Port
DEV. - SIL.			Keyser Tonoloway
SILURIAN			Wills Creek
	7		Bloomsburg Mifflintown

The Eastern Middle Anthracite Field

The Eastern Middle field is the smallest of the four major anthracite fields. Its maximum length is 26 miles and its maximum width is 10 miles, but coal-bearing rocks underlie only about 30 mi^2. Most of the field occupies a high plateau centered on the city of Hazleton (population 27,000). The highest altitudes (1800-2000 feet) occur on the steep escarpments bordering the plateau and along several northeast-southwest trending anticlinal ridges that have local relief of 200 to 300 feet.

Anthracite in the Eastern Middle field occurs in a number of generally elongate, disconnected basins (Figure 19). Only the Hazleton and Jeansville basins exceed a depth of 1000 feet to the bottom coal. In the other basins, the lowest minable coal lies well above sea level and could be completely mined out by open-pit methods under the proper economic conditions. The most important of about a dozen mined coalbeds are the basal Buck Mountain and, 300 feet higher in the Llewellyn Formation, the Mammoth.

The mined coal is generally of very good quality (Socolow and others, 1980). Fixed carbon (dry, ash-free basis) increases in a roughly west-east direction from 94 percent in the McCauley Mountain basin to 98 percent at the east ends of the Hazleton and Jeansville basins. Sulfur content ranges from less than 0.5 percent to somewhat over 1.0 percent. Heating value depends on the amount of volatiles and increases from 14,400 Btu/lb in the east to 15,300 Btu/lb in the west (dry, ash-free basis). Coalification models indicate that anthracite probably formed at temperatures between 200 and 250^{O} C (390 and 480^{O} F) at a burial depth of 5 miles (Levine, 1986).

No longer "King" as it was seventy-five ago, anthracite is now a dwindling part of the diversified economy of the Hazleton area. Production first passed a million tons per year in 1855, reached a peak of 8.9 million tons in 1914 and by 1948 had declined to 4.5 million tons (Ash and others, 1950,

Table 7. Generalized description of lithologic units in the map area of Figure 18.

PENNSYLVANIAN
Llewellyn Formation: Siltstone and shale with abundant plant debris, fine- to coarse-grained sandstone and conglomerate with quartz and quartzite pebbles and less abundant chert and metamorphic rock fragments as much as three inches in diameter with numerous anthracite coal beds. 1,500 ft (460m) thick
Pottsville Formation: Conglomerate, sandstone, and less abundant shale and siltstone with plant debris; and several beds of anthracite coal. 1,500 ft (460 m) thick
MISSISSIPPIAN
Mauch Chunk Formation: Gray conglomerate and sandstone, Red sandstone, shale and siltstone. 3,000 ft (915 m) thick
Pocono Formation: Sandstone, and minor siltstone and shale with abundant plant debris. 600 ft (183 m) thick
MISSISSIPPIAN AND DEVONIAN
Spechty Kopf Formation: Polymictic conglomerate, sandstone, siltstone and shale. 0-500 ft (0-150 m) thick
DEVONIAN
Catskill Formation: Red and gray, sandstone, siltstone, and shale. Duncannon Member: 1,100 ft (335 m) thick Sherman Creek Member: 2,500 ft (762 m) thick Irish Valley Member: 1,800-2,000 ft (550-610 m) thick
Trimmers Rock: Gray siliceous siltstone, silty shale, and sandstone. 2,500 ft (762 m) thick
Harrell Formation: Grayish-black shale. 100 ft (30 m) thick
Mahantango Formation: Gray shale and argillaceous limestone. 1,150-1,260 ft (350-384 m) thick
Marcellus Shale: Dark gray shale and silty shale. 300 ft (90 m) thick
Onondaga and Old Port Formations: Shale, limestone, and chert. 200-325 ft (60-100 m) thick
DEVONIAN - SILURIAN
Keyser Formation: Nodular limestone. 125 ft (38 m) thick
Tonoloway Formation: Laminated limestone. 200 ft (60 m) thick
SILURIAN
Wills Creek Formation: Calcareous shale and limestone. 600-700 ft (183-213 m) thick
Bloomsburg Formation: Red mudstone and siltstone. 500 ft (152 m) thick

Fig. 1). Based on an original resource base of $6.84x10^8$ tons, approximately $1.73x10^8$ tons of anthracite remain to be mined in the Eastern Middle field; of this $1.67x10^8$, lies within 1000 feet of the surface (Stingeln and others, 1984).

Current production in the field depends on large and small stripping operations in formerly deep-mined areas and reclamation of old breaker waste piles. Deep mining ceased entirely by 1970. The major operators include Jeddo-Highland Coal Company (strippings at Ebervale (Stop 4, Day 1), Highland, and Jeansville), Coal Contractors, Inc. (strippings at Gowen and Fern Glen), and Amscot Coal Company (bank reprocessing at Lattimer).

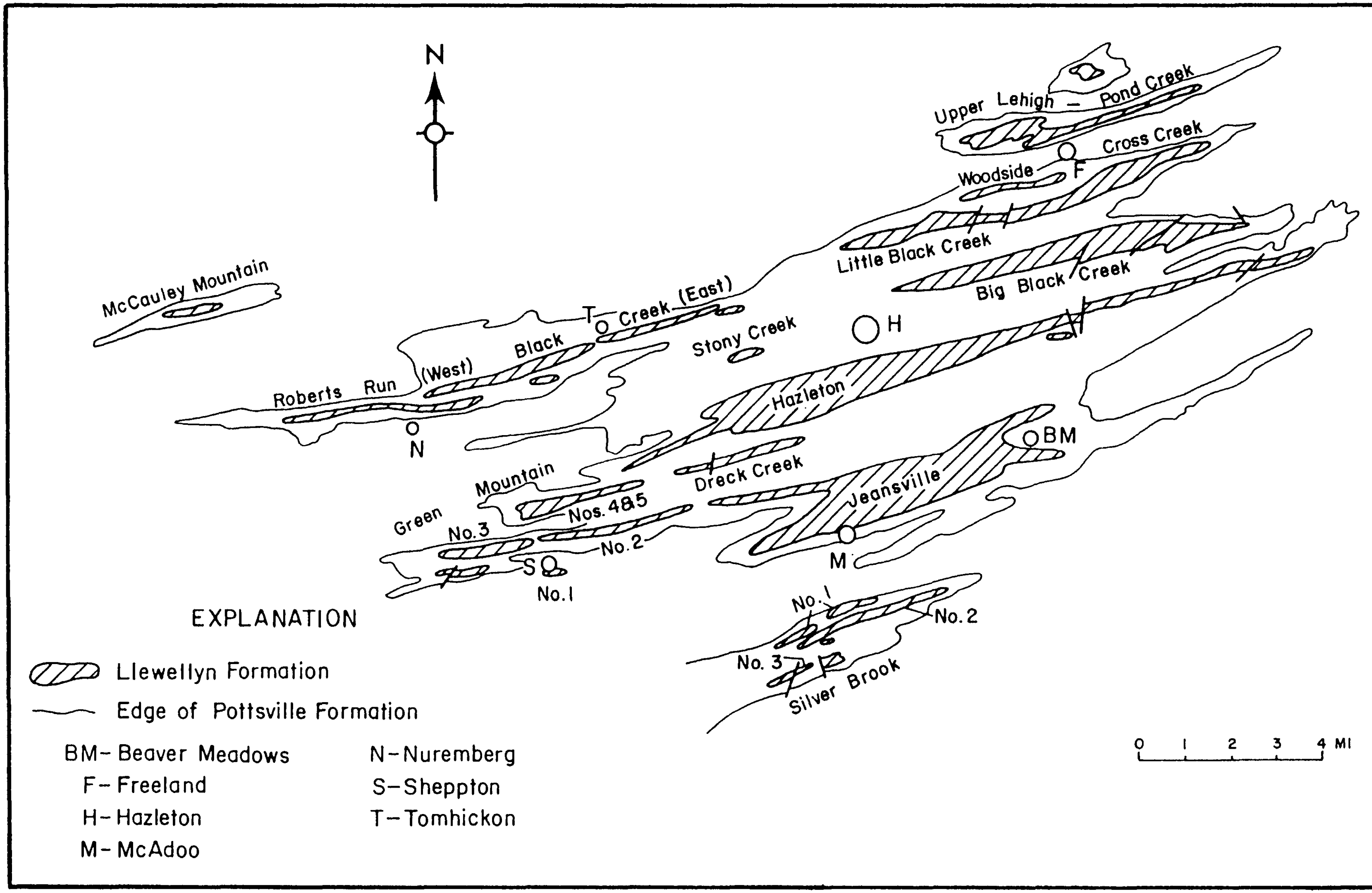


Figure 19. Anthracite basins of the Eastern Middle field. (Although similar in size to basins further north, the Silver Brook basins are geologically more related to the Western Middle field.) (From Inners, 1988.)

BEDROCK GEOLOGY SOUTH OF THE EASTERN MIDDLE COAL FIELD

by Jack B. Epstein, U. S. Geologic Survey,
MS 926, Reston, VA 22092

This year's field trip traverses the Valley and Ridge province of eastern Pennsylvania, comprising rocks of Ordovician through Pennsylvanian age (Fig. 20; Tables 8 and 9). The northeast topographic grain reflects the trends of folds and faults in rocks of varying vulnerability to erosion. The ridges are held up by resistant sandstone, quartzite, conglomerate and siliceous siltstone; the valleys are mostly underlain by weaker shale and limestone. Glacial erratics are mostly derived from the coarser clastic rocks, although cherty limestones of Middle Devonian age supply some large erratics to the drift in places. Table 8 lists the Paleozoic stratigraphic units exposed in the field trip area.

The structural geology of the Valley and Ridge province is dominated by an imbricate thrust system and northwest verging upright and overturned folds (cross section, figure 20). The thickness and heterogeneity of individual packages of stratigraphic units has determined the character of the resultant deformation. These lithotectonic units are believed to be separated by zones of detachment within incompetent rocks, such as the Marcellus and Martinsburg Formations. The deformation in Silurian and younger rocks is entirely Alleghanian. The Tuscarora Formation of Silurian age rests with angular unconformity upon various older rocks. These Cambrian and Ordovician rocks, underlying the Great Valley between Blue Mountain on the north and the Reading Prong highlands on the south, were deformed mildly to severely during the Middle to Late Ordovician Taconic orogeny. Summaries of the structural geology are given in Epstein and Epstein, 1969; Wood and others, 1969; Wood and Bergin, 1970; Epstein and others, 1974; Epstein, 1980; and Lash and others, 1984.

Regional bedrock geologic maps have been published by Berg and Dodge, 1981; Berg and others, 1980, and Lyttle and Epstein, 1987. Detailed 7.5-minute quadrangle coverage includes the maps of Epstein and Lyttle, 1993; Lash, 1985, 1987; Lyttle and others, 1986; and Wood, 1973 and 1974 a,b,c).

In general, older rocks will be encountered as we traverse south from Hazleton. The area between Hazleton and Snyders contains rocks of units 1-5. These rocks are intricately faulted and contain overturned folds; the major folds have wavelengths averaging several miles. The Pottsville Formation (unit 1) contains the most resistant rocks in this stretch. These hold up the ridges in the Hazleton area, including Spring, Broad, and Locust Mountains, and Nesquehoning and Pisgah Mountains in the Tamaqua area. Other resistant rocks include the Pocono Formation (unit 3), underlying Broad Mountain north of Tamaqua and Wildcat Mountain and Mauch Chunk Ridge south of Tamaqua. Conglomerate- and sandstone- rich members of the Catskill Formation(unit 4) hold up parts of Wildcat Mountain, Mauch Chunk Ridge, and lower hills to the south. The Trimmers Rock Formation (unit 5), underlies a moderately high ridge above the valley north of Snyders.

The next older sequence of rocks to the south consists of a heterogeneous assemblage of thin formations comprising unit 6. Some of the sandstones form ridges that rise to 300 feet (90 m) in a few places, but in many areas the rocks have little topographic expression due to extensive pre-Wisconsinan leaching and erosion.

These rocks are succeeded to the south by Silurian clastic units (unit 7) dominated by the ridge-forming Tuscarora Sandstone that holds up Blue Mountain. Within the Great Valley, the Shochary Sandstone of unit 9 forms a moderately high ridge south of New Tripoli.

The Silurian rocks rest with slight angular unconformity on highly contorted rocks of the Martinsburg Formation (unit 8). Most of the deformation in the Martinsburg is Alleghanian, although minor Taconic folding has been recognized.

The rocks of units 9 and 10 were severly deformed during the Taconic orogeny. The Shochary Sandstone and New Tripoli Formation (unit 9) lie within a large overturned, entirely fault-bounded syncline, that is not reflected in overlying rocks. Much of the deformation in that sequence may have been Alleghanian. These rocks are faulted upon the Martinsburg Formation and lie

Table 8. Generalized description of lithologic units in the day 2 field trip area.

PENNSYLVANIAN
Llewellyn Formation: Siltstone and shale with abundant plant debris, fine- to coarse-grained sandstone and conglomerate with quartz and quartzite pebbles and less abundant chert and metamorphic rock fragments as much as three inches in diameter with numerous anthracite coal beds. 3,000 ft (900m) thick
Pottsville Formation: Conglomerate, sandstone, and less abundant shale and siltstone with plant debris; and several beds of anthracite coal. 1,500 ft (460 m) thick
MISSISSIPPIAN
Mauch Chunk Formation: Red and gray sandstone, shale and siltstone. 2,200 ft (670 m) thick
Pocono Formation: Conglomerate, sandstone, and minor siltstone and shale with abundant plant debris. 800 ft (240 m) thick
MISSISSIPPIAN AND DEVONIAN
Spechty Kopf Formation: Polymictic conglomerate, sandstone, siltstone and shale 375-900 ft (115-275 m) thick
DEVONIAN
Catskill Formation: Red and gray, sandstone, conglomerate, siltstone and shale. 8,000 ft (2,400 m) thick
Trimmers Rock Formation: Gray siliceous siltstone, sandstone, and some silty shale. 1,000 ft (300 m) thick
Mahantango Formation: Gray shaly siltstone and silty shale and minor sandstone. 2,000 ft (600 m) thick
Marcellus Shale: Dark gray shale and silty shale. 300 ft (90 m) thick
Selinsgrove Limestone, Palmerton Sandstone, Schoharie-Esopus Formations, undivided, Oriskany Group, and Stormville Formation: Sandstone, conglomeratic sandstone, siltstone, and shaly limestone. 330 ft (100 m)
SILURIAN
Andreas Red Beds, Decker Formation, Bossardville Limestone, and Poxono Island Formation: Gray shaly limestone, shale and dolomite, red shale and siltstone and sandstone, dolomitic limestone. 750 ft (230 m) thick
Bloomsburg Formation: Red and gray shale, siltstone, sandstone, and minor conglomeratic sandstone. 1,500 ft (460 m) thick
Clinton Formation: Gray and some red sandstone, siltstone and shale. 1,400 ft (430 m) thick
Tuscarora Sandstone: Gray, partly conglomeratic quartzite and minor siltstone and shale. 250 ft (75 m) thick
ORDOVICIAN
Martinsburg Formation: Dark gray slate, and graywacke. 12,000 ft (3,700 m) thick
Shochary Sandstone: Gray graywacke, slate, and minor conglomerate. 3,000 ft (900 m) thick
New Tripoli Formation: Gray calcisiltite, shale, and graywacke. 3,900 ft (1,200 m) thick
Windsor Township Formation: Part of Hamburg klippe. Gray shale and siltstone with minor graywacke and red shale. At least 4,600 ft (1,400 m) thick

The clastic rocks of Ordovician age are underlain and succeeded to the south by several thousand feet of limestone, dolomite, and minor sandstone of Cambrian and Ordovician age. South of these rocks, in the Durham and Reading Hills, are gneiss and granite of Proterozoic age.

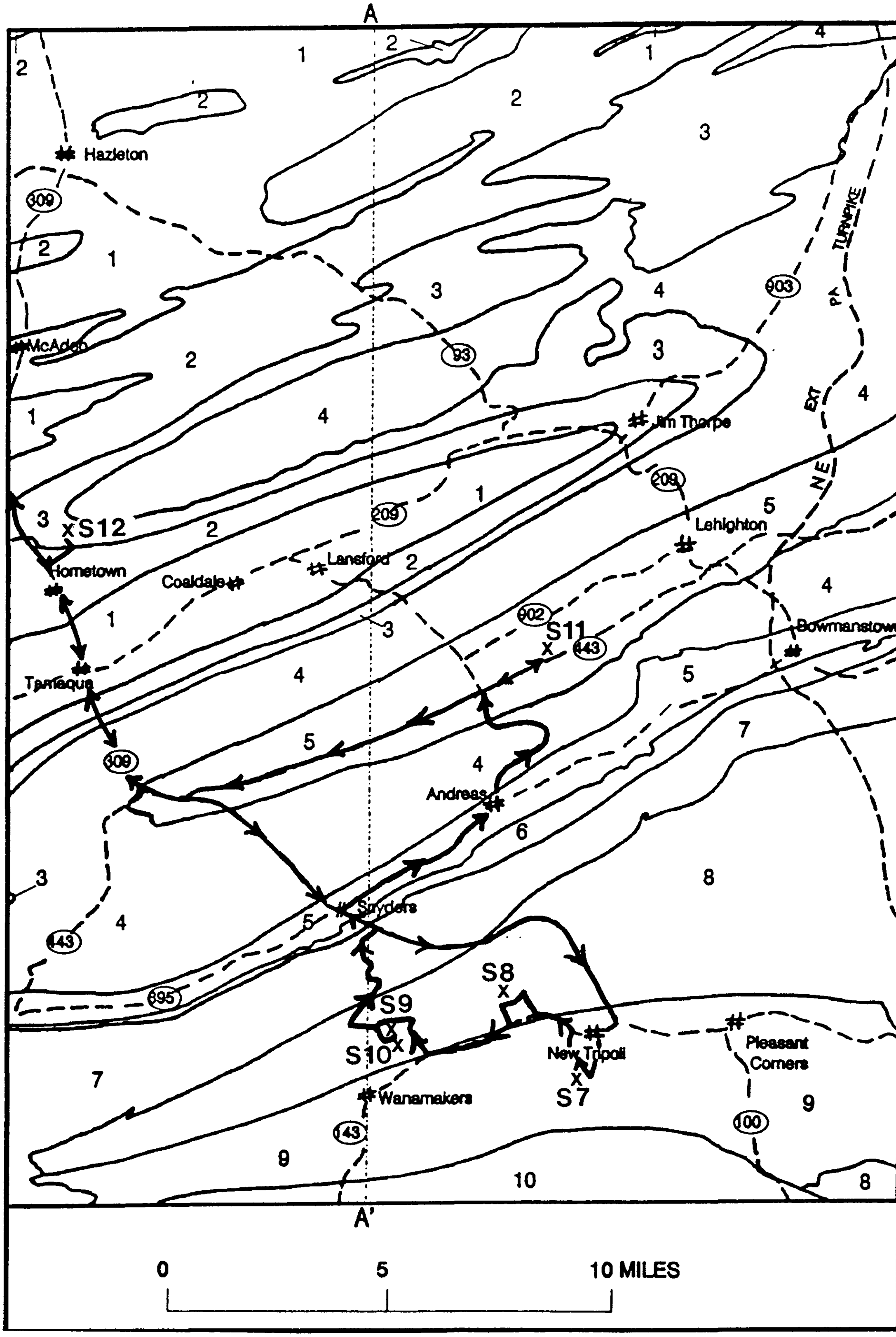


Figure 20. Field trip day 2 generalized bedrock geologic map and cross-section (facing page) showing map units described in table 8 and listed by number in table 9. "p" in cross-section indicates rocks older than the Martinsburg Formation. Field trip route marked by a series of arrows and stops by a letter S followed by a number. Map and section modified from Lyttle and Epstein, 1987.

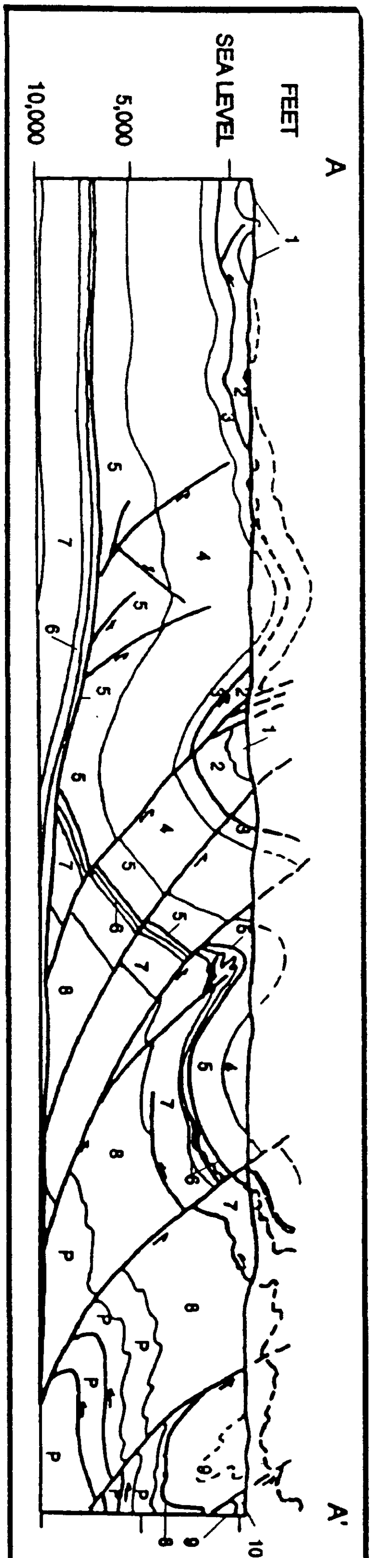


Table 9. Bedrock units exposed in the day 2 field trip area. Formations are grouped into map units shown in figure 20. A general description of the formations is given in table 8.
R, major ridge; r, minor ridge.

PERIOD	MAP UNIT	RIDGE	FORMATION
PENNSYLVANIAN	1	R	Llwewlyn Pottsville
MISSISSIPPIAN	2		Mauch Chunk
	3	R	Pocono
?			Spechty Kopf
DEVONIAN	4		Catskill
	5	r	Trimmers Rock Mahantango Marcellus
	6	r r	Selinsgrove Ls. Palmerton Ss. Schoharie Esopus Oriskaney Gr. Stormville
SILURIAN			Andreas Red Beds Decker Bossardville Ls. Poxono Island
	7	R R	Bloomsburg Clinton Tuscarora Ss.
ORDOVICIAN	8		Martinsburg
	9	r	Shochary Ss. New Tripoli
	10		Windsor Township

structurally beneath the Hamburg klippe, composed in largely of the Windsor Township Formation (unit 10). The Hamburg klippe, a complex sequence of far-travelled rocks similar to the Taconic allochthon of New York, was emplaced during Taconic plate convergence and later modified by Alleghanian folding and faulting (Lash and others, 1984). In general, rocks of the Martinsburg Formation and older carbonate rocks within the Great Valley have been thrust onto rocks of the Hamburg klippe during the Alleghanian orogeny.

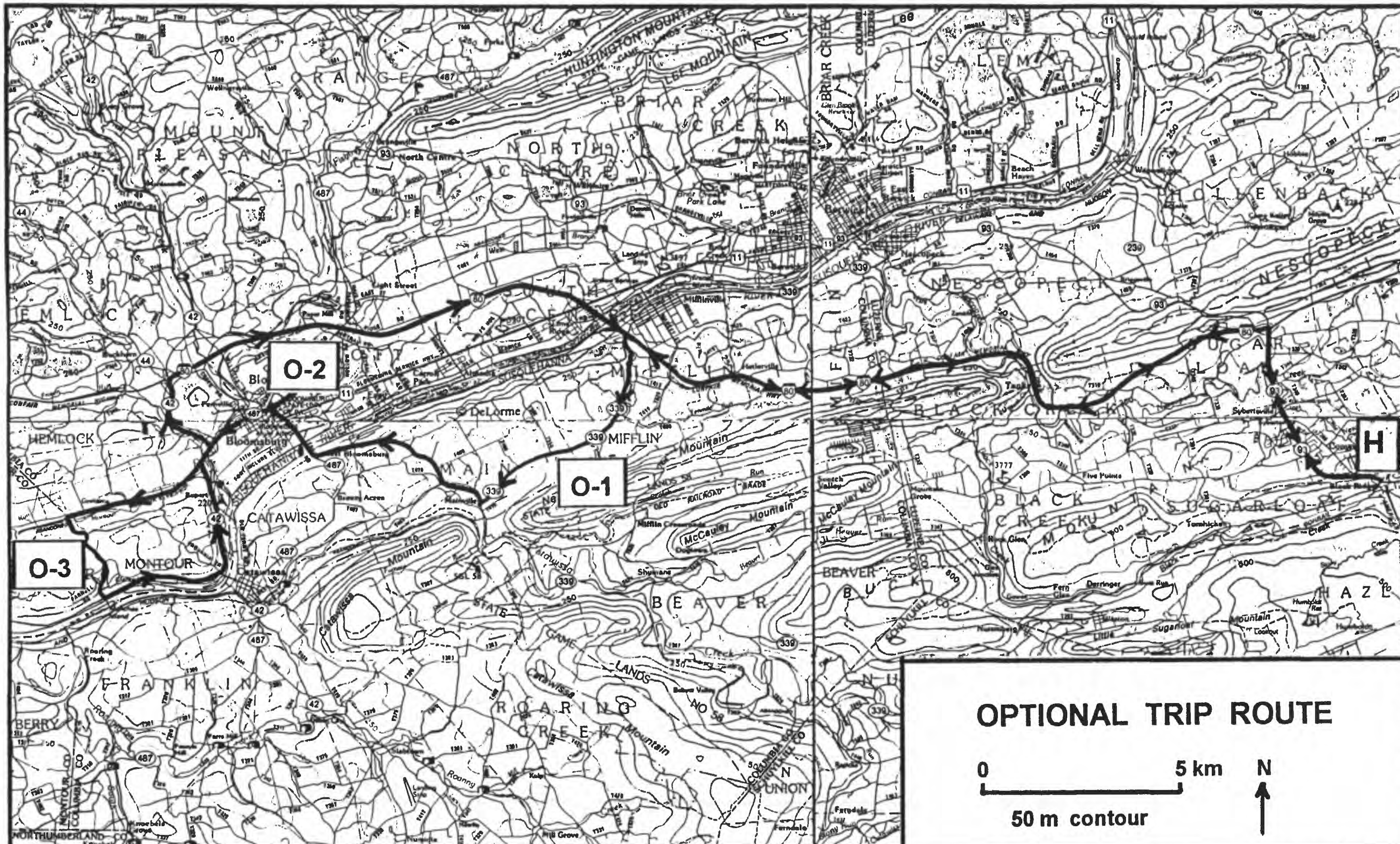


Figure 21. Route of optional trip on Friday to the Rupert water gap to examine the lack of evidence for catastrophic flooding.

OPTIONAL FRIDAY FIELD TRIP ROAD LOG

Mileage Inc	Cum	
0.0	0.0	**RIGHT ONTO 93** from hotel. Travel down scarp face of Buck Mtn., cross late-Illinoian limit within next 1/2 mile but boulder colluvium overlies any glacial material.
1.6	1.6	Straight through light. On skyline ahead, the Pocono Formation holds up Nescopeck Mountain. You are crossing a strike valley on Mauch Chunk redbeds (Conyngham valley); glacial drift patches on hilltops and in first order hollows.
1.9	3.5	Cross Nescopeck Creek. It often runs gray or black from underground coal mine drainage. Upstream of here, the Jeddo drainage tunnel discharges as much 70 MPG of acidic water into the creek. The six to seven mile (10-11 km) long Jeddo tunnel system under-drains all the mines under tomorrow's Stops 3, 4, and 5
0.8	4.3	**BEAR RIGHT ONTO RAMP FOR I-80 WEST.**
0.3	4.6	**MERGE ONTO I-80 WEST.** Run along strike with the Nescopeck Mountain dip-slope to right and the Buck Mountain scarp-slope on the skyline to left.
0.6	5.2	Boulder colluvium floors the forested area on both sides of I-80 for the next 1.3 miles.
2.0	7.2	Curve left away from Nescopeck Mountain.
0.7	7.9	View ahead and to left of Buck Mountain. At its base a late Illinoian "morainic loop", actually a head-of-outwash, starts swinging across the Conyngham strike valley.
0.3	8.2	View ahead of nose of flat topped synclinal McCauley Mountain. The "morainic loop" crosses the center of the strike valley at the base of MaCauley Mountain. On right, outcrops of Mauch Chunk redbeds.
0.5	8.7	Enter large outcrop of Mauch Chunk redbeds.
0.3	9.0	On left, Nescopeck Creek incised meanders.
0.7	9.7	Cross Nescopeck Creek and enter water gap through Nescopeck Mountain. The late Illinoian "morainic loop" attaches to Nescopeck Mountain west (left) of here. Ice filled the water gap and buried Nescopeck Mountain east (right) of here.
0.9	10.6	On right, rock slide on road cut in dip-slope of Mauch Chunk.
0.3	10.9	Next 0.5 mile, cross high road embankment that cuts off an incised meander loop of Nescopeck Creek. On left is the meander undercut scarp-slope of Nescopeck Mountain that exhibits a series of first order hollows cut in the steep slope. On right, deep artificial channel cutting off the incised meander loop.
1.6	12.5	Crest of hill. I-80 will now run westward along strike on rolling uplands underlain by the Catskill Formation redbeds. On left will be boulder colluvium along the base of Nescopeck Mountain. On right will be occasional late Illinoian till patches on hilltops and in heads of hollows. There is no constructional landform, the deposits are identified by soil type and outcrop exposures.
1.4	13.9	On right, road cut at overpass. At top of cut is a thin layer of late Illinoian till, yellowish red to reddish yellow (5YR 5-6/6-8), exposed in woodchuck holes.
0.4	14.3	Go past Rest Area.
2.0	16.3	On right at hilltop more till is exposed in woodchuck holes.
0.5	16.8	Start descent into the North Branch Susquehanna River valley, on right is Catskill Fm. outcrop.
0.7	17.5	To left at the Mifflinville sign is the broad valley of underfit Ten Mile Run, the abandoned valley of the North Branch Susquehanna River. Ahead and to right is the present North Branch Susquehanna valley.
0.3	17.8	Cross under high tension line. On right is an overgrown pit in a late Illinoian recessional kame that blocked the abandoned valley to the left and imponded a shallow proglacial lake. Varves floor the valley to the left, not slackwater deposits as would be expected if catastrophic floods had come down the River.
0.4	18.2	**BEAR RIGHT ONTO EXIT 37** for Rt.339 and Mifflinville. Late Wisconsinan terraces to the right.
0.5	18.7	**TURN LEFT** onto Rt.339 South to Mainville and pass under I-80.

0.1	18.8	**Bear left** staying on Rt.339 South. Pee Wee Hill road goes right.
0.2	19.0	On left is a bedrock outcrop followed by a sharp curve to the right. This is post-glacial valley of Ten Mile Run and the now incised outlet to the proglacial lake mentioned before.
0.7	19.7	To left is view of the underfit Ten Mile Run on the floor of the abandoned North Branch Susquehanna River valley.
0.6	20.3	Ahead, the abandoned Susquehanna valley starts curving to the right to parallel Nescopeck Mountain. Note first order hollows on the mountain face where valley undercuts it, just like where Nescopeck Creek undercut the mountain along I-80.
0.6	20.9	View ahead and to left of the low drainage divide of Ten Mile Run.
0.6	21.5	**PARK ON RIGHT SHOULDER OF RT.339.**

STOP O-1: HEAD OF OUTWASH BLOCKING THE ABANDONED VALLEY OF THE NORTH BRANCH SUSQUEHANNA RIVER

Leader: Duane D. Braun

At this site, a low divide separates the Tenmile Run Valley from an another valley that extends westward along the base of Nescopeck Mountain (Fig. 23). Well data and seisimic refraction profiles show that the divide is underlain by 60 m of sand and gravel, capped by 2-3 m of weathered and disrupted sand and gravel that is texturally a diamict. The divide is asymmetric, the northeast side has a steep slope and the southwest has a very gentle slope (Fig. 22). This is the typical longitudinal profile of a kame fan or head-of-outwash, the steep northeast side is the ice contact face and the gentle southwest side is the outwash surface sloping away from the ice front. The bedrock floor beneath the divide slopes continuously southwestward from the present North Branch Susquehanna River southwestward to Catawissa Creek (Fig. 23). The overall scale and slope of the partially buried bedrock valley indicate that it was cut by the North Branch Susquehanna River (Braun and others, 1984).

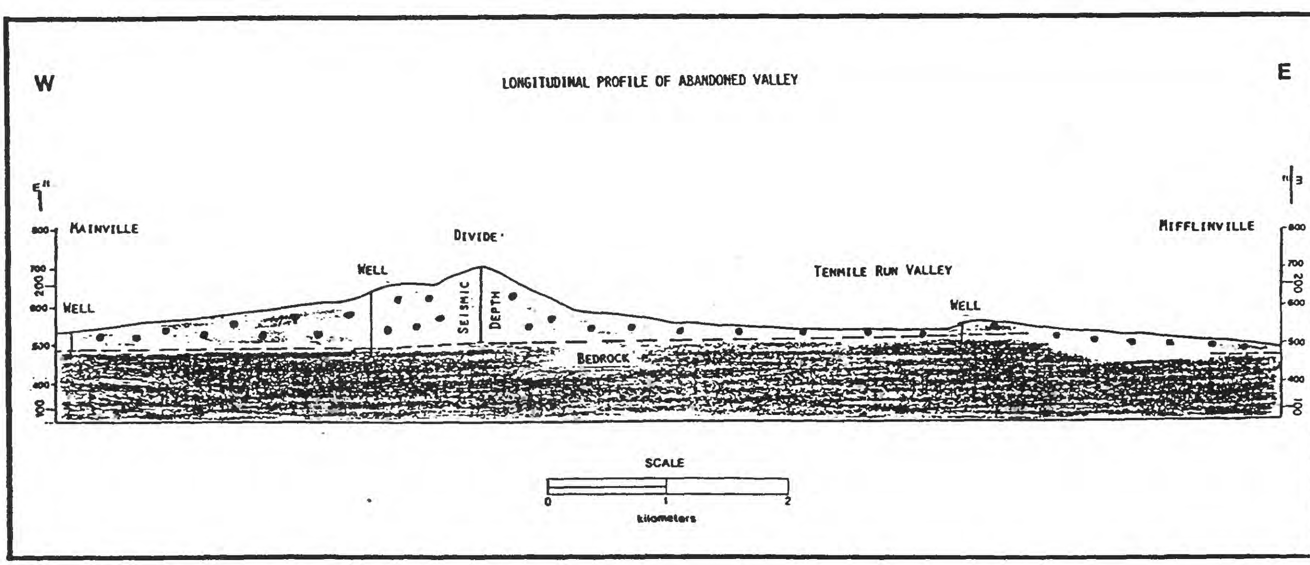


Figure 22. Cross-section along the center of the abandoned valley of the North Branch Susquehanna River showing a continuous westward slope of its bedrock floor under the glaciofluvial deposits (stippled) that form the present divide in the valley.

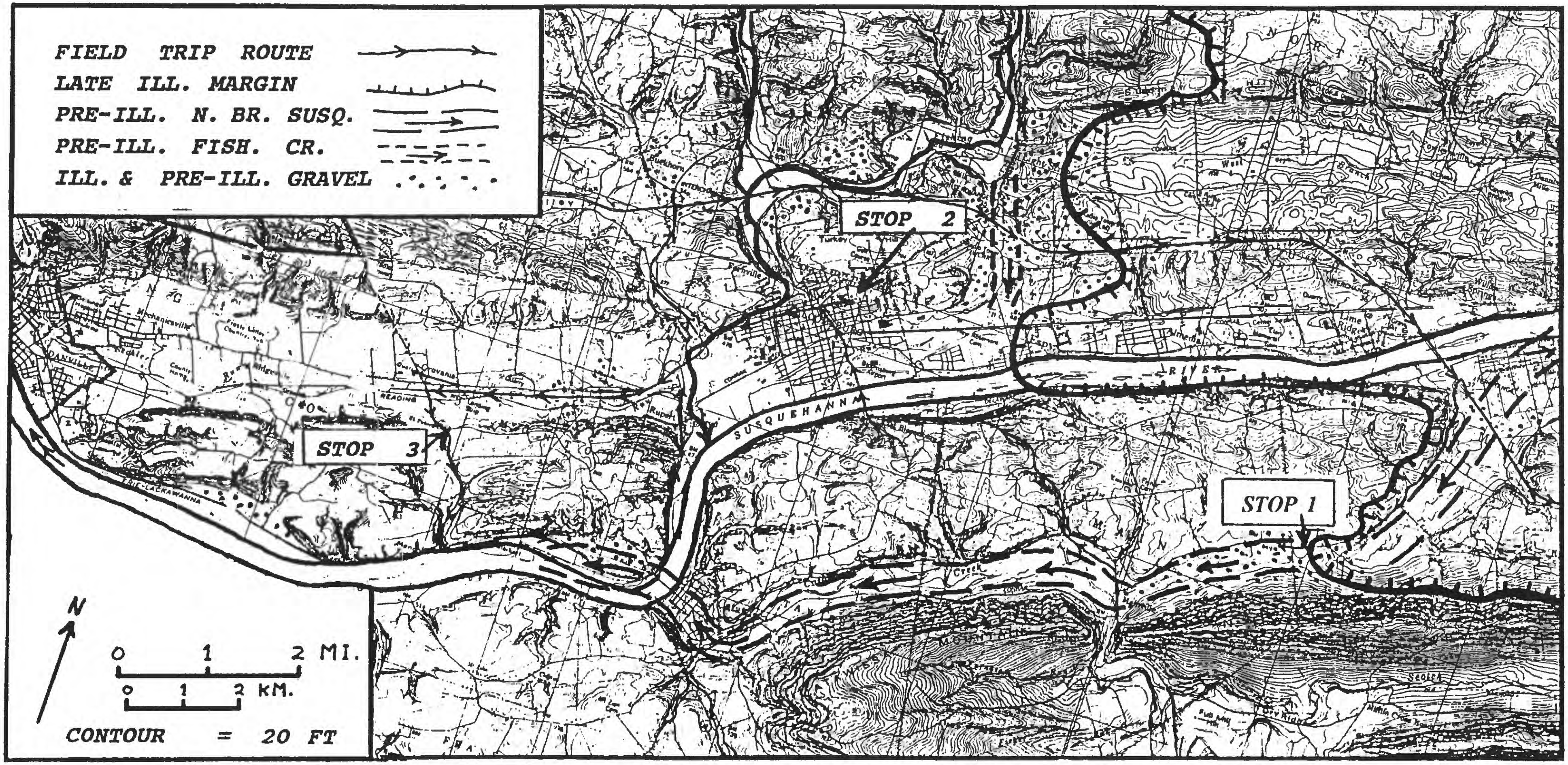


Figure 23. Topographic map of the field trip stops showing the late Illinoian terminus and associated drainage derangements. Note that the abandoned course of the Susquehanna River (right side of map) is much wider than its present course through the Rupert water gap (at the 90° turn in the river). Immediately to the west (left) of Rupert gap is a broad strike valley that provides an alternate route for hypothesized catastrophic floods. Both valleys show no evidence of such floods.

The strike valley in easily eroded shales and carbonates that the North Branch Susquehanna River presently follows provided an alternate route for the river at the end of the glaciation that deposited the kame fan, so that the river abandoned rather than reexcavated its partially filled original course (Fig. 23). Presumably a similar kame fan occupied the strike valley but at a slightly lower elevation that permitted the river a lower post-glacial drainage-way. This is the only known site where glacial deposits have forced the North Branch Susquehanna River to abandon a segment of its preglacial course.

The deposits are part of the late Illinoian terminus, the Bloomsburg ice margin. The thickness and volume of deposits are equivalent to that at the late Wisconsin terminal margin 15 km (10 mi) to the east (Stop 1- tomorrow morning). This implies a similar length of time of ice front stability and supports the contention that the deposit marks the terminal position of the late Illinoian ice.

With regard to the catastrophic flood hypothesis, this site was abandoned by the North Branch Susquehanna River upon recesssion of the late illinoian ice at about 150 Ka and at that time the Rupert water gap was first occupied by the river. The Rupert gap had previously been cut by Fishing Creek, a 100 to 200 foot wide tributary of the river. That creek presently is "aimed" right at the Rupert gap before it joins the river at the upstream entrance to the gap (Fig. 23). The Rupert gap is the smallest gap on the river because it is the "youngest" gap along the river.

This divide is at an elevation of 690 feet (210 m) , at least 100 feet lower than the hilltops to either side of the Ruper water gap, and only 65 feet (20 m) above the level of the strike valley bypass for flood water next to the Rupert gap. As will be emphasized at the next stop, the catastrophic flood cannot fit through the Rupert gap at a water depth of 100 meters (assuming a flow velocity of 10 m/s). The top elevation of that 100 meter deep flow is 800 feet (244 m), 110 feet (33.5 m) above the top of this divide. That much water should have left some scour features at and downstream of this sand and gravel divide. No such features have been observed.

The key question here is: What evidence should one look for to demonstrate catastrophic flooding?

		Continue ahead on Rt.339
0.6	22.1	To left is a house in the center of the valley whose well penetrates 150 feet of sand before reaching bedrock. A seismic refraction line was done across the valley here.
0.4	22.5	To left is the water gap of Catawissa Creek through Nescopeck Mountain (called Catawissa Mountain beyond the gap).
1.2	23.7	**TURN RIGHT** towards Bloomsburg onto the Mainville Road or Rt.2009. (Rt.339 goes left through the water gap). The road now follows the Holocene floodplain of Catawissa Creek, a larger but also underfit stream on the abandoned Susquehanna valley.
0.1	23.8	**Bear left** at Y intersection.
0.8	24.6	Leave floodplain and enter a narrow valley with Catskill Formation outcrops. Occasional cuts show shaly colluvium with a strong downslop-directed fabric.
0.4	25.0	On right outcrop of shaly colluvium with a strong downslope-directed fabric.
1.3	26.3	Crest of ridge underlain by sandstone and shale of the Trimmers Rock Formation. Start descending scarp face undercut by North Branch Susquehanna River. To right is the river and the town of Bloomsburg.
1.3	27.6	**TURN RIGHT AT** T intersection onto Rt.487 North.
0.1	27.7	Cross North Branch Susquehanna River. Rupert Water Gap to left where river disappears around a 90 degree bend as it enters the water gap. The low water line of the river here is at the 450 foot elevation.
0.2	27.9	At the Bloomsburg end of the bridge is an outwash terrace 30 feet above the river, the Binghamton terrace of Peltier (1949). The area is one foot above the "100 year" flood and two feet below the level of the 1972 Hurricane Agnes flood.
0.3	28.2	Cross swale marking old channel position on the Binghamton terrace.
0.2	28.4	Rise up onto the 40 to 50 foot high late Wisconsinan maximum Olean terrace.

0.1	28.5	**Straight through** traffic light continuing on Rt.487 North. **Get into left lane.**
0.1	28.6	**Straight ahead** at next traffic light. Combine with US 11 after crossing railroad tracks.
0.1	28.7	**Straight through** traffic light continuing on US 11 and Rt.487. **Stay in right lane.**
0.2	28.9	**TURN RIGHT** onto College Hill Street. (US 11 goes left; 487 straight). Carver Hall of Bloomsburg University ahead.
0.2	29.1	**TURN RIGHT AT T** in front of Carver Hall and then - **TURN LEFT** immediately before parking garage. Go up hill between campus buildings.
0.2	29.3	On left Hartline Science Center almost reaches edge of the street.
0.1	29.4	**TURN LEFT** into a small parking lot and **park**.

STOP O-2: OVERVIEW OF THE RUPERT WATER GAP AND THE ADJACENT STRIKE VALLEY
Leader: Duane D. Braun

Stop O-2A Rooftop vantage point

Directly below (south) is the North Branch Susquehanna River valley and outwash terraces that the trip crossed to reach this site (Fig. 23). The ridge across the valley, underlain by the Trimmers Rock sandstone and shale dipping to the south, is steeper than normal due to the river undercutting its toe. On the skyline to the south is Catawissa Mountain, underlain by the Pocono sandstone dipping to the south. The mountain ends to the right (west) where the Pocono sandstone wraps around the nose of a eastward plunging syncline. Out of sight at the base of the mountain is the underfit Catawissa Creek valley, the abandoned North Branch Susquehanna valley the trip had followed earlier.

To the east (left) is the strike valley underlain by shale and limestone that is now occupied by the North Branch Susquehanna River (Fig. 23). The Trimmers Rock ridge rises above the strike valley and on the distant skyline is Nescopeck Mountain (Pocono sandstone ridge) where I-80 follows its toe-slope. A housing development across the river on the face of the Trimmers Rock ridge is underlain by deeply weathered outwash that marks the eroded remnants of the Bloomsburg ice margin. Further to the east (right) along the ridge, at about the second notch in the ridge, is the probable location of the pre-glacial low divide in the strike valley that separated the North Branch Susquehanna River from Fishing Creek. Pre-glacial Fishing Creek once passed through the anticlinal ridge behind us (north) in a water gap 2 km (1.3 mi) east of here that is now completely buried by the Bloomsburg ice margin deposits (Fig. 23). The Trimmers Rock ridge in the distance to the east appears to become distinctly lower. That apparent change in height is where the 1.5 km (1 mi) wide abandoned North Branch Susquehanna valley cuts through the ridge. The higher part of the ridge is on the near side of the abandoned valley while the lower part of the ridge is on the far side of the abandoned valley.

To the southwest, the North Branch Susquehanna River turns a 90^0 angle and enters the Rupert water gap. A rail road bridge, now abandoned, crosses the river at that point. The Rupert gap is only 0.5 km (1650 ft) wide, one-third the width of the abandoned North Branch Susquehanna valley cut in the same bedrock. The present river channel itself is 0.3-0.4 km (1000-1300 ft) wide. The narrowness of the gap is due to the fact that the gap was originally cut by 30 m (100 ft) wide Fishing Creek and has been occupied by the river only since the late Illinoian. Fishing Creek today is aimed right at the Rupert gap and joins the river at the entrance to the gap (Fig. 23).

To the west (right) of the Rupert water gap is the continuation of the Trimmers Rock ridge whose face is gentler since the river is no longer undercutting it. Beside that ridge is a lowland, the westward continuation of the strike valley that is bounded on the north by the anticlinal ridge behind us. Just out of sight is the low divide in that strike valley (the Grovania divide, the next stop) that is only 50 m (160 ft) above the floor of the Rupert gap. A few km beyond that divide at Danville, PA., the river re-enters the strike valley. The strike valley is a "straight ahead" bypass for supposed catastrophic floods at the Rupert gap (Fig. 23).

The Rupert water gap is also exceptionally shallow, a maximum of 140 m (450 feet) below the highest adjacent hilltops, as compared to other water gaps along the river. This is because the river is cutting the Trimmers Rock ridge, a lower secondary strike ridge when compared to the higher strike ridges like nearby Catawissa Mountain. Due to this combination of a low strike ridge and a uniquely youthful age, the Rupert water gap has less than one-half the cross-sectional area (0.74×10^5 m^2) than any water gap upriver of it. Upriver of this site the cross-sectional area of five water gaps in the Valley and Ridge and seven valley constrictions in the Appalachian Plateau up to the New York State line were measured. The average cross-sectional area of those sites is 3.1×10^5 m^2 with a range from 1.71 to 5.22×10^5 m^2. Thus the "undersized" Rupert water gap is the critical "choke point" for hypothesized catastrophic floods and should show dramatic evidence of scabland-like erosional features. Also the low level bypass for the flood water, the continuation of the strike valley to the west that the river leaves as it enters the Rupert gap, should show significant scabland-like features.

A "conservative" catastropic flood peak discharge of 3.2×10^6 m^3/s has been estimated for this area (Braun, 1990 and introductory text in this guidebook). At an assumed 10 m^2/s velocity, only 0.74×10^6 m^3/s of the discharge would fit through the Rupert gap. The remaining 2.46×10^6 m^3/s discharge would flow down the adjacent strike valley and across the hilltops surrounding the Rupert gap. The Grovania divide in the strike valley would have had a flow depth of 50 to 100 m. The combined cross-section of the Rupert gap and the strike valley at Grovania, 1.31×10^5 m^2, is still the smallest section on the North Branch Susquehanna River and should show scabland-like erosional features. No such features have been observed. A much smaller "catastrophic" flood, 50 m deep and 0.32×10^6 m^3 discharge, would have been ponded behind the Grovania divide and deposited slackwater material. No such material has been observed. Instead of flood features, undisturbed deeply weathered pre-Illinoian glacial deposits mantle the valley floor (also noted by Leverett (1934) and Peltier (1949)), often capped either by Wisconsinan colluvium or Wisconsinan loess. There is no evidence for catastrophic flooding.

Questions here:

1. What evidence should one look for to demonstrate catastrophic flooding?
2. What evidence would negate the catastrophic flood hypothesis?
3. Why is this site not a "crucial test" of the catastrophic flood hypothesis?

Stop O-2B **Poster session** review of topics discussed at the rooftop vantage point.

	29.5	**TURN RIGHT**, exit from parking lot and head back down hill.
0.2	29.7	**TURN RIGHT** at T at lower end of hill and then - **TURN LEFT** immediately in front of Carver Hall.
0.2	29.9	**Straight through** at traffic light onto US 11 South (Main Street). The area is a late Illinoian outwash terrace 90 feet above present river level. Continue through Bloomsburg on US 11.
0.5	30.4	Descend onto the Holocene floodplain of the North Branch Susquehanna River (to left) and Fishing Creek (to right).
0.7	31.1	To left is the Rupert water gap.
0.1	31.2	Go under Rt.42 North bridge.
0.3	31.5	Cross Fishing Creek.
0.1	31.6	Go under Rt.42 South bridge. To right the road cut once exposed late Wisconsinan loess deposits (not slack water) over weathered late-Illinoian gravels. Now following the strike valley that was an ideal slackwater deposit site and the alternate route for flood waters once Rupert gap full enough. Trimmers Rock sandstone ridge to left, shale and limestone in the valley, and Rosehill "iron sandstone" ridge to right.
1.6	33.2	To either side in over-grown road cuts are weathered glacio-fluvial deposits of pre-Illinoian (B?) age.
1.4	34.6	**TURN LEFT** onto Grovania Drive, an abandoned asphalt plant is on right after the turn. Cross strike valley floor (catastrophic flood channel?) nearly at the divide (to right in field).
0.4	35.0	Sharp curve to left and then ascend scarp face of Trimmers Rock ridge. Road cut at curve, when ditch is cleaned out, exposes late Wisconsinan shaly colluvium underlain by pre-Illinoian drift.
0.1	35.1	**BEAR RIGHT AT Y** onto County Line Drive (St. Peters Lutheran Church on right). Immediately **TURN RIGHT** into church parking lot. Walk back down road to vantage point.

STOP O-3: THE GROVANIA DIVIDE, A LOW LEVEL BYPASS FOR HYPOTHESIZED CATASTROPHIC PLEISTOCENE FLOODING

Leader: Duane D. Braun

This stop is a vantage point overlooking a low-level drainage divide in a broad strike valley at Grovania, Pennsylvania. The North Branch Susquehanna River follows this strike valley both to the east and west of this site (Fig. 23 & 24). Just a few miles east of here the river abruptly turns and leaves the strike valley at the Rupert water-gap to cut across a rolling upland. The strike valley here is a potential alternate route for flood waters coming down the North Branch Susquehanna River.

Shaw (1989) proposed that the Susquehanna valley be examined for evidence of catastrophic discharges on the order of 10^6 m^3/s. Such discharges are hypothesized to come from subglacial floods that formed drumlins and other features in New York State and Ontario (Shaw, 1989; Shaw and Gilbert, 1990). Shaw's technique of estimating drumlin-forming discharges yield a peak discharge of 4.8×10^6 m^3/s. Several of these floods supposedly occurred from 14 to 16 Ka, when the terminus of the Wisconsinan ice sheet was near either the Binghamton or the Valley Heads ice margins.

The smallest valley cross-section along the North Branch Susquehanna, less than one-half of any other, is the Rupert water-gap near Bloomsburg. Assuming a 10 m^2/s velocity, only a 0.74×10^6 m^3/s discharge would fit through the Rupert gap. The hypothesized drumlin-forming flood would spill about 2.46×10^6 m^3/s down the strike valley below this site and across the surrounding hilltops (Braun, 1990).

The Rupert water gap is so small because the river has occupied that gap only since about 150 Ka, the late Illinoian glaciation (Braun and others, 1984; Braun, 1988). The Rupert gap originally had been cut by Fishing Creek, a tributary of the Susquehanna. Deposits at the late Illinoian terminus (Stop 1) diverted the North Branch Susquehanna down a nearby strike valley and into the Rupert gap (Fig. 23). Immediately to the west of where the Susquehanna makes a 90 degree turn to enter the Rupert gap, the strike valley continues westward to provide an alternate low level route for the river to return to its original course (Fig. 23 & 24). A low divide in the strike valley at Grovania is only 50 m above the present bed of the river in the Rupert gap and has a cross-section larger than the Rupert gap.

It is this unique combination of a extremely small constriction and an adjacent low level alternate flood route, that makes the Rupert-Grovania site the most critical test site for the catastrophic subglacial flood hypothesis (Braun, 1990). Any flood greater than 50 m deep and discharge greater than 0.32×10^6 m^3/s, would spill over the Grovania divide. The 2.46×10^6 m^3/s flood would over-top the Grovania divide by 100 m, completely fill the cross-section, and spill across the adjacent hilltops. The steep slope of the strike valley floor to the west of the divide (9.14×10^{-3}), more than ten times that of the present river, should have permitted spectacular scour features to be cut in the valley. Instead of scour features, there are undisturbed pre-Wisconsinan glacial deposits and residuum exhibiting pre-Wisconsin-aged Ultisol soil development and Wisconsinan periglacial deposits showing inceptisol development (Fig. 24). Neither Wisconsinan nor pre-Wisconsinan catastrophic floods of greater than 0.32×10^6 m^3/s discharge could have passed through the region without disturbing these deposits (Braun, 1990).

Catastrophic floods of less than 50 m depth and discharges of less than 0.32×10^6 m^3/s would have been ponded in the strike valley between the Grovania divide and the Rupert gap. This would have been an ideal site for eddy bar deposits and slack water deposits, but no such deposits have been observed. What is observed are undisturbed pre-Wisconsinan glacial deposits having ultisol development and Wisconsinan periglacial deposits and loess displaying inceptisol development (Fig. 24). Non-flood deposits extend down to Wisconsinan outwash terraces that lie 15 m above the bed of the river at the entrance to the Rupert water-gap. Coincidentally, this 15 m height is the elevation of the calculated probable maximum flood for the present river, a flood having a 2 m/s average velocity and a 1.32×10^4 m^3/s peak discharge (U.S. Army Engineer Corps, 1974). The largest historic peak discharge, a "350 year" event, is 1.05×10^4 m^3/s (U.S. Army Engineer Corps, 1974). These 10^4 scale discharges are the only ones that can be supported by the evidence at the Rupert-Grovania site (Braun, 1990). Such discharges are 100's of times smaller than the "drumlin-forming" discharges proposed by Shaw (1989).

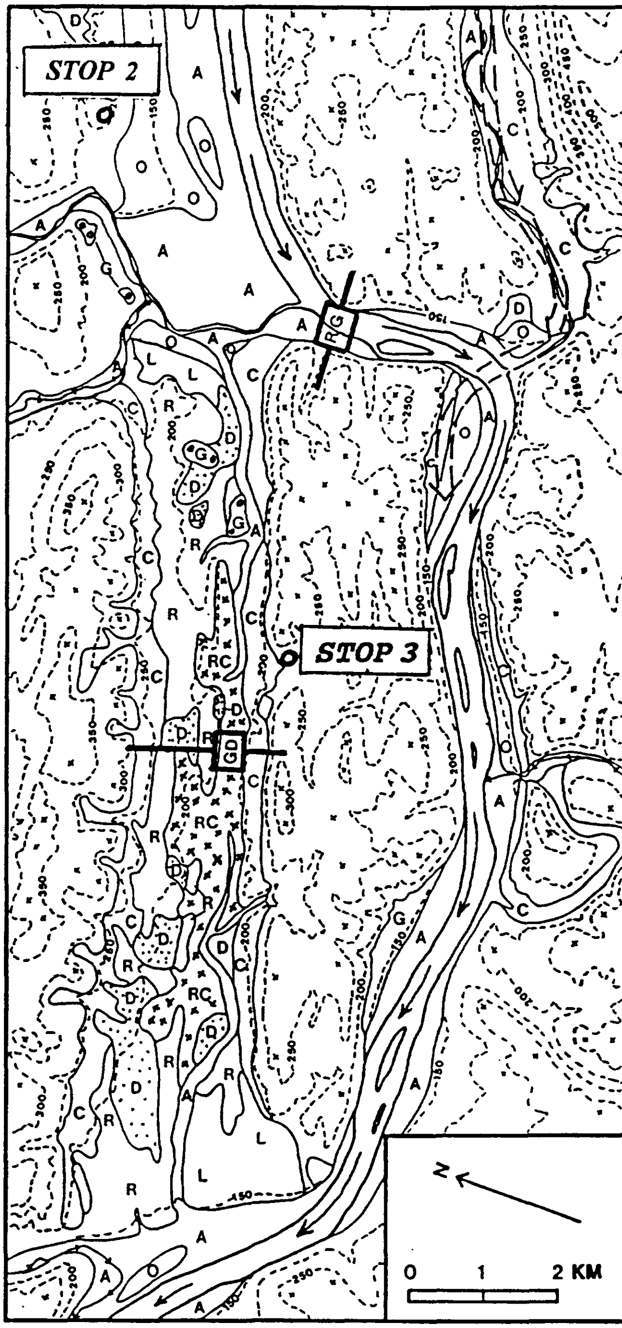


Figure 24. Surficial geology map of the North Branch Susquehanna River valley and adjacent strike valley that provides a by-pass channel for potential catastrophic floods. Pre-Wisconsinan-aged glacial and residual materials mantle much of the strike valley floor. Late Wisconsinan colluvium covers the flanks of the valley. Late Wisconsinan loess mantles either end of the valley where the North Branch Susquehanna River leaves and then again enters the strike valley. GD, Grovania divide; RG, Rupert water gap

	35.1	**TURN RIGHT** exiting from parking lot onto County Line Drive.
0.2	35.3	On left is a first order stream hollow where remnants of pre-Illinoian drift are overlain by shaly colluvium. The road will follow the valley as it narrows and deepens approaching the North Branch Susquehanna River valley.
0.9	36.2	**TURN LEFT AT T**
0.3	36.5	**TURN LEFT AT T** onto Legion Road towards Catawissa. Catskill outcrops to left, North Branch Susquehanna River on right.
1.2	37.7	To left is a gentle slope on eroded Illinoian terraces in a re-entrant cut by the river before it was diverted into the Rupert water gap.
0.6	38.3	Cross indistinct late Wisconsinan (Olean) terrace. On right are latest Wisconsinan and Holocene terraces.
0.5	38.8	On left is an abandoned gravel pit that steepens the Olean terrace riser.

0.2	39.0	**TURN LEFT AT T** onto Rt.42 North. To right is the bridge across the river to Catawissa. Enter Rupert water gap cut through the interbedded shale and sandstones of the Catskill and Trimmers Rock Formation. Other gaps through the same sequence are much wider than this one.
0.9	39.9	Start passing through the Trimmers Rock Formation.
0.5	40.4	On right is the village of Rupert where railroad yards used to handle the coal trains from the anthracite coalfields.
0.7	41.1	**BEAR RIGHT onto ramp for US 11**. **Stay in right lane,** cross Fishing Creek and
0.3	41.4	**BEAR RIGHT onto Rt.43 North to I-80**.
0.6	42.0	Follow Rt.42 through water gap in anticlinal Montour Ridge (Berwick anticline).
1.4	43.4	**BEAR RIGHT ONTO I-80 EAST RAMP.**
0.4	43.8	**MERGE ONTO I-80 EAST.** Across road, gully exposes deeply weathered pre-Illinoian sands and gravels.
0.5	44.3	Fishing Creek bridge. To right is the present Fishing Creek water gap.
0.8	45.1	To either side the adjacent hill tops are capped by weathered late Illinoian gravel.
0.7	45.8	At the Bloomsburg sign, look straight ahead to a flat topped hill on the near skyline. That is the uneroded frontal kame or head-of-outwash marking the late Illinoian limit, the so called Bloomsburg margin.
0.9	46.7	Enter I-80 cut through the Bloomsburg margin.
0.4	47.1	Interchange 35 for Bloomsburg, stay on I-80 East.
0.2	47.3	To left, the Fishing Creek valley is pointed right at I-80.
0.3	47.6	Cross the completely buried Fishing Creek valley. There are 60 m (200 ft) of late Illinoian sands and gravels under this site.
1.1	48.7	Hill crest where I-80 is running right down the axis of the Berwick anticline . To right is the North Branch Susquehanna River Valley, then the Trimmers Rock ridge, and the higher Pocono sandstone ridge (Nescopeck Mountain) on the skyline. To left are the Trimmers Rock ridge and then the Pocono sandstone ridge (Knob Mountain).
0.7	49.4	To right at 238 mile sign beside the road is a large Pottsville erratic next to a tree.
1.6	51.0	At bridge over side road, to left of silos on the left is late Illinoian loess, mentioned in 41st FOP Guidebook (Marchand and others, 1978, p.45). Start descent into North Branch Susquehanna River valley.
0.5	51.5	Go through interchange 36, continuing on I-80 East.
0.6	52.1	On right is a mobile home park on the Olean terrace.
0.3	52.4	Cross Susquehanna River. Low Holocene terraces on the downstream side of the far side of the bridge.
0.5	52.9	Go through interchange 37, the one we had turned off of earlier to go to Stop 1, continuing on I-80 East.
0.8	53.7	Cross recessional late Illinoian kame that partly blocked the entrance to the abandoned North Branch Susquehanna valley that lies to the right.
1.9	55.6	Have ascended back onto rolling upland on the Catskill redbeds and will now run along strike at the base of the colluvial toeslope of Nescopeck Mountain.
1.2	56.8	Pass rest area.
3.6	60.4	Cross artificially cutoff incised meander loop of Nescopeck Creek, "oxbow lake" on right.
0.7	61.1	Enter Nescopeck watergap, on right is a boulder-mantled slope.
0.8	61.9	Cross Nescopeck Creek
0.5	62.4	Enter roadcuts in Mauch Chunk redbeds.
1.2	63.6	On right are active and abandoned incised meander loops of Nescopeck Creek.
3.0	66.6	**BEAR RIGHT onto ramp for Exit 38, Rt.93.**
0.6	67.2	**TURN RIGHT onto Rt.93 South.**
0.6	67.8	Cross Nescopeck Creek again.
2.3	70.1	Start ascending Buck Mountain.
1.3	71.4	**TURN LEFT into hotel.**

FIELD TRIP ROAD LOG - DAY 1, SATURDAY

Mileage Inc	Cum	
0.0	0.0	From motel, **TURN RIGHT ONTO Rt.93**, descend the scarp face of Buck Mountain and cross late Illinoian limit within next 0.5 mile. There is no direct evidence of glaciation on this boulder colluvium covered slope; the limit is projected from sites several miles to either side of here.
1.6	1.6	Straight through traffic light. Nescopeck Mountain on the skyline is topped by the Pocono Formation. You are now crossing Conyngham valley, a strike valley underlain by Mauch Chunk redbeds. There are glacial drift patches on hill tops and in first order hollows.
1.9	3.5	Cross Nescopeck Creek. The creek often runs gray or black from underground coal mine drainage. Upstream of here, the Jeddo drainage tunnel discharges as much as 70 MPG of acidic water into the creek. The six to seven mile (10-11 km) long tunnel system under-drains all the mines under tomorrow's Stops 3, 4, and 5.
0.8	4.3	Cross under I-80, continue on Rt.93, and ascend the Nescopeck Mountain dipslope.
0.3	4.6	On both sides of road is boulder colluvium that originates at the ridge crest.
0.6	5.2	Bedrock outcrop; shallow-to-rock boulder mantle on top. Boulder mantle continues the rest of way to mountain crest.
0.5	5.7	Nescopeck Mountain crest. Descend the scarp slope with the North Branch Susquehanna valley ahead. Prominent ridge further ahead marks the other limb of the large breached anticline that the river follows in this region.
1.4	7.1	On both sides, as the slope becomes gentler, is boulder colluvium.
0.3	7.4	To either side the fields are on residuum/colluvium from the Catskill Redbeds.
0.3	7.7	Cross Hollow Road and a narrow valley that was a sluiceway for late Wisconsinan terminus. That margin runs just to the right (east) and subparallel to Rt.93 (Fig. 25). It displays an "indistinct" morainic landform across the rolling uplands underlain by the Catskill redbeds.
2.0	9.7	Descend scarp face of Trimmers Rock sandstone ridge into the lowland on the breached Berwick Anticline and the North Branch Susquehanna valley. Ahead, the first ridge across the lowland is the north limb of the anticline again underlain by the Trimmers Rock.
0.7	10.4	**TURN RIGHT** into Honey Hole Gravel Pit. Knob and kettle topography on both sides of the entrance to the pit.

STOP 1: HONEY HOLE GRAVEL PIT: THE LATE WISCONSINAN TERMINAL KAME FAN OR HEAD OF OUTWASH

Leaders: Duane D. Braun and Jon D. Inners

This site will be used as a model of the landforms and deposits that characterize the terminus of the Laurentide ice in the Valley and Ridge of eastern Pennsylvania each time the ice terminated in this region. The site is in a broad anticlinal valley underlain by a thick sequence of shale units with some thin limestone units (Inners, this guidebook). The North Branch Susquehanna River enters the valley on its northeast side and then turns and follows the south side of the valley to Bloomsburg. The valley is bounded by ridges held up by interbedded sandstone and shale units of the Trimmers Rock Formation that form the edge of a rolling upland underlain by red sandstone and shale units of the Catskill Formation. Rising above the rolling upland are the strike ridges underlain by the Pocono sandstone and conglomerate.

The Honey Hole sand and gravel operation is in the frontal kame fan of the 20 Ka late Wisconsinan terminus (Fig. 25). The overall sequence of deposits coarsens upward and becomes enriched in locally derived clasts as ice gradually approached the site (Fig. 26). Basal coaly sands

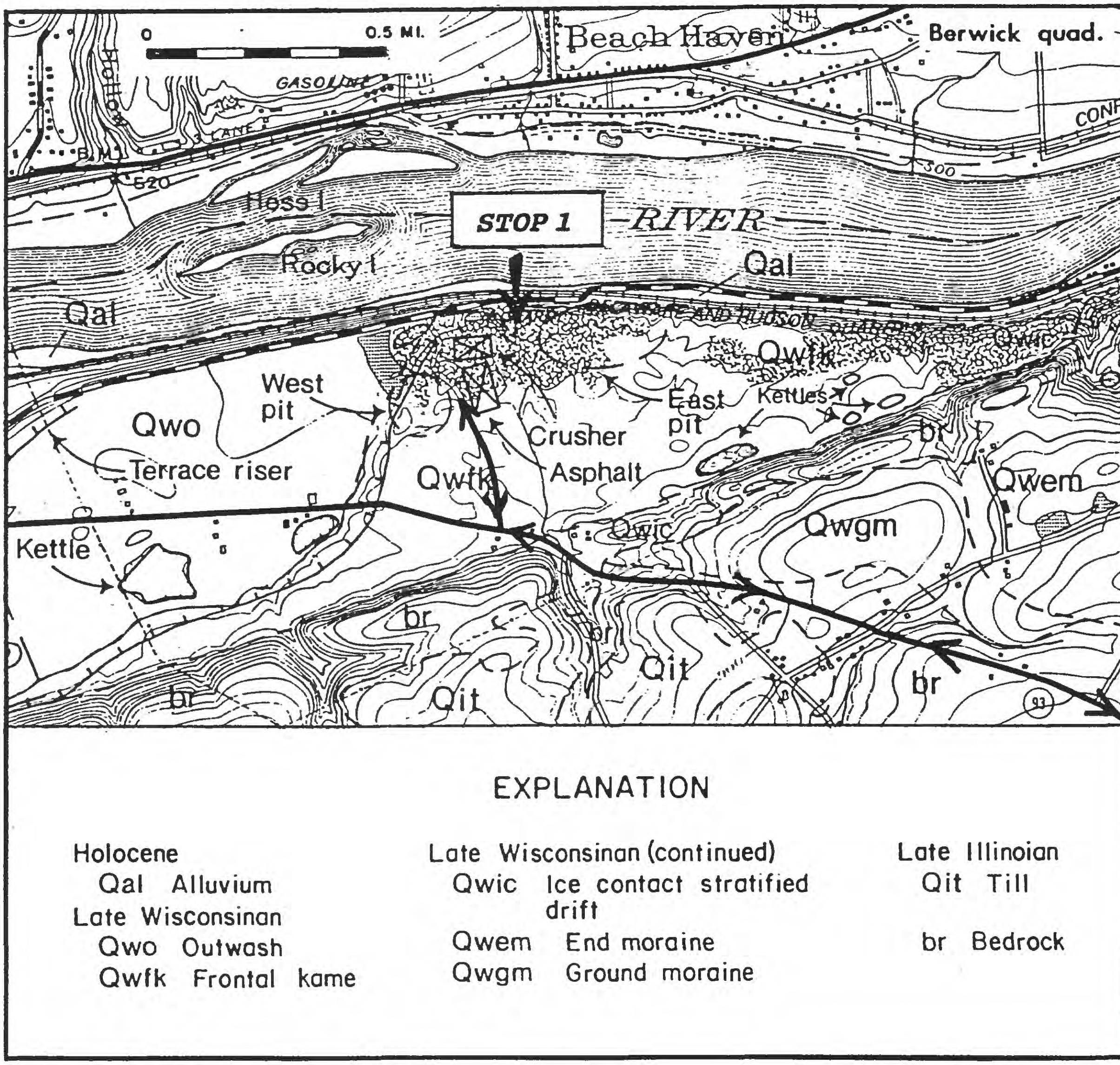
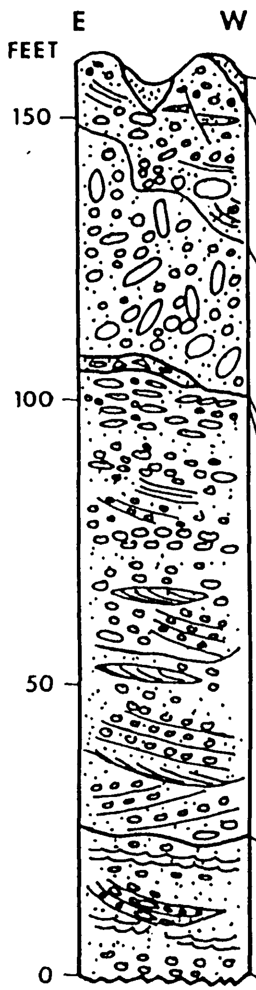


Figure 25. **Topographic and surficial geology map of the area around Stop 1 (Inners, 1978).**

are overlain by gravels that are in turn overlain by diamict and then ice contact stratified drift. Ice advanced at most about 1.0 km (0.7 mi) beyond the west edge of the site before retreating (Inners, 1978). To the southwest of the pit is a kame terrace remnant left by the farthest advance of the ice, part of Peltier's (1949) 4th Olean kame terrace (Fig. 27). The ground surface at the pit exhibits distinct knob and kettle topography with a thin mantle of loess.

Upon ice recession, the moraine was incised and the broad outwash plain at 183-189 m (600-620 ft) was formed immediately west of the pit. The 183-189 m surface is the uppermost and most extensive terrace level at and immediately downstream of the terminus,Peltier's (1949) 3rd Olean kame terrace (Fig. 27) . This suggests that it represents a period of stability in the ice front after ice had retreated a short distance to the east of the terminus. Alternately, the 183-189 m surface represents a stable outlet of the proglacial lake that developed behind the terminus as the ice retreated northeasterly (Braun, 1990).

The kame fan was gradually incised by glacial meltwater until the ice retreated north of the Valley Heads Moraine at the south end of the Finger Lakes in New York at about 14 Ka. This 35 m



Description	Interpretation
Surface landform -local relief 10'; rounded knobs separate closed depressions.	Late Wisconsinan terminal moraine.
Fine sand and silt; discontinuous, 0-2'; in places 5-10'; thick bowl-shaped masses project down from surface; in places mixed with cobbles from underlying material.	Windblown loess deposited concurrently and after underlying material; cryoturbation and surface erosion - redeposition produced thickness variation and mixing with underlying material.
Pebble, cobble and boulder gravel with lenses and some more continuous beds of coarse sands; choatic bedding often inclined 30° to 90°; abrupt changes in grain size vertically and laterally; small scale faults, displace some beds.	Ice contact stratified drift; material washed short distance from active ice out onto stagnant ice; collapse structures from melt out of buried ice.
Clast dominated, cobble to boulder, silty sand matrix diamict; gray where unoxidized; clasts dominantly rounded; many tabular clasts are on end (long axes are vertical); pockets of open framework cobble gravel project down into diamict.	Ablation "till"; debris flow deposits developed as stagnant ice releases debris; lack of distinct inclined fabric argues against lodgement; vertical clasts suggest sediment flows.
Discontinuous, 0-3' thick; compact, gray, mottled, stony, silty sand matrix, diamict; well developed fabric in places; tabular clasts dip 5 to 20° N30-70E; large lodged clasts at contact; contact sharp; iron cementation at contact in places.	Lodgement till; material plastered on bed as glacier advanced and retreated across site.
Cobble to pebble gravel, horizontally stratified; upper 10' dominated by tabular local gray sandstone clasts; next 20' dominated by >10 cm equidimensional, more travelled clasts, occasional lenses of only >10 cm clasts; lower 50' has fewer >10 cm clasts, dominantly pebble gravels; coal clasts common; gently inclined bedding marking channel scours.	Proximial kame fan - braided outwash plain (sandar) coursening upward with enrichment of local clasts as ice front approaches.
Pebbly coarse to medium sand; interbedded with pebble gravel in upper 10'; pebble to cobble cross-bedded coal lenses; sand rippled and festoon cross-bedded throughout; 5 to 10' exposed presently, at least 30' exposed when pit is dry.	Distal kame fan, braided sandy outwash plain.

Figure 26. Stratigraphic column of the deposits at the late Wisconsinan terminus at the Honey Hole gravel pit.

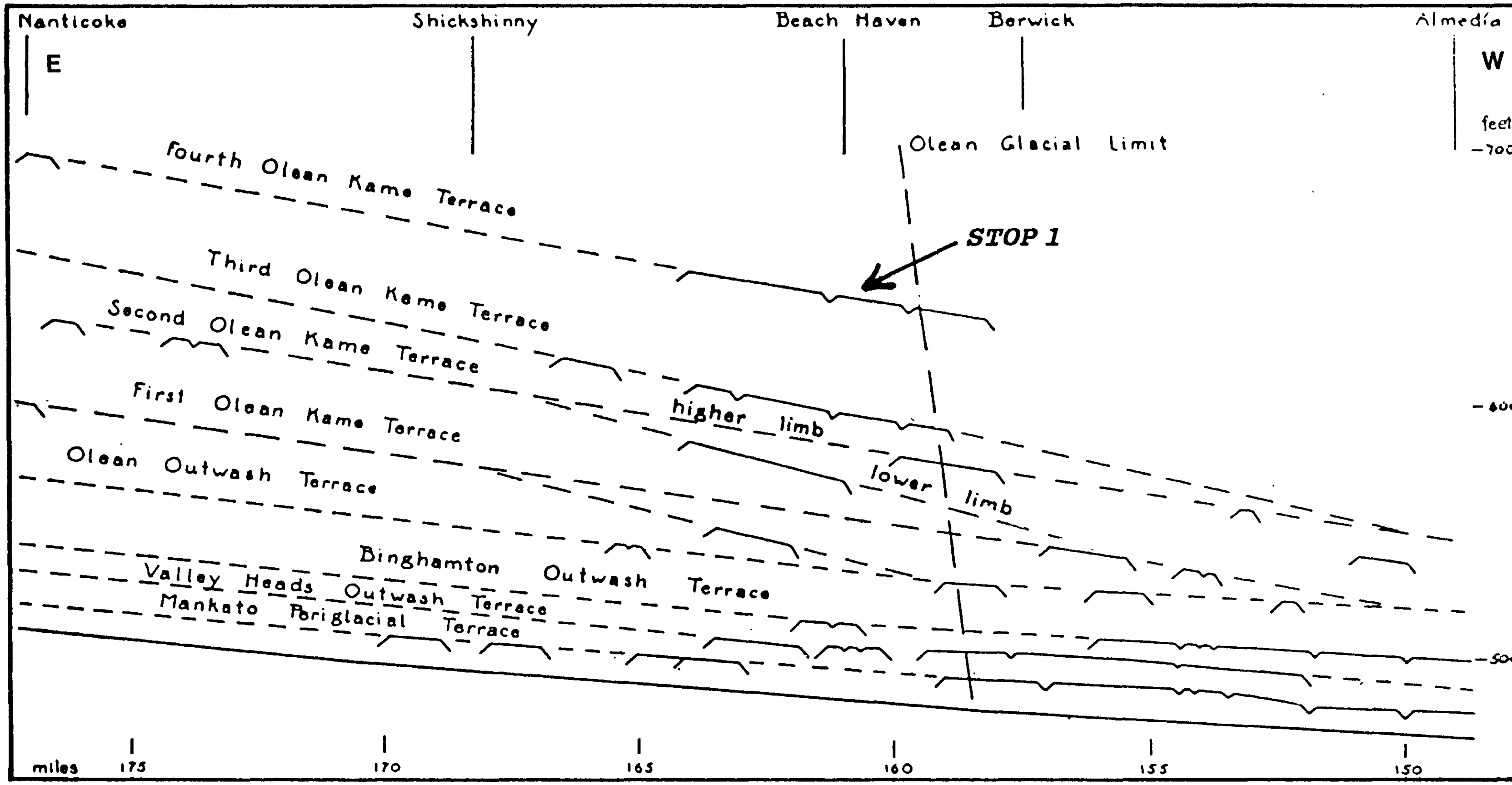


Figure 27. Terrace profiles at the late Wisconsinan (Olean) glacial limit (Peltier, 1949, Fig. 33). The terrace locations shown are for the right (north) bank of the North Branch Susquehanna River.

(115 ft) incision of the river was accentuated by a progressive increase in meltwater discharge caused by a continuous increase in the length of ice front drained as the ice receded to the head of the basin. There was also a progressive increase in non-glacial drainage and a series of glacial lake drainage events that would have assisted the incision of the channel. Terraces only 6 m (20 ft) above the present North Branch Susquehanna channel, at about 152 m (500 ft), have been correlated with the Valley Heads Moraine (Fig. 27) (Peltier, 1949) From 14 Ka to present the river has cut down only 6 m to about the 146 m (480 ft) elevation and is now incising bedrock.

This site should have been covered by ten's of meters of flood water from John Shaw's hypothesized sub-glacial flood that supposedly formed the drumlins north of the Finger Lakes. As of yet, no evidence of catastrophic flood features have been found at this site. Could the knob and kettle topography at the top of the pit actually be giant ripples ?! Is the ice-contact-stratified drift actually catastrophic flood deposits?? Please examine the top of the pit with these alternate interpretations in mind.

This site also illustrates the degree of weathering and soil development produced by a single interglacial stage. Clast surfaces are generally unweathered to slightly weathered with the development of thin weathering rinds. A few clasts are weathered throughout and are probably recycled from older glacial materials eroded by the last ice advance. The Chenango gravelly loam soil series is mapped on the gently sloping kame surfaces and the Wyoming gravelly loam series is mapped on the more steeply sloping terrace risers. Both soils are classified as Inceptisols, Typic Dystrochrepts, loamy-skeletal, mixed, mesic. The solum (A and B horizons) thickness is generally 2 to 2.5 feet. The B horizon color ranges from 7.5YR to 5 Y , depending on parent material color, and has a matrix texture of silt loam grading downward to sand. The near surface silt enrichment is from a loess cap that is well exposed at this site.

Questions at this site:
1. What other features could be added to the descriptions?
2. Are there additional units within the already described units?
3. What other interpretations might be plausible for the origin of each of the units, especially a catastrophic flood origin for the upper most unit?

		Coffee and donuts at vans before leaving. Restart road log.
	0.0	**TURN LEFT** onto Rt.93 and travel back upslope on the scarp face of Trimmers Rock sandstone ridge.
0.2	0.2	Farm on left is on the highest kame of late Wisconsinan terminus.
0.4	0.6	On right near crest of ridge are a sandstone/shale outcrop and conglomerate erratics from late Illinoian glaciation. For the next two miles you will again cross rolling uplands on Trimmers Rock and then Catskill. Late Wisconsinan margin in forests to left (east); patches of late Illinioan weathered and erosional drift on hilltops to either side of road.
2.0	2.6	Cross late Wisconsinan sluiceway and start climbing toeslope of Nescopeck Mountain scarp face.
0.5	3.1	Where forest starts is boulder colluvium.
1.0	4.1	At curve to right, road cut in Catskill redbeds. After that thin-to-bedrock boulder mantle covers surface to the crest of the ridge.
0.6	4.7	Crest of Nescopeck Mountain with view ahead and to right of the Conyngham valley and then Buck Mountain. The notch in Buck Mountain is where the Hampton Inn is located. On the skyline beyond the notch is a conical waste pile that is near today's last stop.
0.1	4.8	On left for the next 0.2 miles are large boulders at the top of the road cut that are in transit downslope. Are they moving under the present climate conditions?
0.8	5.6	Curve to right, to both sides, boulder colluvium in cleared areas.

0.4	6.0	Go under I-80. **BEAR RIGHT ONTO I-80 EAST RAMP.**
0.3	6.3	**MERGE ONTO I-80 EAST.** Run along strike with Nescopeck Mountain on left and Buck Mountain to right across the Conyngham valley.
0.6	6.9	Road cut in Mauch Chunk redbeds. Then just beyond are wooded areas to either side of I-80 for the next 1.5 miles that are covered by boulder colluvium on a very gentle slope. Land use reflects presence or absence of boulder colluvium.
2.8	9.7	Stay on I-80 West through the I-81 interchange. Ramp sign for I-81 South marks Late Wisconsinan terminus. No morainic landform but a distinct change in soil type.
0.3	10.0	Cross over I-81. Wooded slope on left where I-81 starts upslope has well developed morainic topography on the lower mountain slope.
0.5	10.5	Cut in Mauch Chunk redbeds.
0.8	11.3	Both sides, indistinct moraine for next 0.5 miles in forest.
0.6	11.9	Both sides, distinct moraine for next 0.5 miles in forest.
0.7	12.6	**BEAR RIGHT ONTO EXIT 39 RAMP** for Rt.309 South. Ahead and to right is Green Mountain, marked by a strip mine dump scar. Green Mountain marks the north rim of the Middle Eastern Anthracite coal field. A series of tors on Pottsville conglomerate are at its crest. The late Wisconsinan terminus is midway up the mountain side.
0.4	13.0	**TURN RIGHT** onto Rt.309 South to Hazleton. Cross an area essentially lacking morainic topographic expression.
1.3	14.3	Start climbing Green Mountain. Both sides show distinct moraine of the late Wisconsian terminus.
0.6	14.9	Just before curve to left and crest of hill, rise above late Wisconsinan limit.
0.2	15.1	Hill crest outcrop of Mauch Chunk redbeds. Cross a synclinal axis.
0.1	15.2	**Get into left lane.** Ahead is Buck Mountain.
0.3	15.5	**TURN LEFT** at traffic light onto RT 221 to Freeland. Proceed up hill.
0.1	15.6	On right is a large Pottsville conglomerate boulder from the crest of Green Mountain. Over the next mile or so, you will gradually ascend Green Mountain. Ahead Green Mountain wraps around to right and merges with Buck Mountain at the nose of a eastward plunging anticline.
0.4	16.0	Woods to both sides of road for the next 0.7 miles are floored by boulder colluvium.
0.4	16.4	On left is an outcrop of boulder colluvium.
0.5	16.9	On left is an outcrop of conglomerate in the upper Mauch Chunk Formation.
0.4	17.3	**Get into left lane to prepare for left turn.**
0.1	17.4	**TURN LEFT** onto an unnamed local road . Pottsville conglomerate outcrops to either side.
0.2	17.6	Cross under large power line with view to left of the boulder-mantled surface.
0.2	17.8	Sharp dog-leg.
0.2	18.0	Sharp curve to left. Note abandoned strip mines. Cross shallow synclinal valley.
0.5	18.5	On right is coal breaker waste or culm bank.
0.2	18.7	**BEAR RIGHT** through village. On right are Culm banks behind the houses. Many abandoned homesite foundations show that this "mine mouth" village was once much larger during the peak of anthracite mining activity.
0.9	19.6	Sharp turn to right, recross synclinal valley. Freeland ahead on hilltop.
0.3	19.9	**PARK ON RIGHT SHOULDER.**

STOP 2: LATE ILLINOIAN (PRE-ILLINOIAN B?) TERMINUS

Leaders: Duane D. Braun and Robin Koeberle

This strip mine site is in the center of a shallow synclinal valley , the Upper Lehigh - Pond Creek coal basin (Fig. 19). The syncline has only two workable coal beds, the upper one of which has been stripped out to form the pit that exposes the glacial material (Fig. 29). The valley is bounded on the south by an anticlinal ridge, a geographic continuation of Buck Mountain, the ridge at the conference hotel. Where the ridge is occupied by the town of Freeland, it is underlain by the Pottsville conglomerate and the non-red conglomeratic uppermost Mauch Chunk. The ridge to the north of the valley, Green Mountain (Fig. 28), is underlain by the Pottsville and the lower-most part of the coal-bearing Llewellyn Formation. Beyond that ridge to the north is the broad Conyngham strike valley underlain by the Mauch Chunk redbeds, the valley we followed before climbing Green Mountain to this site.

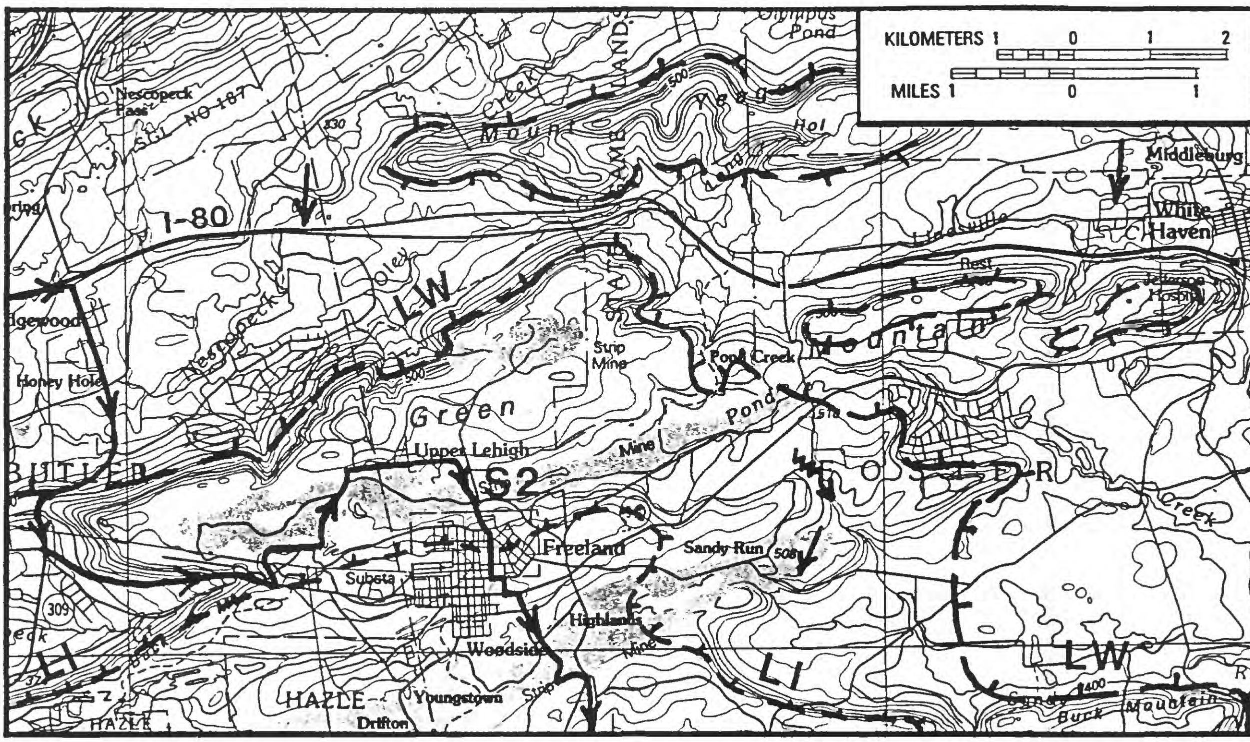


Figure 28 . Topographic map of the area surrounding Stop 2. Ice flow from the north blocked the easterly draining Pond Creek to impound a shallow proglacial lake at this site. Late Illinoian ice had to be on the order of 100-200 m thicker than late Wisconsinan ice to reach this site. LW, late Wisconsinan terminus; LI, late Illinoian terminus. Late Wisconsinan sluiceway shown by zig-zag arrow. Bedrock striations shown by short arrows. Field trip route shown by continuous line of arrows. Contour interval = 10 meters.

The northwesterly trending late Wisconsinan terminous is at the base of Green Mountain directly north of this site and surmounts Green Mountain at a lower elevation (50-100 m) about 4 km to the east of this site (Fig. 28). For ice to reach this site, it had to be on the order of 200 to 300 meters thicker than the late Wisconsinan ice. The ice first reached the site as a lobe moving westward in the east-draining strike valley, a situation that should have impounded a proglacial lake in the valley.

The late Illinoian (pre-Illinoian B ?) border has been placed near the crest of the ridge immediately south of Stop 2 at Freeland for three reasons. First, the continuity and thickness of the deposits in the valley are similar to other sites along the late Illinoian Bloomsburg margin and is

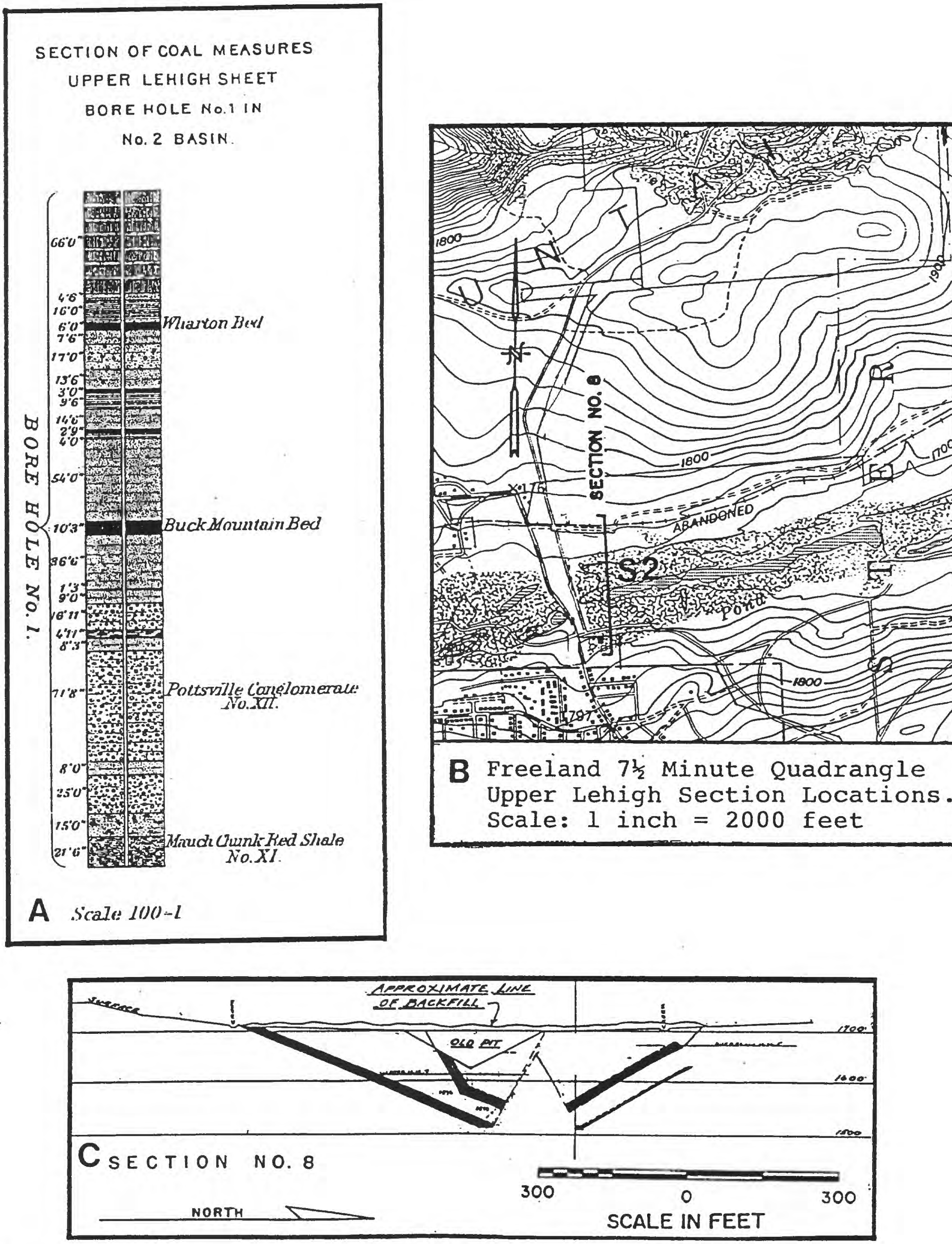


Figure 29. Bedrock geology of the Upper Lehigh - Pond Creek syncline at Stop 2.
A. Stratigraphic column of the Llewellyn Formation under the site.
B. Map showing the location of bedrock cross section.
C. Bedrock cross section at the eastern end of the strip pit.

is distinctly greater than at other coal basins to the south of this site (Stop 4 and 5). Second, there are remnant glacial deposits near the crest of Green Mountain, a situation that occasionally occurs within the Bloomsburg margin but not outside it. On the ridge to the south at Freeland, no such glacial deposits or colluvium derived from such deposits have been observed, as is the case for all other ridge crests further south in Middle Eastern Anthracite field. Third, placing a border at this site correlates well with the overall trend of the Bloomsburg margin both east and west of this area.

Stop 2A Overview of pit and north face outcrop

From this vantage point, during times of low strip mine pool elevation, one can view 5 to 10 meters of steeply dipping bedrock overlain by about 20 meters of surficial deposits that are in turn overlain by 5 to 15 meters of waste dump material. The mine pool has no surface outlet and has relatively constant subsurface discharge through mines and waste material at the eastern end of the pool. The pool level fluctuates seasonally, reaching a high in April or May and a low in September or October. If the mine pool is exceptionally high, as it may well be during the field trip, the bedrock and the lower 5 to 10 meters of surficial deposits may be flooded. If the water level has receded from its seasonal high, one or more small scale hanging deltas will be present at the base of the gully cut in the opposite face slighty to the east (to right) of the vantage point.

The uppermost part of the stratigraphy has been truncated by a early phase of the strip mining that produced the second line of waste piles behind the edge of the pit. A relatively thin layer of boulder mantled colluvium derived from the Pottsville conglomerate on Green Mountain originally capped the glacial deposits that presently form the top of the in-place section. A distinct horizontal line can be seen that marks the boundary between the in-place and dumped materia.

The overall stratigraphy is shown as a series of alternating lighter and darker brown bands that dip gently westward (to left) across the outcrop face. The darker bands are the two glacial diamict units that are interpreted to represent oscillations of a single glaciation (Table 10, units 9 and 4). The lower dark band wedges out to the west end of the outcrop. The uppermost lighter brown layer (Table 10, unit 10) is the upper part of the diamict of the second glacial advance. The two lower lighter brown bands represent more freely draining sands and gravels (Table 10, units 6-8 and 1-3).

The waste piles we are standing on are a composite of all materials exposed. In places, the surface is littered with rounded Mauch Chunk erratics from the other side of Green Moutain. Striated clasts can be collected with varying degrees of weathering dependent on their original depth in the deposits. A wide variety of lithlogies from the Mauch Chunk, Pottsville, and Llewellyn Formations and the overlying surficial deposits are present on the dumps.

There are two alternate routes to the north face outcrop. For the most agile outcrop climbers, take the abandoned access road down into the west end of the pit where there are outcrops of lake silts. Then cross the stream cascading into the pit and circle the end of the pit to the eastern end of the north face where a stratigraphic column has been cleaned off. For the less agile outcrop climbers, follow the abandoned haulage road that runs parallel to the present highway to the north until a dirt track comes in from the left. There, turn right onto a foot trail that runs beside an over grown drainage ditch and follow it for a couple of hundred meters to where the drainage ditch water cascades down into the strip pit. A stratigraphic column has been cleaned off on the east side of the gully. For the least agile outcrop climbers, it would be best to stay at the vantage point, examine the material there, watch the other outcrop climbers across the way fall into the mine pool, and return directly to the vans.

Table 10 . Stratigraphic column from the northeastern end of the strip pit at Stop 2.

M	COLUMN	DESCRIPTION	INTERPRETATION
6-9	11	Reddish brown; diamict with clasts of rock, tree, and anthropogenic debris; lower contact is a truncated ground surface with the original colluvial mantle removed.	Strip mine waste pile
3	10	Reddish brown; bouldery, clayey sand to clayey silt matrix diamict; compact with tabular clasts showing slight N-NE dip; clasts are dominantly Mauch Chunk redbed material, angular to subrounded, striated, and many show some weathering, boulder concentration in lower 5 ft. ; lower contact is sharp.	Lodgement till, final readvance of late-Illinoian ice to terminus
3.5	[illegible]	Reddish yellow to reddish brown; pebbly to cobbly silty to clayey sand diamict grading downward into a pebbly to cobbly compact sand ; matrix in upper part displays a well developed near horizontal fissility or play structure, becomes blocky below; no boulders; no striated clasts; Mauch Chunk clasts common but not dominant; lower contact transitional over 2 to 3 cm.	Sub-lacustrine debris flows from advancing late Illinoian ice
1.5	8	Light brown in upper 3 ft., light brownish gray below; gravelly sand; friable; sparse redbed clasts; forms top surface of a bench; forms light brown stripe across the outcrop face; lower contact is sharp.	Sub-lacustrine outwash, initial ice readvance
1	7	Reddish brown; clayey silt; massive to laminated; at lower contact the material is draped over underlying boulders.	Proglacial lacustrine sediment
1.8	6	Reddish yellow to reddish gray with black laminae; bouldery sand to silty sand; silty material more grayish; boulder concentration near base; forms a lighter brown stripe across the outcrop face; lower contact gradational over 2 - 4 cm.	Sub-lacustrine outwash
2.7	5	Reddish brown to reddish gray; interbedded bouldery sand, sand, and silty sand; silty material more grayish with more redbed grains; cross-bedding dips SW; gray sandstone clast surfaces strongly oxidized, quartz conglomerate and redbed clast surfaces less oxidized; a few striated clasts; bedding curves under some clasts; lower contact gradational over 5 - 10 cm.	Sub-lacustrine outwash
1.5	4	Reddish brown; pebbly to cobbly, clayey sand to clayey silt matrix diamict; matrix becomes sandier upward; redbed clasts dominant near base; a few striated clasts; thickens to 18 f.t eastward 400 ft.; lower contact transitional over 2 - 5 cm.	Till from first late Illinoian advance to terminus
0.5	3	Reddish gray; clayey sand to clayey silt matrix diamict; forms a grayish band across the outcrop face; clasts are dominantly pebble sized; lower contact transitional over 5 -10 cm.	Debris flows from advancing ice
2.4	2	Reddish gray to brown; pebble gravel with sand stringers; rounded to platy clasts; horizontal strata with clast imbrication indicating flow from N-NE; redbed pebbles dominant at base, subequal to quartz and sandstone pebble upward; Black bands and patches of Fe & Mn cemenation; lower contact transitional over 2 cm.	Colluvium deposited primarily by slope wash, lower part from removal of pre-Illinoian drift
0.6	1	Light gray to pinkish gray; pebbly to cobbly, clayey silt matrix diamict; moderate red to pale red purple, angular tabular clasts from underlying bedrock; light to medium gray quartz and sandstone clasts from upslope; lower contact sharp.	Mixed colluvium and residuum of pre-Illinoian age
		Micaceous sandstone, siltstone, claystone, & coal; dips steeply to SE, upper surface weathered with widened joints.	Llewellyn Formation bedrock

Stop 2B Examination of north face outcrop

The units in the stratigraphic column are labeled on the outcrop near its eastem end and in the gully cut to the west. The trip leaders will "hang out" at either site for individual or small group discussions of the stratigraphy.

Questions at this site:
1. What other features could be added to the descriptions?
2. Are there additional units within the already described units?
3. What other interpretations might be plausible for the origin of each of the units?
4. Do the silts actually indicate the presence of a lake?
5. Is there one, two, or more glaciations represented here?
6. Are the lower gravels colluvium and could TL or cosmogenic dating techniques be of use in constraining their age?

	19.9	Reload the vans and **continue ahead** upslope to enter the town of Freeland.
0.4	20.3	**BEAR RIGHT AT Y FORK.** Do Not Enter signs on left fork. Continue uphill on Birkbeck Street.
0.3	20.6	**TURN LEFT AT T** onto Front Street.
0.1	20.7	**TURN RIGHT AT T** onto Graham Street (Shurfine Store on right).
0.1	20.8	**TURN LEFT AT T** with South Street (Rt.940). SUNOCO station on left.
0.1	20.9	**TURN RIGHT** at small sign for Eckley Miners' Village (Mike Stower Coal and Oil on right). Immediately cross over anticlinal ridge crest.
0.2	21.1	Enter strip mined area in shallow synclinal valley.
0.5	21.6	Sharp curve to right. To left is view of an active pit. Start ascending south side of synclinal valley.
0.1	21.7	Sharp curve to left with boulder colluvium on right.
0.3	22.0	On left is the active strip operation again and the coal haulage road.
0.2	22.2	Crest of anticlinal ridge; start descending into next synclinal valley and next set of strippings.
0.4	22.6	Straight ahead at T- intersection.
0.3	22.9	Coal haulage road crosses highway we are traveling on.
0.1	23.0	Valley floor "silt flat" covered by coal wash from now defunct coal breakers.
0.3	23.3	Curve to left; on right is Eckely Miners' Village.
0.8	24.1	**TURN RIGHT** into parking lot for Eckely Miners' Village Visitor Center.

STOP 3. LUNCH AT ECKLEY MINERS' VILLAGE (1 hour time limit- eat lunch first)

If it is not raining, we will eat lunch next to the Visitor's Center. If it is raining, we will continue to the right around the Visitor's Center and go down the main street (Church Street) to a Church at the other end of the village. The bottom floor of the church has been renovated to use as a lunch room. Rest rooms are available at the Visitor's Center and single one at the church. After eating, retum to the Visitor's Center for a 15 minute film and to view the exhibits. Time permitting, one can go on a self guided tour of the village.

Bedrock geology, mining history, and "The Molly Maguires" in the Eckley-Jeddo area
by: Jon D. Inners

Eckley, first named Fillmore, was founded in 1854 by Sharpe, Leisenring, and Company to supply housing for the company's Council Ridge Colliery. It was soon after renamed Eckley, after Eckley B. Coxe, a mine owner, mining engineer and philanthropist who lived in Drifton from 1869 till his death in 1895, and whose company acquired the Council Ridge mines in 1886. While other company towns founded in the area about the same time (Jeddo, Drifton, Japan, Oakdale, Highland,

and Sandy Run) can still be recognized as old "patch towns", their houses have been modernized and the overall village appearance changed considerably. Eckley, however– because it remained in private hands longer than the others and because it was singled out for special restoration efforts, as noted below– is now very near to its original appearance.

The interested reader is referred to Landis (1988a and 1988b) for further information on Eckley, its founders, and its history. Much of what follows is abstracted from these two articles and from a poster paper prepared for the 27th Annual Meeting of the Northeastern Section Geological Society of America (Inners and others, 1991).

Bedrock Geology. The Eckley-Jeddo area lies along the eastern margin of the Eastern Middle Anthracite field and includes parts of the Cross Creek, Big Black Creek, and Hazleton basins (Figure 31). Bedrock units outcropping in the area range from the late Mississippian-aged Mauch Chunk Formation to the middle to late Pennsylvanian-aged Llewellyn Formation (Figure 30). The Mauch Chunk (the "red rock") is characterized by an abundance of red and green claystone,

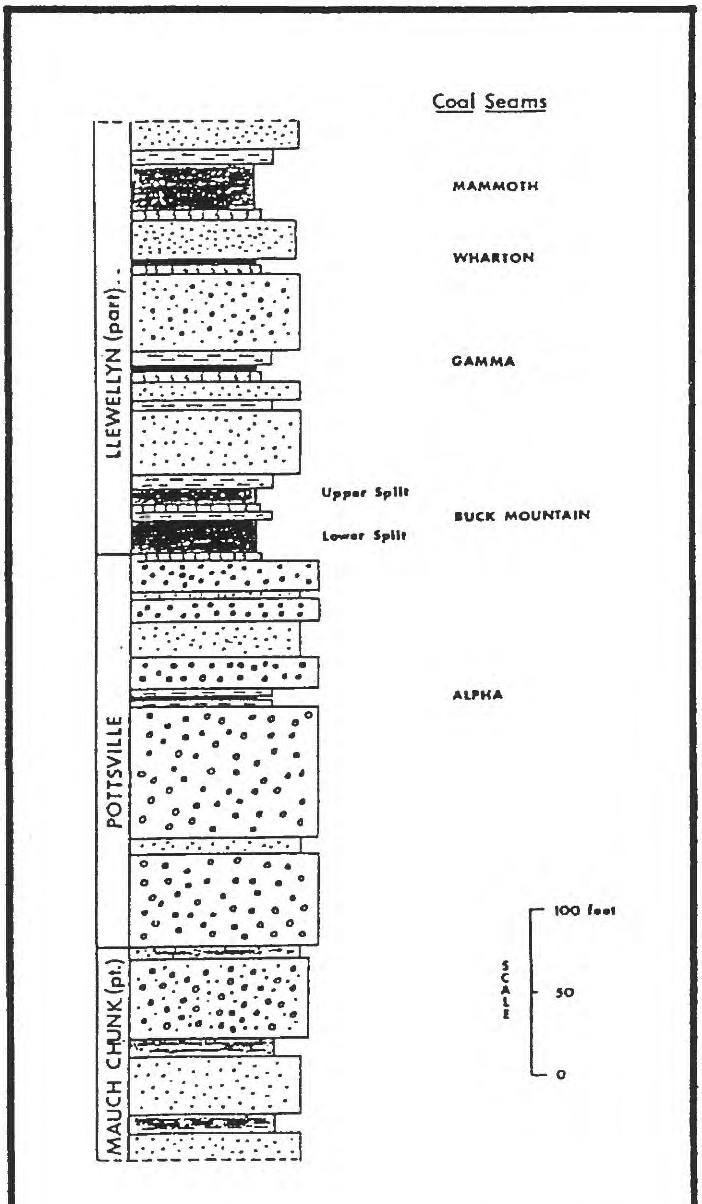


Figure 30. Generalized bedrock stratigraphic column in the Eckley-Jeddo area.

interbedded with thick orangish-weathering conglomerates in the upper part. Massive, light-gray quartz conglomerate and sandstone constitutes nearly all of the erosionally resistant Pottsville Formation. The Llewellyn Formation (the "coal measures") consists predominantly of dark-gray sandstone, siltstone, and shale interbedded with numerous seams of glassy anthracite.

The synclinal coal basins of the Eckley-Jeddo area are relatively shallow. Local maximum thicknesses of anthracite-bearing strata are approximately 500 feet in the Cross Creek basin, 600 feet in the Big Black Creek basin, and 300 feet in the Hazleton basin. Most important of the mineable coal seams are the Buck Mountain (locally more than 20 feet thick and divided into an upper and lower split at the east end of the Big Black Creek and Hazleton basins) and the Mammoth (30 feet or more thick near Drifton and Jeddo).

Structurally the basins range from v-shaped (Hazleton basin at Buck Mountain) to nearly flat-bottomed (Big Black Creek basin near Oakdale). The flatter basins are commonly complicated by localized anticlines, or "whalebacks," that may have amplitudes of 50 feet or more (Inners and Lentz, 1989). Some such anticlines may be overturned and may grade laterally into thrust faults (e.g., the Eckley overturned-anticline and thrust fault).

Anthracite in the Eckley-Jeddo area is typically hard, glassy, "white ash" coal. Fixed-carbon on a dry, ash-free basis may run as high as 98 percent. Typical heat values are in the range of 14,000 to 14,500 btu/lb. In the 19th century such coal was highly valued for use in anthracite-iron blast furnaces and as a locomotive fuel.

Mining History. Anthracite mining in this part of the Eastern Middle field began in 1839 with the opening of the "old" Buck Mountain colliery (Buck Mountain seam) in the Hazleton basin. Asa Foster engineered a gravity tramroad system that consisted of two planes and a tunnel driven 200 feet through solid rock to transport coal from the colliery to an outlet on the Lehigh Canal at Rockport, a distance of about five miles (Landis, 1988). The first shipment of Buck Mountain anthracite to market took place in November of 1840. The Buck Mountain Coal Company soon expanded operations into the east end of the Big Black Creek basin to the north.

Development of the area was slow for the next 10 years, but the pace picked up considerably in the 1850's. In 1854 Sharp, Leisenring and Company opened the Council Ridge (Eckley) colliery in the Big Black Creek basin (just west of the new Buck Mountain operations) and shipped their first coal (from the Buck Mountain seam) the next year. Three years later (1858) G. B. Markle and Company began mining of the Mammoth seam at Jeddo. (Shipment of Jeddo and Eckley coal to market was greatly facilitated by completion of the Jeddo-Hazle Brook railroad tunnel in 1859.) Within a few years the Markle Company moved into the Cross Creek basin and began mining the Buck Mountain at Highland.

The next two decades saw the opening of the Coxe Brothers' Drifton colliery (February 1865) and M. S. Kemmerer and Company's Sandy Run colliery (1876), both on the Buck Mountain seam in the Cross Creek basin.

Operation of these collieries (and ownership as well) was fairly stable over the years. Jeddo, Highland, Drifton, and Sandy Run were managed by the same companies for decades. The Eckley (Council Ridge) and Buck Mountain mines changed hands several times. (Often this involved merely a change of partnership arrangements.) For example, between 1854 and 1905, the colliery at Eckley was operated by: Sharp, Leisenring and Co. (1854-1860), Sharpe, Weiss and Co. (1860–1874), J. Leisenring and Co. (1874–1886), and Coxe Brothers Coal Co. (1886–1905). The Lehigh Valley Railroad took over from the Coxe's in 1905, but retained the Coxe name. In 1954 Jeddo-Highland assumed ownership and continued to operate the mines until 1962 when the property was sold to George Huss.

At all of these collieries the main type of underground mine opening was the slope--only a few drifts and shafts were used. Most eventually exploited multiple seams. For example, by the turn of the century the Buck Mountain, Gamma, Wharton, and Mammoth seams were mined at the Drifton colliery. Strip mining began about 1890, and by 1900 fully half of the production of the Eckley colliery came from stripping operations.

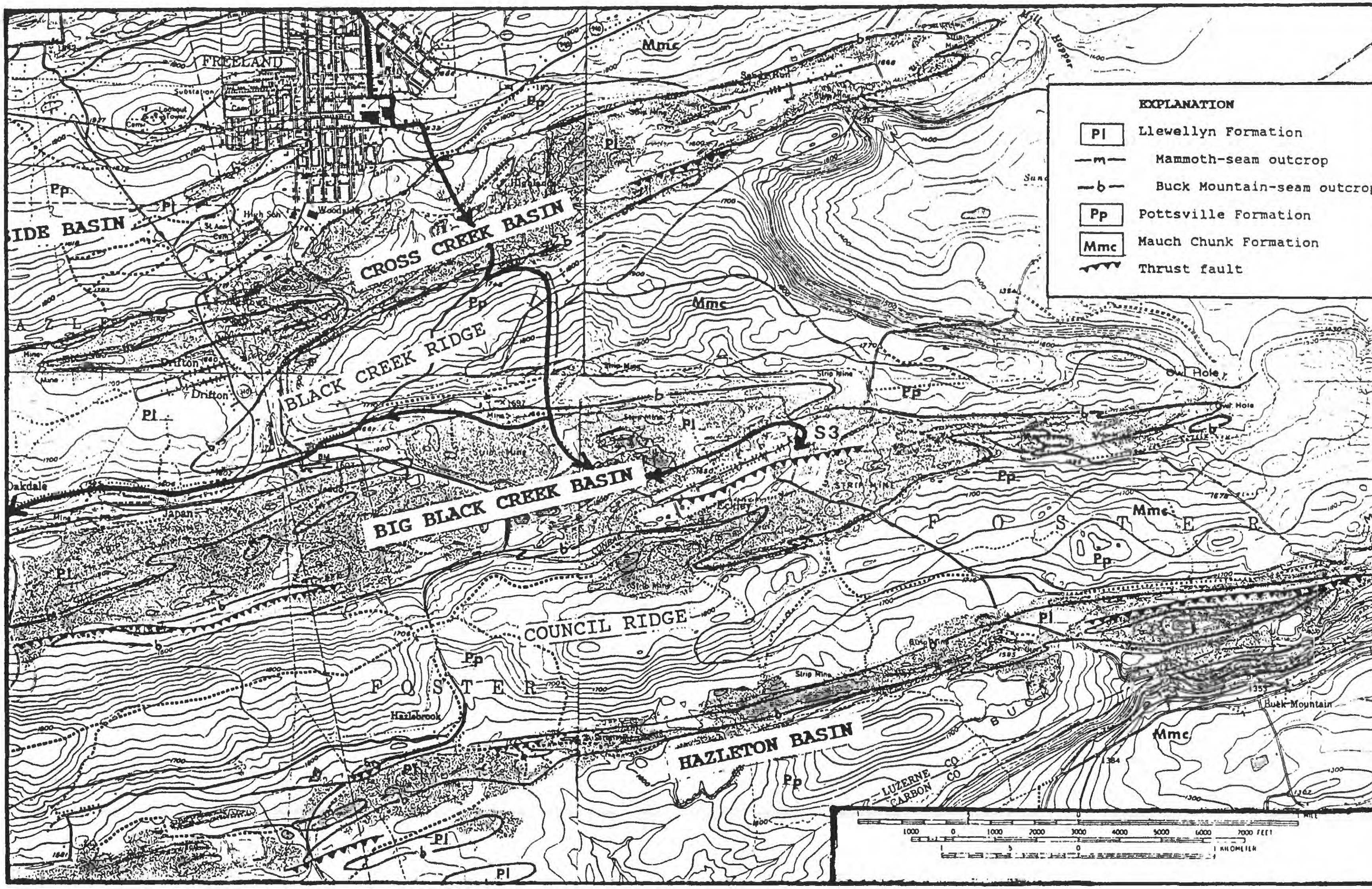


Figure 31 . Simplified bedrock geologic map of the Eckley-Jeddo area (after Inners and others, 1992). Line of arrows marks the field trip route to and from Stop 3.

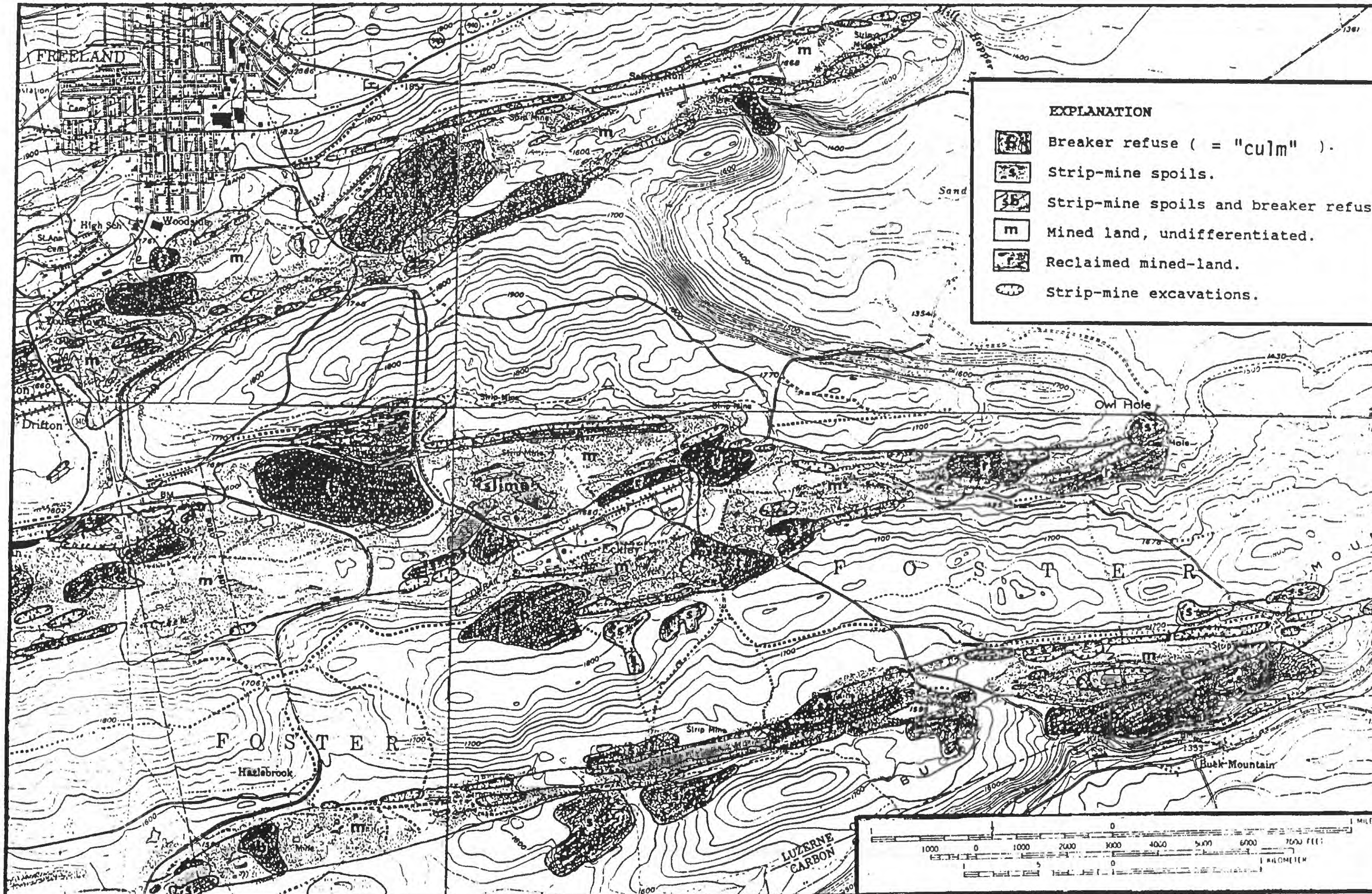


Figure 32. Surficial "geologic" map of the Eckley-Jeddo area (after Inners and others, 1992).

Production in the area peaked at approximately 2,200,000 tons in 1919 and then entered a period of "fluctuating" decline. Many slope mines closed in the 1940's and early 50's. Deep mining ceased at Eckley in 1963, and the last underground mine in the entire Eastern Middle field, the Drifton no. 1 mine of the J. F. Lee Mining Co., ceased operations in 1970. Today, the Jeddo-Highland Coal Co. (successor to the G. B. Markle Co.) has large stripping operations at Highland(s) and Oakdale; and Beltrami Enterprises striped intermittently in the vicinity of Eckley and Buck Mountain until recent bankruptcy proceedings halted operations. Total production from currently operating strip mines is probably about 500,000 tons per year. (See Inners, 1988b; Inners and Lentz, 1989.)

A century and a half of mining in the Eckley-Jeddo area has left a blasted "cultural landscape" characterized by innumerable open strip pits, barren "culm banks," featureless silt and slime "ponds," and ugly piles of stripping spoils (Fig. 32). Since most of the breakers, headframes, shafts, slopes, and fan houses are gone - victims of surface mining, reclamation, or scavenging, these unsightly "surficial deposits" are all that is left to evoke the somewhat less than halcyon days "when anthracite was king." Most striking locally is the "silt flat" - capped by irregular piles of coarser waste - that fills the valley of Black Creek north of Eckley. This deposit is about 20 feet thick in the middle of the valley (Sisler and others, 1928), and its build-up has undoubtedly necessitated raising of the Eckley-Highland road on several occasions.

The "culm" and silt deposits commonly have some economic value today as sources of fuel for cogeneration plants. During the early years of the anthracite industry, two factors combined to lead to a great amount of processing waste: fine coal sizes had no market and methods of separating coal from shale and rock were primitive. As a result, much coal, particularly the small sizes went onto the waste piles. Although nearly all "culm" and silt banks have been reprocessed at least once over the past century, some still retain sufficient heat value (generally at least 3000 btu/lb) to serve as suitable fuel in fluidized-bed cogeneration plants. Recent studies indicate that the extensive Highland No. 5 bank (located just south of the road from Jeddo to Eckley) has good potential for use as "cogen" fuel (W. E. Edmunds, 1993, personal communication).

"The Molly Maguires" and Preservation of the Village. George Huss owned Eckley when Paramount Studios chose the site to film the movie "The Molly Maguires" in 1968. Because the village houses had never been sold to the tenants, the structures had changed little, making Eckley a prime location for filming. A breaker, company store, and mule barn were reconstructed; and the houses and streets were make to look as they would have in the 1870's. At the conclusion of the filming, a group of Hazleton businessmen formed the Anthracite Historic Site and Museum, Inc., to attempt to preserve the town as an historic site. They purchased the village from George Huss and, in 1971, turned it over to the Pennsylvania Historical and Museum Commission (PHMC).

At the time of aquisition by the Commonwealth, about 250 residents lived in Eckley, some for over seventy years. Today, of that number only thirteen remain. The Commission allows only those residents who resided there in 1971 to live in village houses. Once a house is vacated, it is not reoccupied.

In 1980 the Village's Visitors' Center was opened to the public. The permanent exhibits reflect the daily and seasonal activities of the anthracite miner's family. The traditional work week of household chores is illustrated through artifacts, photos, graphics, and quotes taken from oral histories, as are the seasonal chores and the social, religious, and educational activities of patch-town residents. A miners' double house has been restored in the village to illustrate the period of 1880—1890, showing living conditions of the immigrant miner, his family and boarders,and the improvements these workers would have made in their environment as they continued to work for the coal company. Eckley is being preserved as an example of the hundreds of anthracite "patch towns" that sprang up in the region in the nineteenth century. The Miners' Village tells the story of the men, women, and children who worked to supply the most important fuel for America's mid- to late-nineteenth-century industrial expansion.

	0.0	Restart road log and leave Eckely Miners' Village parking lot.
	0.0	**TURN LEFT**, retracing route around village. The late Illinoian limit is just on the other side of ridge that forms skyline on the right.
0.9	0.9	Coal wash or "silt flat" area again. To the right are some redbed erratics on waste banks rising above the coal wash.
0.4	1.3	On right is a brownish red waste bank of pre-Illinoian till.
0.1	1.4	**TURN LEFT AT T-intersection.** On right is a reclaimed area.
0.3	1.7	Culm banks on left and on right is the remains of a coal breaker.
0.2	1.9	On right is boulder colluvium composed of large Pottsville conglomerate boulders.
0.5	2.4	Enter the borough of Jeddo.
0.1	2.5	**Bear left** continuing on main road.
0.4	2.9	**STRAIGHT AHEAD AT Y-INTERSECTION** onto RT 940. **Careful**: blind curve to right side on Rt.940.
0.7	3.6	Enter village of Oakdale.
0.5	4.1	**TURN LEFT**. Active pit ahead. Cross Railroad tracks.
0.1	4.2	**TURN LEFT** into coal haulage road to pit.

STOP 4: PRE-ILLINOIAN (B or D ?) DEPOSITS AND OVERLYING COLLUVIUM EXPOSED IN THE EBERVALE OPEN-PIT MINE OF JEDDO-HIGHLAND COAL CO.

Leaders: Robin Koeberle, Jon D. Inners, and Duane D. Braun

The Ebervale open-pit mine is situated near the middle of the 7-mile-long Big Black Creek basin (Fig. 19). The basin occupies a broad synclinal valley drained by west-flowing Black Creek and is bounded by anticlinal Black Creek Ridge on the north and anticlinal Council Ridge on the south (Fig. 34). The ridges are underlain by the Pottsville conglomerate and the conglomeratic upper Mauch Chunk. The valley is underlain by the Llewellyn Formation, a repeating sequence of sandstone, shale, and anthracite coal.

The high wall of the pit exposes interbedded dark-gray sandstones, siltstones, and "slates" between the Little Orchard (top) and Mammoth (bottom) seams. Dark-gray, hackly, rootworked and fossiliferous claystone that represent the seatrock of the Mammoth bed forms the inclined footwall along the south side of the pit. The Mamoth coal is 25 to 28 feet thick and of excellent quality. The Big Black Creek basin in the vicinity of the Ebervale mine is an asymmetrical, flat-bottom syncline with the steep limb on the south (Fig. 33).

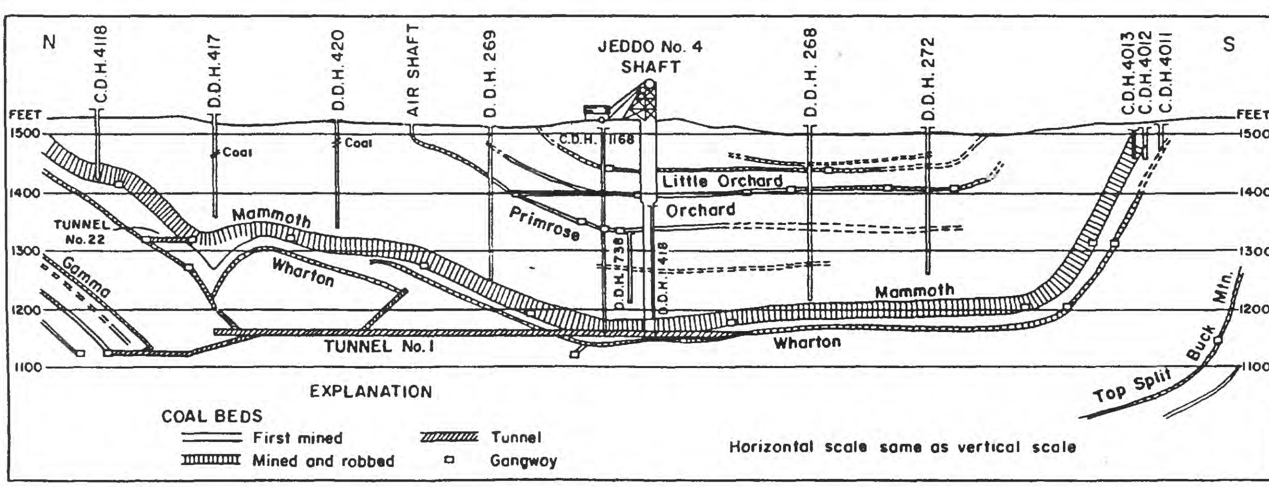


Figure 33. Geologic cross section of the Big Black Creek basin west of the present open pit.

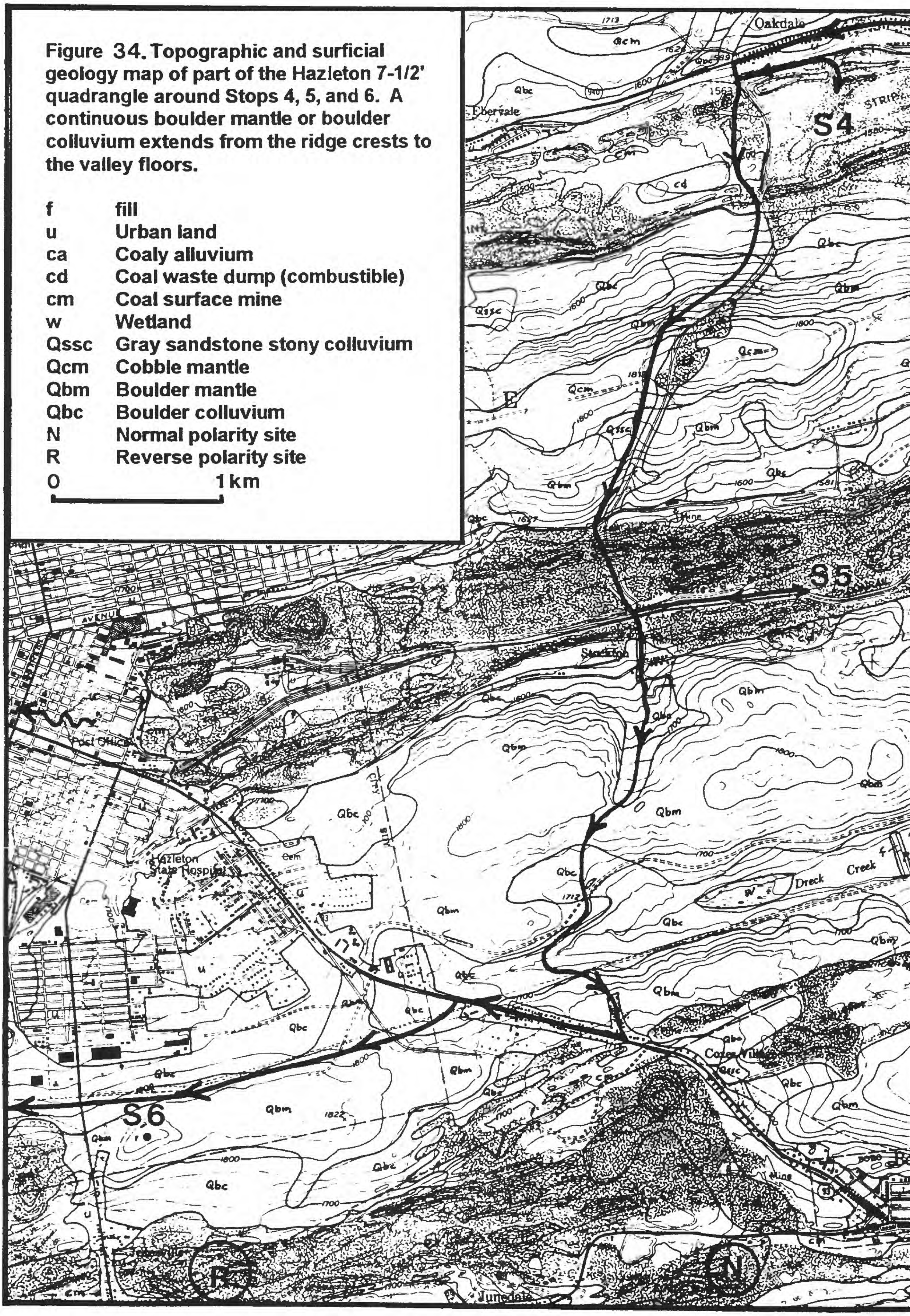


Figure 34. Topographic and surficial geology map of part of the Hazleton 7-1/2' quadrangle around Stops 4, 5, and 6. A continuous boulder mantle or boulder colluvium extends from the ridge crests to the valley floors.

f	fill
u	Urban land
ca	Coaly alluvium
cd	Coal waste dump (combustible)
cm	Coal surface mine
w	Wetland
Qssc	Gray sandstone stony colluvium
Qcm	Cobble mantle
Qbm	Boulder mantle
Qbc	Boulder colluvium
N	Normal polarity site
R	Reverse polarity site

0 1 km

Deep mining in this part of the Big Black Creek basin commenced in 1858 and continued until the late 1950's or early 1960's. Shallow surface mining probably began about the turn of the century. The current surface mine activities were initiated in 1966 after the Jeddo- Highland Co. acquired drag-lines big enough to go after the pillars of Mammoth coal remaining in the underground workings at the center of the basin. The largest dragline in the anthracite coal fields works the lower level of this pit. It is a walking dragline with a 300 foot boom length and a bucket capasity of 85-yd^3 (128 tons). A second somewhat smaller dragline makes the final lift of waste rock to the ground surface from the 300 foot deep pit. (condensed from Inners, 1988, Stop 11)

Weathered bedrock, pre-Illinoian glacial deposits, and boulder colluvium (Table 11) are usually exposed by this active strip pit. The section given below was exposed several years ago at the toe of the slope at the south side of the valley. That area today is west of the present pit and is undergoing reclamation.

Table 11 . Stratigraphic column of the Pleistocene deposits at Stop 4.

M	COLUMN	DESCRIPTION	INTERPRETATION
3		Black; sand and silt sized coal fragments; stratified; buries 1 meter in diameter hemlock stumps that project from underlying material; sharp planar lower contact.	Breaker coal waste settling pond.
2.3		Light gray to browish gray; quartz pebble, clayey sand matrix diamict; well developed spodsol soil profile; lower contact gradational over 0.25 meter.	Late Wisconsinan colluvium
6		Reddish brown to gray; prominant iron and manganese staining; cobbly, clayey to silty sand matrix diamict; few boulder clasts; Mauch Chunk redbed clasts subequal in number to non-redbed clasts; striated, rubified, and/or rounded clasts common; sharp planar lower contact.	Pre-Illinoian till
1.5		Yellow brown to gray brown; prominant iron and manganese staining; quartz pebble, clayey sand matrix diamict; many strongly rubefied sandstone clasts; a few conglomerate cobbles; lower contact gradational 2-3 centimeters.	Pre-Illinoian colluvium
1		Yellow brown to gray brown; prominant iron and manganese staining; clayey sand; a few very rubified pebbles; sharp lower boundary.	Pre-Illinoian colluvium or residuum.
10		Gray; coarse grained lithic sandstone; steeply dipping to the north; bottom of pit.	Pennsylvanian bedrock.

Questions at this site:

1. What other features could be added to the descriptions?
2. Are there additional units within the already described units?
3. What other interpretations might be plausible for the origin of each of the units?

Return to the entrance to the pit and

0.0 **TURN LEFT** upon leaving haulage road and restarting road log. Cross synclinal valley. Active pit on left; partly reclaimed area on right.

0.5 0.5 Start ascending next anticlinal ridge; boulder colluvium and then thin to bedrock boulder mantle extends nearly to crest.

0.8	1.3	At crest to left is partly hidden shale pit in upper-most Mauch Chunk green mudstone. This material produces a cobbly but not bouldery mantle at the top of the ridge here.
0.3	1.6	On right is boulder colluvium where Pottsville conglomerate starts outcropping.
0.3	1.9	Curve to left staying on main road through intersection. Immediately enter synclinal valley and strip mines.
0.2	2.1	On left reclamation in progress on valley floor.
0.1	2.2	**TURN LEFT** at Railroad tracks (just after crossing small bridge) and drive down along the left side of the tracks.

STOP 5: PRE-ILLINOIAN (B OR D ?) DEPOSITS, MULTIPLE COLLUVIAL DEPOSITS, AND PALEOSOLS AT HAZLE CREEK

Leaders: Duane D. Braun, Edward J. Ciolkosz, and Robin Koeberle

This site is located in a synclinal valley east of the city of Hazleton, the so called Hazleton anthracite coal basin (Fig. 19). This syncline is asymmetric with a gently dipping north limb and a steeply dipping south limb (Fig. 35). Hazle Creek cascades into the abandoned strip mines and exposes both the Pleistocene stratigraphy and the underlying steeply dipping bedrock of the south limb of the fold. The first thin coal seam above the thick Mammoth vein, the Holmes vein, is exposed on the north bank of Hazle Creek.

This area is a few kilometers farther southwest of the late-Illinoian (pre-Illinoian B ?) border than Stop 4 (Fig. 34). The anticlinal ridge that was crossed to reach this site has had all evidence of glaciation and a significant amount of bedrock eroded from it since it was last glaciated. This site differs from the last one in that the stream in the synclinal valley, Hazle Creek, is east draining rather than west draining and was damned by the advancing ice. The resulting proglacial lake discharged over a low col to the west at what is now downtown Hazleton at an elevation of 1620 feet (Fig. 34, sinuous arrow). At Stop 5 the lake surface was about 30 m (100 ft) above the level of the present exposure along Hazle Creek.

At this site Hazle Creek has produced one of the best exposures in the region of the a complete bedrock to ground surface section of Pleistocene deposits (Table 12). The exposure shows weathered bedrock and pre-glacial colluvium overlain by pre-Illinoian till. The till is overlain by lacustrine silts that are in turn overlain by colluvium derived from upslope pre-Illinoian till. A paleosol in that colluvium is overlain by colluvium of late Wisconsinan age derived solely from bedrock upslope. There appears to be at least one paleosol at this site (Fig. 36) developed on the "colluviated till" (Table 12, unit 5) This site was found at the end of last summer's field season and has not yet been thoroughly described or sampled. The site will be sampled for paleomagnetism determination this summer.

Warning: Hazle Creek is the main storm drain from the eastern part of the city of Hazleton. During rain events the stream will rise several feet in a few minutes, trapping the unwary on the north side of the channel. The one meter-sized boulders on the channel bed move in each major storm ! Also, as with any urban storm drain system, there is a significant component of sewage in the drainage so beware of getting into the "gray water". If the temperature is warm enough, there will be a distinct odor of sewage.

Questions at this site:

1. What other features could be added to the descriptions?
2. Are there additional units within the already described units?
3. What other interpretations might be plausible for the origin of each of the units?
4. What is the age significance of the paleosol in the colluvium above the lake silts?
5. Could the colluvium above the silts be a second in place till unit?
6. Do the silts actually indicate the presence of a lake

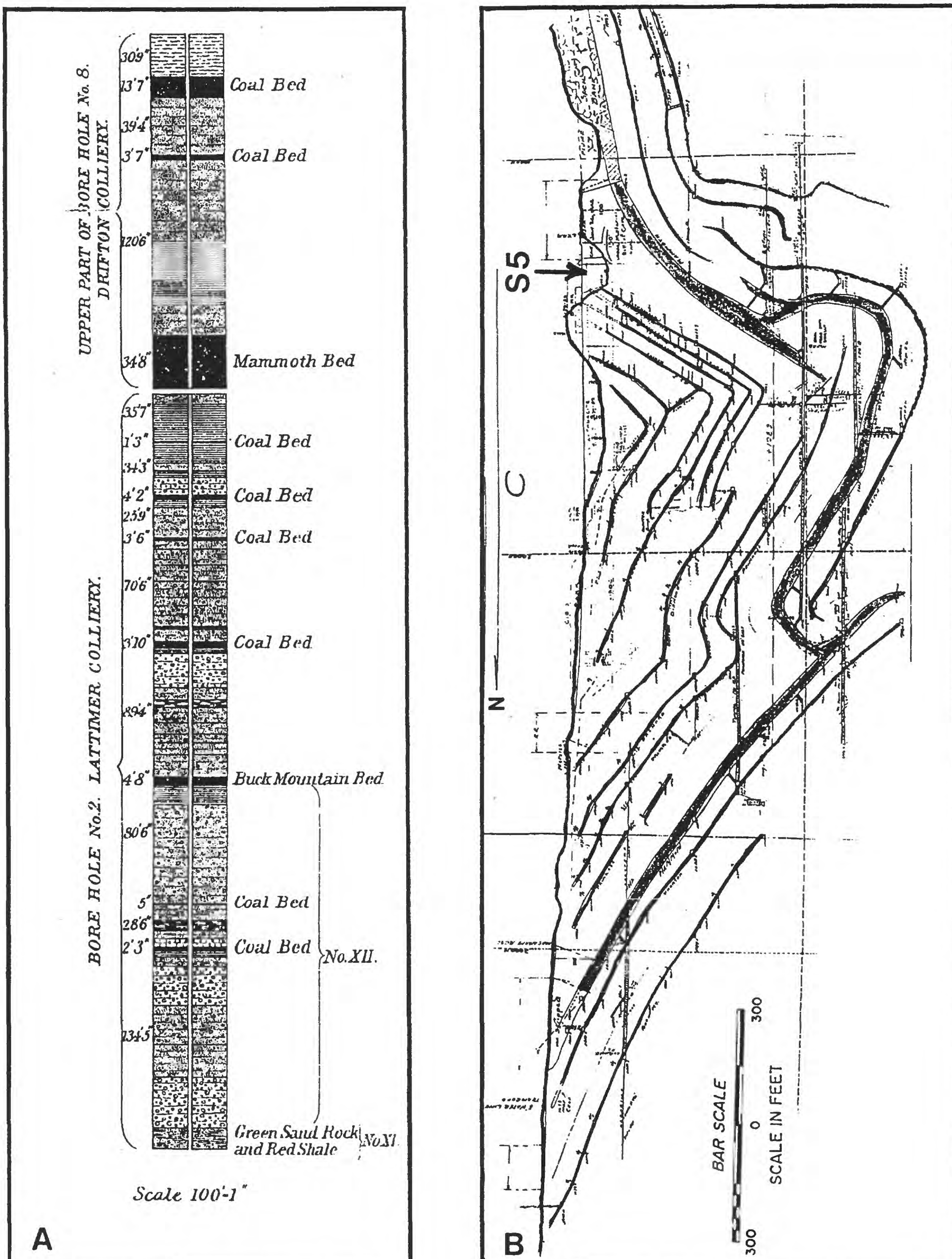


Figure 35. Bedrock geology of stops 4 and 5. **A.** Stratigraphic column of the Llewellyn Formation underneath Stop 4 and 5 **B.** Cross section of the syncline just east of Stop 5.

Table 12. Composite north and south bank section of Pleistocene deposits at the Hazle Creek site.

M	COLUMN	DESCRIPTION	INTERPRETATION
0-.5	7	Black; angular anthracite coal and coaly shale fragments; sharp and contorted lower contact.	Historic coal waste under a railroad bed, contortions from emplacement of coaly material.
.5-1	6	Gray-brown; conglomerate boulder mantled (boulders removed at this site); quartz pebble and quartz conglomerate cobble clast, clayey sand matrix diamict; distinct wavy lower contact.	Late Wisconsinan colluvium.
.5-1	5	Dark grayish red, grading downward to reddish brown; bouldery, sandy silt to clayey silt matrix diamict; some Mauch Chunk redbed clasts; weak downslope fabric;	Pre-Wisconsinan paleosol in colluvium derived from pre-Illinoian till.
0-1	4	Reddish-yellow; clayey silt and silt; stratified; sharp wavy lower contact.	Pre-Illinoian lake sed.
2-4	3	Reddish brown; bouldery, silty sand matrix diamict; Mauch Chunk red mudstone and sandstone 5 to 30 % of clasts depending on where sampled; striations on most clasts; many clasts are rubified; sharp to gradational lower contact showing inclusions of underlying material.	Pre-Illinoian till, drag and inclusion of underlying material suggests lodgement till.
0-1	2	Gray-brown; quartz pebble clast, clayey sand diamict.	Pre-Illinoian or pre-glacial colluvium.
0-1	1	Weathered sandstone, coaly shale, and coal; bright red-yellow to yellow red clayey residuum retaining the structure of the bedrock.	Pre-glacial saprolite.

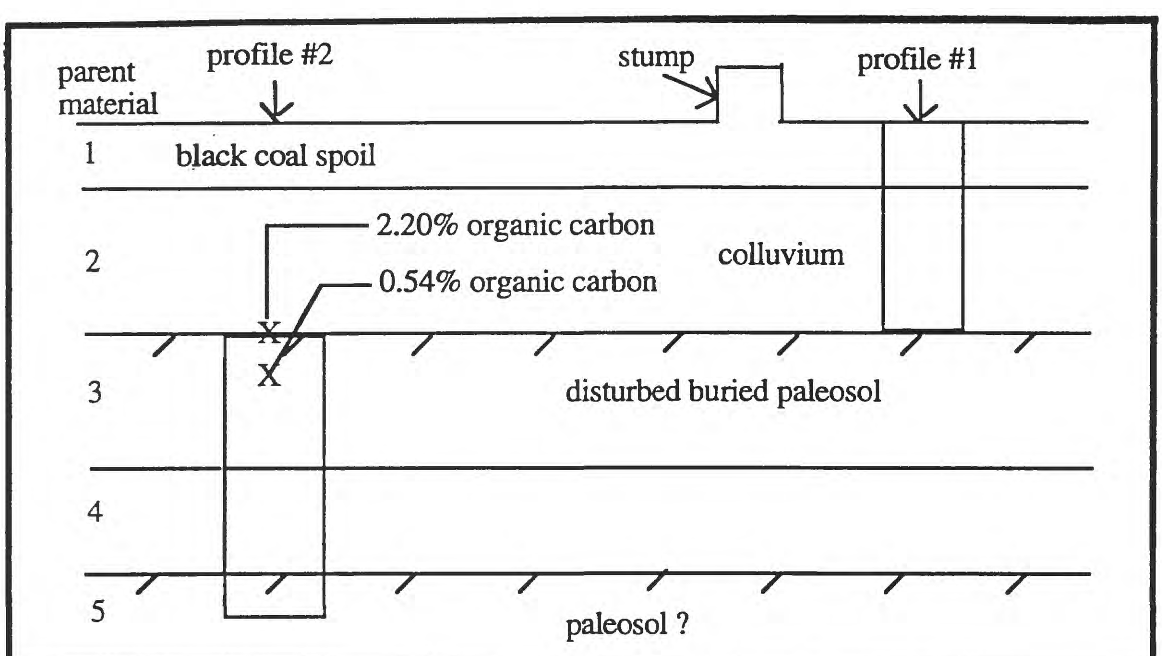


Figure 36. Schematic cross-section of the south face of the gully cut by Hazle Creek showing the basic stratigraphy and the relative location of the soil profile descriptions given below.

Surface soil and subsurface soil (paleosol) descriptions at the Hazle Creek Site
By: Edward J. Ciolkosz

Profile #1 (Stump site)

1. 0-18" Black (N 2/0) coal and shale spoil material.

2. 18-35" Mixed black coal shale (N 2/0) and pebbly yellowish brown (10YR 5/4) mottled sandy clay loam material.

3. 35-38" 2Ab Black (10YR 3/1) silt loam; friable to firm; massive; non-sticky, non-plastic; 15% pebbles; abrupt wavy boundary.

4. 38-45" 2Egb Grayish brown (10YR 5/2) and dark brown (10YR 3/3) sandy clay loam; firm; weak subangular blocky structure; slightly sticky, slightly plastic; 15% pebbles; abrupt wavy boundary.

5. 45-52" 2Btgb Grayish brown (10YR 5/2) with many brownish yellow (10YR 6/8) mottles, sandy clayloam; firm ; weak subangular blocky structure; slightly sticky, slightly plastic; few thin and moderately thick clay films in pores; 15% pebbles; clear wavy boundary.

6. 52-58" 2Bxlb Yellowish brown (10YR 5/4) with many light brownish gray (10YR 6/2) and strong brown (7.5YR 5/6) mottles, heavy loam; very firm and brittle; moderate very coarse prismatic structure which parts to very weak subangular blocky strucutre; slightly sticky, slightly plastic; few thin clay films in pores; 15% pebbles; clear wavy boundary; prism faces are 10YR 6/2 with a 10YR 5/6 iron oxide zone interior to the face.

7. 58-72" 2Bx2b Yellowish brown (10YR 5/4) with common light yellowish gray (10YR 6/2) and strong brown mottles; heavy loam; very firm and brittle; moderate very coarse prismatic structure with massive interiors; slightly sticky, slightly plastic; 15% pebbles; clear wavy boundary; prism face colors and iron oxide same as previous horizon.

8. 72-98" 2Bx3b Dark yellowish brown (10YR 4/4) all other characteristics the same as the previous horizon.

Profile #2 Abridged description of the paleosol

1. 0-5" 3Alb Black (10YR 2/1) loam; 10% rock fragments. Organic carbon content 2.20%.

2. 5-11" 3A2b Very dark grayish brown (10YR 3/2) silty clay loam; 10% rock fragments. Organic carbon content 0.57%.

3. 11-18" 3Btb Strong brown (7.5YR 4/6) silty clay loam; 20% rock fragments with some red shale.

4. 18-25" 3BCb Red (2.5YR 4/6) heavy silt loam; 20% rock fragments.

5. 25-34" 3C1b Weak red (2.5YR 4.2) heavy silt loam; 20% rock fragments.

6. 34-57" 4C2b Yellowish red (5YR 4/6) silt loam; 15% rock fragments, many red shale and some sandstone.

7. 57-63" 4C3b Yellowish brown (10YR 5/6) coarse sandy loam. Appears to be a coarse sandy layer in the dominantly silty 4th parent material. No rock fragments.

8. 63-113" 4C4b Yellowish red (5YR 4/6) silt loam; no rock fragments.

9. 113"+ 5C4b Reddish brown (2.5YR 4/4) heavy loam to silty clay loam; 40% rock fragments with many of them being very large.

Notes
1. Parent material #2 appears to be colluvium. The pebbles appear to be quartz or quartzite.
2. Parent material #3 may be till or more likely also colluvium (colluviated till?). The material has red shale rock fragments and there is no red shale upslope of this site.
3. Parent material #4 appears to be lacustrine or possibly loess. The sandy layer may be slope wash.
4. Parent material #5 appears to be an oxidized till. It has many very large rock fragments.

		Return to vans and drive back along the railroad and
	0.0	**TURN LEFT** exiting from railroad tracks and restart road log. Ascend next anticlinal ridge with its associated boulder colluvium mantle.
0.8	0.8	At crest is Pottsville conglomerate unlike previous ridge that exposed the underlying Mauch Chunk.
0.7	1.5	Curve to left across a shallow synclinal valley and then start up next anticlinal ridge.
0.3	1.8	Ridge crest with brief view ahead of culm banks and then the south edge of Eastern Middle coal field.
0.1	1.9	**TURN RIGHT** onto Rt.93. Proceed back over the same anticlinal ridge.
0.7	2.6	**TURN LEFT** on to Arthur Gardner Highway (to Rt.309). Highway runs down strike along the flank of the anticlinal ridge crest that lies to the left.
1.2	3.8	**PARK ON RIGHT SHOULDER** across from entrance to Water Tank on left. Walk across road and up access road towards the Water Tank.

STOP 6: RIDGETOP TOR, BOULDER COLLUVIUM, AND REGIONAL OVERLOOK
Leaders: Duane D. Braun and Jon D. Inners

Stop 6A Regional overlook and ridgetop tor

REGIONAL OVERVIEW - View to the northwest is shown as dashed line on Figure 3, page 6.

Immediately to the north is a shallow synclinal valley underlain by the coal-bearing Llewellyn Formation and occupied by downtown Hazleton. The next ridge to the north, occupied by residential Hazleton, is another anticlinal ridge underlain by the Pottsville conglomerate. Slightly to the northwest (left) of Hazleton is another ridge with a shallow notch in it. That ridge marks the north limb of the Eastern Middle Coal field and the limit of late Illinoian or pre-Illinoian B ice. The notch is where the I-81 & Rt.93 intersection and conference hotel is located. Beyond that and not directly in sight is the Conyngham strike vally underlain by the Mauch Chunk redbeds. That valley was filled by the late Illinoian or pre-Illinoian B glacier. The next more distant ridge is Nescopeck Mountain, underlain by the Pocono sandstone, and a feature that the field trip crossed twice this morning. On the other side of Nescopeck Mountain, the late Wisconsinan terminus crosses the North Branch Susquehanna River valley. On the far distant skyline, 40 miles (60 km) away, is the flat top of the southern edge of the Appalachian Plateau (North Mountain) where it just projected from the late Wisconsinan terminus at 20 Ka .

Immediately to the west (left) of the vantage point is a deep mine breaker refuse pile or culm bank. Immediately to the south, past the water tank, is another synclinal valley underlain by anthracite coal and covered by abandoned strip mines. The site with the reversed polarity magnetism is in the valley immediately below the water tank (Fig. 34). The site with the normal polarity material is in the same valley 2.5 km to the east (Fig. 34). The ridgetop that marks the skyline to the south is the south limb of the Eastern Middle coal field, a homoclinal ridge underlain by the Pottsville conglomerate.

RIDGETOP TOR

This tor site displays the weathering and erosion of a strike ridge that has occurred since the area was glaciated sometime in the pre-Illinoian (Fig. 37). Strip mines in adjacent valleys showing striated bedrock pavement indicate that the pre-Illinoian ice was erosive and should have removed any pre-glacial tors. The ridge is a broad anticlinal crest underlain by the Pottsville conglomerate and sandstone. A series of tors occupy the crest of the ridge where the most resistant member of the Pottsville, the Sharp Mountain conglomerate, outcrops. The tors are elongate along strike and bounded by strike and dip joint faces that are several meters apart. At the north side of the ridge crest where the intact bedrock dip should be slightly northward, the tors are back rotated into the slope and display bedding dips of 20^0 to 25^0 to the south. The top surfaces of the tors show a variety of weathering pits, pans, and runnels (Fig. 37). Downslope the tors progressively become more disrupted and form the head of the boulder colluvium that extends the rest of the way down to the valley floor under the city of Hazleton.

The volume of colluvium downslope of the tors indicates at least 10 to as much as 30 m of material has been eroded off the ridgecrest since pre-Illinoian ice covered the site. Two one-meter long cores have been taken out of the largest most highly projecting tor. The samples are currently being processed for cosmogenic isotopes. Hopefully in the near future there will be some quasi-absolute dates to constrain the length of time that these tors have been exposed at the ground surface. The younger the tors are, the more effective is the post-glacial erosional process that produces the boulder colluvium. As argued by Braun (this guidebook) the most reasonable process is gelifluction under periglacial conditions.

Tor questions:

1. Age of tor features; post pre-Illinoian, post late Illinoian, or what??
2. Rate of formation of small scale weathering features and do they form more readily under interglacial or periglacial conditions?

Figure 37 . Photograph of the Tors and the gentle slope extending down to the city of Hazleton.

Retrace route back to the highway, cross the highway, and continue downslope, bearing to the left under some electric lines, to an excavation behind a building that lies farther downslope.

Stop 6B: Boulder colluvium outcrop

To either side of the ridge are synclinal valleys underlain by the coal-bearing Llewellyn Formation. Under the axis of the valley floors are pre-Illinoian glacial deposits. A layer of colluvium mantled by large Pottsville conglomerate boulders starts at the ridge crest tors and extends down both flanks of the ridge to the valley floors (Fig. 34). Progressing downslope, the colluvium thickens over the shattered bedrock, then overlies older colluvium, and finally reaches the valley floor where it overlies both older colluvium and glacial deposits (as observed at the previous stop). The boulder colluvium is a sheet of transported material that covers essentially all slopes in this region. Between the tors at Stop 6A and the excavation at Stop 6B, large Pottsville boulders have been transported down a 0.08 (8.0% or 4.6^0) slope angle. In nearby areas such boulders have been transported to the center of the valleys, down slopes as gentle as 0.01(1.0% or 0.6^0).

From the tors to just downslope of the highway, the soil series mapped is the Dekalb extremely stony sandy loam. That series is usually underlain by bedrock at depths of less than 1 meter (3 ft.) and often at a depth of less than 0.5 m (1.5 ft.). The series is classified as an Inceptisol, Typic Dystrochrept, Loamy-skeletal, mixed, mesic.

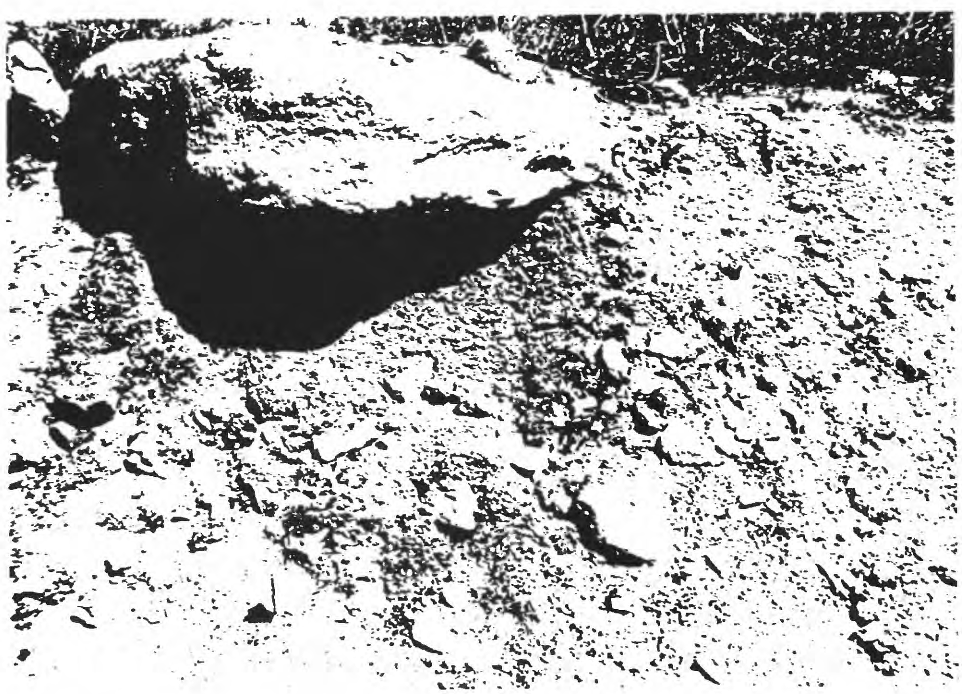


Figure 38 . Photograph of the boulder colluvium outcrop showing a large Pottsville conglomerate boulder at the surface underlain by cobble sized clasts floating in a pebbly clayey sand matrix.

The excavation is capped by large Pottsville conglomerate boulders (Fig. 38). Under the boulders is a pebbly to cobbly, clayey sand matrix diamict. The clasts are exclusively Pottsville, the pebbles are disaggregated (recycled) Pottsville conglomerate. The soil series mapped at the site is the Pocono extremely stony sandy loam. That series is usually underlain by bedrock at depths of more than 2 meters (6 ft.). It is classified as an Ultisol, a Typic Hapludult, Loamy-skeletal, mixed, mesic. The Ultisol nature of this soil is essentially because of the low base saturation of the parent material.

Both soil series at the site indicate at least several thousand years of slope stability to permit the development of the observed profiles (Ciolkosz and others, 1985). This suggests that the movement of the boulder colluvium is not taking place under present conditions- the interglacial temperate climate. The only present process that might be capable of moving such large clasts is some sort of debris flow mechanism. But the gentle slopes and the ubiquituous sheet-like nature of this material argues against the debris flow mechanism. Also, the greatest historic rainfalls in this region had absolutely no affect on these gentle boulder-mantled slopes. The only reasonable way to reduce the shear strength of this material so that it could even creep significantly is to have significant amounts of melting ground ice in a gelifluction process.

Colluvium questions:
1. Process of transport and rate of transport?
2. If not periglacial, how is the material being moved now?
3. Does cosmogenic dating of the surfaces of the largest boulders projecting from the ground surface yield a minimum date for land surface stability??

	3.8	Continue ahead. Boulder colluvium on both sides.
0.2	4.0	Straight through intersection.
0.3	4.3	Culm bank on left.
0.9	5.2	**TURN RIGHT AT T** onto Rt.309 North. Abandoned breaker ahead.
0.8	6.0	Go over another anticlinal crest and descend into synclinal valley occupied by downtown Hazleton (Stop 5 is to the left in the same valley).
0.6	6.6	Pass under railroad tracks.
0.2	6.8	**Get in left lane** after crossing another set railroad tracks at street level.
0.1	6.9	**TURN LEFT** at traffic light onto 4-lane Broad Street (Rt.93 West and Rt.924), the "main street" of Hazleton. **Get into right lane and stay there**; you will eventually be bearing right.
1.0	7.9	**TURN RIGHT** continuing on Rt.93 & 924. Either lane is all right now; there will be no more turns until the Hampton Inn.
3.1	11.0	**TURN RIGHT** into the Hampton Inn.

TRIP ROAD LOG - DAY 2, SUNDAY

Mileage Inc	Cum	
0.0	0.0	**START.** Leave Hampton Inn parking lot and at traffic light **GO STRAIGHT ACROSS RT 93.**
0.1	0.1	**TURN LEFT** onto I-81 South entrance ramp.
0.3	0.4	**MERGE** onto I-81 South. To right are abandoned strip mines in a synclinal valley that expose as much as 20 meters of pre-Illinoian till, lacustrine sediment, and colluvium.
0.4	0.8	Cross Black Creek and then cut across a small anticlinal ridge where the Pottsville conglomerate is exposed.
0.7	1.5	To right are reclaimed strip mines of the Stony Creek basin.
0.3	1.8	On right are exposures of the Pottsville conglomerate and then the lower Llewellyn Formation on the south limb of the Hazleton basin.
0.7	2.5	Interchange 40, continue on I-80 South.
0.4	2.9	Pass under Rt.924 bridge.
0.2	3.1	To right are dark colored culm banks or coal breaker waste piles.
1.1	4.2	Deep cut through the upper Mauch Chunk Formation at the axis of the Pismire Ridge anticline. Polymict conglomeratic sandstone caps the underlying reddish siltstones and sandstones.
0.6	4.8	To right are the water filled-strippings of the Honeybrook basin. Along strike to the west the bedrock is overlain by 3 meters of reddish till derived from the Mauch Chunk Formation that is in turn overlain by 3 to 5 meters of boulder covered sandy quartz pebble colluvium derived from the Pottsville Formation on the slopes.
0.4	5.2	To left is a strip mine waste bank.
0.3	5.5	Pass under coal haulage road bridge.
0.6	6.1	Enter Schuylkill County and cross the southern part of the Honeybrook basin. The area is covered by abandoned strip mines or "orphan lands".
0.6	6.7	View ahead of the Spring Mountain road cut. Spectacular exposure of a thrust fault in the upper Mauch Chunk and lowest Pottsville interval at the core of the Spring Mountain syncline.
0.4	7.1	Enter Spring Mountain cut.
0.4	7.5	To right is a view of the anticlinal Ringtown Valley rimmed by the Pottsville Formation. The pre-Illinoian(G?) limit lies near the crest of the south limb of the anticline, the skyline where the radio towers are. Beyond that crest is the Western Middle Anthracite basin where extensive strip mines expose no evidence of glaciation.
0.2	7.7	**BEAR RIGHT** onto exit ramp at Interchange 39 for Rt.309 and McAdoo. Cuts expose Mauch Chunk redbeds and conglomerate. You will be making a 180 degree turn to follow the anticlinal valley eastward and pass under I-81.
1.5	9.2	**BEAR RIGHT** onto ramp for Rt.309 south to Tamaqua.
0.2	9.4	**MERGE** onto Rt.309. To left are culm banks that mark the Silver Brook basin and a culm-burning cogeneration plant of 45-megawatt capacity. On the far skyline ahead is the ridge marking the north limb of the Southern Anthracite coal field. Start descending a steep grade to leave the Eastern Middle coal field.
1.0	10.4	Curve to left. Outcrop of Pottsville dipping northward. Ahead is a broad anticlinal valley eroded out of the Mauch Chunk Formation. Very discontinuous pre-Illinoian deposits exist on top of the broadest hilltops and on some first order hollow floors.
0.9	11.3	Continue through traffic light. Straight ahead is Broad Mountain, an anticlinal ridge underlain by the Pocono Formation.
1.1	12.4	Hill crest. On left is an over-grown road cut that exposes the anticlinal axis and the upper Pocono - lower Mauch Chunk transition.
0.6	13.0	Continue through traffic light for the Hometown Mall. On left in the forest is boulder

		colluvium from Broad Mountain.
0.7	13.7	Continue through traffic light for Rt.54. Straight ahead is Nesquehoning Mountain, the north limb of Southern Anthracite coal field, "held up" by the Pottsville conglomerate. Start descending a hill with cuts exposing the Mauch Chunk redbeds.
0.9	14.6	On left is an exposure of Mauch Chunk conglomerate.
0.2	14.8	Enter watergap through Nesquehoning Mountain cut by the Little Schuylkill River.
0.4	15.2	Bear right after crossing the Little Schuylkill River and enter Tamaqua. There are several levels of deep coal mines under the city. Continue through Tamaqua, staying on Rt.309 South.
0.5	15.7	Continue through Rt.209 intersection, staying on Rt.309. Strip mine exposures show that pre-Illinoian (G?) ice extended across this area and continued a few kilometers farther to the west.
0.4	16.1	Cross Little Schuylkill River. Its often reddish color is from oxides precipitating from the acid drainage from the abandoned deep mines. Thanks to this mine drainage, the Schuylkill river, each year, carries more tons of dissolved load than particulate load. On left is the bright reddish remains of a burned culm bank and acidic water discharging from a collapsed mine opening.
0.1	16.2	Go through the watergap in Pisgah Mountain, "held up" by the Pottsville conglomerate, the south limb of the coal field.
0.2	16.4	On left, outcrops of Mauch Chunk redbeds.
0.7	17.1	Go through the watergap in Mauch Chunk Mountain, "held up" by the Pocono sandstone and conglomerate.
0.6	17.7	On left, outcrops of Catskill Formation redbeds. On right for the next 1.5 miles are culm banks from coal breakers that no longer exist.
1.2	18.9	Continue through Rt.443 intersection, staying on Rt.309. Low ridge ahead marks the north rim of the Mahoning valley anticline and is underlain by thin sandstone and conglomerate units in the lower Catskill and Trimmers Rock Formations.
0.3	19.2	On left is outcrop of Trimmers Rock Formation.
0.4	19.6	On left is a low col that functioned as sluiceway for a proglacial lake impounded in the anticlinal Mahoning valley by pre-Illinoian ice that advanced from the northeast.
0.2	19.8	Route 443 turns left into Mahoning Valley. This breached anticlinal valley (Stop 5 later today) is underlain by shale with a veneer of fan gravel and isolated hilltop patches of glacial outwash.
0.3	20.1	To left is a view down the axis of the anticlinal Mahoning valley. On right is the anticlinal nose where the ridge wraps around the west plunging fold axis.
0.4	20.5	Start ascending ridge "held up" by Trimmers Rock and lower Catskill sandstone and conglomerate. This is the south limb of the Mahoning anticline.
0.1	20.6	On both sides are outcrops of the Trimmers Rock sandstone and shale.
0.4	21.0	At hill crest the lower Catskill Formation contains a thin conglomerate unit. On left by VFW Post 5069 is localized bouldery colluvium.
0.2	21.2	Descend slope, going down-section in the Catskill Formation.
0.8	22.0	Route 309 enters an along-strike tributary valley within the Catskill Formation outcrop belt. This synclinal area of rounded hills consists of interbedded shales and sandstones with no distinct strike ridges. There are no identifiable pre-Illinoian deposits remaining in this area.
0.6	22.6	Route 309 now follows a valley that cuts across strike through the Catskill Formation.
1.0	23.6	On left are outcrops of the Catskill Formation that dip to the north. From here to Stop 7, Rt.309 will proceed down-section from the Devonian-aged Catskill Formation to the Ordovician-aged Martinsburg Formation.
0.8	24.4	Straight through Route 895 intersection staying on Rt.309. This strike valley is in the same shale sequence as the just traversed Mahoning valley and the North Branch Susquehanna valley at Stop 1 yesterday. Ahead, the tower on the skyline marks the ridge crest "held up" by the Tuscarora Formation (Blue Mountain).

0.1	24.5	Cross Lizard Creek.
0.5	25.0	To either side the Palmerton sandstone creates low secondary ridges on the toeslope of Blue Mountain.
0.1	25.1	On right is a hollow that contains boulder colluvium from mountain ahead.
0.7	25.8	Ahead and on left is a view of the slope profile of Blue Mountain. The gentle lower slope is covered by thick boulder colluvium and the steeper upper slope is covered with a thin boulder mantle. More to the left are the rolling hills underlain by the Trimmers Rock and Catskill Formations. Immediately to the right, a few Palmerton sandstone erratics indicate the presence of pre-Illinoian ice.
0.3	26.1	Road cut through the Bloomsburg redbeds.
0.2	26.3	Cross first order hollow mantled by boulder colluvium that continues upslope.
0.5	26.8	On right is a topographic nose with shallow-to-bedrock boulder mantle.
0.1	26.9	Cross another shallow hollow mantled by boulder colluvium.
0.4	27.3	Crest of Blue Mountain, mantled by boulders. This part of the crest probably projected from the pre-Illinoian ice that left erratics on both slopes of the mountain a few hundred feet in elevation below this point.
0.2	27.5	To right is a partly obstructed view southward across the Great Valley to the Reading Prong (Blue Ridge equivalent). To left is a Tuscarora sandstone outcrop.
0.9	28.4	On both sides is boulder colluvium with particularly large boulders.
1.3	29.7	In fields on both sides are erratic-containing pre-Illinoian drift remnants overlying slate.
0.6	30.3	Cross a small stream valley and ascend into rolling hills underlain by slate of the Martinsburg Formation.
0.4	30.7	Cross a first order hollow floored with a pre-Illinoian lag of erratic clasts in a slaty silt matrix diamict, the last stage in the removal of pre-Illinoian deposits from the landscape.
0.1	30.8	Crest of hill. Straight ahead is Shochary Ridge. Erratic containing pre-Illinoian deposits stop midway up the slope.
0.2	31.0	To right on the far skyline is Hawk Mountain, a plunging anticline-syncline zig-zag of the Blue Mountain crest. Also to the right, on the intermediate skyline, rolling slate hills mark where the pre-Illinoian ice margin crossed the slate lowlands and started obliquely ascending Blue Mountain where Rt.309 descends it.
1.1	32.1	**TURN RIGHT ONTO RT. 143 south** where 309 bears left. Follow Ontelaunee Creek valley.
0.3	32.4	Decorative landscaping erratic of Tuscarora sandstone.
0.3	32.7	**TURN LEFT** and go uphill past the Blue Ridge Inn on the road to Lynnville.
0.3	33.0	On both sides of road Tuscarora erratics are commonly for the next 0.4 mile .
0.4	33.4	**TURN RIGHT** onto Zeisloft Road and curve right around a red barn.
0.1	33.5	**STRAIGHT AHEAD** at T intersection onto Oswald Road. **PARK ON RIGHT SIDE OF ROAD**

STOP 7: THE SOUTHERN LIMIT OF THE LAURENTIDE ICE

Leaders: Duane D. Braun and Jack B. Epstein

This site is on the north facing slope of Shochary Ridge with a view north across the Ontelaunee valley, the rolling hills underlain by slate, and Blue Moutain on the skyline (Fig. 39). This area is the southern edge of where erratics from the north side of Blue Mountain can be found. Examine the fields around this site and every few paces one will observe redbed, white chert, and quartz sandstone erratics. Most of the sandstone erratics are Tuscarora sandstone from Blue Mountain. The clasts are in a shaly matrix of residual and colluvial material from the underlying

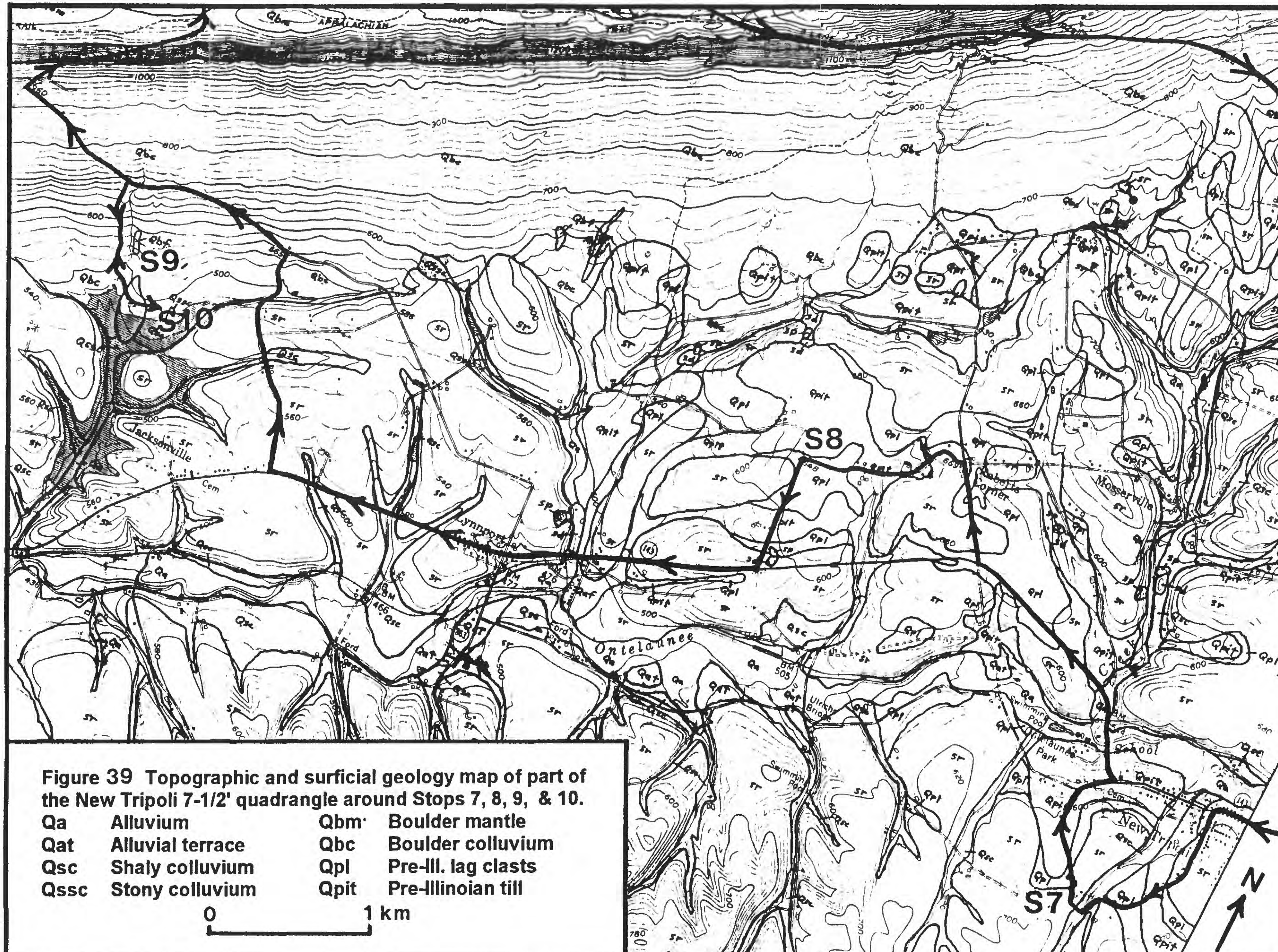


Figure 39 Topographic and surficial geology map of part of the New Tripoli 7-1/2' quadrangle around Stops 7, 8, 9, & 10.

Qa	Alluvium	Qbm	Boulder mantle
Qat	Alluvial terrace	Qbc	Boulder colluvium
Qsc	Shaly colluvium	Qpl	Pre-Ill. lag clasts
Qssc	Stony colluvium	Qpit	Pre-Illinoian till

0 —— 1 km

shale and are a relict lag of once much thicker glacial deposits (Fig. 39, Qpl). Downslope, shale fragments become less common at the surface and a colluvial mantle derived from glacial material begins and thickens to several meters at the toe of the slope. There is no soil series defined in this area for this less shaly to non-shaly "colluviated till" material, so one must criss-cross the area on foot to map the distribution of remnants of glacial material. The area of glacial material remnants trends northwesterly from here to the base of Blue Mountain (Fig. 39).

To the south of here one can walk the fields for miles and not see any erratics except for anthropogenic ones. Today, large "erratic" clasts are popular landscaping items and local property owners need to be quered as to whether or not the clast is theirs. Also Tuscarora sandstone from Blue Mountain has been used for over 150 years to build farm buildings throughout the Great Valley and care must be taken at sites where such structures have been long since demolished. The most subtle anthropogenic "erratics" are Tuscarora sandstones clasts used in stone lined field (french) drains. One must be wary of "stone stripes" in the slate and shale areas. Due to agricultural soil erosion and the use of larger mechanized equipment, some of the stone drains are now being ripped up and scattered across the fields. Many of the cobbles used in the drains and buildings have been "mined" out of the colluvium from the base Blue Mountain (Stop 9) and the clasts can show considerable weathering. For an area to be mapped as glacial (Fig. 39, Qpl & Qpit), erratics of at least two lithologic types must be found at least every few tens of paces across every acre of a field. A few Tuscarora clasts in a several acre field are almost assuredly anthropogenic.

What should be emphasized at this site is the total lack of landform associated with glacial activity. The distribution of glacial material on top of only the broadest hilltops and accumulation of colluvium derived from glacial material in first order and larger hollows (Fig. 39, Qpl & Qpit) indicates that erosion is in the final stages of completely removing evidence of glaciation from the landscape. Using the late Wisconsinan margin as a model for the thickness and type of deposits at a glacial terminus, there was once a head-of-outwash filling the Ontelaunee valley to or above the level of this site (50 m (160 ft) above the valley). In other words there has been on the order of 50 or more meters of glaciofluvial material removed from this area in post glacial times. This is expectable if the glacial limit here is as old as the 800+ Ka pre-Illinoian G advance, as suggested by the reversed magnetism sites.

Questions here:

1. How could one determine that this area had been glaciated without the erratic material from the north side of Blue Mountain?
2. How could any intact pre-Illinoian G to present weathering profile be preserved here?
3. What slope erosion processes leave behind a few lag clasts of erratic material while removing the matrix material?

		Coffee and other refreshments at the Vans before leaving.
		Continue ahead, downslope.
0.4	33.9	**TURN RIGHT** onto Allemaengel Road.
0.2	34.1	**TURN LEFT** onto Kings Highway (Rt. 143). Church across road.
0.1	34.2	Cross Ontelaunee Creek.
0.5	34.7	Ascend slope where to either side there is a lag of pre-Illinoian erratic clasts.
0.3	35.0	**TURN RIGHT** onto Gun Club Road at sign for Ontelaunee Gun Club. At intersection is a lag of pre-Illinoian erratic clasts.
0.2	35.2	Hill top where there is slate with no erratics.
0.1	35.3	On right in patch of trees are erratic boulders from the pre-Illinoian lag.
0.2	35.5	**TURN LEFT** onto Spring House Road.
0.3	35.8	To either side is pre-Illinoian drift remnants in the head of a first order hollow.
0.1	35.9	**PARK ON RIGHT SIDE OF ROAD**

STOP 8: MORAINIC LANDFORM AT THE GLACIAL LIMIT, GLACIAL OR PERIGLACIAL ?

Leaders: Duane D. Braun and Jack B. Epstein

This site is just inside the glacial limit on a broad hilltop, 0-3 % slope, underlain by slate (Fig. 39. The area is covered by several meters of glacially derived diamict and there is an abundance of erratics on the ground surface and in stone walls at the edge of the fields. Outside of the glacial limit no such stones or stone walls are observed. There are a series of shallow closed depressions, typically less than a meter deep and 10's of meters wide (Fig. 40). Some depressions hold perennial wetlands while others are farmed but are often too wet to plow or harvest. Patches of woods on these hilltops often are there because of these depressions. This area exhibits better knob and kettle or "morainic" topography than much of the late Wisconsinan terminus ! Leverett (1934) noted similar features southeast of here near Allentown and northeast of here in the central Susquehanna valley. He thought that such features were constructional and marked the Illinoian limit. This site is part of Leverett's Illinoian limit. But if this glacial limit is pre-Illinoian G (800+ Ka) in age and there has been on the order of 50+ m of erosion since then (as noted at the last stop), how can such features represent a constructional surface of a moraine?!

Figure 40 . Photograph of the seasonally water filled shallow depressions on hilltop underlain by a few meters of pre-Illinoian G drift. The depressions are 10's of meters across but less than a meter deep.

Shallow depressions in areas of non-soluble rock are actually observed both within and outside the glacial limit in this region. George Crowl and Ben Marsh have shown me similar features in the central Susquehanna region outside of the areas described by Leverett (but within the glacial limit). The common denominator with all the sites is that they are gentle slope hilltop or strike valley floor divide positions with more than 2 meters of unconsolidated material on top of bedrock. The material is usually colluvium, both outside and inside the glacial limit, and occasionally glacial drift or "colluviated drift" within the glacial border. This distribution suggests that the depressions are unrelated to glaciation and only require a significant thickness of unconsolidated material on a hilltop. A process that might form such a landform is the repeated freezing and thawing of ground ice lenses in material of variable conductivity under a periglacial environment. These depressions could be considered a type of relic patterned ground.

Questions at this site:
1. What other processes could form depressions at this site?
2. Does this material look as weathered as one would expect pre-Illinoian G material to be?
3. Have any of you observed similar features in Pennsylvania or elsewhere?

		Continue ahead and then
	36.1	**TURN LEFT** at T onto Quarry Road (gravel). To right the hollow floor is mantled with pre-Illinoian drift while slaty colluvium/residuum mantles adjacent slopes.
0.2	36.3	On left is a hedgerow that marks the start of the next pre-Illinoian drift-floored hollow that extends downslope around the farm ahead and to the right.
0.1	36.4	On left is a slate pit and dump on a slate hilltop having no pre-Illinoian deposits.
0.1	36.5	**TURN RIGHT** onto King's Highway (Rt. 143).
0.1	36.6	On left in a patch of woods in a hollow is the southwestern most remanent of pre-Illinoian drift. Southwest of here, the direction we are traveling, there is no evidence of glaciation. The area ahead is either slate residuum on the hills or slaty colluvium in the hollows.
0.7	37.3	On right, in Lynnport, an excavation exposes slate and slaty residuum .
1.1	38.4	**TURN RIGHT** onto Ontelanee Road. Jacksonville Boro line. Blue Mountain ahead.
0.2	38.6	On right is landscaping boulder or anthropogenic erratic. These are popular landscaping items in this region so care must be taken in mapping supposed glacial erratics. Also field drains composed of buried lines of sandstone cobbles produce anthropogenic erratics on poorly drained soil sites throughout the slate belt.
0.4	39.0	**STRAIGHT AHEAD** onto gravel Swamp Road at T intersection with sign for Leaser Lake. Then curve sharply right while a descending wooded slope.
0.1	39.1	Cross small stream and enter boulder colluvium area.
0.2	39.3	**TURN LEFT** at T intersection. Ahead on right are small pits dug into the boulder colluvium. Continuing ahead the road will go over several step-and-riser features a few meters high and tens of meters wide that trend obliquely across the slope. These features have been suggested to be from nival processes associated with wind oriented snow accumulation (Marsh, in Clark and others, 1992).
0.7	40.0	**TURN LEFT** onto Ontalaunee Road and descend slope across a step-and-riser feature.
0.3	40.3	**PARK ON RIGHT SIDE OF ROAD** just before a small stream crosses the road. Then walk back up the road 50 meters and go right (east) following an abandoned road way. Within 50 to 100 meters enter a series of pits dug into the toe of a boulder field.

STOP 9: BOULDER FIELD AND BOULDER COLLUVIUM EXPOSURE

Leaders: Duane D. Braun and Jack B. Epstein

This site is an unvegetated strip of boulder colluvium (boulder or block stream) on the toe slope of Blue Mountain (Fig. 39). The unvegetated portion of the block stream is about 400 feet (120 m) long in the downslope direction and 50 to 100 ft (15-30 m) wide. The slope angle here is 10 % or 5.7^0 and forested block streams farther downslope are on angles of only 2 % or 1.15^0. This is a small scale block stream as compared to others that have been described else where in Pennsylvania (Peltier, 1949; Smith and Smith, 1945; Smith, 1953; Ciolkosz, 1974; Sevon, 1967, 1975; Epstein, 1974; Clark, 1991). This one (Fig. 41 and 42) shows the same sort of surface features as at other block streams; on edge tabular blocks, crude sorting of different sized clasts into stone "circles", and a surface microtopography of pits and mounds (Clark, 1991). These features

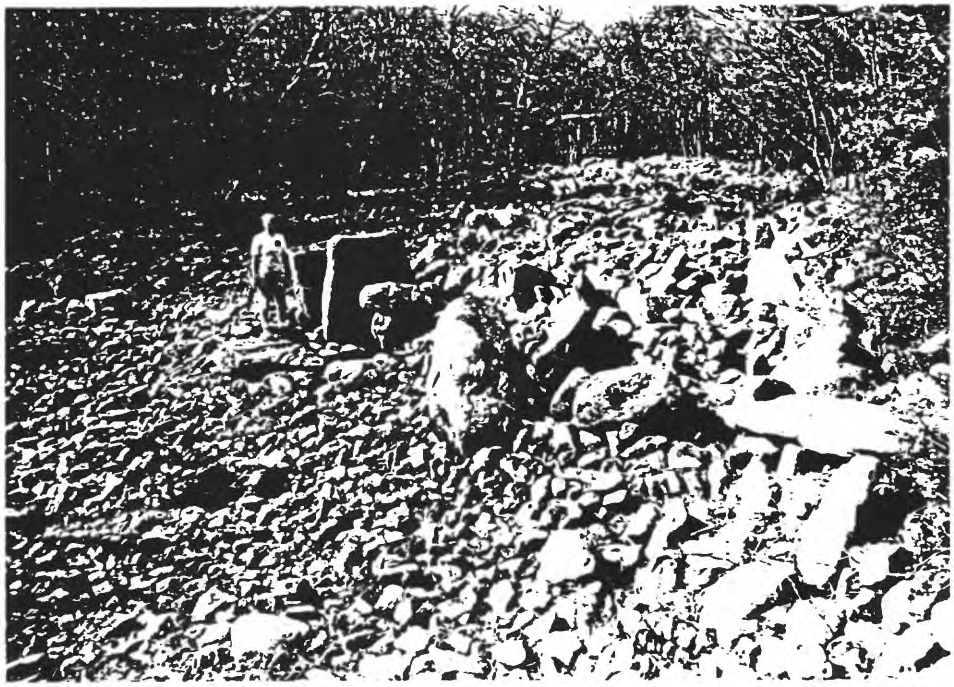


Figure 41 . Photograph of pit dug into the boulder field that shows the large on edge surface blocks underlain by smaller, more rounded clasts.

Figure 42. Photograph of the surface of the boulder field showing subtle block patterns and surface irregularities.

suggest a transport process with a strong heave component. Also as with other described block streams, the top surfaces of blocks are etched and pitted by weathering and surface blocks are splitting or disintegrating in place (Sevon, 1975; Clark, 1991). These features indicate that the blocks have not moved for a long time, probably the entire Holocene, and that they are relict features. The only other Pleistocene climate that has affected this area is a colder climate, near to or at periglacial conditions.

At this site abandoned pits in the block stream, dug to obtain building and field-drain stone, expose the material under the large surface blocks. That material is about 2 meters of open framework, subrounded to rounded, "polished surface", sandstone and conglomerate cobbles and small boulders (10-30 cm long axis). At the floor of the pits, about 2.5 meters from the suface, sandy matrix fills the spaces between the clasts. A remnant of block stream material between two of the pits exposes matrix supported material immediately below the surface clasts. The soil developed in that material has a thin, 5-15 cm, light gray (10YR 7/2) sandy A horizon underlain by brownish yellow (10YR 6/8) loamy B horizon. An Andover very stony sandy loam is typically mapped in these wet sites along the base of Blue Mountain. That soil is poorly drained, typically mottled to within 15 cm of the surface, and has a clay loam textured B horizon. The soil in the face of the pit is better drained than that because the pit lowered the water table.

That the now matrixless boulder streams once had matrix when they were in transport is suggested by two lines of evidence. The first line of evidence is that the long, narrow boulder streams are always located in the axis of zero or first order drainage ways. Immediately to either side of the matrixless material is a surface veneer of blocks underlain by either matrix-supported material or clast-supported material with matrix. During wet periods water can be observed running between the blocks of the block stream and springs, often "boiling sand" springs, issue from their downslope toes. This suggests that during the Holocene, when the blocks were locked in place, the sandy matrix was gradually "piped" out from between the blocks where a strong upward groundwater flow component existed, the axis of drainageways.

The second line of evidence that suggests the the block stream once had matrix is the smoothing, even polishing, of the entire surface of the clasts. While the "ball and socket" effect explains the polishing at boulder contact points, considerable "tumbling" of the clasts would be necessary to polish the entire clast surface by this mechanism. The polishing would be more readily accomplished by having more tools available, a sandy matrix rubbing against the clasts in addition to the other adjacent clasts.

It should be noted that there are other unvegetated matrixless block slopes higher on the mountain sides near here and elsewhere. They have a wide sheet form with bedrock either immediately undemeath or outcropping above them. Such areas are talus that never had significant matrix.

These unvegetated block streams have probably been over emphasized in discussions of the origin of the boulder colluvium in this region. The unvegetated sites represent a tiny fraction of the boulder-mantled mountain slopes in Pennsylvania and southward. The continuous sheet-like nature of the now forested boulder colluvium along all the strike ridges in Pennsylvania is what needs to be explained, not the few exceptional sites without forest cover. What Holocene process is capable of transporting and depositing a sheet of boulder colluvium everywhere along the strike ridges - an infinite number of side by side debris flows?? As discussed in the introductory material, available evidence for the genesis of the colluvium points towards a relict form that dates from late Wisconsinan and older periglacial conditions. A process that is reasonably capable of transporting such material is gelifluction, as pointed out by Louis Peltier 45 years ago.

Questions here:

1. Is this a place to do cosmogenic dating of the block surfaces?
2. Is there any temperate climate process that could transport these materials found everwhere along the base of the strike ridges in this region?
3. How could one tell if these blocks were moving today?

4. Aren't there analogous boulder colluvium deposits in today's periglacial zone, such as the strike ridges of the Brooks Range in Alaska?
5. Are such boulder colluvium deposits described somewhere in the literature on modern periglacial forms?

		Reload vans and **continue ahead**.
0.2	40.5	**TURN RIGHT** into parking lot for Leaser Lake park.

STOP 10: LUNCH AT LEASER LAKE PARK

Boulder colluvium underlies the picnic area and extends under the lake to the center of the valley now flooded by the lake. Upslope of the entrance to the park are fields on a slate bedrock knob that projects above the boulder colluvium (Fig. 39). The hill is like the prow of a ship, diverting the boulder colluvium to either side of it.

		Reload vans and assemble at entrance to parking lot.
0.1	40.6	**TURN LEFT** and retrace route past Stop 3.
0.5	41.1	**TURN LEFT** onto Spring House Road and continue obliquely up the boulder colluvium-covered slope of Blue Mountain.
0.6	41.7	**TURN RIGHT** onto Leaser Road and continue upslope.
0.1	41.8	On right is a pit exposing the boulder colluvium. Continue up the steepening slope where the boulder colluvium thins to a shallow to bedrock boulder mantle.
0.6	42.4	Crest of Blue Mountain where the Appalachian Trail crosses the road.
0.4	42.8	On left at sharp curve to right is small exposure of non-bouldery colluvium derived from adjacent slope underlain by the Bloomsburg Formation. Have just crossed a hollow with boulder colluvium.
0.5	43.3	Enter lower slope area where the boulder colluvium contains erratic Palmerton sandstone and conglomerate clasts. From here northward we will proceed "deeper" under the pre-Illinoian ice but observe very little evidence of its passage.
0.4	43.7	Enter area of fields that mark the downslope edge of the boulder colluvium. On left in the fields are shallow depressions developed in colluvial material.
0.1	43.8	**TURN RIGHT** onto Blue Mountain Drive.
0.5	44.3	**TURN LEFT** onto Rt 309 from Blue Mountain Drive.
0.8	45.1	**TURN RIGHT** onto Rt.895 towards Bowmantown.
0.1	45.2	Road cut through knob exposes pre-Illinoian erratic clast lag and yellowish red weathering penetrating into the shale. For the next few miles colluvium, derived from the Trimmers Rock sandstones and shales, floors hollows and lower toe slopes of this strike valley. Thin shaly residuum mantles the knobs.
3.0	48.2	On right is waste dump that hides a large mine pit that produces road aggregate .
1.0	49.2	**TURN LEFT** onto Andreas Road and enter narrow valley cut in the Trimmers Rock Formation.
0.4	49.6	**BEAR RIGHT** up a tributary hollow floored by colluvium derived from gray interbedded shale and sandstone of the Trimmers Rock Formation.
1.6	51.2	**TURN LEFT** towards Normal Square and go directly uphill. Church on right before turn is made.
0.2	51.4	**BEAR LEFT AT Y** staying on the main road now called Church Hill Road.
0.4	51.8	**STRAIGHT AHEAD AT Y** staying on Church Hill Road. Descend into a hollow where both slopes are shallow-to-bedrock residuum and colluvium with only a narrow alluvium strip on the floor of hollow.
0.8	52.6	**BEAR LEFT** staying on Church Hill Road.
0.3	52.9	**BEAR RIGHT** at stop sign onto Fritz Valley Road towards Normal Square.
0.3	53.2	Enter Mahoning valley, the core of the breached anticline crossed this morning.

0.3	53.5	**TURN RIGHT** onto Rt. 443 East at blinking traffic light at Normal Square.
0.1	53.6	On right is a shale pit showing only 20-30 cm of soil on top of bedrock.
1.0	54.6	Enter shallow road cut with a few rounded clasts on the left (pre-Illinoian lag).
0.1	54.7	Start crossing pre-Illinoian outwash, bright red color in small exposures on the right.
0.2	54.9	**TURN LEFT** into Grant's Dairy Bar and park on the side of the lot towards Rt.443.

STOP 11: PRE-ILLINOIAN (G ?) INTENSELY WEATHERED OUTWASH

Leaders: Duane D. Braun and Edward J. Ciolkosz

The region that the field trip has been traversing since crossing Blue Moutain has been glaciated but the only remaining evidence of that glaciation is scattered erratics on the north facing slope of Blue Mountain and in the Lizard Creek strike valley. The trip has just crossed the rolling upland underlain by the Trimmers Rock and Catskill Formations where even isolated erratics are lacking. Just before arriving at this site, the trip has traveled down the broad, gently sloping anticlinal Mahoning valley underlain by the Mahantango shale. The Mahoning Creek drainage is asymmetric, with a series of much larger tributaries heading at a major strike ridge to the northwest and much smaller tributaries starting in the rolling hills to the southeast (Fig. 43). The larger tributaries from the northwest have left a series of alluvial fan features on the floor of the Mahoning valley and have "pushed" the east draining Mahoning Creek towards the southeast side of the valley. These gravel mantled surfaces have previously been mapped as glacial outwash (Wood, 1974b). In actuality much of those surfaces is alluvial fan lag, rounded sandstone clasts from the strike ridge to the northwest in a shaly matrix derived from the underlying shale (Fig. 43, Qfl). Locally as much as a meter of weathered gravels remain with imbrication showing depositon from the northwest. In a few places in the center and southeast side of the Mahoning valley are rounded clast lags that cannot be directly related to the tributaries from the northwest and those sites have been mapped as pre-Illinoian lag (Fig. 43, Qpl). Much of the alluvial fan lag may be the last remnants of a period of alluvial fan construction from the stripping of the pre-Illinoian glacial deposits from the mountain slopes to the northwest.

At this site is the only known pre-Illinoian glacial deposit exposed at the ground surface in the entire Nesquehoning 7-1/2' quadrangle. The only other outcrops of glacial deposits are in northern part of the quadrangle where strip mines in the eastern end of the Southern Coal field expose up to a few meters of till buried under up to 20 meters of colluvium. Other pre-Illinoian glacial deposits are probably buried under the boulder colluvium that covers the lower slopes of all the strike ridges in the region. The intensely weathered material here is exposed in an over grown several hundred meter long and several meter deep road cut along Rt.443. The cut is in a very gently sloping hilltop that projects from the south flank of the Mahoning valley and whose top surface is about 60 feet above present stream level. Woodchuck holes are the primary "outcrops" on the face of the cut.

The stratigraphic section and associated Allenwood soil profile can be broken into three basic units. From the surface to 25-30 cm is a dark brown (7.5YR 4/4) silt loam with 10 % or more of rounded sandstone gravel clasts (A_p horizon). Below that to 50-60 cm is a red (2.5YR 4/6) silty clay loam to clay loam with a blocky ped structure and thick continuous clay films on ped and clast surfaces (upper B horizon). Rounded sandstone gravel and subrounded shale clasts make up 10 to 30 % of the material. Most clasts are strongly rubified, including the shale clasts. Beneath that is a yet brighter red (2.5YR 4-5/8) with specks of white or gray where weathered shale clasts have been cut by the shovel (lower B horizon). The matrix texture is silty to clayey loam with 10-30 % rubified and clay film-covered gravel clasts as above. There is some relict stratification below one meter primarily shown as sandier clast deficit bands a few to 10's of cm thick. Local shale clasts become more common towards the base. At 2.7-3.3 m (9-11 ft) a irregular and strongly weathered shale bedrock surface is reached. The contact is often gradational over 10's of cm with red clay films penetrating between disrupted and rubefied bedrock fragments. The soil B horizon is

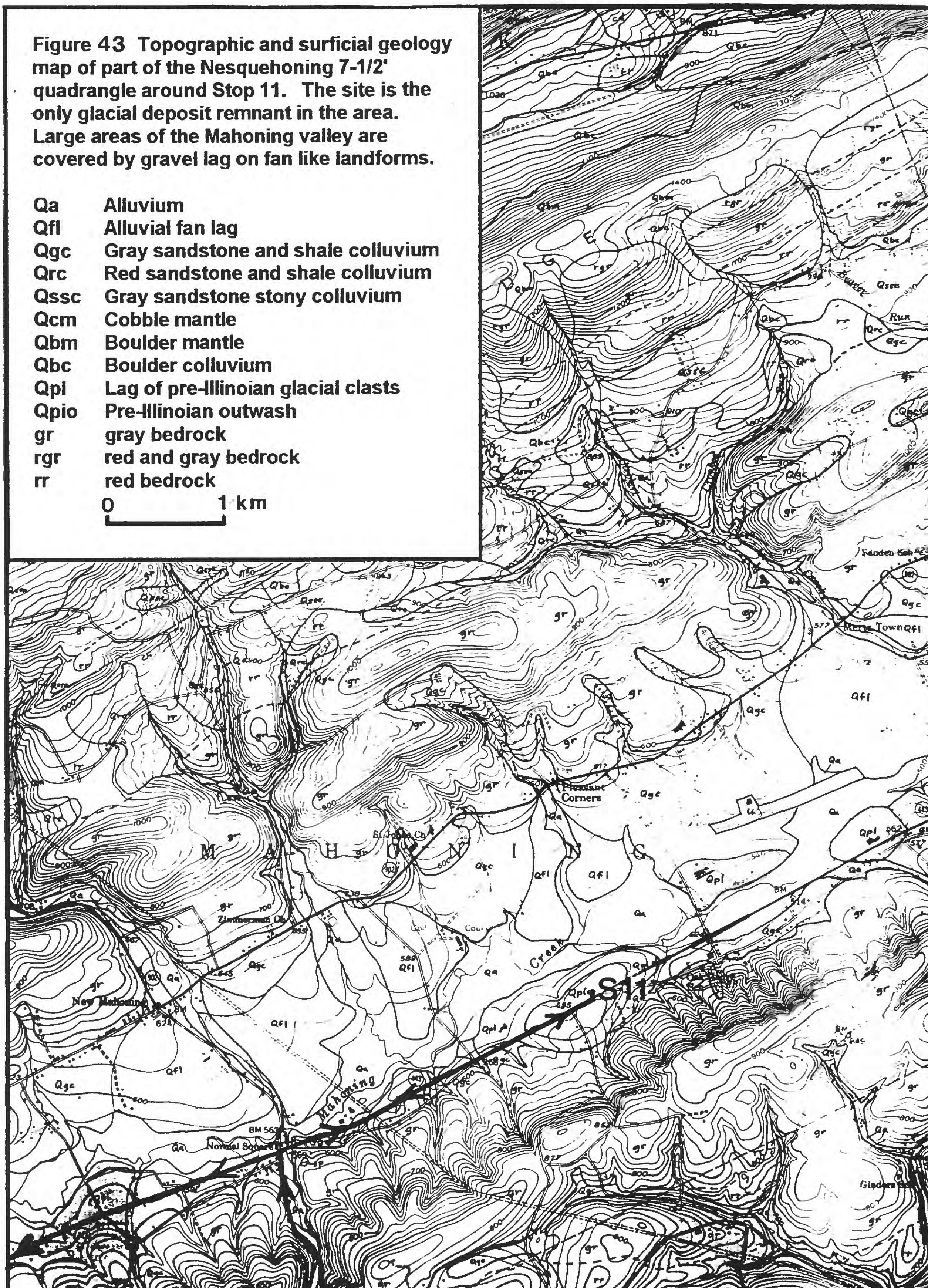


Figure 43 Topographic and surficial geology map of part of the Nesquehoning 7-1/2' quadrangle around Stop 11. The site is the only glacial deposit remnant in the area. Large areas of the Mahoning valley are covered by gravel lag on fan like landforms.

Qa	Alluvium
Qfl	Alluvial fan lag
Qgc	Gray sandstone and shale colluvium
Qrc	Red sandstone and shale colluvium
Qssc	Gray sandstone stony colluvium
Qcm	Cobble mantle
Qbm	Boulder mantle
Qbc	Boulder colluvium
Qpl	Lag of pre-Illinoian glacial clasts
Qpio	Pre-Illinoian outwash
gr	gray bedrock
rgr	red and gray bedrock
rr	red bedrock

0 1 km

penetrating through the glacial deposit and into the bedrock. Down grade at the eastern and deeper end of the cut, fresh bedrock is encoutered 2-3 meters below the bedrock surface.

This material is interpreted as pre-Illinoian outwash deposited in a proglacial lake impounded in the Mahoning valley by ice advancing from the northeast. The diamict-textured material is interpreted to be weathered sand and gravel rather than glacial till (Kaktins and Braun, 1986). One small woodchuck hole exposure shows clast imbrication dipping east, suggesting flow from the east, opposite to that of the present creek. No erratics have yet been identified and a wide variety of lithilogies can be derived from the Mahantango through Pocono Formations underlying the drainage basin of the creek. The lowest outlet to the lake is at the west end of the Mahoning valley at an elevation of 855 ft (260 m) (mile 19.6 this morning). Subglacial drainage of the lake is unlikely in that the next lower "sublet" is the Lehigh River gorge, 7 km directly up ice gradient from this site. Also, late Wisconsinan ice repeatedly impounded lakes with high extra-glacial outlets in similar topographic positions throughout northeastern Pennsylvania (Braun, 1989a). This site is at an elevation of 595 feet (181 m), 260 feet (80 m) below the lake surface. The material is either sublacustrine outwash or the basal remnant of a valley side kame that reached near to or above lake level.

The degree of weathering observed here is not significantly different than what has been observed in similar parent material and topographic positions at the Bloomsburg ice margin (late Illinoian ?). It has been argued that in the moderate relief landscape of eastern Pennsylvania, Pleistocene erosion has removed at least 10's of meters to as much a 100 meters of material from the overall landscape (Braun, 1989c). If there has been that scale of erosion, it is highly unlikely that any present day ground surface is anywhere near the 800+ Ka pre-Illinoian G ground surface. In essense, nearly all soils on older than late Pleistocene (late Wisconsinan) materials are truncated profiles. An exception to this may be a few isolated constructional surface remnants of late middle Pleistocene (late Illinoian) age such as at Bloomsburg, Pennsylvania.

Questions at this site:

1. How could this material be dated?
2. What features can be added to the weathering and soils descriptions?
3. Is this a truncated soil profile?
3. Could this material just be a high terrace remnant of Mahoning Creek?
4. Is the diamict glacial till?
5. Is this the maximal degree of weathering from the pre-Illinoian G to present or is this a deeply truncated profile of the last remnant of glacial material in the region?
6. Was there ever a lake here at all?

		Rest stop at the dairy bar before continuing
0.1	55.0	Leave Grant's Dairy Bar and **TURN RIGHT** onto Rt.443 West.
1.4	56.4	Straight through Normal Square, staying on Rt.443. For next several miles on right the toeslopes have some areas of pre-Illinoian erratic-clast-lag but are mostly colluvium derived from the interbedded shale and sandstone.
5.6	62.0	On left is a shale pit at a farm machinery dealer showing that bedrock is essentially at the ground surface on the hillslopes.
1.5	63.5	**TURN RIGHT** onto Rt.309 north. To right is the only available sluiceway out of the Mahoning valley, but no gravel deposits remain.
0.2	63.7	Descend into the Little Schuykill valley.
0.7	64.4	Straight ahead on Rt.309 where Rt.443 turns left.
1.5	65.9	Water gap through ridge underlain by the Pocono Formation.
0.8	66.7	Water gap through ridge underlain by the Pottsville Formation. On the next "Pottsville" ridge ahead are grassy areas that mark reclaimed coal strip mines.
0.5	67.2	Enter Tamaqua and cross the Little Schuylkill River.
0.3	67.5	Cross Rt. 209 (third traffic light) and immediately **BEAR RIGHT** following Rt.309 North where it becomes one way.

0.1	67.6	**BEAR LEFT** at traffic light, continuing on Rt.309 North
0.5	68.1	Cross Little Schuylkill River again and leave Tamaqua.
0.3	68.4	Water gap through Nesquehoning Mountain, underlain by Pottsville conglomerate tilted near vertical. Then start ascending hill with Mauch Chunk outcrops on the right.
0.8	69.2	**GET INTO RIGHT LANE** approaching traffic light ahead.
0.3	69.5	**STRAIGHT THROUGH** traffic light at Rt.54 intersection, staying on Rt.309 north.
0.8	70.3	**TURN RIGHT** at traffic light for the Hometown Mall and continue straight ahead on the industrial park road. Do not turn left into the Hometown Mall.
0.2	70.5	**TURN LEFT** at the truck entrance to the Hometown Mall.
0.2	70.7	Stop behind the Hometown Mall buildings.

STOP 12: CROSS-SECTION OF BOULDER COLLUVIUM WITH STONE STRIPES

Leaders: Duane D. Braun and Edward J. Ciolkosz

This site is at the toe slope of the plunging nose of anticlinal Broad Mountain (Fig. 45). A wide outcrop area of Pocono sandstone and conglomerate upslope of the site provides an abundance of boulder sized material to be transported down the gentle anticlinal nose. The slope angle across the site is 6.7 % or 3.8^0 . On the ground surface above the outcrop are a series of eight distinct boulder concentrations having a centerline spacing of 14.6 m (range 12 to 18.5 m). The boulder concentrations vary in width from 2 to 5 m except for an 11 m wide band that overlies a bedrock high in the exposure. There are three well exposed boulder concentrations in the center of outcrop exposure. Those concentrations show a distinct V shape in cross-section (Fig. 44). The larger tabular boulders tend to be tilted on edge. These features look identical to stone stripes described in areas of on-going periglacial activity.

Figure 44 . Photograph of the outcrop of one of the regularly spaced boulder concentrations. The concentrations tend to narrow to a V downward and many of the tabular clasts are tilted edgewise.

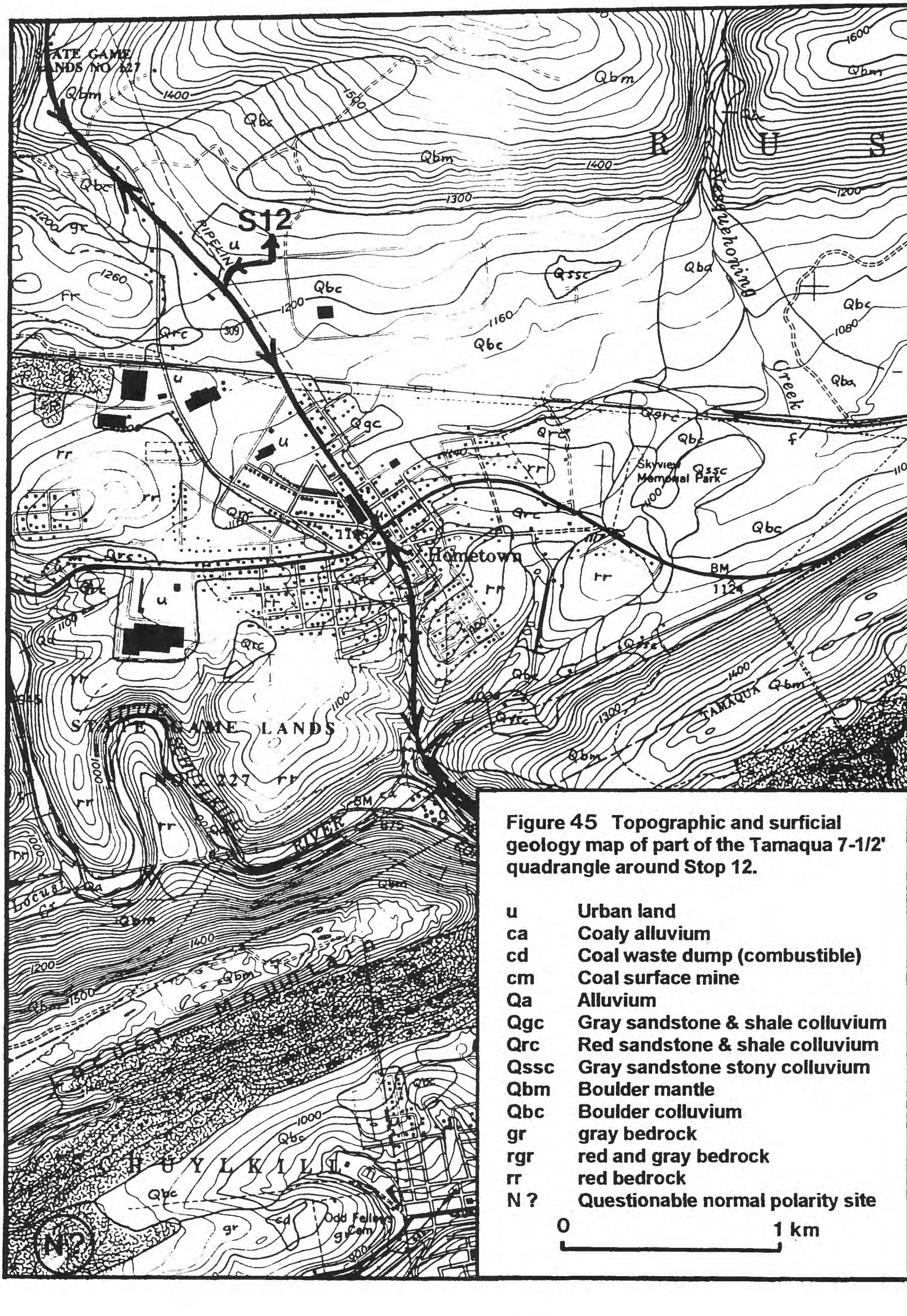


Figure 45 Topographic and surficial geology map of part of the Tamaqua 7-1/2' quadrangle around Stop 12.

u	Urban land
ca	Coaly alluvium
cd	Coal waste dump (combustible)
cm	Coal surface mine
Qa	Alluvium
Qgc	Gray sandstone & shale colluvium
Qrc	Red sandstone & shale colluvium
Qssc	Gray sandstone stony colluvium
Qbm	Boulder mantle
Qbc	Boulder colluvium
gr	gray bedrock
rgr	red and gray bedrock
rr	red bedrock
N ?	Questionable normal polarity site

0 — 1 km

A deep and moderately well drained Buchanan soil is mapped over large areas of these colluvial toe-slopes. The Laidig soil is mapped at dryer sites and the Andover soil (Stop 9) is mapped at wetter sites. The Buchannon soil usually has a thin organic horizon (1-10 cm, black to dark gray, 10YR 2-4/1) at the surface and at this site a thin sandy eluvial horizon under the organics (1-5 cm, light gray, 10YR 5-6/6). The upper B horizon is usually about 40-60 cm thick, yellowish brown (10YR 5/6), and has a gravelly to sandy clay loam texture. The lower B horizon is a fragipan that is typically about 60-80 cm thick, yellowish brown (10YR 5/4) with gray and stong brown mottles, and has a gravelly to sandy clay loam texture. Below the boulder concentrations at the surface, the amount of clasts is highly variable ranging from 5 to 60 %. Large boulder clasts are most common at the ground surface.

The degree of soil profile development, especially fragipan development, indicates landscape stability during the Holocene (Ciolkosz and others, 1986). In other words, this is a relict landscape left over from Wisconsinan (isotope stage 2-4) periglacial mobilization of the landscape. The soil development is far less than that at Stop 11and this material does not contain remnant masses of the older glacial material. This indicates that the older glacial deposits have already been completely eroded off the mountain side in pre-Wisconsinan time. These colluvial deposits, as exposed at Stop 5 yesterday and in strip mines in the adjacent southern coal field, bury older colluvial deposits and pre-Illinoian glacial material in valley floor topographic positions.

Questions here:
1. If these are not periglacial stone stripes, what are they?
2. Could this soil have developed in the Holocene?
3. What process could have transported these materials on this gentle slope?

		Circle around and retrace route.
0.2	70.9	**TURN RIGHT** onto industrial park access road.
0.2	71.1	**TURN RIGHT** at traffic light onto Rt.309 north
1.8	72.9	**STRAIGHT THROUGH** traffic light staying on Rt.309 north.
0.5	73.4	Ascend ridge "held up" by the Pottsville conglomerate at the south limb of the eastern middle coal field.
0.9	74.3	To right are culm banks and cogeneration plant.
0.3	74.6	Ramp for I-81 Interchange 39, continue on Rt.309 North.
0.3	74.9	Go under bridge for I-81 Interchange 39.
0.3	75.2	Shallow anticlinal valley underlain by the Mauch Chunk redbeds.
0.4	75.6	Pass under railroad bridge, Mauch Chunk outcrops on the left.
0.2	75.8	Crest of Spring Mountain, enter the town of McAdoo. Continue on Rt.309 through McAdoo, passing through three traffic lights.
0.8	76.6	Leave McAdoo and enter abandoned strip mine "orphan lands".
0.2	76.8	On left is an abandoned coal breaker surrounded by culm banks.
0.5	77.3	On right for the next 0.5 mile is another culm bank area surrounding a burned out breaker structure.
1.1	78.4	Go past Arthur Gardner Highway on right, the road traveled to Stop 6 yesterday. We are now retracing yesterday's route to the Hampton Inn. Abandoned breaker on left.
0.8	79.2	Go over another anticlinal crest and descend into synclinal valley occupied by downtown Hazleton (Stop 5 is to the right in the same valley.)
0.6	79.8	Pass under railroad tracks.
0.2	80.0	**Get in left lane** after crossing another set of railroad tracks at street level.
0.1	80.1	**TURN LEFT** at traffic light onto 4-lane Broad Street (Rt. 93 West and Rt. 924), the "main street" of Hazleton. **Get into right lane and stay there;** you will eventually be bearing right.
1.0	81.1	**TURN RIGHT** continuing on Rt. 93 & 924. Either lane is all right now; there will be no more turns until the Hampton Inn.
3.1	84.2	**TURN RIGHT** into the Hampton Inn.

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FRIENDS OF THE PLEISTOCENE
ROSTER OF CONFERENCES
1934-1994

Conference	Leaders	Area
1. 1934 M25-27	George White and J. Walter Goldthwait	Durham to Hanover, NH
2. 1935	Dick Flint	New Haven to Hartford, CT
3. 1936	Kirk Bryan	S. RI to Cape Cod, MA
4. 1937 M21-23	J.W. & Dick Goldthwait and Dick Lougee	Hanover to Jefferson, NH
5. 1938 M6-8	Charlie Denny and Hugh Raup	Black Rock Forest, NY
6. 1939 M20-21	Paul MacClintock and Meredith Johnson	Drifts, N. NJ
7. 1940 M18-19	Kirtley Mather and Dick Goldthwait	W. Cape Cod, MA
8. 1941 M23-25	John Rich	Catskill Mts., NY
1942 - 1945	War Years	
9. 1946 J1-2	Lou Currier and Kirk Bryan	Lowell-Westford area, MA
10. 1947 M23-25	Earl Apfel	E. Finger Lakes, NY
11. 1948 M21-23	D.F. Putnam, Archie Watt, Roy Deane	Toronto to Georgian Bay, ONT
12. 1949 M20-22	Paul MacClintock and John Lucke	'Pensauken' problem, NJ
13. 1950 M26-28	O.D. Von Engeln	Central Finger Lakes, NY
14. 1951 M26-27	John Hack and Paul MacClintock	Chesapeake soils and stratigraphy, MD
15. 1952 M23-25	Dick Goldthwait	Tills, central OH
16. 1953 M22-24	Lou Currier and Joe Hartshorn	Outwash sequences, Ayer quad., MA
17. 1954 M21-23	Charlie Denny and Walter Lyford	Wellsboro-Elmira-Towanda, PA-NY
18. 1955 M20-23	Paul MacClintock	Champlain lake & sea, NY
19. 1956 M25-27	Nelson Gadd	St. Lawrence lowlands, QUE

20. 1957 M24-26	Paul MacClintock and John Harris	St. Lawrence seaway, NY
21. 1958 M23-25	John Hack and John Goodlett	Appalachians, Shenandoah, VA
22. 1959 M15-17	Alexis Dreimanis and Bob Packer	Lake Erie till bluffs, ONT
23. 1960 M20-22	Ernie Muller	Cattaraugus Co., W. NY
24. 1961 M19-21	Art Bloom	Marine clay & ice margins, SW. ME
25. 1962 M18-20	Cliff Kaye and Phil Schafer	Charlestown moraine & vicinity, RI
26. 1963 M24-26	Hubert Lee	Lower St. Lawrence, QUE
27. 1964 M22-24	Cliff Kaye	Lower St. Lawrence, QUE Marthas Vineyard, MA
28. 1965 M21-23	Joe Upson, Les Sirkin	Northern Long Island, NY
29. 1966 M20-22	Nick Coch and Bob Oaks	Scarps & stratigraphy, SE. VA
30. 1967 M19-21	Hal Borns	Marine & moraines, E. ME
31. 1968 M24-26	Carl Koteff, Bob Oldale, Joe Hartshorn	E. Cape Cod, MA
32. 1969 M23-25	Nelson Gadd and Barrie McDonald	Sherbrooke area, QUE
33. 1970 M22-24	Dick Goldthwait and George Bailey	Mt. Washington region, NH
34. 1971 M19-21	Gordon Connally and Les Sirkin	Upper Hudson, Albany, NY
35. 1972 M19-21	Art Bloom and Jock McAndrews	Central Finger Lakes, NY
36. 1973 M18-20	Don Coates and Cuchlaine King	Susquehanna & Oswego Val., NY-PA
37. 1974 M17-19	Bill Dean and Peter Duckworth	Oak Ridges-Crawford Lake, ONT
38. 1975 M9-11	George Crowl, Gordon Connally, Bill Sevon, Les Sirkin	Lower Delaware Valley, PA
39. 1976 J4-6	Bob Jordon and John Talley	Coastal Plain, DE
40. 1977 M20-22	Bob Newton	Ossipee quad., NH

41. 1978 M5-7	Denis Marchand, Ed Ciolkosz, Milena Bucek, and George Crowl	Central Susquehanna Valley, PA
42. 1979 M	Jesse Craft	NE Adirondack Mts., NY
43. 1980 M	Bob LaFleur and Parker Calkin	Upper Cattaraugus, Hamburg, NY
44. 1981 M15-17	Carl Koteff and Byron Stone	Nashua Valley, MA
45. 1982 M28-30	Pierre LaSalle, P.P. David, M.A. Bouchard	Drummondville, QUE
46. 1983 M14-15	Woody Thompson and Geoff Smith	Ice margins, central ME
47. 1984 M18-20	Peter Clark and J.S. Street	St. Lawrence lowland, Massena-Malone, NY
48. 1985 M3-5	Ed Evenson, Jim Cotter, Dave Harper, Carl Koteff, Jack Ridge, Scott Stanford, Ron Witte	Great Valley, NJ-PA
49. 1986 M23-25	Tom Lowell and Steve Kite	Northernmost ME
50. 1987 M8-10	Carl Koteff, Janet Stone, Fred Larsen, Joe Hartshorn	Lake Hitchcock Valley, CT-MA
51. 1988 M27-29	E.H. Muller, D.D. Braun, W.J. Brennan, R.A. Young	Genesee Valley, NY
52. 1989 M	P. LaSalle, W.W. Shilts, M. Lamothe, D. Demers	Mid-St. Lawrence Valley, QUE
53. 1990 M25-28	Robert Mott and Ralph Stea	Halifax to Canso Strait, NS
54. 1991 M17-19	Jack Ridge	Western Mohawk Valley, NY
55. 1992 M15-17	R. Dineen, E.L. Hansen, R. LaFleur, David Desimone	Lower Mohawk Valley, NY
56. 1993 M	Carol Hildreth and Richard Moore	Contoocook, Souhegan Piscataquoag area, NH
57. 1994 M20-22	D.D. Braun, E.J. Ciolkosz, J.D. Inners, J. Epstein	Eastern PA

LIST OF PARTICIPANTS OF 1994 FOP CONFERENCE

David Barclay	Dept. of Geology, Univ. of Buffalo, Buffalo, NY 14260
James Black	2305 Bayward Blvd., Wilmington, DE 19802
A.L. Bloom	Dept. of Geol. Sci., Cornell University, Ithaca, NY 14853
Duane and Ruth Braun	Geog. & Ear. Sci., Bloomsburg Univ., Bloomsburg, PA 17815
Donald H. Cadwell	NYS Geologyical Survey, Albany, NY 12230
Parker E. Calkin	Dept. of Geology, Univ. at Buffalo, Buffalo, NY 14260
Steven C. Carlisle	582 Sherwood Rd., Aurora, NY 13026
Edward Ciolkosz	Penn State Univ., University Park, PA 16802-3504
John G. Clark	Wetlands Advisory Group, Inc., Selinsgrove, PA 17870
Mike Clark	Dept. of Geol. Sci., Univ. of Tennessee, Knoxville, TN 37996
Nancy Craft	RD #1, South Lake Rd., DeRuyter, NY 13052
David L. Cremeens	GAI Consultants, Inc., Monroeville, PA 15146
Richard Cronce	P.O. Box 363, Bainbridge, PA 17502
Bob Dineen & Martey Costello	P.O. Box 197, Geigertown, PA 19523
Rick Dunn	Dept. of Geology, Univ. of Delaware, Newark, DE 19716
Gary D'Urso	1134 Sabraton Ave., Morgantown, WV 26505-5829
Jack Epstein	USGS, Reston, VA 22092
John Galbraith	26 Buffalo Rd., Brooklondale, NY 14817
Tom Gardner	Penn State Univ., University Park, PA 16807
Dru Germanoski	Dept. of Geology, Lafayette College, Easton, PA 18042
Bill Goodman	Earth Dimensions Inc., Elma, NY 14059
Jon D. Inners	PA Geological Survey, Harrisburg, PA 17105-8453
George C. Kelley	343 West Lake Rd., DeRuyter, NY 13052
Gregory Kirby	39 Valleyview Rd., Merrimack, NH 03054
Steve Kite	Dept. of Geol. and Geog., West Virginia Univ., Morgantown, WV, 26506-6300
Charles L. Klinger	360 E. Lamb St., Bellefonte, PA 16823
Robin B. Koeberle	129 E. Dewart St., Shamokin, PA 17872-4803
Carl and Dianne Koteff	USGS, Reston, VA 22070
Jeff Leberfinger	R.E. Wright Associates, Middletown, Pa.
Lawrence D. Lyon	Cornell Univ., Ithaca, NY 14853-1504

Bill Manley	INSTAAR/Dept. of Geol. Sci., Univ. of Colorado, Boulder, CO 80309-0450
Ben Marsh	Dept. of Geog., Bucknell Univ., Lewisburg, PA 17837
Hugh H. Mills	TTU, Cookeville, TN 38505
Ernest H. Muller	Geol. Dept., Syracuse Univ., Syracuse, NY 13244-1070
Magda Okulicz	3 Leroy Ave., Lansing, NY 14882
Don Owens	Earth Dimensions Inc., Elma, NY 14059
Don Pair	Dept. of Geol., Univ. of Dayton, Dayton, OH 45469-2364
Steven Parisio	241 Sheldon Hill Rd., Olivebridge, NY 12461
Frank Pazzaglia	Dept. of Geol. & Geophysics, Yale Univ., New Haven, CT 06520
Louis Cook Peltier	6021 Southport Dr., Bethesda, MD 20814
Joan Ramage	Dept. of Geosci., Penn State, University Park, PA 16802
Allan D. Randall	USGS, Albany, NY 12201
Jack Ridge	Dept. of Geology, Tufts Univ., Medford, MA 02155
Mauricio Roma	27 Victor Dr., Albany, NY 12203
E. Lester Rothermel	RD #1 Box 142A, Klingerstown, PA 17941
Gregory Rubino	112 B South Water Street, Hummelstown, PA 17036
Ira Sasowsky	Nittany Geoscience, Inc., State College, PA 16801
William D. Sevon	PA Geological Survey, Harrisburg, PA 17105-8453
Thomas W. Small	Dept. of Geog., Frostburg State Univ., Frostburg, MD 21532
Scott Stanford	New Jersey Geological Survey, Trenton, NJ 08625
Robert A. Stewart	242 Newell Ave., Bristol, CT 06010
John M. Stiteler	116 ASI Bldg., University Park, PA 16802
Greg Tucker	Dept. of Geosci., Penn State Univ., University Park, PA 16802
John Wehmiller	Dept. of Geology, Univ. of Delaware, Newark, DE 19716
Edgar A. White	USDA-SCS, Harrisburg, PA 17111
Ron Witte	NJ Geol. Survey, Trenton, NJ 08625

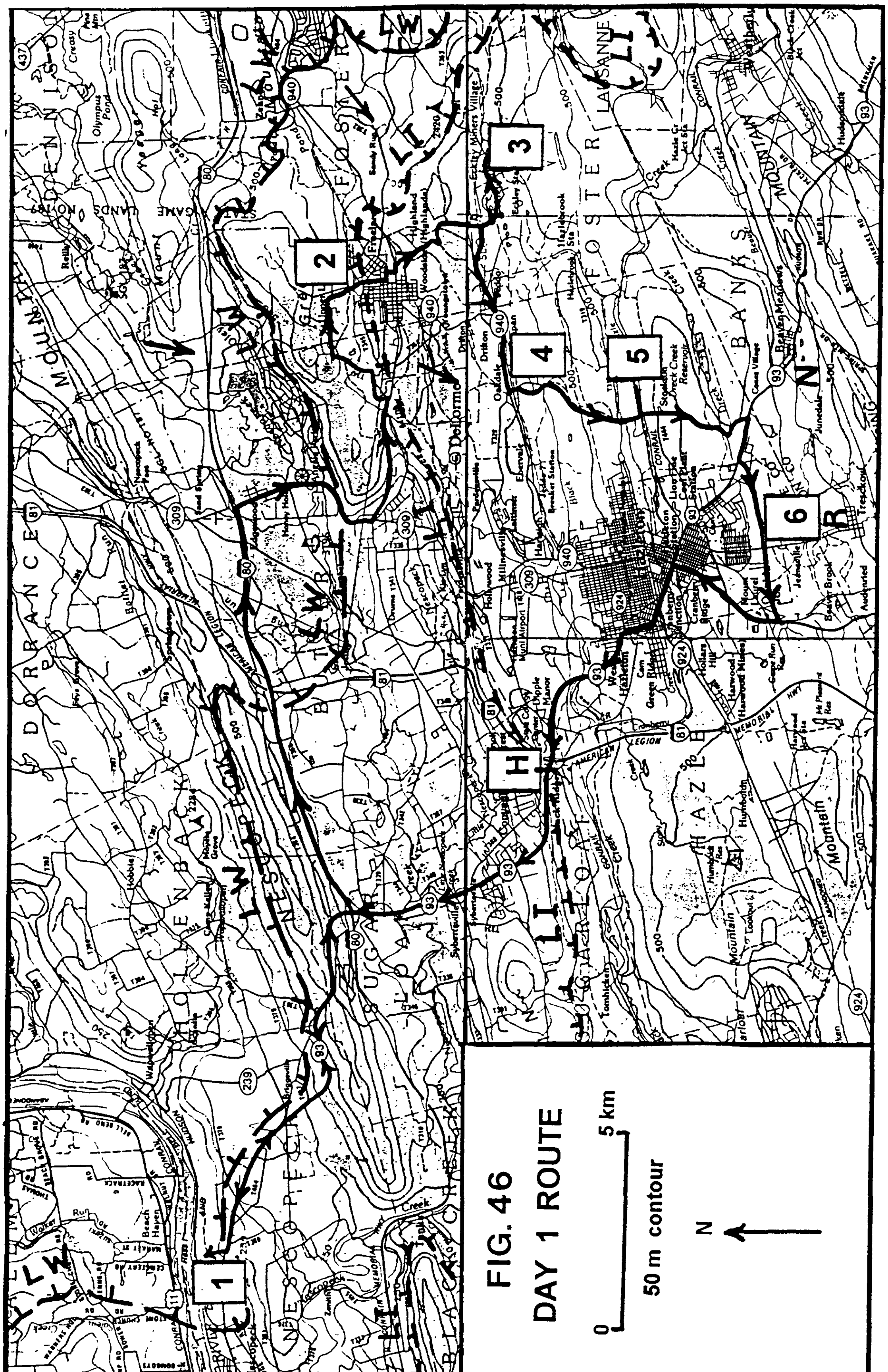

FIG. 46
DAY 1 ROUTE

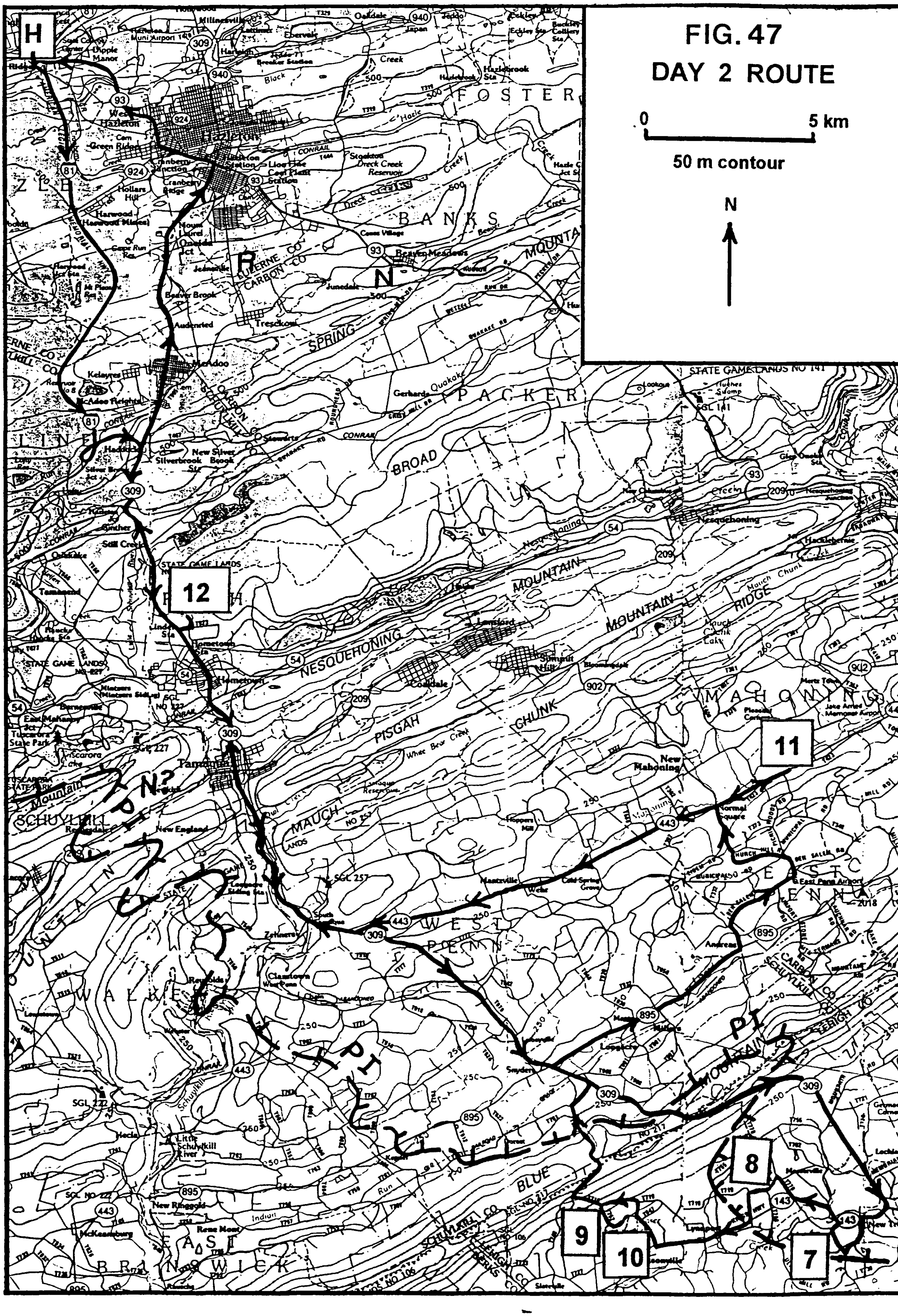

FIG. 47
DAY 2 ROUTE