Guidebook for the

74th Annual Reunion of the Northeastern Friends of the Pleistocene and Harrisburg Area Geologic Society / Association of Environmental & Engineering Geologists (NEFOP/HAGS/AEG) Field Trip

June 3-5, 2011, Mansfield, Pennsylvania

Glacial deposits, landforms, and drainage diversions in the "Grand Canyon" region of north-central Pennsylvania

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Acknowledgements

The organizers of the field trip would like to thank those who gave permission to enter private property and /or took time to open gates for us: Robert Murphy of Wellsboro; Earl Dean, Galeton pit; Mr. Dunbar, Cowanesque pit; Richard Giles, GOH Lawrenceville pit, and Chad Cross, Cross Excavating, Lawrenceville pit. The organizers of the field trip would also like to thank the Geography and Geosciences Department, Bloomsburg University, for assistance with printing the guidebook and for providing van transportation on Sunday.

This guidebook is dedicated to those Pleistocene geologists that have published on this area before

Pre-Friends		
H. Carvill Lewis	1	1884
William C. Alden and Myron L. Fuller (in Fuller)		1903a, 1903b
Past Friends		
Charles S. Denny		1956
Charles S. Denny and Walter H. Lyford		1954, 1963
George H. Crowl		1981
Present Friends		
Donald R. Coates		1966, 1974
William D. Sevon (with George H. Crowl)		1980
William D. Sevon (with Duane D. Braun)		1997a, 1997b

Overview of Pleistocene events in north central Pennsylvania

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The overall history of glacial advances into Pennsylvania

There is evidence of at least four Pleistocene glacial advances into Pennsylvania (Shepps *et al, 1959;* White *et al., 1969;* Marchand *et al., 1978;* Braun, 1994, 1999, 2011; Braun *et al.,* 2008,). From oldest to youngest (using the terminology of Richmond and Fullerton, 1986), these glaciations are: Early Pleistocene (Pre-Illinioian G - Marine Isotope Stage [MIS] 22 or older); Early middle Pleistocene (probably Pre-Illinoian D - MIS 16); Middle middle Pleistocene (Pre-Illinioan A or B - MIS 10 or 12) or late middle Pleistocene (MIS 6 - Illinoian); and Late Pleistocene (MIS 2 - Late Wisconsinan). The Early Wisconsinan did not reach Pennsylvania (Braun, 1988; Ridge *et al.,* 1990). The oldest glaciation extended the farthest south and each younger glaciation extended less far to the south (Fig. P1). The trace of each successive glacial advance's maximum limit is remarkably similar (essentially parallel) to that of the previous advance (Fig P1.).

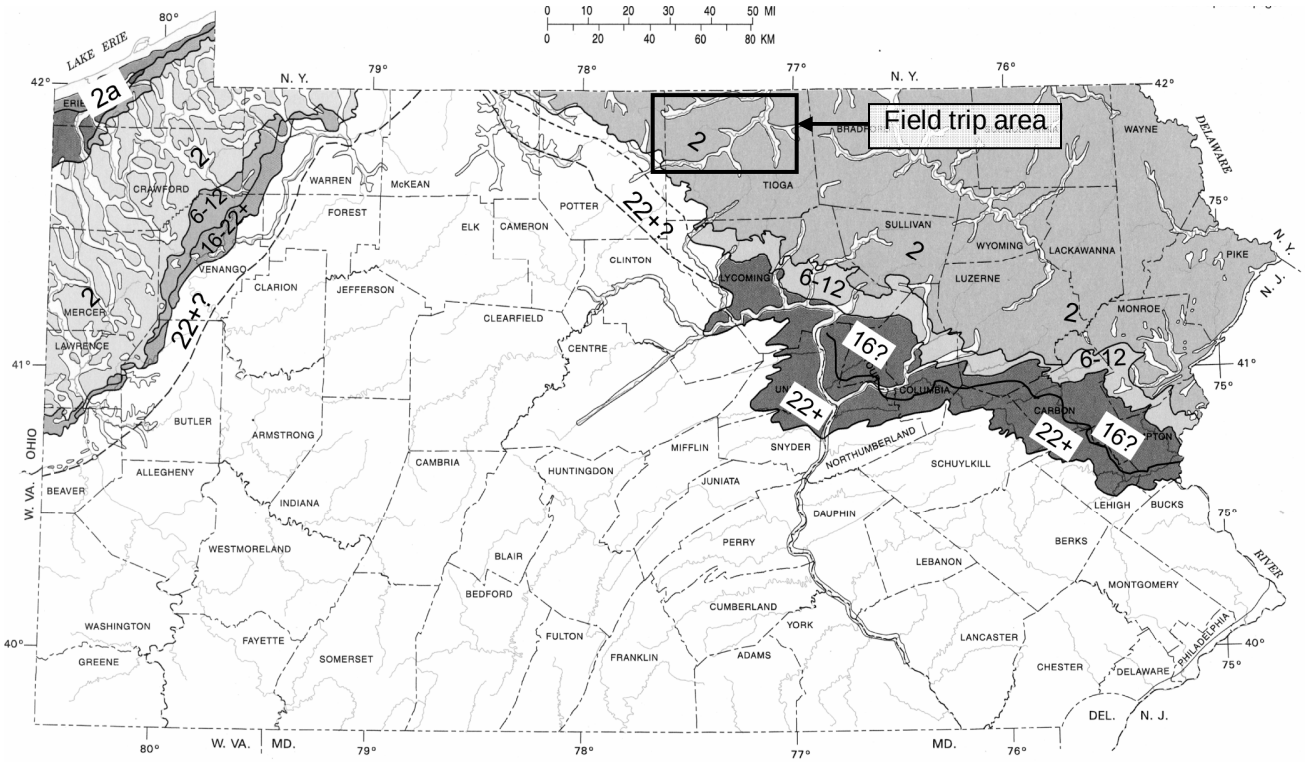

Figure P1 Map of the extent and age of glacial deposits in Pennsylvania. Ages given by Marine Isotope Stage (MIS) number: (Braun, 2011, Figure 1, modified from Sevon and Braun, 1997b).

Each time the Laurentide ice sheet advances, it enters Pennsylvania from the north-west (the Erie-Ontario lobe) and the north-east (the Lake-Champlain-Hudson River lobe) (Crowl & Sevon, 1999) (Fig. P2). The triangular shaped unglaciated area between the two lobes is called the Salamanca Reentrant (Fig. P2). In northwestern Pennsylvania the relatively low relief (100 to 300 meters), shaly bedrock, and abundant debris from the Great Lake basins produced a dominance of deposition over erosion in each glacial advance. This produced multiple till sheet sequences whose layers can be separated by partly preserved weathering profiles. These till sequences are well exposed in open pit coal mines (White *et al.,* 1969). The pre-glacial stream drainage in north-western Pennsylvania was to the north-west and was blocked and diverted to the south-west by glaciation to form an ice marginal drainage system, the present Allegheny-Ohio River system (Carll, 1880; Leverett, 1902, 1934; Kaktins & Delano, 1999). The Early Pleistocene and probably Early-middle Pleistocene glaciations impounded large proglacial lakes, especially glacial lake Monogahela, whose deposits are widespread in valleys in southwest Pennsylvania and northern West Virginia (White, 1896; Campbell, 1903; Leverett, 1902, 1934; Lessig, 1963; Jacobson *et al.*,

1988). By mid Pleistocene time at the latest, the Allegheny-Ohio system was fully integrated into its present form (Leverett, 1934).

In northeastern Pennsylvania the moderate relief (300 to 500 meters) on sandstone bedrock produced a dominance of erosion over deposition in each glacial advance. Each successive advance almost entirely removed the deposits of the previous advance and some bedrock (Braun, 1989b, 1994, 2006). Described multiple till sites are few and those are all questionable sites (Braun, 1994). Older glacial advances are recognized only because remnants of their deposits are present south of the younger advances. Each advance left thick glacial and proglacial deposits in the valleys while adjacent ridge crests are essentially bare bedrock. Only in partly to completely buried valleys transverse to ice flow would older deposits be expected but none have yet been delineated on the basis of preserved weathering profiles. Portions of the pre-glacial stream drainage were to the east or north-east and the first glacial advance blocked that drainage to form glacial lake Lesley (Williams, 1902; Ramage, *et al.*, 1998) and Glacial Lake Packer (Williams, 1894) at the Early Pleistocene glacial limit. Other proglacial lakes were impounded along that limit, along younger limits (Fuller, 1903a, 1903b; Braun, 1988), and north of the Late Pleistocene limit as ice receded from Pennsylvania (Willard, 1932, Coates, 1966b; Gardner *et al.*, 1993; Braun, 1989a, 1997, 2002). Part of the northeasterly drainage was diverted south to form the 'Grand Canyon' of Pennsylvania, a 750 ft deep bedrock gorge (Fuller, 1903a, 1903b; Crowl, 1981).

The Marine Isotope record and the North American glacial deposit record (Richmond & Fullerton, 1986) suggest that at least ten Pleistocene glacial advances should have approached close to PA and caused periglacial climate conditions there (Braun, 1989b,1994). The area south of the Late Pleistocene glacial limit is characterized by extensive colluvial deposits and other features of paleoperiglacial origin (Peltier, 1949; Denny and Lyford, 1963, Ciolkosz *et al.*, 1986; Clark & Ciolkosz, 1988; Clark *et al.*, 1992; Marsh, 1999; Braun, 1989b, 1994, 2006a). Most of the material that was mapped as Early and Middle Pleistocene glacial deposits is actually colluvium derived from glacial deposits (Braun, 1994, 1999, Braun *et al.,* 2008). Even north of the Late Pleistocene limit there are significant periglacial effects and production of colluvium on the slopes (Peltier, 1949; Denny, 1956, Denny and Lyford, 1954,1963; Coates & King, 1973; Braun, 1997, 2002, 2006).

Direction of ice flow

The first state wide report on the glacial deposits in Pennsylvania (Lewis, 1884) noted that in northeastern Pennsylvania the overall ice flow direction was to the southwest and was perpendicular to the northwesterly trend of the "terminal moraine" (Lesley letter of transmittal, p.xix in Lewis, 1884). Lewis noted that "striae in a valley are sometimes at high angles to those on the summit of a mountain (p.11)." "The high-level striae, which are uniform over large districts, are the only ones that indicate the general flow of the glacier" (p.11). Figure P2 is a more recent compendium of ice flow direction (Crowl and Sevon, 1980, 1999).

Alden and Fuller (in Fuller,1903b) noted that the most recent or Wisconsin glacier generally flowed northeast to southwest across the region. Local ice flow was "dependent upon the configuration of the surface over which the ice passed, and varied from S10^{0}W to due west (p. 7)". The ice was deflected more westerly by east-west stream valleys with the maximum influence during opening and closing stages of glaciation when the ice sheet was relatively thin. The larger topographic features were from streams, with "ice producing the beautifully flowing contours of the broader areas of Chemung rocks (today's Lock Haven Formation) (p.8)." Those rolling hill-top, shaly rock areas have only a thin till mantle so that the rounding was from erosion of the bedrock by the flowing ice rather than by deposition of till.

In north central Pennsylvania Sevon and Braun (1997a) noted that 64 glacial striation measurements had an average direction of S46^{0}W, showing that the glacier was flowing obliquely across most of the major river valleys and strike ridges in the region. Forty one of the striations were in the range of S30°W to S60°W with a second groupin g of 12 striae at S70°W to due west. As noted previously by Fuller and Alden (1903), the ice flow direction was quite variable and was influenced by local topography.

When the regional distribution of all the striation sites noted in Fuller and Alden (1903) and Sevon and Braun (1997a) is examined, it is seen that the ice had a lobate margin as it flowed across

the landscape of north-central Pennsylvania. The ice formed broad protruding lobes a mile or two long and five or more miles wide in the lower elevation areas of rolling hills (breached anticlinal lowlands) with the center and north side of each lobe having a strongly westerly flow direction. Such lobation explains the glacial striation site that lies at the center of the rolling hill lowland near Mansfield that has two directions recorded, the regional flow direction of S46^{0}W and a S62ºW direction (Fuller and Alden, 1903). The regional S46ºW direction was from thick ice flowing all the way to the terminus about 20 miles southwest of the site. The more westerly S62ºW flow direction was from when the ice was receding across the site and the thinner ice was shaped into a lobate form by the topography.

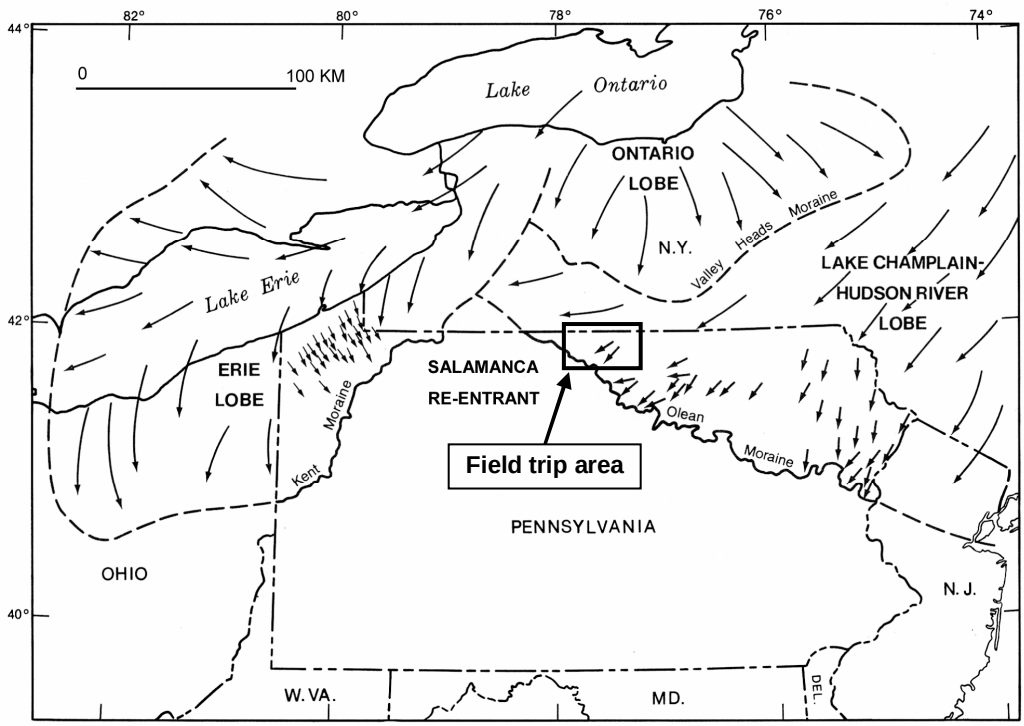

Figure P2 Map showing the ice flow directions into northeastern and northwestern Pennsylvania to either side of the Salamanca Re-entrant (from Crowl and Sevon, 1999, Fig. 15-2).

The southward diversion of northeast draining preglacial Pine Creek

The term Pine Creek gorge has been used by previous workers for the entire narrow, deep Pine Creek valley from the entrance to the gorge at Ansonia, Pennsylvania (Stop 1) to 50 miles south to the edge of the Appalachian Plateau (Denny, 1956, Denny and Lyford, 1963, Crowl, 1981). The discussion here will be focused on the upper most few miles of the gorge where it crosses the regional north-south preglacial drainage divide (Fig. P3, and the green line on digital terrain map of the field trip route at the start of the field trip log). That is the critical breached divide part of the gorge where the preglacial northeast draining upper Pine Creek was diverted to the south by glacial blockage. South of that Pine Creek is simply occupying the southwest and south trending course of preglacial Babb Creek (digital terrain map of the field trip route at the start of the field trip log).

In their discussion of the glacial history Alden and Fuller (in Fuller 1903b) emphasized that drainage in the region was obstructed and deflected by the first glacier to advance across the region (pre-Kansan or Kansan age). They thought that the Pine Creek gorge had been cut deep enough to afford the “easiest outlet for the waters of upper Pine Creek” (p. 7) by the end of the Kansan stage. They considered the time interval since the Kansan to be many times longer than postglacial Wisconsin time with multiple glaciations not reaching past the Wisconsin limit. Muller (1957) argued that the glaciation-induced drainage diversions that cut large gorges in the region, like Pine Creek gorge in Pennsylvania, dated from early Pleistocene times. Crowl (1981) emphasized that workers in northwestern Pennsylvania and in Ohio had all concluded that the glacial drainage diversions that

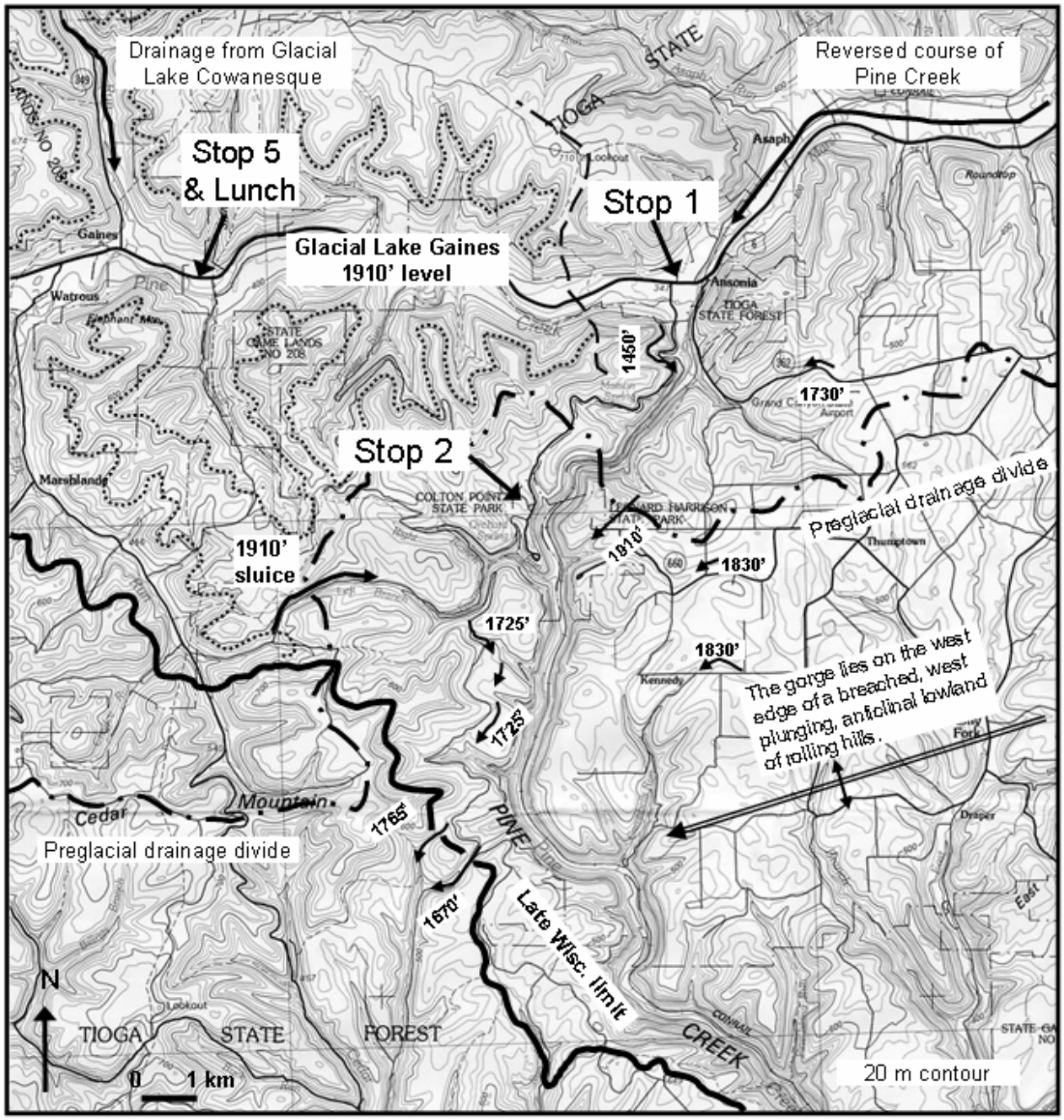

Figure P3 The Pine Creek gorge has been produced by the glacial diversion of northeast draining Pine Creek to its present south course through the 800 ft deep gorge. Three pre-Wisconsinan glacial advances crossed the site and continued to the southwest. Those earlier events cut the gorge to near its present level. The late Wisconsinan advanced just across the gorge. At the late Wisconsinan limit, Glacial Lake Gaines initially drained across a sluiceway near the head of the Pine Creek drainage at an elevation of 1960 ft (shown on Stop 4 & 5 Fig.). As the glacier started to retreat the 1910 ft elevation sluiceway opened (Stop 3) and drainage started along the west flank of the Pine Creek gorge in a series of sluiceways just within and across the terminus (arrows with elevation numbers).

formed the Allegheny-Ohio River system had started in the early Pleistocene and had been finished by the mid Pleistocene (Illinoian) time. So it would be reasonable that the Pine Creek diversion would also have been initiated in the early Pleistocene and completed in the mid Pleistocene.

Denny (1956) noted that Wisconsinan terminus deposits mantle the floor of Pine Creek valley where the south trending diverted reach of the creek joins the southwest trending Babb Creek (see middle bottom of digital terrain map of the field trip route). That indicated to him that the gorge was

cut to its present depth prior to the Wisconsinan. But that site is 14 miles (22.5 km) downstream of the breached divide site and does not necessarily indicate that the gorge at the divide was cut to its present level prior to the Wisconsinan. It only indicates that the valleys south of the divide, part of the south and southwest preglacial drainage system, had been cut to about their present levels in pre-Wisconsinan time. Denny thought that the Pine Creek diversion probably dated from a pre-Illinoian glacial advance because another 10 miles (16 km) farther downstream at Slate Run there are pre-Wisconsin (Illinoian ?) gravels with a well developed paleosol. That site is where Braun (2011) has placed the Illinoian or older glacial terminus (Fig. P1). Also, those gravels are only 40 feet or so above present creek level and indicate that the valley had been cut to near its present level in pre-Illinoian times. But again, while such sites do indicate that the south and southwest draining valleys of the preglacial drainage system had been cut to near their present levels in pre-Illinoian time, they do not directly show that the gorge at the divide had been cut to near its present level in pre-Illinoian time.

Evidence that the Pine Creek gorge was cut to near its present level at the breached divide site in pre-Wisconsinan times lies north of the entrance to the gorge. Northeast of the head of Pine Creek gorge is a 10 mile long reach of the broad, gentle gradient and underfit southwest draining Marsh Creek. That creek is flowing in a reversed segment of the northeast draining preglacial Pine Creek (Fig. P3 and digital terrain map of field trip route). Bedrock at the divide at the head of Marsh Creek on the floor of preglacial Pine Creek is at about 1150 feet (ground surface at 1180, Fig. P3), the bedrock at the entrance to the gorge is at 1120 feet, so there has been a minimum of 30 feet of erosion of the floor of the preglacial Pine Creek valley to form the reversed course Marsh Creek valley. But preglacial Pine Creek must have had some northeast gradient through the 10 mile long March Creek segment. That northeast gradient can be conservatively estimated using the gentlest gradient of present Pine Creek upstream of the entrance to the gorge. That gentlest gradient reach is between Galeton (Stop 3) and Gaines (Stop 5) and has a gradient of 11.2 ft/mi. Applying that gradient to the Marsh Creek reach yields a total depth of erosion of 142 feet for the preglacial Pine Creek valley just east of the entrance to the gorge. Well data and seismic profiling (Williams *et al.*, 1998) show that the bedrock floor of the Marsh Creek valley has a broad U shaped cross-section and the overall trend of the valley is subparallel to regional ice flow. That suggests that much of the erosion of the reversed segment of the preglacial Pine Creek valley has been by the three more extensive older glacial advances (Fig. 1, MIS 22+, 16?, 6 or 12) with final deepening by the Wisconsinan advance.

That the diversion of Pine Creek is of pre-Wisconsinan age is also indicated by the incision of present Pine Creek into the broad floor of the Pine Creek valley upstream of the entrance to the gorge (Stop 5). There is a 30-50 feet deep, 400-700 feet wide inner gorge cut into the glacially eroded 2000-3000 ft wide overall floor of the Pine Creek valley. The overall floor is covered by 10's to more than a 100 feet thick layer of Wisconsinan glacial deposits that Pine Creek has cut through to reach the bedrock. This inner valley starts about two miles (3 km) upstream of the entrance to Pine Creek gorge (Stop 1), continues for about seven miles (11 km), gets gradually shallower upstream, and ends at Gaines. The gradient in this reach is 13.1 ft/mi, somewhat steeper than the 11.2 ft/mi of the next reach upstream from Gaines to Galeton (used in the calculations above). Upstream of Gaines the valley floor is entirely mantled by glacial deposits with a broad floodplain like that at Galeton (4.5 miles upstream at Stop 3). This inner gorge or valley is interpreted to be a knickpoint migrating upstream from Pine Creek gorge at Stop 1. The presence of such a knickpoint suggests that the Pine Creek gorge was cut to near its present level in pre-Wisconsinan times, was deepened some in Wisconsinan times, and then a knickpoint migrated upstream from near the gorge entrance in late to post glacial times.

The specific location of the Pine Creek gorge can be related to the overall broad synclinal mountain and rolling hill, breached anticline topography of the area (Figure P3 and digital terrain map of the field trip route). The Pine Creek valley upstream of the gorge is deeply incised in a broad synclinal upland capped by resistant sandstones. The gorge site lies at the western down plunge end of a breached anticline with a rolling hill landscape that is cut in interbedded shale and sandstone. The rolling hilltops average 300-400 feet lower than the adjacent synclinal uplands. So the north draining tributary valley at the site of the gorge would have had a distinctly lower saddle

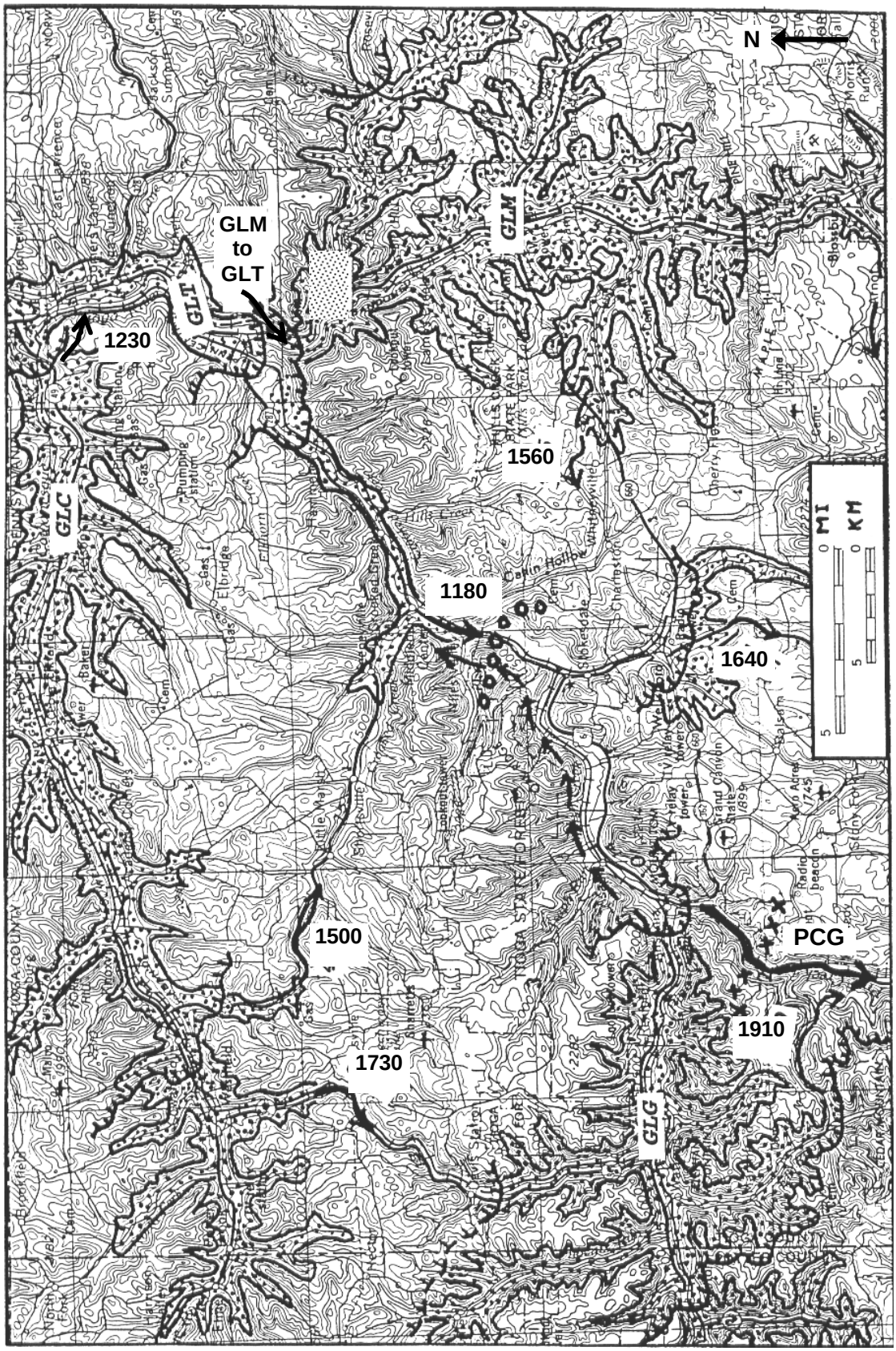

Figure P4 Topo map showing the extent of the various proglacial lakes in the region. Sluiceway floor elevations in feet. GLC - Glacial Lake Cowanesque, GLG - Glacial Lake Gaines, GLM - Glacial Lake Mansfield, GLT - Glacial Lake Tioga, PCG - Pine Creek Gorge, .x.x.x.x Preglacial divide at Pine Creek gorge, o o o o Present Marsh Creek - Crooked Creek divide (modified Fig. 14, Gardiner *et al., 1993).*

over the divide to its opposing south draining valley than any other such tributary to the west carved entirely in the upland. The glaciers, each time they receded northeasterly up the preglacial Pine Creek valley would find a distinctly lower outlet for the sizable lake in front of them when they reached this point. The discharges from the lowering of those lakes would help to carve the saddle here deeper than at other equivalent saddles to the east in the rolling hills lowland that would open up as the ice continued to recede.

Erosion of features by large scale to catastrophic glacial lake outburst floods has been prominent in the Pleistocene geology literature in recent years. It would be expectable that the Pine Creek gorge was at least partly cut by such floods in that a deep, extensive Glacial Lake Gaines was present just upstream of the entrance to the gorge (Fig. P4 and P5). When the gorge was initiated in the early Pleistocene the difference in elevation of the more headward outlet (Fig. P5, arrow labeled 1960 and blue arrow on digital terrain map of the field trip route) and the gorge would have limited the maximum lake outburst flood. The headward outlet is estimated to have started at an elevation of about 2150 feet, and the gorge is estimated to have started at an elevation of about 1850-1900 feet, yielding a potential maximum lake level drop of around 300 feet. An intermediate sluiceway just west of the gorge (Fig. P3 and T1 and discussed in the Stop1 text) that started out at about the 2100 feet, reduced the maximum lake level drop to about 250 feet. The early Pleistocene initiation of the gorge would have produced the most rapid down-cutting of any of the later glacial advances due to the relatively narrow upper part of divide and the steep gradient on the downstream side of the divide. After two more glaciations (MIS 16? and MIS 6 or 12?, Fig. P1) the gorge should have been cut to near its present level and the potential elevation differential for a glacial lake outburst would have approached 700 feet or so. So the potential for catastrophic lake outburst floods was present, but was it realized?

The Pine Creek gorge is transverse to ice flow and a thick “till shadow” should have partly to completely filled the gorge as each glacier advanced across the site. The partial to complete burial of valleys transverse to ice flow is a common feature in north central and northeastern Pennsylvania (Braun, 1997, 2004, 2006) and many individual quadrangle maps by Braun). At some sites as much as 400 feet of glacial till and other glacial deposits remain as eroded remnants on the sides of deep, narrow valleys like the Pine Creek gorge. The late Wisconsinan glacier coincidently terminated at the gorge and probably was the most effective in filling the gorge with glacial deposits. So in each glaciation, when the receding ice front opened up the Pine Creek gorge, the gorge was not actually there until the lake outburst floods eroded out a 10 mile length of glacial deposits in the gorge at and below the divide. This would tend to maximize the durations and minimize the discharges of such floods.

Glacial lake outburst floods down Pine Creek gorge may have also been limited if the “glacial plug” in the gorge started leaking and lowering lake level as the glacial recession approached the gorge. The Pine Creek gorge is also transverse to the ice surface profile and the meltwater hydraulic head gradient within the glacier. So water levels within the glacier at the gorge would be higher than the proglacial Lake and that should stop leakage until ice front receded past the entrance to the gorge. That would tend to increase rather than reduce the potential for a large flood.

Evidence for large scale floods in the Pine Creek valley south of the gorge at the divide has been looked for without success. Only evidence that limits the scale of such floods has been observed so far (Denny, 1956, Denny and Lyford, 1963, Braun, unpublished field work). That evidence was noted above in the context of determining the age of the gorge, the presence of weathered pre-Wisconsinan gravels near the floor of Pine Creek valley south of the gorge at the divide. To preserve those gravels, the maximum Wisconsinan lake outburst flood height could have been at most 40 feet in the Pine Creek valley downstream of the breached divide gorge. Exposure of the gravels is poor and may even extend downslope to only 20 feet above present stream level. One could calculate a maximum discharge by estimating a maximum cross-section and assuming a maximum water velocity at the site but so far I have been reluctant to calculate such an “assumed flood”.

Subglacial flow cutting tunnel valleys in the direction of ice and meltwater flow has also been prominent in the Pleistocene geology literature in recent years. Such subglacial flow could be called upon to help cut the Pine Creek gorge but again the gorge is transverse to ice flow, ice surface

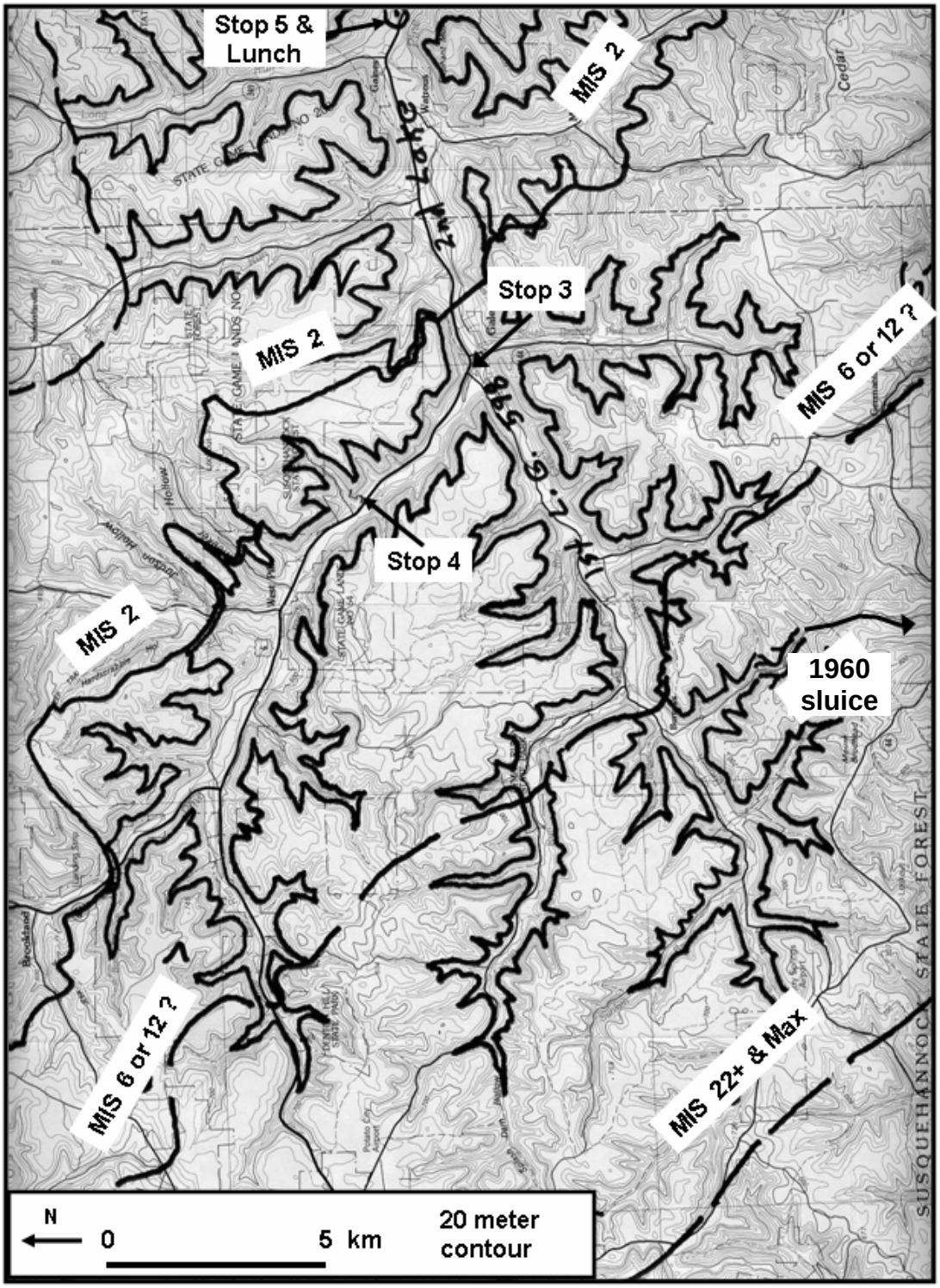

Figure P5 Topo map showing the extent of the highest two phases of Glacial Lake Gaines, the highest level outlet at 1960 ft, the proposed glacial limits of different ages crossing the region, and the field trip stop locations that underlie the lake.

profile gradient, and meltwater gradient in the ice. So it would not be expected that subglacial flow played a role in cutting the gorge. Subglacial flow may have assisted the cutting of those sluiceways that run near parallel to ice flow but probably did not play a dominant role as indicated by the two prominent sluiceways from Glacial Lake Cowanesque (Fig. P6, Stop 6 and 8). Both sluiceways have similar depth but the more headward one (Fig. P6, Stop 6, Fig. T5, Stop 6) is near parallel to ice and meltwater flow while one to the east is transverse to ice and meltwater flow (Fig. T6, Stop 8).

Also the one to the east that is transverse to ice flow is wider and has a longer channel across the divide, though that may simply be from longer term discharge through that sluiceway.

Glacial lake outburst floods must have assisted in cutting the Pine Creek gorge, because as noted above, its unique position at the west end of the rolling hill lowland. Three other valleys in the rolling hills lowland parallel to and east of the Pine Creek gorge were also available to be used as a diversion to the south (Digital terrain map of the field trip route). Each of them was used for local drainage to the south during each ice recession but they did not become deeply incised. This is because the high level headward part of Glacial Lake Gaines had already dropped to the level of the first saddle in the rolling hill lowland, the Pine Creek gorge. The valleys to the east did not have the advantage of the glacial lake outburst flows but only the local flows to cut them. The question at this point is how could one attempt to reasonably estimate how large the Glacial Lake Gaines outburst floods actually were? Maybe features indicative of large scale flooding such as large scale boulder bars have yet to be discovered in the Pine Creek valley downstream of the gorge and in the West Branch Susquehanna valley. It is time to scour the new Lidar imagery of the region for clues.

The large-deep proglacial lakes with multiple outlets and levels

Most early workers on the glacial geology of the region did not visualize that large lakes existed in front of the glacier. Lewis (1884) was focused on tracing the terminus across Pennsylvania and talked only of kettle lakes in the moraine. Where he traced the terminus across the Pine Creek valley near Galeton he made no mention of a lake in front of the ice. Fuller and Alden (1903) described approximately level-topped or gently sloping deposits of stratified drift that they called morainal and frontal terraces in the Tioga valley from Tioga to the New York State line (Stops 9, 10, 13, and 14). They thought the terraces had “generally been deposited in standing water ponded between the ice and adjacent slope” (p. 4). Such terraces “rise in nearly every case to an elevation of about 1180 feet, and were evidently deposited in slack water, the level of which was determined by the height of the divide between Niles Valley and Stokesdale Junction” (p.4) (the present divide between Pine Creek and Crooked Creek-Tioga River drainages). But they thought that the valley for long distances was occupied by stagnant ice with only localized pondings beside the ice instead of a long proglacial lake.

Lohman (1939) noted extensive lake clays in borehole records in the northeast draining valleys in the region and did envision large proglacial lakes in those valleys. Willard (1932) noted extensive lake clays in the Cowanesque River valley and named the lake Glacial Lake Cowanesque.

Denny (1956) noted all of the high level sluiceways of Glacial Lake Gaines but left it at that. In the headwaters of the north draining Genesee valley in Pennyslvania he made no mention of a proglacial lake or of lake sediments. Denny and Lyford (1963) noted that “Although the Tioga River drains northward and appears favorably located for the development of a large melt-water lake, no extensive lake sediments were found there (p. 12)”. They did note though, when describing the extensive colluvial cover in the area, a site where colluvium overlaid lake sediments (their Fig. 10 E) in a tributary valley to the Tioga valley in Mansfield area.

Coates (1966b) noted a number of lake sediment localities in the Cowanesque River valley and its tributaries but thought, due to the variety of elevations at which they were found, that they represented a series of small, disconnected proglacial lakes. He did note extensive lake sediments under the floor of the Cowanesque valley, in places overlain by till, that suggested a larger proglacial lake there that had been overridden by the advancing ice.

Braun (1989a) emphasized that there were a number of large proglacial lakes, each with several levels, in north-central Pennsylvania. Sevon and Braun (1997a) mapped (in the middle 1980’s) the glacial deposits in the region but the mapping should still be considered reconnaissance type mapping. The field maps were done at 1:24,000 or 7.5’ scale and then compiled and published at 1:100,000 scale. Many more of the smaller stream valleys were walked as compared to the earlier mapping by Alden and Fuller (1903) and Denny and Lyford (1963). In those smaller valleys, outcrops of clayey to sandy glacial lake sediments at the toes

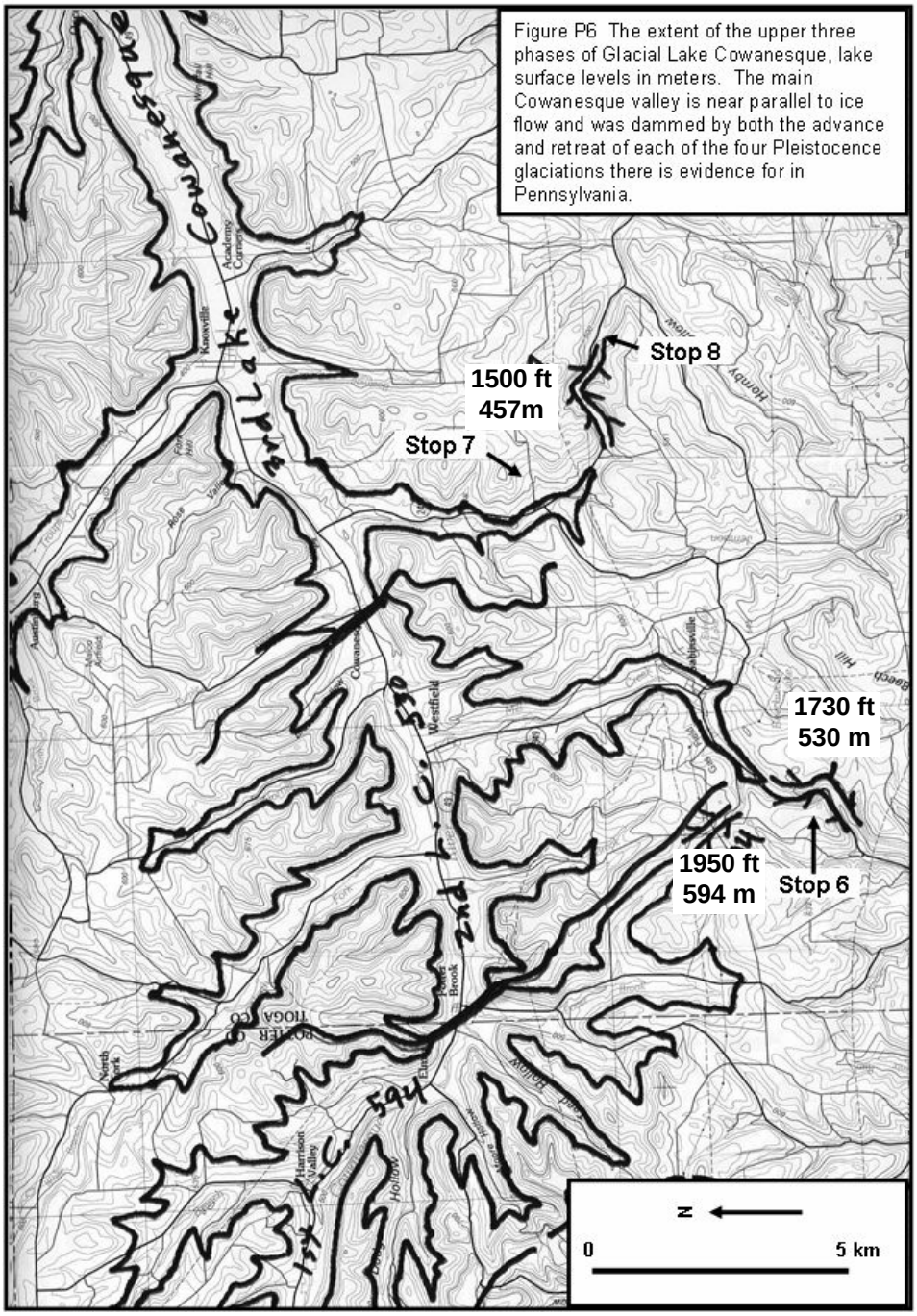

Figure P6 The extent of the upper three phases of Glacial Lake Cowanesque, lake surface levels in meters. The main Cowanesque valley is near parallel to ice flow and was dammed by both the advance and retreat of each of the four Pleistocence glaciations there is evidence for in Pennsylvania.

of active slumps were common and showed that glacial lakes were far more numerous and extensive than thought by previous workers. Usually the lake sediments were overlain by colluvium or other glacial deposits, hence units on the map like Glacial Till and Lake Sediments undivided (TL).

Williams and others (1998) studied the hydrogeology of the glaciated valleys in the region using a large number of well records and some geophysical profiling. That work showed that glacial lake sediments form a near continuous layer, locally more than 100 feet thick, underneath the floodplains of all the major valleys in the region. They further noted that glacial lake sediments form the bulk of the deposits under the major valleys and that the greater the depth to bedrock the greater the thickness of lake sediments. They mentioned that in places the lake sediments are interbedded with coarser-grained deltaic deposits.

The large proglacial lakes in north central Pennsylvania had multiple outlets and levels because the entire northeasterly drainage system slopes in the downstream direction, both the valleys and

the interfluves. Ideally as one proceeds northeasterly, saddles in the interfluves get progressively lower, providing progressively lower sluiceways for the lakes. This ideal case occurs if the bedrock geology is homogeneous along the length of the interfluve and the crest of the interfluve remains in about the same position between the two valleys along the length of the interfluve. The interfluve between the Cowanesque valley and the Pine Creek-Crooked Creek valleys most closely follows this ideal with the interfluve divide being composed of the same interbedded shale and sandstone unit along its length and the divide stays approximately in the same position between the two valleys. Thus Glacial Lake Cowanesque had three sluiceways of progressively lower elevation to the northeast (Fig. P6, Stop 6 and 8, Fig. P4, 530, 457, 375), none of which has incised enough to divert the drainage southward.

The interfluve between the preglacial Pine Creek valley and the preglacial Babb Creek valley to the south had a marked change in geology and elevation at the site of the Pine Creek gorge (as was discussed above when talking about the specific location of the gorge). While Glacial Lake Gaines had progressively lower sluiceways up to and including the Pine Creek gorge, the progression did not continue farther down the preglacial Pine Creek valley because erosion became focused at the site of the Pine Creek gorge.

In the north draining Tioga River valley the incision of a single sluiceway between the sluiceway near the head of the drainage and the confluence with Crooked Creek (preglacial Pine Creek) took place due to change in geology going downstream (Figure P4. and digital terrain map of field trip route between Stop 12 and Stop 9). The Tioga River starts in a synclinal upland, crosses a breached anticlinal lowland, and then crosses another synclinal upland to join Crooked Creek at the Tioga-Hammond damsite (Figure P4, GLM to GLT; Figure T7, ice margin 2). The sluiceway for Glacial Lake Mansfield was in the center of the rolling hills, breached anticline lowland, hundreds of feet lower than any possible outlet saddle in the adjacent synclinal uplands and so it remained as the sole outlet for the lake. The sluiceway drained westward to the Pine Creek gorge.

Glacial Lake Tioga occupied the Crooked Creek valley (preglacial Pine Creek valley) and the Tioga River valley upon drainage of Glacial Lake Mansfield to the Glacial Lake Tioga (Fig. P4, GLM to GLT). That lake had no interfluve ridge crossing sluiceways. Instead it had a single, low altitude sluiceway on the floor of the Crooked Creek valley at the very head of the lake (Fig. P4). That low sluiceway was due to the cutting of the Pine Creek gorge and the southward diversion and partial reversal of the Pine Creek drainage (Marsh Creek). The sluiceway crosses the low altitude divide between northeast draining Crooked Creek (preglacial Pine Creek) and Marsh Creek (the reversed course portion of preglacial Pine Creek)(Fig. P4, line of circles; digital terrain map of the field trip route, bright green line). In the early Pleistocene stages of the cutting of the Pine Creek gorge, reversal and down cutting of the Pine Creek drainage in the Marsh Creek reach would have been only been partially completed. An early Pleistocene Glacial Lake Tioga in the abandoned course of Pine Creek would have reached almost to the Pine Creek gorge and would have been on the order of 100 to 150 feet deeper than the late Wisconsinan Glacial Lake Tioga.

Overall, the specific location of the sluiceways in the region was combination three factors. First was the general direction of drainage in the region. Second was the topography of the interfluves as controlled by the bedrock geology, primarily the resistance to erosion of the strata and the fold structure of the region. Third was the glacial history of the cutting of the sluiceways themselves, particularly the cutting of the Pine Creek gorge and the southward diversion of the upper Pine Creek drainage. Future glaciations will produce further incision of the existing sluiceways until the entire drainage of the region is diverted southward through the Pine Creek gorge.

Constraining ice margin positions through glacial lake sluiceway locations

Due to the moderate relief of the Appalachian Plateau of north-central Pennsylvania glacial deposits are discontinuous across the region and individual ice margins (recessional moraines) cannot be traced continuously from valley to valley across the region. Braun (1989a) used sequences of successive proglacial lake and sluiceway elevations to delineate ice margin positions across north central Pennsylvania (Fig P7 and P8). A number of proglacial lakes (four greater than 100 km^2, 25 less than 60 km^2) and sluiceways (four major and more than 70 minor) existed in front of the retreating ice. In any particular area the sluiceway elevations and associated lake levels

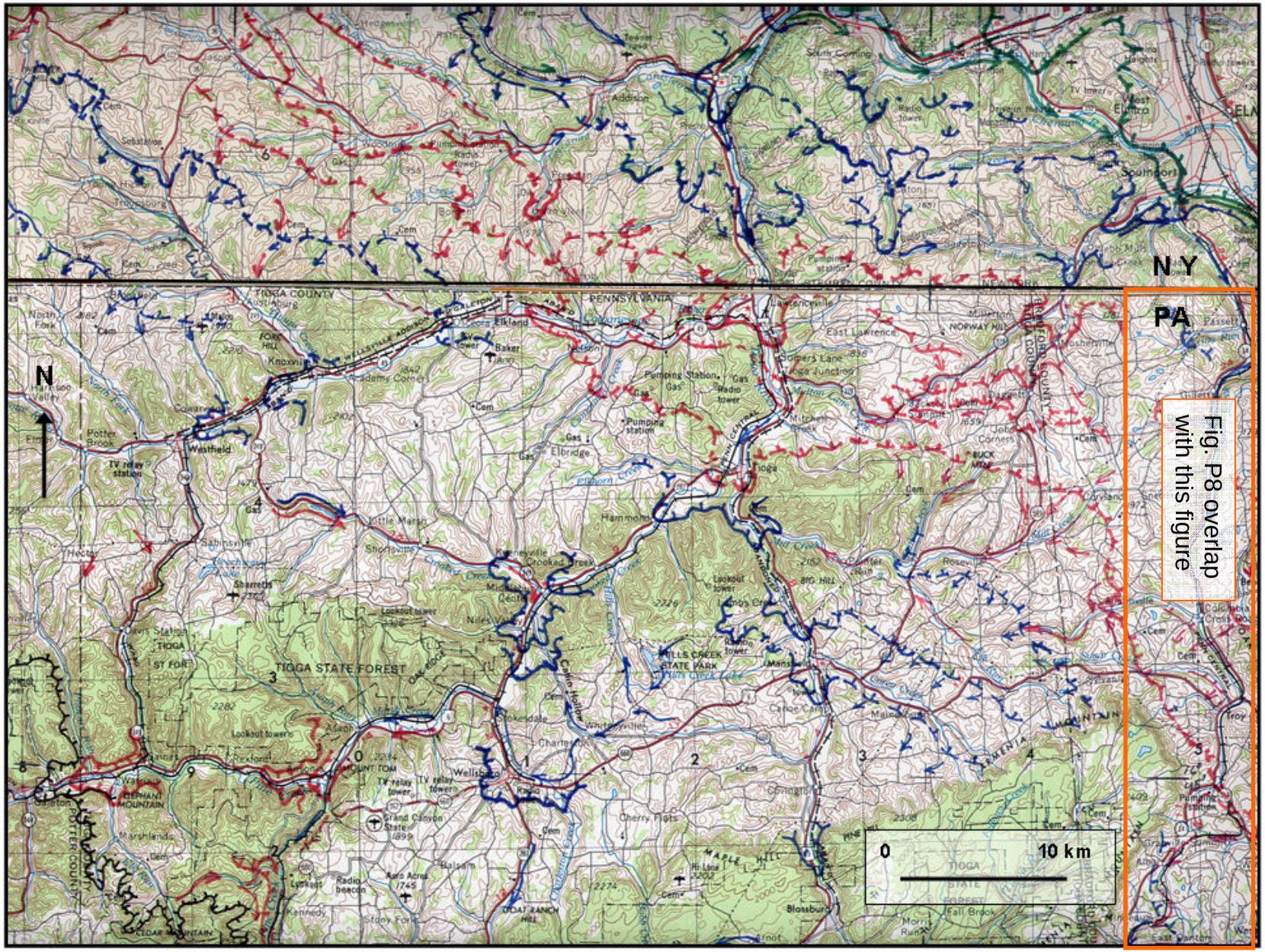

Figure P7 Ice margin positions in north central Pennsylvania and southern New York State determinded from sequences of successive proglacial lake suiceway levels. Arrows mark sluiceways.

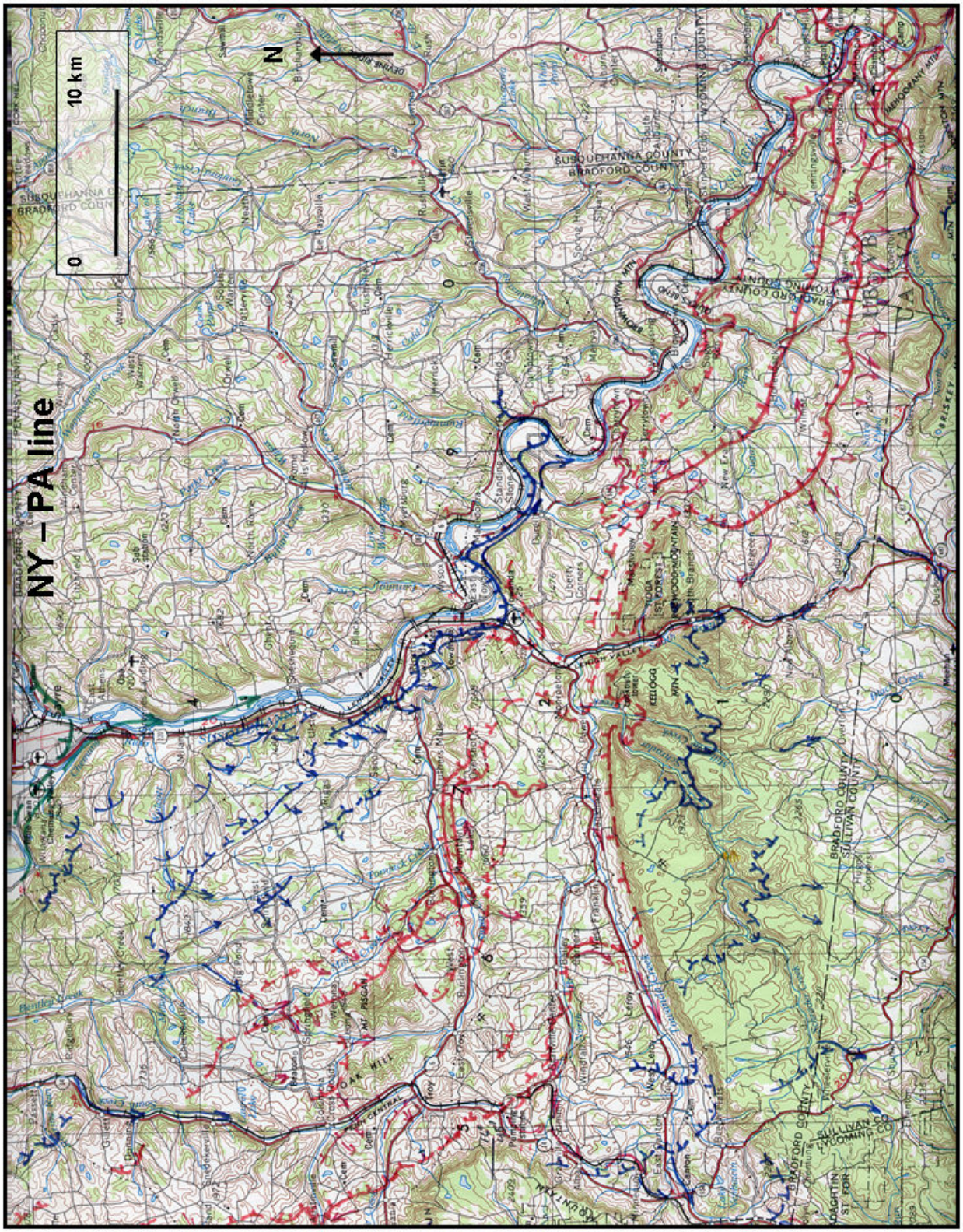

Figure P8 Ice margin positions in north central east of Figure P7 determined from sequences of successive proglacial lake sluiceway levels. Margins of similar age are grouped by color.

constrain where the ice front could be to permit the functioning of individual sluiceways or series of sluiceways across an interfluve or series of interfluves.

The ice margins so delineated trend generally southeast-northwest and are near parallel to the N 60^0 W trend of the Wisconsinan glacial terminus and near perpendicular to glacial striation directions. The margins can be correlated most accurately from valley to valley where sluiceways are so closely space that ice recession of only a few hundred meters opens new outlets between or

out of the valleys. In places the ice margins can be correlated across a series of valleys due to ice marginal sluiceways that cross the intervening uplands. Sequences or groups of 2-4 parallel margin segments, each a few miles apart, could be traced 10 to 30 miles (20-50 km) across the region. The groups of margins that are considered to be correlative or of similar age are given the same color on Figures P7 and P8. The longest set of parallel margin segments that was delineated ran for about 100 miles (160 km) from near Mehoopany, Pennsylvania to near Canisteo, New York (the grouping of red colored margins from the lower right of Figure P8 northwesterly to the upper right of Figure P7).

While no single ice margin can be continuously delineated across the region, it is thought that groups of sequential ice margin segments can be used as a generalized ice margin trend that can be correlated across the region. That means than in any grouping of ice margin segments (a group of margin segments having the same color on Fig. P& and P8), as one goes northwesterly across the region at least one of the margins in the group at the southeast end of the trend will correlate with at least one of the margins at the northwest end of the trend. In other words, when going northwesterly along the trend of the margins, the southernmost margin of a group of segments may correlate with the middle margin of the group or even the most northerly margin of the group when one arrives at the northwest end of the trend. Thus, while one has little confidence in correlating a single ice margin across the entire region, one has a much higher likelihood of being correct that one of the margins of a group correlate with one of the other margins of the group as one goes across the region.

It should be noted that Figures P7 and P8 are compilations of the ice margins on 1:250,000 scale maps that have been further reduced to fit on the 8.5x11 inch page. The individual ice margin segments were first constructed using 1:24.000 scale maps where details of the sluiceway form and elevation were quite apparent. Then all the 1:24,000 scale map margins were compiled and correlated on 1:100,000 scale maps that retained nearly all the 1:24,000 scale information. Finally, for display purposes, the 1:100,000 scale information was compiled onto 1:250,000 scale maps (Figs. P7 and P8).

Periglacial and parglacial deposits overlie glacial deposits

Denny and Lyford's 1954 Friends of the Pleistocene trip in north central Pennsylvania primarily dealt with the Olean versus Binghamton till question and presence of periglacial deposits on top of the glacial deposits. About one-half of the stops looked at Olean ("drab colored", local sedimentary clasts) deposits and Binghamton ("bright colored", some crystalline and limestone erratic clasts) deposits. The Olean-Binghamton question was resolved in the 1960's with the realization that the difference in clast lithology did not have an age significance but only a source significance. The glacier was picking up outwash material that had been washed out ahead of the advancing glacier to produce a "bright" drift in major valleys. Adjacent uplands contained only "drab" drift. The other one-half of the stops looked at periglacial deposits overlying glacial till, outwash, and varves. They used the term congeliturbate for the periglacial deposits but apparently the USGS editors at the time did not approve of that term so in their USGS reports (Denny, 1956; Denny and Lyford, 1963) they used the term rubble or periglacial deposit or colluvium.

Denny (1956) argued that intense periglacial activity took place immediately after ice withdrawal and was the causative agent that produced the colluvium that blanketed the region. Denny and Lyford (1963) emphasized that throughout the region extensive colluvial deposits overlaid every type of older deposit - drift, residuum, and bedrock. No weathering of the older material under the colluvial blanket was observed, indicating that the colluvium was generated immediately upon retreat of the glacier.

In all three publications Denny and Lyford noted fans on the floors of major valleys that were deposited by upland tributaries. They originally thought that much of the fan was deposited immediately after glaciation with additional material added during the Holocene (Denny and Lyford, 1954; Denny, 1956). But in their 1963 report they only noted the fans as being of recent age (USGS editors again?). Such fans are ubiquitous throughout north central and northeastern Pennsylvania (Sevon and Braun, 1997a). Rare exposures of the entire fan thickness show that the lower part and bulk of the deposit contains no macroscopic organic matter and is overlain by material that contains

organic matter as individual particles or entire layers. That suggests that much of the fans were deposited shortly after local deglaciation when there was a scarcity of organic matter to be deposited.

In recent years the term paraglacial or paralglacial activity has been coined (Renwick, 1992) for accelerated erosion and deposition of materials when the landscape is free of vegetation immediately upon recession of the glacier from an area. The fans discussed above are such paraglacial features from rapid erosion of the uplands with deposition starting immediately after glacial lake drainage and/or meltwater cessation in the major valleys. During this paraglacial phase of accelerated erosion, probably much of the deep incision into the glacial deposits occurred in upland valleys and locally across glacial deposit blockages of the major valleys. This rapid incision should have initiated large scale slumping in the glacial lake deposits that continues at a reduced rate today.

The entire landscape of the region, from upper middle slope to valley floor margin positions is covered by a blanket of periglacial and paraglacial colluvial deposits formed during the several thousand year period when the glacier was retreating through and north of this area. In calibrated calendar years, the glacier was at its terminus here about 25,000 BP, started receding from the terminus at about 23,000 BP, had retreated to the Pennsylvania-New York line in the Tioga River valley by about 20,000 BP, and was at the southern end of the Finger Lakes at about 17,000 BP (extrapolation of dates from Ridge, 2003). Vegetation didn't become well established in northern Pennsylvania until about 15,000 BP (calendar years) (Dalton and others, 1997). So that gives a 5000 to 8000 year period during which periglacial and paraglacial processes modified and buried the glacial deposits and landforms. That has produced the relatively smooth sloped landscape that helped lead early workers in the region to conclude that the Wisconsinan deposits here were older than the late Wisconsinan. Also the colluvial mantle in this region, as much a 30 feet (10 m) thick on the toeslopes, buried almost all lake sediments and lead early workers to conclude that glacial lake sediments were rare here.

Large scale slope failures in varves deposited on steep mountain sides.

Alden and Fuller (1903) noted and had a photograph of glacial clays deformed by post glacial slippage exposed in tributaries to the Tioga River near Mansfield. Delano and Wilshusen (1999, Plate 1) mapped the distribution of slope failures throughout north central Pennsylvania and delineated the Tioga valley and its adjacent east side tributary valleys as a zone of high susceptibility to landsliding (mostly slumps). They mapped 100's of slump areas in this region, the largest occupying a square mile or so. We will visit a large scale slump site floored by varves at Stop 11 and talk about a recent major slump/slide event caused by placing road fill on varves at Stop 9. Anywhere within the areas that were covered by the proglacial lakes in the region (Fig. P4, P5, and P6) there is a potential for varves to be present. The varves are nearly always buried by colluvium or by glacial deposits and colluvium. The deepest seated and usually largest slumps occur where varves were overridden by the advancing glacier and were deeply buried by glacial till. Shallower seated and generally smaller slumps occur where the varves were deposited as the glacier was receding and are exposed at the ground surface (rarely) or are covered by colluvium (usually). On these steep mountainside it would have been expected that all the varves slid/slumped to the valley floor during paraglacial times. But varves have been observed on the mountain sides as high as 500 feet above the floors of the major valleys (the glacial lakes here were as deep as 750 feet). That means that any human modification of slopes in this region should take into account the possible presence of varves anywhere there were glacial lakes (as will be emphasized at Stop 9).

Problems for Marcellus shale gas drilling activity in glaciated terrain

In north central Pennsylvania, the glacial lake varves are most significant problem/hazard for the gas drilling industry. The only areas where slumping in varves are naturally active today are where streams are presently undercutting a slump area. Slopes underlain by varves with inactive slumps "terracing the hillsides" or even with no apparent past slumping can be reactivated or destabilized by human excavation into the slope. Cut and fill operations such as putting in a drilling pad could

readily reactivate a slump if the head of the slump is loaded by pad fill. A drilling rig slumping down the slope while in the process of drilling could be a bit hazardous. Likewise if a pad starts to head downslope during a fracturing operation, it could cause significant hazards on and off site.

It is the network of gas pipelines that are being installed in the region that will most likely run into problems with the extensive varve deposits. The worst scenario would be a pipeline crossing obliquely across a slump area with the pipeline bedded in material, like sand, with a relatively high hydraulic conductivity. In that situation groundwater flow from upslope would become channeled down the pipeline and into the preexisting slump slippage zones. That would raise the porewater pressure in the slippage zones and tend to reactivate the slumps. Presently most gas pipelines are made of plastic that connect wells to compressor stations where the gas is injected into the large diameter, high pressure steel gaslines for export from the area. Those plastic gaslines are somewhat resistant to breakage by slump movement but there are limits as to how far one can deform the plastic before it ruptures. If you rupture the gaslines on these mountain sides, the gas readily flows downhill to pool in the valleys until a spark sets the gas off. So there is a significant potential for hazardous events due to slumping along the pipelines. It would be useful for the gas companies to avoid or be especially careful in the placement of their gaslines in areas underlain by varves, particularly in those areas with preexisting slump failures that could be reactivated. Probably we will have to have at least one gasline failure to have this issue brought to the attention of the gas industry, the regulators, and the public. Then again it is possible, but unlikely, that there will never be a slope failure that causes a rig or pipeline failure.

The other significant problem/hazard for the gas drilling industry in north central Pennsylvania is the rapid migration of well drilling or fracturing fluids in high hydraulic conductivity glaciofluvial deposits. This could cause both groundwater and surface water contamination problems at and at a considerable distance from the gas well site. This will be considered in more detail by Jennifer Whisner in a companion article in this guidebook concerning water quality problems with the gas drilling activity.

REFERENCES FOR OVERVIEW AND FIELD TRIP STOPS

Braun, D. D., 1988, Glacial geology of the Anthracite and North Branch Lowland Regions: *in*, Inners, J. D., ed., Bedrock and Glacial Geology of the North Branch Susquehanna Lowland and the Eastern Middle Anthracite Field: Annual Field Conference of Pennsylvania Geologists, 53rd, Hazleton, PA, Guidebook, p. 3-25.

Braun, D. D., 1989a, The use of proglacial lake and sluiceway sequences to determine ice margin positions on the Appalachian Plateau of north-central PA: Geological Society of America Abstracts with Programs, v. 21, no. 2, p. 6.

Braun, D. D., 1989b, Glacial and periglacial erosion of the Appalachians: *in,* Gardner, T. W. & Sevon, W. D., eds., Appalachian Geomorphology: Geomorphology, v.2, no. 1-3, p. 233-258.

Braun, D. D., 1994, Late Wisconsinan to pre-Illinoian (G?) glacial events in eastern Pennsylvania: *in*, Braun, D. D., ed., Late Wisconsinan to pre-Illinoian (G?) glacial and periglacial events in eastern Pennsylvania, Guidebook for the 57th Field Conference, Friends of the Pleistocene, U.S. Geologic Survey Open File Rpt. 94-434, p.1-21.

Braun, D. D., 1997, Physiography and Quaternary history of the Scranton/Wilkes-Barre area: *in,* Inners , J. D., ed., Geology of the Wyoming-Lackawanna Valley and its Mountain Rim, Northeastern Pennsylvania; Annual Field Conference of Pennsylvania Geologists, 62nd, Scranton, PA, Guidebook, p. 1-15.

Braun, D. D., 1999, Pleistocene geology of the Allentown Area: *in,* Sevon, W. D. and Fleeger, G. M., eds., Economic and environmental geology and topography in the Allentown-Bethlehem Area; Annual Field Conference of Pennsylvania Geologists, 64th, Allentown, PA, Guidebook, p. 31-40.

Braun, D. D., 2002, Quaternary history of the Tunkhannock-Great Bend Region: *in,* Inners, J. D. and Fleeger, G. M., eds., Bedrock, surficial, economic, and engineering geology of the “Endless Mountains” region, Susquehanna and Wyoming Counties; Annual Field Conference of Pennsylvania Geologists, 67th, Harrisburg, PA, Guidebook, p. 32-38.

Braun, D. D., 2004, The Glaciation of Pennsylvania, U SA: *in,* Ehlers, J and Gibbard, P. L., eds., Quaternary glaciations extent and chronology, Part II: North America, Developments in Quaternary Science, Elsevier, p. 237-242.

Braun, D. D., 2006, Deglaciation of the Appalachian Plateau, northeastern Pennsylvania—till shadows, till knobs forming “beaded valleys”: Revisiting systematic stagnation-zone retreat; *in,* Fleisher, P. J., Knuepfer

P. L .K., and Butler, D.R., eds., Ice sheet geomorphology - past and present processes and landforms: Geomorphology, v.75, no.1-2, p. 248-265.
Braun, D. D., 2009a, Surficial geology of the Jackson Summit 7.5-minute quadrangle, Tioga County, Pennsylvania and Steuben County, New York: Pennsylvania Geological Survey, 4th ser., Open-File Report OFSM 09–07.0 24 p., Portable Document Format (PDF).
Braun, D. D., 2009b, Surficial geology of the Mansfield 7.5-minute quadrangle, Tioga County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFSM 09–08.0, 25 p., Portable Document Format (PDF).
Braun, D. D., 2010a, Surficial geology of the Blossburg 7.5-minute quadrangle, Tioga County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFSM 10–06.0, 24 p., Portable Document Format (PDF).
Braun, D. D., 2010b, Surficial geology of the Liberty 7.5-minute quadrangle, Tioga and Lycoming Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFSM 10–0.0, 20 p., Portable Document Format (PDF).
Braun, D. D., to be released Aug. 2011, The Glaciation of Pennsylvania, U SA: *in,* Ehlers, J, Gibbard, P. L., and Hughes, P. D. eds., Quaternary glaciations extent and chronology, A closer look: *in*, Vandeer Meer, J., series ed., Developments in Quaternary Science, v. 15, Elsevier, Amsterdam, Netherlands, chap. 40.
Braun, D. D., Ciolkosz, E., and Drohan, P., 2008, The Pleistocene record in the middle and lower Susquehanna River basin and the longer term evolution of the Susquehanna Basin landscape: Guidebook for the 20th Biennial Meeting of the American Quaternary Association, State College, Pennsylvania, 58 p.
Cameron, C. C., 1970, Peat deposits of northeastern Pennsylvania: U.S. Geological Survey, Bulletin. 1317- A, 90 p.
Campbell, M. R., 1903, Geographic development of northern Pennsylvania and southern New York: Geol. Soc. of Am. Bull., v.14, p. 553-573.
Carll, J. F., 1880, Geology of the oil regions of Warren, Venango, Clarion and Butler Counties: Pennsylvania Geological Survey, 2nd ser., Report, III, xxxiv, 482 p.
Chamberlin, T. C., 1883, Terminal moraine of the Second Glacial Epoch: U.S. Geological Survey, 3rd Annual Report, p. 291-402.
Ciolkosz, E. J., Cronce, R. C. and Sevon, W. D., 1986, Periglacial features in Pennsylvania: Pennsylvania State University, Agronomy Series, v. 92: 15 p.
Clark, G. M., Behling, R. E., Braun, D. D., Ciolkosz, E. J., Kite, J. S., and Marsh, B., 1992, Central Appalachian periglacial geomorphology: International Geographical Congress, 27th, Guidebook, Pennsylvania State University Agronomy Series, v. 120, 248 p.
Ciolkosz, E. J., Cronce, R. C., Sevon, W. D., and Waltman, W. J., 1995, Genesis of Pennsylvania's limestone soils; Pennsylvania State University, Agronomy Series, v. 135, 28 p.
Coates, D. R., 1966a, Glaciated Appalachian Plateau: till shadows on hills: Science, v.152, p.1617-1619.
Coates, D. R., 1966b, Geomorphology of the Cowanesque Basin Pennsylvania: Report for U.S. Army Engineer District, Baltimore, 44 pp.
Coates, D. R., 1966c, Discussion of K. M. Clayton "Glacial Erosion in the Finger Lakes Region (New York, U.S.A.)": Z. Geomorphol., v. 10, p. 469-474.
Coates, D. R., 1974, Reappraisal of the glaciated Appalachian Plateau: *in,* Coates, D. R., ed., Glacial geomorphology: Annual geomorphology Symposia, 5th, Publications in Geomorphology, SUNY at Binghamton, New York, p. 205-244.
Coates, D. R. and King, C. A. M., 1973, Glacial geology of Great Bend and adjacent region: *in,* Coates, D. R., ed., Glacial geology of the Binghamton-Western Catskill region; Contribution 3, Publications in Geomorphology, SUNY at Binghamton, New York, p. 3-30.
Coates, D. R. and Kirkland, J. T., 1974, Application of glacial models for large scale terrain development: *in,* Mahaney, W. C., ed., Quaternary environments, proceedings of a symposium: York, Ontario, p. 99-131.
Crowl, G. H., 1981, Glaciation in north-central Pennsylvania and the Pine Creek Gorge: *in,* Berg, T. M. and others, eds., Geology of Tioga and Bradford Counties, Pennsylvania: Annual Field Conference of Pennsylvania Geologists, 46th, Wellsboro, PA, Guidebook, p. 39-44.
Crowl, G. H. and Sevon, W. D., 1980, Glacial border deposits of late Wisconsinan age in northeastern Pennsylvania: Pennsylvania Geological Survey, 4th ser., General Geology Report 71, 68 p.
Crowl, G. H. and Sevon, W. D., 1999, Quaternary: *in,* Shultz, C. H. ed., The Geology of Pennsylvania: Pennsylvania Geological Survey and Pittsburgh Geological Society, Special Publication 1, p. 225-231.
Dalton, T. S., Carson, B., and Meltzer, A. S., 1997, Holocene stratigraphy of Lake Lacawac: Establishment of a framework for paleoecological and paleoclimatic studies: Geological Society America, Abstracts with Programs, v. 29, no.1, p.39.
Delano, H. L. and Wilshusen, J. P., 1999, Landslide susceptibility in the Williamsport 1 by 2 degree quadrangle, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Environmental Geology Report 9,

200 p.
Denny, C. S., 1956, Surficial geology and geomorphology of Potter County, Pennsylvania: U.S. Geological Survey Professional Paper 288, 72 p.
Denny, C. S. and Lyford, W. H., 1954, Wellsboro-Elmira-Towanda Region, Pennsylvania - New York: Friends of the Pleistocene, 17th, Guidebook, Ithaca, N.Y., 27 p.
Denny, C. S. and Lyford, W. H., 1963, Surficial geology and soils of the Elmira-Williamsport region, New York and Pennsylvania: U.S. Geological Survey Professional Paper 379, 60 p.
Fuller, M. L., 1903a, Description of the Gaines quadrangle, Pennsylvania: U.S. Geological Survey Atlas of U.S. Folio 92, 6 maps, 7 p.
Fuller, M. L. and Alden, W. C., 1903b, Pleistocene geology of the Elkland and Tioga quadrangles: *in,* Fuller, M.L., Description of the Elkland-Tioga quadrangles, Pennsylvania; U.S. Geological Survey Atlas of U.S. Folio 93, 6 maps, 9 p.
Gardner, T. W., Braun, D. D., Pazzaglia, F. L., and Sevon, W. D., 1993, Late Cainozoic landscape evolution of the Susquehanna River Basin: International Geomorphology Conference, 3rd, Post-Conference Field Trip Guidebook, 288 p.
Jacobson, R. B., Elston, D. P., and Heaton, J. W., 1988, Stratigraphy and magnetic polarity of the high terrace remnants in the upper Ohio and Monongahela Rivers in West Virginia, Pennsylvania, and Ohio: Quaternary Research, v. 29, p. 216-232.
Kaktins, U. and Delano, H. L., 1999, Drainage basins: *in,* Shultz, C. H. ed., The Geology of Pennsylvania; Pennsylvania Geological Survey and Pittsburgh Geological Society, Special Publication, 1, p. 378-390.
Koteff, C., 1974, The morphologic sequence concept and deglaciation of southern New England: *in,* Coates, D.R., ed., Glacial Geomorphology: Annual Geomorphology Symposia, 5th, Publications in Geomorphology, S.U.N.Y. at Binghamton, New York, p. 121-144.
Koteff, C. and Pessl, F., Jr., 1981, Systematic ice retreat in New England: U.S. Geological Survey Professional Paper 1179, 20 p.
Lessig, H. D., 1963, Calcutta Silt, a very early Pleistocene deposit, upper Ohio Valley: Geological Society of America Bulletin, v. 74, p. 129-139.
Leverett, F., 1902, Glacial formations and drainage features of the Erie and Ohio Basins: U.S. Geological Survey Monograph 41, 802 p.
Leverett, F., 1934, Glacial deposits outside the Wisconsin terminal moraine in Pennsylvania: Pennsylvania Geological Survey, 4th ser., General Geology Report 7, 123 pp.
Lewis, H. C., 1884, Report on the terminal moraine in Pennsylvania and western New York: Pennsylvania Geological Survey, 2d ser., Report on Progress, Z, 299 p.
Lohman, S. W., 1939, Ground water in north-central Pennsylvania, with analyses by E. W. Lohr: Pennsylvania Geological Survey, 4th ser., Bulletin W6, 219 p.
MacClintock, P. and Apfel, E. T., 1944, Correlation of the drifts of the Salamanca Reentrant, New York: Geological Society of America Bulletin, v. 55, p. 1143-1164.
Marchand, D. E., 1978, Quaternary deposits and Quaternary history: *in,* Marchand, D. E., Ciolkosz, E. J., Bucek, M. F., and Crowl, G. H., eds., Quaternary deposits and soils of the Central Susquehanna Valley of Pennsylvania: Friends of the Pleistocene Field Conference, 41st , Pennsylvania State University Agronomy Series 52, p. 1-19.
Marine, J. T., 1997, Terrace deposits associated with ancient Lake Monongahela in the lower Allegheny drainage, western Pennsylvania: Unpublished M.S. thesis, University of Pittsburgh, 182 p.
Marsh, B., 1999, Paleoperiglacial landscapes of central Pennsylvania: Friends of the Pleistocene, 62nd, Guidebook, Lewisburg, Pennsylvania, 69 pp.
Muller, E. H., 1957, Filled bedrock gorges in the drainage basin of Cayuga Lake, New York: Geologic Society of America Bulletin, Abstract, v. 12, pt. 2, p. 1771.
Peltier, L. C., 1949, Pleistocene terraces of the Susquehanna River: Pennsylvania Geological Survey, 4th ser., General Geology Report 23, 158 p.
Ramage, J. M., Gardner, T. W., and Sasowasky, I.D., 1998, Early Pleistocene Glacial Lake Lesley, West Branch Susquehanna River valley, central Pennsylvania: Geomophology, v. 22, p. 19-37.
Rayburn, J. B. and Braker, W. L., 1981, Soil Survey of Tioga County, Pennsylvania: U.S. Department of Agriculture, 94 p., 66 plates.
Renwick, W. H., 1992, Equilibrium, disequilibrium, and nonequilibrium landforms in the landscape: Geomorphology, v. 5, p. 265-276
Richmond, G. M. and Fullerton, D. S., 1986, Summation of Quaternary glaciations in the United States of America: *in,* Sibrava, V., Bowen, D. Q., and Richmond, G. M., eds., Quaternary Glaciations in the Northern Hemisphere; Quaternary Science Reviews, v.5, p. 183-196.
Ridge, J.C., 2003, The last deglaciation of the northeastern United States: A combined varve, paleomagnetic, and calibrated ^{14}C chronology: *in,* Cremeens, D. I. and Hart, J. P., eds., Geoarchaeology of landscapes in

the glaciated northeast; New York State Museum Bulletin 497, p. 15-45.
Ridge, J. C., Braun, D. D., and Evenson, E. B., 1990, Does Altonian drift exist in Pennyslvania and New Jersey?: Quaternary Research, v. 33, no. 2, p. 253-258.
Sevon, W. D. and Braun, D. D., 1997a, Surficial geology of the Wellsboro and Towanda 30x60 minute quadrangles, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report 96-19 and 96-20, 37 p., 2 maps.
Sevon, W. D. and Braun, D. D., 1997b, Glacial deposits of Pennsylvania, 2nd ed.: Pennsylvania Geological Survey, 4th ser., Map 59, scale 1:2,000,000.
Shepps, V. C., White, G. W., Droste, J. B. and Sitler, R. F., 1959, Glacial geology of northwestern Pennsylvania: Pennsylvania Geological Survey, 4th ser., General Geology Report 32, 59 p.
U.S Army Engineer Corps, 1972, Tioga-Hammond Lakes Design Memorandum No.14, Geology, Soils, and Embankment: Baltimore, Maryland, 42 p. and 69 plates.
White, I. C., 1896, Origin of the high terrace deposits of the Monongahela River: American Geologist, v. 18, p. 368-379.
White, G. W., Totten, S. M. and Gross, D. L., 1969, Pleistocene stratigraphy of northwestern Pennsylvania: Pennsylvania Geological Survey, 4th ser., General Geology Report 55, 88 p.
Willard, B., 1932, Glacial Lake Cowanesque: Geological Society America Bulletin, v. 43, p. 441-448.
Williams, E. H., Jr., 1894, The age of the extramoraine fringe in eastern Pennsylvania: American Journal of Science, v. 47, p. 34-37.
Williams, E. H., Jr., 1902, Kansan glaciation and its effects on the river systems of northern Pennsylvania: Wyoming Historical and Geological Society, Proceedings, v. 7, p. 21-28.
Williams, J. H., Taylor, L. E., and Low, D. J., 1998, Hydrogeology and groundwater quality of the glaciated valleys of Bradford, Tioga, and Potter Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resources Report 68, 89 p., 2 maps.
Wilshusen, J. P. and Wilson, D., 1981, Tioga-Hammond flood control project: *in,* Berg, T. M. and others, eds., Geology of Tioga and Bradford Counties, Pennsylvania: Annual Field Conference of Pennsylvania Geologists, 46th, Wellsboro, PA, Guidebook, p. 77-89.

Bedrock geology of the "Grand Canyon" region of north-central Pennsylvania

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INTRODUCTION

The northern tier of Pennsylvania, in the Potter-Tioga county area, is divided into three physiographic provinces, consisting of the Glaciated Low Plateaus, the Glaciated High Plateaus, and the Deep Valleys Section (Sevon, 2000). The geomorphology of these regions reflects Pleistocene glacial modification of gently folded, Paleozoic-age, clastic sedimentary rocks. Dissection of the landscape exposes a succession of Upper Devonian – Pennsylvanian rocks with a stratigraphic thickness greater than 700 m arranged in a series of northeast-southwest trending anticlines and synclines (Figure 2). This portion of the guidebook focuses on the stratigraphic and structural framework of the Paleozoic bedrock within the Wellsboro 30x60 quadrangle. Many of the descriptions and observations outlined below are from bedrock mapping of the Mansfield (McLaurin, 2010) and Blossburg (McLaurin and Dodge, in press) 7.5' quadrangles in the area east of Wellsboro Borough.

STRATIGRAPHIC NOMENCLATURE

The stratigraphic units within the Wellsboro area range from Devonian to Pennsylvanian in age, encompassing marine rocks at the base of the section within the Upper Devonian Lock Haven Formation to primarily nonmarine rocks within the Pennsylvanian Allegheny and Pottsville Formations (Figure 1). Geologic studies of the area began as part of the First Geological Survey of Pennsylvania (Rogers, 1858). Rogers (1858) recognized synclinal "mountains or table-lands" with sandstone escarpments, conglomerate and coal characteristically forming "coal-basins" while distinguishing broad anticlinal valleys underlain by calcareous sandstone and shale. The stratigraphic nomenclature utilized by Rogers (1858) for the northeastern district of Pennsylvania, covering the Wellsboro quadrangle, included the Vergent shales, Ponent series, Vespertine series, Umbral red shales, Seral conglomerate/Coal measures. These units are roughly equivalent to the modern designations of Lock Haven, Catskill, Huntley Mountain/Burgoon, Mauch Chunk, and Pottsville/Allegheny units, respectively. Later mapping by Sherwood (1878) as part of the Second Geological Survey of Pennsylvania used a mix of terminology that included the Chemung, Catskill and Pocono Sandstone for the lower units, but adhering to Rogers (1858) use of Umbral, Seral and Coal measures for the uppermost stratigraphic units. Mapping by Fuller (1903) of the Elkland and Tioga 15' quadrangles utilized the term Chemung for the lowermost calcareous sandstone and shale but abandoned the Catskill and Pocono nomenclature in favor of the New York terms Cattaraugus (Clarke, 1902) and Oswayo (Glenn, 1903), respectively. This change in terminology was attributed to Catskill – Pocono stratigraphic "…distinctions which do not hold in the region under consideration…" (Fuller, 1903, p. 2). The modern stratigraphic nomenclature in this region of the Appalachian Plateau has evolved since the time of the First and Second Pennsylvania Geological surveys. The use of the term Chemung for the lowermost "Vergent shales" was questioned by Frakes (1963) who noted it relied on biostratigraphic criteria instead of lithostratigraphic description which is necessary to establish formations according to the North American Stratigraphic Code (NACSN, 2005). Thus, he encouraged restriction of the term "Chemung" to a localized time-stratigraphic unit. Faill and Wells (1977) established the Lock Haven Formation to encompass this largely marine stratigraphic interval between the Brallier and Catskill Formations. Further revisions of stratigraphic nomenclature were proposed by Berg and Edmunds (1979) who defined the Huntley Mountain Formation to cover the Catskill-Burgoon transition. The use of Huntley Mountain Formation is restricted to the northern part of Pennsylvania and replaces the Pocono and Oswayo previously established in this area. The Burgoon Sandstone was originally defined in the within the southeastern areas of the Wellsboro quadrangle (Berg et al., 1980), but recent mapping (McLaurin and Dodge, in press) in the Blossburg quadrangle suggest that the Pottsville Formation disconformably overlies the Huntley Mountain Formation and the Burgoon Sandstone is absent due to erosion or a possible facies change below the sub-Pottsville unconformity (Berg and Edmunds,

1979; Edmunds, 1981). The coal-bearing Pennsylvanian Pottsville and Allegheny groups as defined by Dodge (2010) are separated based on the position of coal bed markers. Due to the difficulty in defining formations within these groups because of poor exposure, the recommendation of Edmunds (1996) is followed reducing the stratigraphic rank of the Pottsville and Allegheny from group to formation. The boundary between these formations as defined by Dodge (2010), is the position of the Cannel coal seam. Accurate mapping of this boundary is difficult considering the poor outcrop, thus the position of this coal marker is based on compilation of drilling records and the location of coal mine workings.

STRATIGRAPHY

The stratigraphic nomenclature applied is largely that used for compilation of the geologic map of Pennsylvania (Berg et al., 1980). The stratigraphic units described below that occur in the Wellsboro quadrangle are the Upper Devonian Lock Haven Formation and Catskill Formations, the Upper Devonian-Mississippian Huntley Mountain Formation, the Pennsylvanian Pottsville Formation and the Allegheny Formation. Although the Mauch Chunk Formation and Burgoon Sandstone have been described within the Wellsboro quadrangle, they are primarily confined to the southeastern portion of the quadrangle (Colton and Luft, 1965; Colton, 1963) and are not discussed in this report. Figure 3 is the geologic map of the Wellsboro 30x60' quadrangle as compiled from Berg et al. (1980)

Upper Devonian

Lock Haven Formation. The Lock Haven Formation is the oldest stratigraphic unit exposed in the Wellsboro area and occupies broad anticlinal valleys of rolling topography. Approximately 130 m of the upper, shaly portion of the unit is exposed which is small compared to the total estimated thickness of 1181 m described in the type section in the Williamsport area by Faill and Wells (1977). Exposure of the Lock Haven is mainly limited to road cuts along US Hwy 15, parallel secondary roads and rare outcropping along creeks. The Lock Haven Formation is characterized by interbedded gray-green shale with thin, flaggy, very fine- to fine-grained, gray to gray-green sandstone (Figure 4) that is often calcareous. The sandstones are massive to laminated and may contain brachiopod fossil lags and bioturbated zones. The sandstone units are typically less than 1 m in thickness, but can occasionally exceed 2 m. Shale intervals are up to 8 m thick, but when interbedded with sandstone are 50 cm to 1 m thick. The Lock Haven Formation is interpreted to represent deposition in a marine foreland ramp setting (Castle, 2000) with distal ramp deposits indicated by gray fissile shale. Proximal ramp deposits are characterized by laminated very fine-grained sandstone. Storm deposits are suggested based on the occurrence of coarse lags of abraded brachiopod fossils (Castle, 2000).

Catskill Formation. The Catskill Formation shows a gradual northward thinning from approximately 600 m in the Williamsport area (Faill and Wells, 1977) to 260 m in the Mansfield quadrangle (McLaurin, 2010). The base of the Catskill is delineated at the first red mudstone/shale that begins an interval of approximately 80 m of alternating red Catskill-type sandstone (Figure 5A) and shale with gray-green Lock Haven-type sandstone and shale. This transitional zone between the Lock Haven and Catskill formations is comparable in lithology and thickness to the Irish Valley Member of the Catskill Formation in the Williamsport area (Faill and Wells, 1977). Within this zone there are five of these red/gray-green couplets that reach a maximum thickness of 24 m. The sandstone within this interval is fine grained with rip-up clasts and cross lamination. Interbedded with the sandstone are heterolithic layers containing 1 to 2 cm thick sandstone and 3 to 5 cm thick shale. The gray-green intervals within the lower Catskill look similar to Lock Haven lithologies but lack the brachiopod fossils.

Overlying the transitional zone is a red interval of very fine to fine-grained, cross laminated sandstone that is arranged in fining-upward cycles that culminate in finer-grained interbedded sandstone and shale. This part of the Catskill Formation is exposed in a road cut along US Hwy 15 N, just north of Blossburg and is approximately 130 m thick. Sandstone units vary in thickness from less than 1 m to greater than 6 m. Sandstone and shale interbedded units are up to 24 m thick. This interval within the Catskill is similar to the Sherman Creek Member, although it is considerably

thinner than has been mapped in other areas (Faill and Wells, 1977).
The uppermost part of the Catskill Formation is characterized by gray-red, fine to medium-grained, cross-bedded sandstone (Figure 5B). These sandstones are occasionally amalgamated into larger sheet sandstones that exceed 10 m in thickness. Interbedded red, clayey silt, and shale units intercalated with thin, very fine-grained sandstones can exceed 10 m thick.
Although there is a tripartite stratigraphy of informally defined members within the Catskill Formation in this area, the overall exposure of these units is poor. Extension of these informal units into parts of the study area away from road cuts is tenuous, thus the Catskill Formation is mapped as a single, undivided unit. The Catskill Formation is interpreted as a prograding delta complex of interbedded alluvial plain deposits and shallow marine muds (Cotter and Driese, 1998) in the lower zones. Within the upper part of the formation, thick crossbedded sandstone units and red shales are indicative of deposition within moderate to high-sinuosity alluvial systems.

Upper Devonian –Mississippian

Huntley Mountain Formation. The Huntley Mountain Formation underlies the higher elevations and forms steeper slopes compared to the underlying Catskill and Lock Haven formations and is much sandier. It is a minimum of 140 m thick. The contact between the Catskill and overlying Huntley Mountain Formation was recognized by Colton (1968) as the top of the highest occurrence of red sandstone within a red-bed succession. The Huntley Mountain Formation is characterized by gray to buff, fine to medium-grained, cross-bedded sandstone (Figure 6A). The sandstone occurs as amalgamated sheets that are up to 10 m thick, and average 5 to 6 m. Gray shale/siltstone intervals are not readily observed in outcrop, but are often present in road cuts where they rarely exceed 3 m thick. There is a succession of green and red shale approximately 60 – 80 m above the base of the Huntley Mountain Formation that is 5 m thick. Sandstones contain a high density of erosion surfaces and where shale intervals are observed they often cannot be traced more than a few meters laterally before they are cut out by overlying channels. The lowermost parts of the Huntley Mountain Formation contain finer-grained zones that are composed of interbedded very fine-grained sandstone and gray siltstone/shale. A distinctive, uppermost interval of Huntley Mountain Formation is 15 – 20 m thick and includes interbedded red shale (Figure 6B) and gray to buff sandstone that occurs as fine- to medium-grained and very fine- to fine-grained beds, 50 – 80 cm thick. Pebbly zones are present in the sandstone and consist of up to 1 cm diameter red chert pebbles. The Huntley Mountain Formation is interpreted as fluvial deposits of low to moderate sinuosity river systems (Berg and Edmunds, 1979). Channel migration and avulsion were frequent processes as indicated by the lack of thick, preserved muddy floodplain deposits. The thicker successions of red mudstone and shale suggest establishment of an extensive, stabilized floodplain with periodic crevasse splay deposition.

Pennsylvanian

Pottsville Formation. The Pennsylvanian units occupy the highest elevations within the synclinal ranges throughout the Wellsboro quadrangle. Many of these areas are covered by Wisconsinan till deposits that average 5 m thick (Braun, 2010). The Pottsville Formation contains buff, medium to coarse-grained sandstone 2 – 3 m thick (Figure 7). The sandstone contains 30 cm thick conglomeratic interbeds with quartz clasts up to 1 cm in diameter. Sandstone units are frequently crossbedded in sets 3 – 20 cm thick and may also contain interbeds of gray shale 10 cm thick. Approximately 20 m above the basal sandstones, gray-black, paper-thin shale ranging in thickness from 1 – 6 m is interbedded with buff-gray, fine to medium-grained crossbedded sandstone. Eight to nine coal seams have been identified within the Pennsylvanian in Tioga County (Rogers, 1858; Sherwood, 1878; Dodge, 2010). The coal seams within the Pottsville Formation are designated the Kidney, Bear Creek, and Bloss and are up to 1 m thick. The Pottsville Formation is at least 48 m thick.

Allegheny Formation. The Allegheny Formation is primarily observed in the eastern areas of the Wellsboro quadrangle along the axis of the Blossburg syncline. It is also present northwest of

Wellsboro Borough within the Pine Creek syncline. The other synclinal ridges within the area do not have Allegheny Formation present due to erosion. The poor quality of Allegheny exposure and the difficulty in identifying the key marker beds such as the Cannel coal seam in outcrop require the use of subsurface drilling records and coal mining reports to accurately map the contact between the Pottsville and overlying Allegheny Formation. Those data indicate that the estimated minimum thickness of the Allegheny Formation is 70 m. The lithologies present within the Allegheny Formation are cyclical successions of sandstone, shale, claystone, conglomerate, and coal. The coal seams that are assigned to the Allegheny Formation are the Cannel, Morgan, Seymour, and Rock.

STRUCTURE

The folding within this area of the Appalachian Plateau Province was noted by Rogers (1858) and Sherwood (1878) as an *en echelon* arrangement of broad folding with the anticlines occupying topographically lower valleys and the synclines represented by higher ridges (Figures 1 and 8). This is in contrast to the steeper dips of bedding within the Valley and Ridge Province to the southeast. The strike and dip of bedding shows that the strata, on average, have a low dip from 4°to 6°. There is some indication based on the field data that the north limbs of the synclines are slightly steeper dipping than the south limbs. This observation is consistent with that of Sherwood (1878) and Rogers (1858) who noted that the anticline limbs show steeper southerly dips and more shallow northerly dips. The orientations of the fold axes show a trend change from more northeasterly in the western part of the Wellsboro quadrangle (56°– 236°) to east-northeasterly (74°– 254°) in the eastern areas.

REFERENCES

Berg, T.M., McInerney, M.K., Way, J.H. and MacLachlan, D.B., 1993, Stratigraphic correlation chart of Pennsylvania : Pennsylvania Geological Survey, 4th ser., General Geology Report 75

Berg, T.M., Edmunds, W.E., Geyer, A.R., Glover, A.D., Hoskins, D.M., MacLachlan, D.B., Root, S.I., Sevon, W.D. and Socolow, A.A., 1980, Geologic map of Pennsylvania: Pennsylvania Geological Survey, 4th ser., Map 1, 2nd ed., 3 sheets, scale 1:250,000.

Berg, T.M. and Edmunds, W.E., 1979, The Huntley Mountain Formation: Catskill to Burgoon transition in north-central Pennsylvania: Pennsylvania Geological Survey, 4th ser., Information Circular 83, 80 p.

Braun, D.D., 2010, Surficial geology of the Blossburg 7.5-minute quadrangle, Tioga County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFBM 10–08.0, 25 p

Castle, J.W., 2000, Recognition of facies, bounding surfaces and stratigraphic patterns in foreland successions: An example from the Upper Devonian, Appalachian Bain, U.S.A.: Journal of Sedimentary Research, v. 70, p. 896-912.

Clarke, J.M., 1902, Preliminary statement of the paleontologic results of the areal survey of the Olean quadrangle, *in* Clarke, J.M., Report of the State Paleontologist, 1901: New York State Museum Bulletin No. 52, p. 524 – 528.

Colton, G.W., 1968, Bedrock geology of the Waterville quadrangle, Lycoming County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Progress Report 174.

Colton, G.W. and Luft, S.J., 1963, Bedrock geology of the Slate Run quadrangle, Clinton, Lycoming and Potter Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Progress Report 167.

Colton, G.W., 1963, Bedrock geology and surface structure of the Cedar Run quadrangle, Tioga and Lycoming Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Progress Report 164.

Cotter, E. and Driese, S.G., 1998, Incised-valley fills and other evidence of sea-level fluctuations affecting deposition of the Catskill Formation (Upper Devonian), Appalachian Foreland basin, Pennsylvania: Journal of Sedimentary Research, v. 68, p.347-361.

Dodge, C.H., 2010, Coal geology of Pennsylvania, *in* 2010 Keystone Coal Industry Manual, Mining Media International, p. 523-544.

Edmunds, W.E., 1996, Correlation chart showing suggested revisions of uppermost Devonian through Permian stratigraphy, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open File Report 96-49, 8 p.

Edmunds, W.E., 1981, Mississippian-Pennsylvanian disconformity and Pennsylvanian stratigraphy, Guidebook for the Annual Field Conference of Pennsylvania Geologists: United States, [Field Conference of Pennsylvania Geologists]: [Harrisburg, PA], United States, p. 33-38.

Faill, R.T. and Wells, R.B., 1977, Bedrock geology and mineral resources of the Salladasburg and Cogan Station quadrangles, Lycoming County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas

133cd, 44 p.
Frakes, L.A., 1963, Stratigraphy of the nonred Upper Devonian across Pennsylvania, *in* Shepps, V.C., ed., Symposium on Middle and Upper Devonian stratigraphy of Pennsylvania and adjacent states: Pennsylvania Geological Survey, 4th ser., General Geology Report G39, p. 183-199.
Fuller, M. L., 1903, Description of the Elkland and Tioga quadrangles, Pennsylvania: U.S. Geological Survey Folio No. 93, 9 p. and plates.
Glenn, L.C., 1903, Devonic and Carbonic formations of southwestern New York, *in* Clarke, J.M., Report of the State Paleontologist, 1902: New York State Museum Bulletin No. 69, p. 967 – 989.
Ingham, A.I., 1951, Geologic structures of the northern plateaus region, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Progress Report 138.
McLaurin, B. T., 2010, Bedrock geologic map of the Mansfield quadrangle, Tioga County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFBM 10–03.0, 11 p.
North American Commission on Stratigraphic Nomenclature (NACSN), 2005, North American Stratigraphic Code: AAPG Bulletin, v. 89, p. 1547-1591
Rogers, H.D., 1858, The geology of Pennsylvania: A government survey: Edinburgh, William Blackwood and Sons, v. 1, 586 p.
Sherwood, A., 1878, Limits of the Catskill and Chemung formations, *in* Report of Progress in Bradford and Tioga counties: Second Geological Survey of Pennsylvania, G, p. 1 – 96.
Sevon, W. D., compiler, 2000, Physiographic provinces of Pennsylvania: Pennsylvania Geological Survey, 4th ser., Map 13, 4th ed., 1:2,000,000 scale.

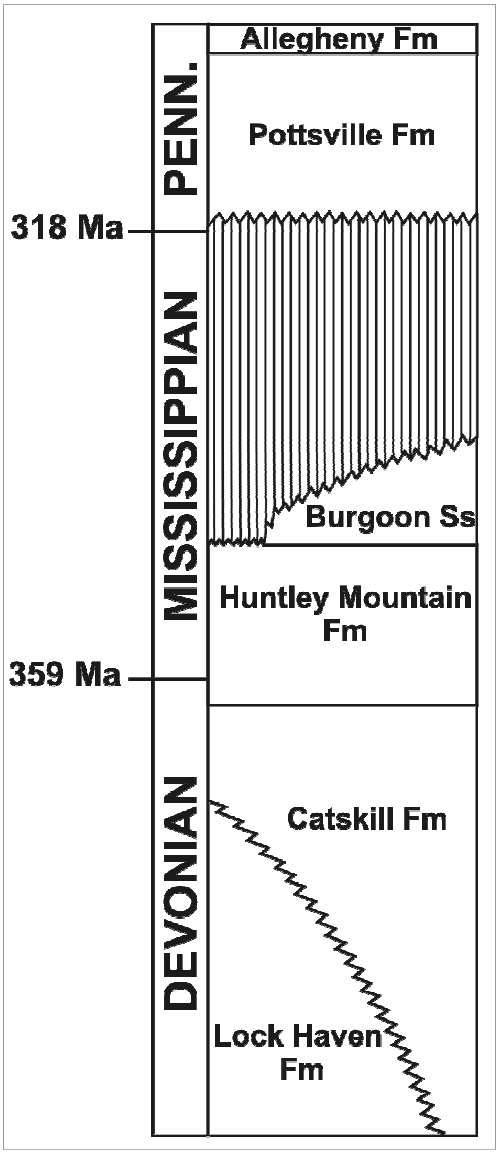

Figure 1. Composite stratigraphic column for the Tioga-Potter county area of the northern tier of Pennsylvania. Modified from Berg et al. (1993).

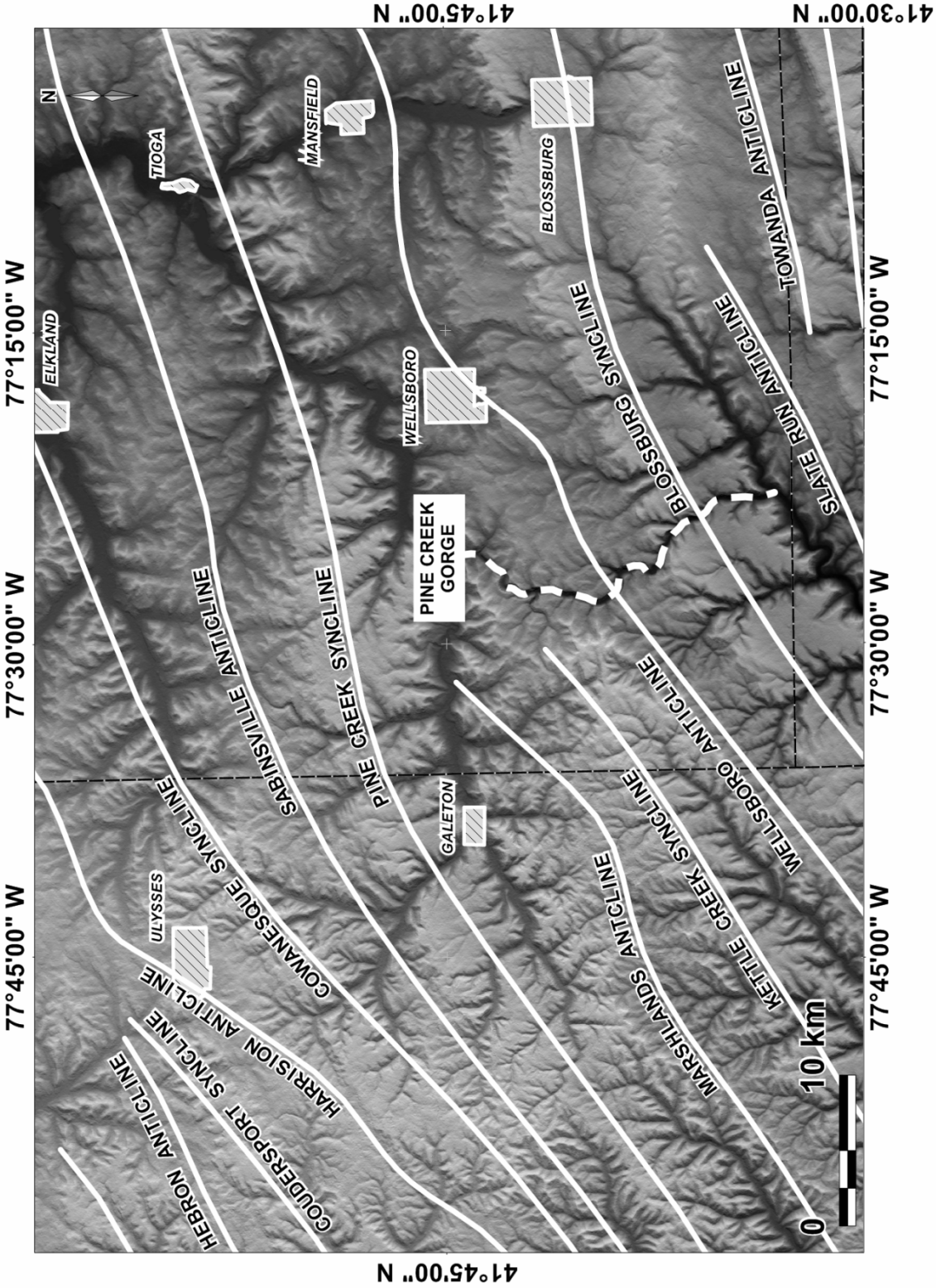

Figure 2. Shaded relief map of the Wellsboro 30x60' quadrangle with the locations of Appalachian Plateau folds delineated. Fold axes modified from Ingham (1951).

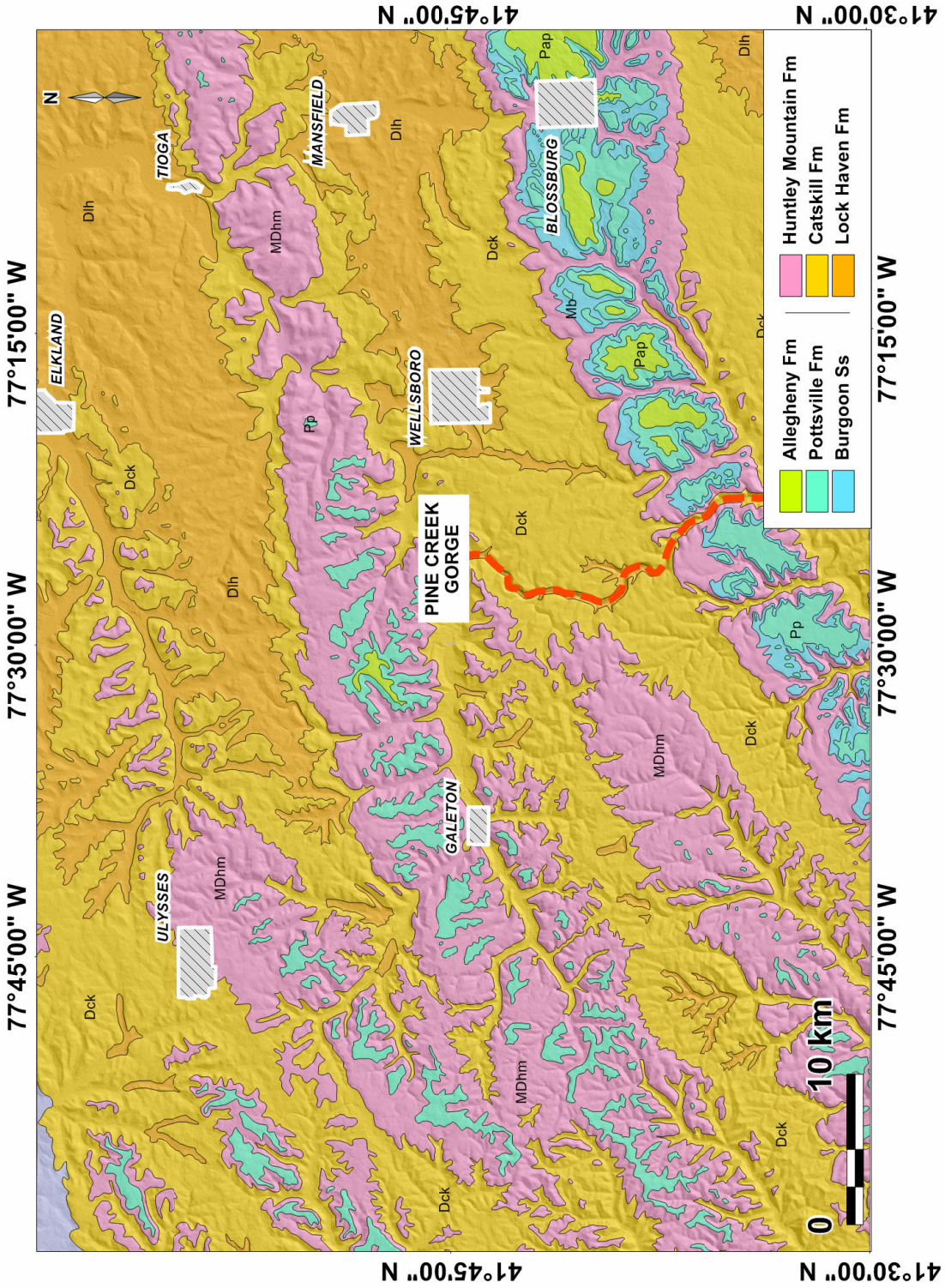

Figure 3. Generalized geologic map of the Wellsboro 30x60' quadrangle. Pine Creek gorge highlighted as red dashed line. Geology was taken from Berg et al. (1980).

Figure 4. Interbedded very fine sandstone and shale within the Lock Haven Formation.

and

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Figure 5. A. Interbedded sandstone shale in the lowermost portion of the Catskill Formation. B. Crossbedded sandstone in lower Catskill Formation.

Figure 6. A. Fine- to medium-grained channel sandstone within the Huntley Mountain Formation. B. Shales in the upper part of the Huntley Mountain Formation, just below the Pottsville Formation.

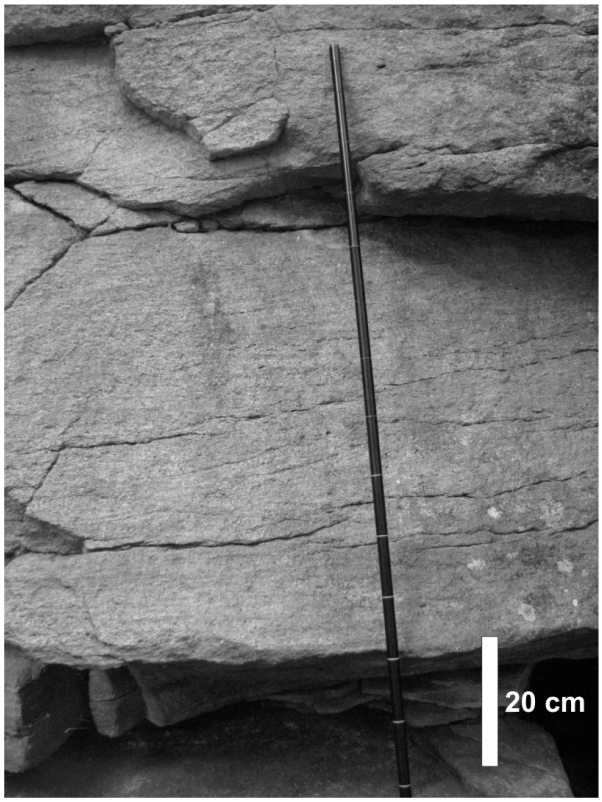

Figure 7. Conglomeratic, crossbedded sandstone from the Pottsville Formation.

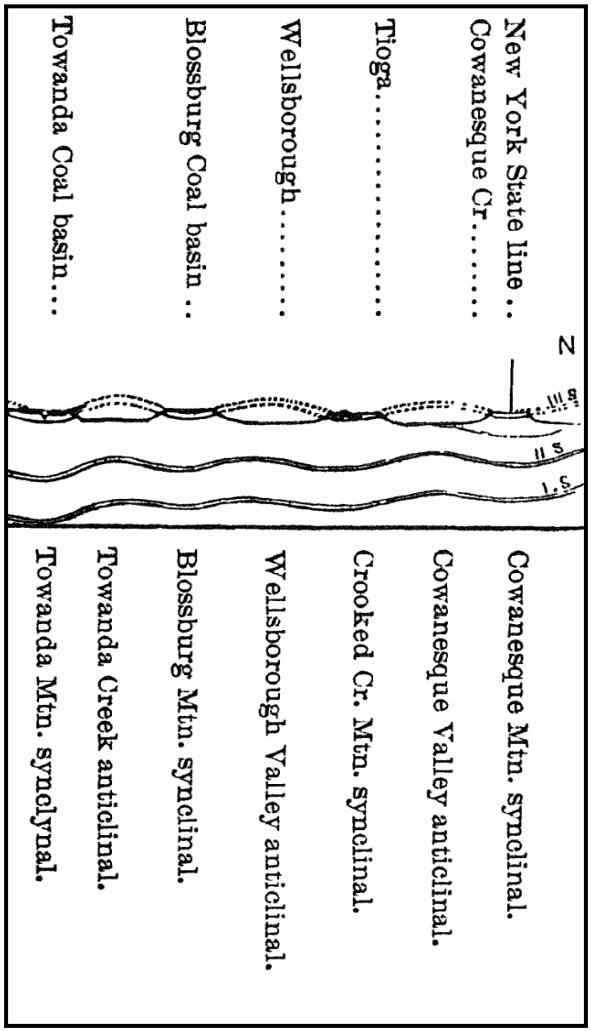

Figure 8. Schematic cross section across the Appalachian Plateau folds beginning at the Towanda Mtn. syncline on the south to the New York State Line on the north. Modified from Sherwood (1878**).**

Gas Drilling in the Marcellus Black Shale

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Overview

Pennsylvania's natural resources have been exploited repeatedly: lumber during the 18th and 19th centuries, coal from the 18th century to the present, and oil since the beginning of the petroleum industry in northwestern Pennsylvania in 1859. Environmental degradation accompanies each wave of natural resource development, and the potential for additional environmental impacts associated with the current push to develop more of Pennsylvania's natural gas resources is causing alarm. Newspaper articles, blogs, even movies play up possible downsides of the new natural gas drilling boom and well stimulation techniques.

More than 1 billion gallons of groundwater are drawn from Pennsylvania aquifers each year, and more than half of that water is used by nearly one quarter of Pennsylvanians living in more than 1 million residences (Swistock, 2007). Groundwater is the primary residential water source in rural areas, including the active drilling areas in northern-central and northeastern Pennsylvania (Fig. 1). Most of these wells are shallow and rarely tested, and many were installed with poor well construction techniques. Pennsylvania does not regulate the construction details of residential water wells [Pennsylvania Department of Conservation and Natural Resources, 2010]). The Susquehanna and its tributaries cover much of the northern extent of the Marcellus Formation, are a source of drinking water for millions of people in New York, Pennsylvania, and Maryland, and provide over half the fresh water to the Chesapeake Bay.

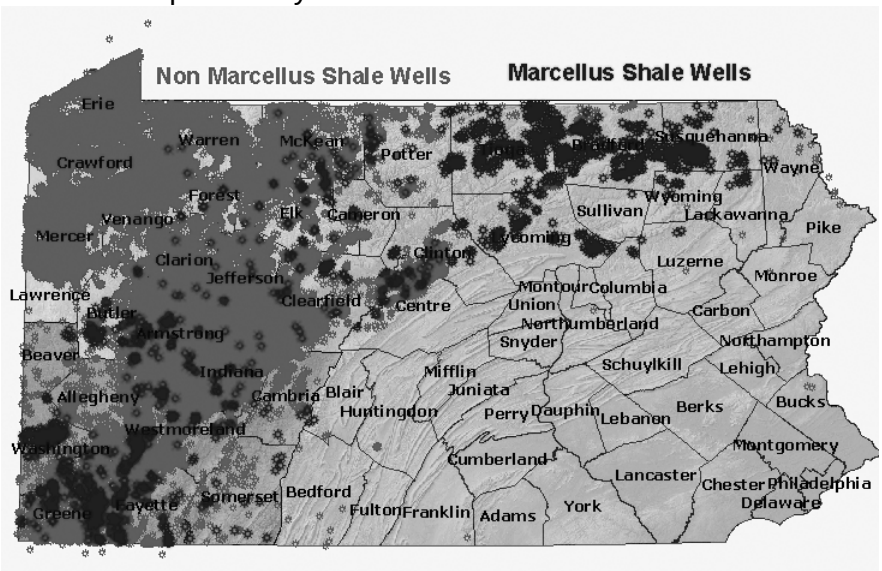


Figure 1. Pennsylvania active oil and gas well locations. Nearly 4000 wells have been drilled in the Marcellus Formation since 2003. Modified from PADEP (2010).

Many chemicals added to the hydraulic fracturing fluid used to stimulate wells, and substances such as brines and radionuclides returned to the surface from fractured rock units are known to be detrimental to human health, and may have adverse impacts on surface water, soil, and groundwater. While residents of the economically depressed Northern Tier appreciate the monetary benefits of leases, royalties, and increased secondary employment, they do have environment-related concerns including: 1) withdrawal of large volumes of fresh water from surface or groundwater sources required for hydraulic fracturing, 2) contamination of soil, surface water or groundwater by constituents of the frackwater (sent down the well for hydraulic fracturing or "fracking") and/or flowback water (water pumped back out of the formation after fracking), and 3) contamination of groundwater with methane. Many homeowners are most concerned with migration of frack and flowback fluids. Cases in which hydraulic fracturing fluids migrated from depth to

useable aquifers have not been documented in Pennsylvania or definitively proven in shale gas plays in other areas of the country. What have been documented, however, are chemical, fuel, frackwater, and flowback water spills, surface and subsurface methane leaks, blowouts, poor gas and water well construction, and poor flowback water disposal practices that almost certainly will have the occasional acute and cumulative adverse affect on water resources in this area.

A little background

Organic-rich shales across the continental United States have recently attracted attention as major gas plays (Fig. 2). The Marcellus Formation in particular is one of a number of Devonian shales that extend along the east side of the Appalachians and across parts of the Michigan and Illinois basins (Figs. 2 and 3). The black, organic-rich Marcellus and its correlatives recorded early development of and deposition in an anoxic, sediment-starved foredeep associated with the onset and southward spread of the Acadian orogeny (Faill, 1985). Younger Devonian shales to the west in Pennsylvania and Ohio represent deposition in the westward-migrating basin Acadian foreland basin during more active sedimentation.

Gas originating from black shales has been produced since the early days of petroleum exploration. For example, Devonian organic-rich shales have been identified as source rocks for hydrocarbons in more porous and permeable units such as the Oriskany (Repetski et al., 2008). Black shales themselves, in areas of high organic content and high natural fracture density, have also been successfully drilled in the Appalachians for many years. In fact, in 1821, the first Devonian gas shale wells were used to light gas street lamps in Fredonia, New York (U.S. Department of Energy [DOE], 2007). Shale units were slow to acquire widespread recognition as both source and reservoir, however, as low porosity and permeability and a lack of understanding of characteristics likely to result in economical, long-lasting production prevented most major oil and gas companies from making large investments. Rising oil prices, the Mideast oil crisis, and fears of dwindling U.S. natural gas reserves spurred the DOE to encourage exploration and exploitation of U.S. energy resources by sponsoring the

Figure 2. Shale gas plays in the lower 48 states. From the Energy Information Administration (2010).

Unconventional Gas Research (UGR) programs in the 1970s and 1980s. The Eastern Gas Shales Project, one of several UGR programs, was developed to conduct basic research on how to encourage better gas recovery from known resources, including initiation of the first directionally drilled well in Appalachian Devonian shales in West Virginia in 1972 (DOE, 2007). Despite major investment and research by the DOE and the industry-founded, surcharge-funded Gas Research

Institute, gas shales still were not widely developed as plays until the economics of the play changed in the early 2000s. Increased natural gas prices and a new efficiencies from combining existing drilling and well stimulation technologies (horizontal drilling and hydraulic fracturing) that proved successful in the Barnett Shale in the Fort Worth basin of north-central Texas (Fig. 2) started the boom. Not until 2003 did activity spread from Texas to other areas such as the Fayetteville Shale in Arkansas and the Marcellus Formation in the Appalachian basin. In 2003, the Marcellus Formation in a Range Resources well in southwestern Pennsylvania showed promise, though it was not the original drilling target. Additional time and money were invested in the horizontal drilling and hydraulic fracturing techniques that had been successfully employed in the Barnett Shale, with great success (Harper, 2008). Since that time, permitting and completion of Marcellus wells in Pennsylvania has increased by leaps and bounds (there were 375 permits for suspected Marcellus wells in 2007 (Harper, 2008); 195 Marcellus wells were drilled in 2008, 768 in 2009, 1386 in 2010, and 515 Marcellus wells drilled by the end of April 2011 (Pennsylvania Department of Environmental Protection (PADEP), 2009, 2010, 2011). Marcellus drilling has been concentrated in two areas: the southwestern corner of the state, primarily in Washington and Greene counties, south of Pittsburgh and in the Northern Tier, especially Tioga, Bradford, and Susquehanna counties

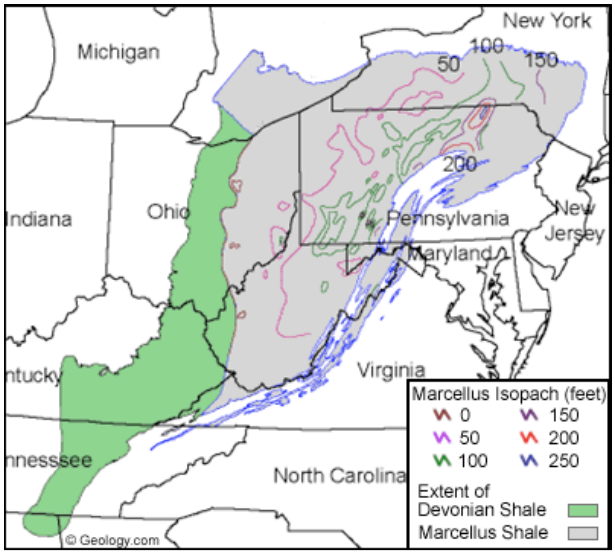


Figure 3. Extent and thickness of Marcellus and other Devonian shales. After Milici and Swezey (2006).

(Fig. 1). In the period from July 2009 through June 2010, slightly more than 4000 active Marcellus wells in Pennsylvania produced more than 194,000 million cubic feet of natural gas (PADEP, 2010), more than twice as much as was consumed in the entire state during 2009 (U.S. Energy Information Agency [EIA], 2011(a)).

The technologies that make drilling tight shales economical also increase potential exploration impacts. Horizontal or directional drilling allows the well bore to extend vertically to the desired depth, then kick off at an angle and extend up to several thousand feet through the area of interest. This technique provides a distinct advantage in tight formations such as the Marcellus by greatly increasing the well drainage area in the reservoir. However, an increased well length in the producing formation requires far more hydraulic fracturing fluid (and therefore greater volumes of added chemicals) than conventional vertical wells, which in turn requires more extensive land disturbance to accommodate more frack-fluid-carrying vehicles. The average pad size for 376 measured well sites was 3.1 acres (Johnson, 2010).

Hydraulic fracturing is a technique that increases the porosity and permeability of otherwise low conductivity lithologies by breaking the rock apart via injection of high-pressure (~70000 kPa) fluid. Chemicals such as thickeners, friction reducers, biocides, scale inhibitors, and surfactants are commonly mixed into the frack water along with proppants such as sand. Many chemicals included in the frackwater are known to be human health hazards (ex: benzene) at the concentrations at

which they are injected. The proppants prop open the newly formed fractures that otherwise would close under the high confining pressures, increase conductivity, and maintain the increased surface area from which natural gas can desorb. Frack fluid additives are expensive for high-volume horizontal fracking, so a less expensive method is preferred for Marcellus wells. Slickwater fracking requires less proppant, uses low friction fluid, and relies on development of natural asperities to keep fractures open (Harper, 2008).

Geology

North-central Pennsylvania lies primarily in the Appalachian Plateau, an area where rocks of Cambrian through Permian age were only mildly deformed in the Alleghanian foreland. Broad, elongate ENE-WSW-trending folds with shallowly dipping limbs fade almost to non-existence from west to east across the region. Devonian to Pennsylvanian rocks are exposed at the surface in north-central Pennsylvania and, in general, outcrop of older (Devonian) strata becomes more scarce to the west, and outcrop of younger (Mississippian and then Pennsylvanian) strata becomes more common. The physiographic provinces also change from east to west across the north-central region from the Glaciated Low Plateau section in the east, characterized by rounded hills and valleys of low to moderate elevation dissected by dendritic drainages in mainly Devonian strata, to higher maximum elevations where Mississippian and Pennsylvanian outcrop in the west. Just west of the Glaciated Low Plateau lies the Glaciated High Plateau, with elongate uplands and shallow valleys. Still further west and surrounding the isolated lozenges of Glaciated High Plateau lie the Deep (unglaciated) Valleys – deep, angular valleys separated by broad to narrow uplands) (Sevon, 2000). The southern part of the region lies in the Appalachian Valley and Ridge, which consists here of elongate ENE-WSW-trending valleys and ridges whose locations and orientations are controlled by lithology and Alleghanian folding (Sevon, 2000).

Northwestern and northeastern Pennsylvania have been covered by glaciers at least four times in the last 2 million years, as recorded by deposits of pre-Illinoian-G (850 Ka) age, late Illinoian (150 Ka) or (pre-Illinoian-B (450 Ka) age, and Wisconsinan (26 Ka) age (Fig. 4) (Braun, 1994, 1999, 2002). In northwestern Pennsylvania, glacial depositional processes dominated, and remnants of

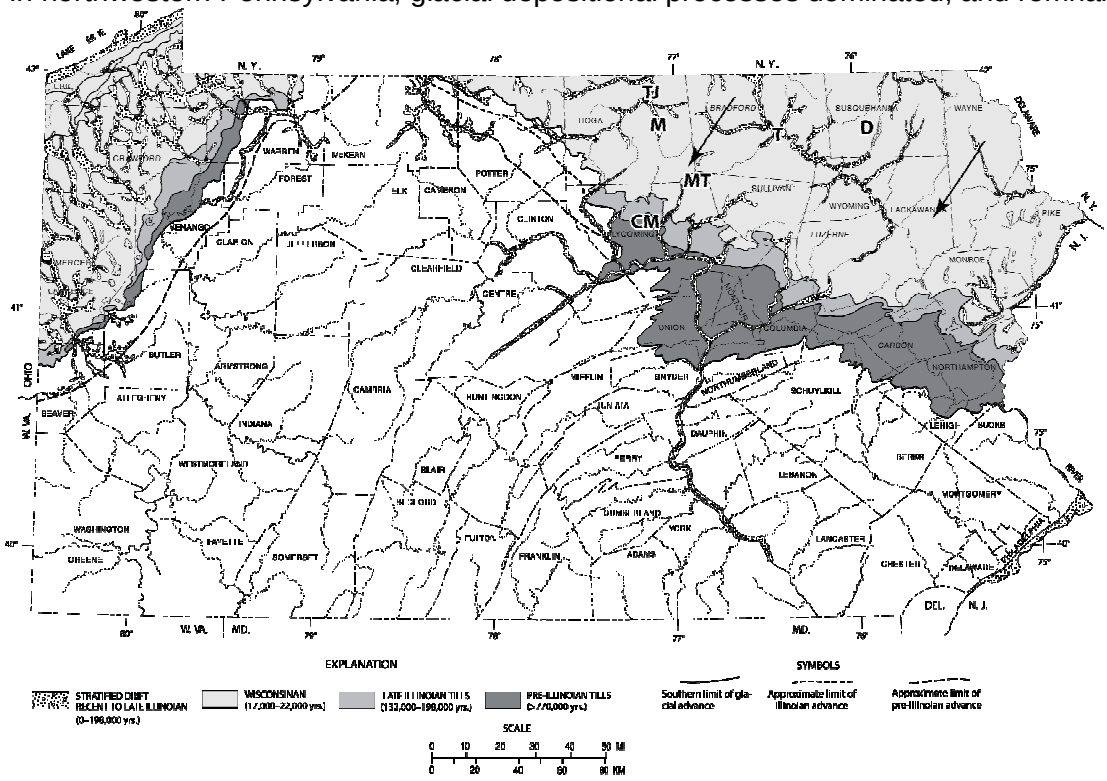


Figure 4. Glacial deposits of Pennsylvania. Arrows show regional ice flow direction. CM – Cogan House and Mifflin Townships, D - Dimock, M – Mansfield, MT – McNett Township, TJ – Tioga Junction, T – Towanda. After Sevon and Braun (1997b) and Braun (2002).

successive glacial advances are preserved in a vertical sequence. In northeastern Pennsylvania,

erosional processes dominated during glaciation, and most older drift was eroded away by successive glaciations; older deposits exist primarily where they extended beyond the terminus of subsequent, younger glaciations (Crowl and Sevon, 1999). Valleys running parallel to ice movement direction (Fig. 4) are more likely to have been scoured; valleys running transverse to ice movement are more likely to preserve accumulated glacial deposits, especially on the up-ice side of the valley (Braun, 2002). The lithology and grain size of the till is largely controlled by bedrock lithology (Williams et al., 1998). Till atop finer-grained clastic bedrock (generally Devonian) tends to be finer grained than till deposited on coarser-grained bedrock (generally Mississippian and Pennsylvanian) (Sevon and Braun, 1997a, 1997b). Evidence of till-rich ice stagnation and retreat is preserved in the form of "beaded valleys" (Braun, 2002), in which flatter areas (now often wetlands), which were covered by stagnant ice during still-stands, are separated by till knobs--the eroded remains of moraines deposited sequentially by active ice atop the ice remaining from the previous still-stand. Stream valleys may also be mantled with ice-contact stratified drift, sometimes in the form of kame terrace deposits laid down by sediment-laden meltwater flowing along valley sides and kame deltas or alluvial fans where tributary streams flow into larger, sometimes lake-filled valleys (Sevon and Braun, 1997a, 1997b; Williams et al., 1998; Braun, 2002,). Many valleys in north-central Pennsylvania are filled with as much as 200 feet of glacial till, ice-contact stratified drift, glacial lake deposits and delta deposits, topped by glacial outwash and finally post-glacial alluvium (Fig. 5) (Williams et al., 1998).

The entire glaciated area was substantially modified by erosion and mass wasting during successive glaciations, and during the periodic intervals of periglacial climate that occurred between glacial advances and retreats. Along the valley sides, and atop the valley bottom deposits lie boulder colluvium and "colluviated till", till that has developed a conspicuous slope-parallel fabric as it slowly wastes downhill (Braun, 2002, 2004).

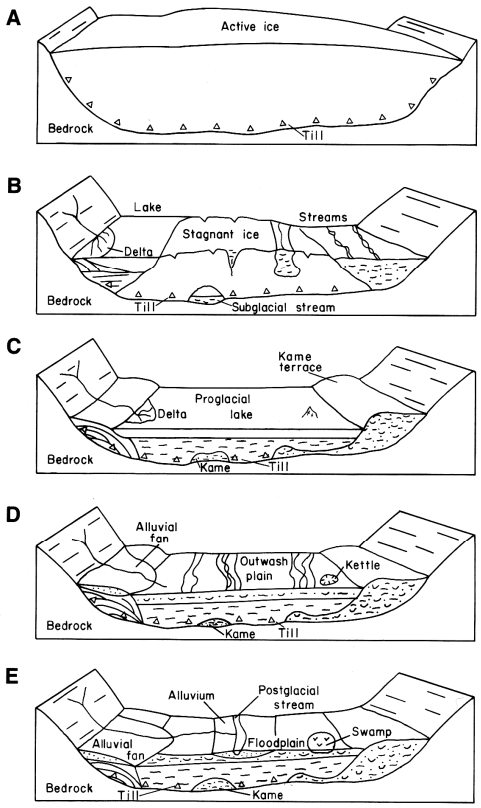


Figure 5. Generalized diagrams of a possible deglaciation sequence, including selected depositional environments and associated deposits. Modified from Reynolds and Brown (1984) in Williams et al. (1998). Note stratified drift deposits below till and glacial lake deposits.

Hydrology

Most of north-central Pennsylvania lies in one of two watersheds: the West Branch Susquehanna, which covers (among others) Lycoming and Tioga counties and parts of adjacent Bradford and Potter counties, and the Upper Susquehanna River basin, which covers. The two branches join just downstream of Northumberland to form the main trunk of the Susquehanna River (Fig. 6).

In 1970, groundwater comprised about half the water used in the West Branch Susquehanna River basin, and about one half of the groundwater (one fourth of the total water use) was used for either public or residential water supply (Taylor et al, 1983). In the Upper Susquehanna River basin, far less groundwater is used for pubic and residential drinking water supplies in terms of percent of total population, however approximately 40% of the population outside the Scranton-Wilkes Barre area relies on groundwater (Taylor, 1984).

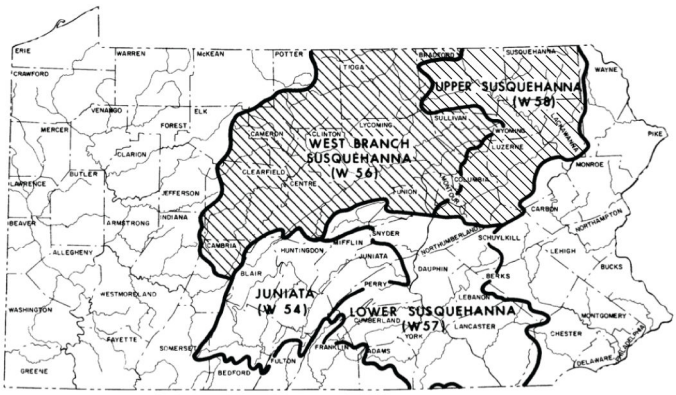


Figure 6. Surface water drainage basins, North-Central PA. Modified from Taylor, et al. (1983) and Taylor (1984).

Aquifers in north-central Pennsylvania consist of bedrock (nearly flat-lying on the plateau and folded to high angles in the Valley and Ridge) that is sufficiently fractured and/or karstic, in carbonate aquifers, to transmit water. In the glaciated region, stratified drift aquifers in stream valleys are the most important water sources (Lohman, 1939; Williams et al., 1998). In the unglaciated region, alluvial aquifers are the largest producers of groundwater (Lohman, 1939). In many larger stream valleys, there are two aquifers, a surficial unconfined aquifer consisting of outwash and/or post-glacial alluvium and a lower ice-contact stratified drift aquifer, separated from the unconfined aquifer by a confining layer of basal till and/or glacio-lacustrine deposits (Fig. 5). Groundwater generally flows from topographic highs to stream discharges in topographic lows. Lloyd and Carswell (1981) estimated that the lowest fresh water circulation in the greater Williamsport region would be no deeper than 350 feet below valley bottoms.

Impacts to surface and groundwater

Surface water withdrawal impacts

Gas shales require a tremendous amount of water for each fracturing stage. The average Marcellus well uses 5.5 million gallons of frack fluids (which is 99.5% water) (Chesapeake Energy, 2010). Although regulations require that the Susquehanna River Basin Consortium (SRBC) issue permits for withdrawals of **any** surface or groundwater for drilling purposes, this requirement is sometimes ignored. In 2008, PADEP had to order Range Resources and Chief Oil to cease their unpermitted water withdrawals from streams in Cogan House and Mifflin Townships in Lycoming County (CM on Fig. 4) (PADEP 2008), because of dangerously low stream levels. Neither company had a permit for their surface water withdrawals. The group American Rivers recently placed the Susquehanna River in the number one spot on its endangered rivers list for 2011, due to the potential impacts to surface water, and suggested that the SRBC deny all permits for withdrawals for drilling purposes (American Rivers, 2011). The SRBC, on the other hand, is confident that with careful planning, gas industry withdrawals can be accommodated (SRBC, 2011)

Large-volume water withdrawals are of concern not only for surface water, but also for groundwater. Although groundwater supply regionally far exceeds demand, Taylor et al. (1983) note that ". . . individual sites that remove large amounts of water may cause severe localized depressions in the water table." Their primary concern was with industrial users and quarries pumping more than 100,000 gallons per day. Although water withdrawals for hydrofracturing would be only a one-time use, considering that each Marcellus well requires 5.5 million gallons of water and that 555 wells have been permitted in the first four months of 2011 in Bradford, Tioga, Potter, Clinton, and Lycoming counties alone (PADEP, 2011) water levels in wells in the vicinity of potential groundwater sources of frackwater should be carefully monitored while drilling activity is high. Drilling activity is expected to taper off over a period of years.

Surface water contamination impacts

Surface water bodies in the Marcellus drilling area have, both on purpose and by accident, become *de facto* drilling waste disposal sites. According to PADEP data, in the last 6 months of 2010, 16 publically owned treatment works (POTWs) accepted 42 million gallons of flowback and production water. PADEP analyses of flowback water from selected Marcellus wells reveal elevated concentrations of some organics such as benzene and glycols, inorganics such as barium, calcium, chloride, iron, lithium, manganese, sodium, and strontium, as well as radionuclides such as radium, lead, and in some cases uranium. Levels of some constituents may be extremely high in some Marcellus wells (Ba >7,000 mg/L, Sr > 4,000 mg/L, Fe > 190 mg/L). This is a concern because disposal methods such as treating flowback water at POTWs, is not completely effective at removing these solutes. Often the brine and radionuclide components of the flowback water cannot be properly treated or sufficiently diluted by the POTW to reach the even the higher effluent concentrations allowed by their grandfathered National Pollutant Discharge Elimination System (NPDES) permits. Frackwater contamination from POTW discharges into the Monongahela was discovered when U.S. Steel noted corroded pipes and equipment (Sapien, 2009). In April 2011, after high levels of total dissolved solids and radioactivity in frackwater were widely publicized, DEP asked gas drillers to voluntarily cease disposing flowback water at POTWs (Ferris, 2011). New, more stringent effluent standards will likely be put in place for the formerly grandfathered POTWs.

Discharges to surface water may not be treated at all. Unintended causes of surface (and potentially groundwater) contamination include blowouts. One blowout occurred in Clearfield County in the East-Central portion of the Pennsylvania, approximately 100 miles southeast of Mansfield. Drillers lost control of a well on the property of the Punxatawney Hunt Club after hydraulic fracturing was completed, and at least 35,000 gallons of flowback water were blown into the air around the well site (Barnes, 2010; USEPA, 2011). More recently, a blowout near a tributary to Towanda Creek in Bradford County (T on Fig. 4) resulted in a declaration of the Maryland Attorney General's intent to file suit against Chesapeake for violations of the Resource Conservation and Recovery Act and the Clean Water Act. In the two days it took to control the leak, thousands of gallons of flowback water flowed across local farmland, into a small creek, to Towanda Creek, and ended up in the Susquehanna. Chesapeake was asked to suspend all Pennsylvania drilling activities pending clarification of the problem at the Towanda area well,

Contaminants also frequently make it to the ground surface via smaller spills, leaking frackwater pits, fires, and tanker accidents. Any substance on the ground surface could be transported through the vadose zone to groundwater. The productive valley bottom aquifers are recharged via precipitation on the valley floor, recharge from overland flow near the valley walls, and sometimes by infiltration from tributary streams (Williams et al., 1998). Harrison (1983) models pathways by which surface contamination could be transported through similar bedrock and glacial aquifers in Northwest Pennsylvania. Figure 7 shows pathways for surface contaminants to reach wells in the valley bottom aquifer system, although, as yet, this has not been clearly documented. Also important, given the lack of oversight of water well installations in Pennsylvania, is the possibility that an improperly constructed well could act as a conduit for contaminants.

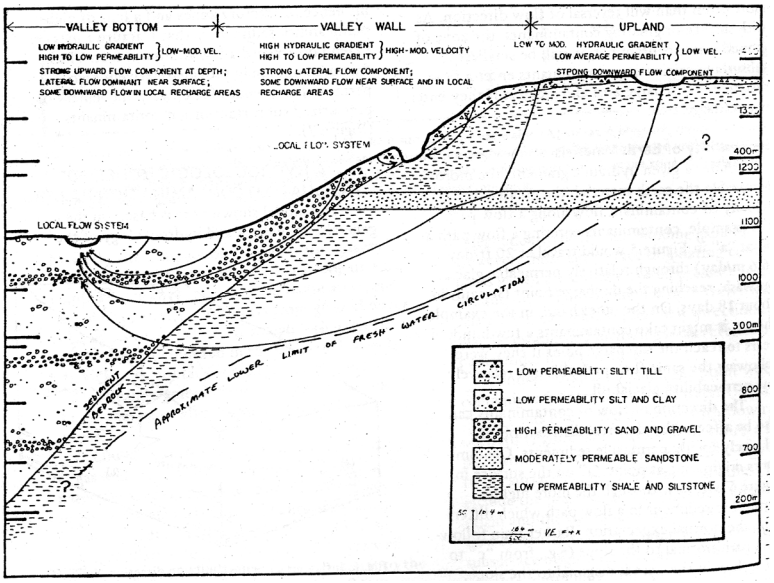


Figure 7. Model of major factors affecting contamination hazard of gas wells in various hydrogeologic zones on the Glaciated plateau. Harrison identified Valley Wall locations as most hazardous due to rapid lateral flow with only low to moderate dilution due to a lower-volume local flow system. From Harrison (1983).

Fracwater/flowback water migration through improperly cased well or interconnected joints

A concern commonly expressed by the public, upon hearing about the large volumes of water and high pressures used for hydrofracturing shale units, is that frackwater or flowback water will migrate upward through fractures into potable aquifers. The oil and gas industry (ex: Halliburton, 2011), and others such as University of Pittsburgh environmental engineering professor Radisav Vidic (Junkins, 2010) suggest that there is little danger to groundwater from hydraulic fracturing, pointing out that hydrofracturing has been used since the late 1940s with no proven groundwater contamination. Few mention that the hydraulic fracturing process used in horizontal gas shale wells is substantially different from the more frequently used vertical well fracks in the amount of water used (and thus the total volume of potentially harmful additives injected into the subsurface).

Engelder (2011) suggested that migration of frack or formation water to useable aquifers would be unlikely due to the large vertical distance (on average 7000 ft) between the Marcellus and potable aquifers and the scarcity of natural interconnected fractures in Carboniferous clastics above the Marcellus. Gas companies have little incentive to overengineer the fracking process and extend the fractures past the unit of interest, as some of the natural gas they are trying to obtain would be lost. A comparison of the vertical extent of hydraulically-induced fractures for wells in the Marcellus and the depth of the deepest well in the county in which each well is drilled indicates at least 3000 feet of vertical separation exists between the top of induced fractures and the bottom of deep water wells (Fisher, 2010). In addition, the generally low velocity of groundwater makes it unlikely that water moving upward via through-going fractures from the Marcellus to higher regional aquifers and regional discharge zones would be detected in residential drinking water very quickly. Drilling locations close to the Marcellus outcrop limit, such as those in southern and eastern Lycoming County are shallower, and closer to aquifers, so contamination and migration may be more of a concern.

Gas migration via the well bore

Although most chemical constituents of frack and flowback water have not been found in well water at concentrations that would make them unambiguously attributable to gas drilling, methane contamination is very common. The gas, however, may be unrelated to drilling. Lohman (1939) notes that the “Chemung” (Lockhaven (?)) formation in North-central Pennsylvania commonly has bubbles of natural gas in it. The town of Tioga Junction, just north of Mansfield, (TJ on Fig. 4) along the Tioga River valley had issues with explosive levels of methane in wells in 2001, long before the shale gas boom began. A study by Breen et al. (2005), suggests that the gas is a mixture of deep

Oriskany Sandstone gas as well as gas) from a gas storage facility just west of town, perhaps percolating up through enhanced fracture system near the hinge of the Sabinsville Anticline.

On the other hand, in many cases drilling does appear to increase methane contamination. For example, in Bainbridge, Ohio, a house exploded because a gas well was not properly cemented prior to fracking, and gas was able to migrate into the drinking water aquifer. Gas wells are supposed to be cased and cemented to a depth below the lowest useable aquifer so gas cannot migrate up the well to aquifers, but this is not always done correctly. PADEP issued 10 violations for improper well casing construction in the last 2.5 years, and just (May 17, 2011) fined Chesapeake more than $1 million for contaminating Bradford County water supplies due to improper well casing. In that case, shallow, non-Marcellus gas was able to migrate into families' wells. In other areas, especially in Western Pennsylvania, water and recent gas wells may be properly constructed, but gas may still migrate up (or contaminants could migrate down) through the thousands of old, improperly abandoned oil and gas wells.

Dimock, Pennsylvania (D on Fig. 4) is another location where improper gas well construction has had an impact on residents. PA DEP stopped Cabot Oil and Gas Corp from drilling near Dimock, citing flaws in for than half of Cabot wells in that area. Improper casing of is allowing gas to percolate through connected pores in the cement, dissolve in people's drinking water and accumulate in their homes (Legere, 2011). Dimock lies in the Glaciated Low Plateau of Susquehanna County, an area that is predominantly covered by Wisconsinan Till (Braun, 2006). Cabot was supposed to have fixed the problem, and paid a $4.6 million dollar fine, but gas is still leaking, so the moratorium is holding

What is the outlook?

Gas shale drilling is here to stay, along with the potentially unpleasant environmental results. Fortunately, no cases of water impact from hydraulic fracturing fluid or flowback water have yet been detected. Much of the impact is coming from surface activities/spills, rather than percolating up from great depth. In fact, to this point, most environmental impacts from drilling in the Marcellus appear to have been avoidable, if individuals and businesses complied with applicable regulations, professional best practices, common sense, and ethical standards. Perhaps they will do so from this point on, and save citizens of the commonwealth a lot of headaches (both literal and figurative), and preserve the good health of another valuable natural resource, clean water.

REFERENCES:

American Rivers, 2011, "Natural gas development putting clean water at risk for million of people", http://www.americanrivers.org/our-work/protecting-rivers/endangered-rivers/endangered-susquehanna.html, May 17, 2011.

Barnes, T., July 14, 2010, Pittsburgh Post-Gazette, "2 drillers fined for Pennsylvania gas well blowout" http://www.post-gazette.com/pg/10195/1072546-454.stm#ixzz1M3vLEopk Accessed May 11, 2011.

Braun, D. D., 1994, Late Wisconsinan to pre-Illinoian (G?) glacial events in eastern Pennsylvania: *in* Braun, D. D., ed. Late Wisconsinan to pre-Illinoian (G?) glacial events in eastern Pennsylvania: Guidebook, 57th Annual Field Conference of Friends of the Pleistocene (Northeastern Section), U.S. Geological Survey Open-File-Report 94-434, p. 1-21,

Braun, D. D., 2002, Quaternary History of the Tunkhannock – Great Bend Region: *in* Inners, J.D., and Fleeger, G.M., eds., Bedrock, surficial, economic and engineering geology of the "Endless Mountains" region, Susquehanna and Wyoming Counties: Guidebook, 67th Annual Field Conference of Pennsylvania Geologists, Tunkhannock, PA.

Braun, D. D., 2006, Surficial geology of the Springville 7.5-minute quadrangle, Susquehanna and Wyoming Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFSM 06–17.0, 13 p., Portable Document Format (PDF).

Braun, D. D., 2006, Surficial geology of the Susquehanna 7.5-minute quadrangle, Susquehanna County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFSM 06–04.0, 18 p., Portable Document Format (PDF) file

Braun, D. D., 2010, Surficial geology of the Blossburg 7.5-minute quadrangle, Tioga County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report OFSM 10–08.0, 25 p., Portable Document Format (PDF) file.

Breen, K. J., Revesz, K., Baldassare, F. J., and McCauley, 2005, Natural Gases in Ground Water near Tioga

Junction, Tioga County, North-Central Pennsylvania-Occurrence and Use of Isotopes to Determine Origins, 2005: U.S. Geological Survey Scientific Investigations Report 2007-5085, 65 p. Portable Document Format (PDF) file.

Chesapeake Energy, Marcellus Water Use Fact Sheet, July 2010, http://www.chk.com/Media/MarcellusMediaKits/Marcellus_Water_Use_Fact_Sheet.pdf, May 5, 2011.

Crowl, G. H. and Sevon, W. D., 1999, Chapter 15 – Quaternary: in The Geology of Pennsylvania, Schultz, C. H. (ed). Geological Survey of Pennsylvania and Pittsburgh Geological Society: Harrisburg; p. 224-231.

Engelder, T., 2011, The distribution of fractures above black shales in the Appalachian basin, Geological Society of America Abstracts with Programs, v. 43, p. 141.

Clarke, C., "Families Blast Gas Exploration", Williamsport Sun-Gazette, April 23, 2011, http://www.sungazette.com/page/content.detail/id/563113/Families-blast-gas-exploration.html, May 12, 2011.

de Witt, W., Jr., 1993. Principal oil and gas plays in the Appalachian basin (Province 131): U.S. Geological Survey Bulletin 1839-I, p. I1-I37.

deWitt, W., Jr., Roen, J. B., and Wallace, L. G., 1993, Stratigraphy of Devonian black shales and associated rocks in the Appalachian basin, in Roen, J.B., and Kepferle, R.C. eds., Petroleum geology of the Devonian and Mississippian black shale of eastern North America: U.S. Geological Survey Bulletin 1909B, p. B1-B57.

Faill, R.T., 1985, The Acadian orogeny and the Catskill delta, in Woodrow, D.L., and Sevon, W.D., eds., The Catskill delta: Geological Society of America Special Paper 201, p. 15-37.

Faill, R.T., 1997, A geologic history of the north-central Appalachians, part 2: The Appalachian Basin from the Silurian through the Carboniferous: American Journal of Science, v. 297, p. 729-761.

Ferris, S., "DEP asks Marcellus drillers to stop taking waste water to treatment plants", Greens County Messenger, 4/21/2011, http://www.heraldstandard.com/gcm/news/local_news/dep-asks-marcellus-drillers-to-stop-taking-waste-water-to/article_d810dd58-fc71-5464-9483-2f3777d8af22.html, May 17, 2011.

Fisher, K., 2010, Data Confirm Safety of Well Fracturing: American Oil and Gas Reporter, online <http://www.aogr.com/index.php/magazine/frac_facts/, May 12, 2011.

Halliburton, "Hydraulic Fracturing 101", http://www.halliburton.com/public/projects/pubsdata/hydraulic_fracturing/fracturing_101.html, May 5, 2011.

Harper, J. A., 2008, The Marcellus Shale—An Old "New" Gas Reservoir in Pennsylvania: Pennsylvanian Geology, v. 38, p. 2-13.

Harrison, S. S., 1983, Evaluating system for ground-water contamination hazards due to gas-well drilling on the glaciated Appalachian Plateau: Ground Water, v. 21, p. 689-700.

Johnson, N., 2010, Pennsylvania Energy Impacts Assessment Report 1: Marcellus Shale Natural Gas and Wind, The Nature Conservancy, 47 pp.

Junkins, C., "Water Gets Cloudier: Pitt Professor Says Fracking Safe, Decades-Old Process", The Intelligencer/Wheeling News-Register, Oct. 3, 2010, http://www.news-register.net/page/content.detail/id/542386.html, May 7, 2011.

Legere, L., "DEP cites persistent flaws in Cabot wells and keeps drilling on hold in parts of Dimock", The Times-Tribune, May 17, 2011, http://thetimes-tribune.com/news/dep-cites-persistent-flaws-in-cabot-wells-and-keeps-drilling-on-hold-in-parts-of-dimock-1.1147813#axzz1MxdUzSwi, May 20, 2011.

Lloyd, O. B. and Carswell, L. D., 1981, Groundwater Resources of the Williamsport Region, Lycoming County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 51, 69 p.

Lohman , S. W.. 1939, Ground water in north-central Pennsylvania; Department of Environmental Resources, Bureau of Topographic and Geologic Survey, 3rd printing, 1973. 219 p., 11 pls., geol. map, scale 1:380,160.

Kucuk, F., and W. K. Sawyer, 1980, Transient flow in naturally fractured reservoirs and its application to Devonian gas shales: Paper SPE-9397, 55th Annual Fall Technical Conference and Exhibition of the Society of Petroleum Engineers of AIME, Dallas, Texas.

Milici, R. C., and Swezey, C. S., 2006, Assessment of Appalachian Basin Oil and Gas Resources: Devonian Shale-Middle and Upper Paleozoic Total Petroleum System. Open-File Report 2006-1237. United States Geological Survey. Accessed accesses May 11, 2011.

ODNR Division of Mineral Resources Management , Oil and Gas, expert Panel Report: Bainbridge Township Subsurface Gas Invasion, Sept. 21, 2008, http://www.dnr.state.oh.us/portals/11/bainbridge/odnr_summary_expert_report.pdf, May 15, 2011.

Pennsylvania Department of Conservation and Natural Resources, Bureau of Topographic and Geologic Survey, 2010, Fact Sheet: Recommendations for Construction of Private Water Wells In Bedrock, www.dcnr.state.pa.us/topogeo/groundwater/water_wells_fs_2010.pdf, May 16, 2011.

Pennsylvania Department of Environmental Protection, July 2010, Punxsutawney Hunting Club 36H Well Report

Pennsylvania Department of Environmental Protection, Bureau of Oil and Gas Management, 2008 Total Wells Drilled, 2009, http://www.dep.state.pa.us/dep/deputate/minres/oilgas/BOGM%20Website%20Pictures/2008/2008%20Wells%20Drilled.jpg, May 11, 2011

Pennsylvania Department of Environmental Protection, Bureau of Oil and Gas Management, 2009 Total Wells Drilled, Jan. 25, 2010, http://www.dep.state.pa.us/dep/deputate/minres/oilgas/BOGM Website Pictures/2009/2009_Wells Drilled.jpg, May 11, 2011

Pennsylvania Department of Environmental Protection, Bureau of Oil and Gas Management, 2010 January-December Wells Drilled, Jan. 5, 2011, http://www.dep.state.pa.us/dep/deputate/minres/oilgas/photogallery/photo13295/2010%20%20Wells%20Drilled.gif, May 11, 2011

Pennsylvania Department of Environmental Protection, Bureau of Oil and Gas Management, Active Marcellus and non-Marcellus Shale wells, 2010 http://www.dep.state.pa.us/dep/deputate/minres/oilgas/BOGM Website Pictures/2010/Operator Active Wells.jpg, May 11, 2011.

Pennsylvania Department of Environmental Protection, Bureau of Oil and Gas Management, Marcellus Permits Issued and Wells Drilled January-April 2011, updated May 3, 2011, http://www.dep.state.pa.us/dep/deputate/minres/oilgas/photogallery/photo13295/Marcellus Wells permitted-drilled April 2011.gif, May 17, 2011.

Pennsylvania Department of Environmental Protection, DEP Detects Elevated Levels of Total Dissolved Solids in Monongahela River, , Aug. 7, 2009, http://www.prnewswire.com/news-releases/pennsylvania-dep-detects-elevated-levels-of-total-dissolved-solids-in-monongahela-river-62184012.html, May 17, 2011.

Pennsylvania Department of Environmental Protection, DEP Orders Partial Shutdown of Two Natural Gas Drilling Operations in Lycoming County, PA DEP Daily News Releases, May 30, 2008, http://www.wvsoro.org/news/2008/05_30.html, May 16, 2011.

Repetski, J.E., Ryder, R.T., Weary, D.J., Harris, A.G., and Trippi, M.H., 2008, Thermal Maturity Patterns (CAI and %Ro) in Upper Ordovician and Devonian Rocks of the Appalachian Basin: A Major Revision of USGS Map I-917-E Using New Subsurface Collections: U.S. Geological Survey Scientific Investigations Map 3006. http://pubs.er.usgs.gov/usgspubs/sim/sim3006, April 20, 2011.

Reynolds, R. J., and Brown, G. A., 1984, Hydrogeologic appraisal of a stratified-drift aquifer near Smyrna, Chenango County, NY: U.S. Geological Survey, Water-Resources Investigations Report 84-4029, 53p.

Robinson, B. G., "Shell Appalachia releases statement on methane in water well in Tioga County", Williamsport Sun-Gazette, April 14, 2011, http://www.sungazette.com/page/content.detail/id/562767/Shell-Appalachia-releases-statement-on-methane-in-water-well-in-Tioga-County.html?nav=5011, May 4, 2011.

Sapien, Joaquin, 2009, "With Natural Gas Drilling Boom, Pennsylvania Faces an Onslaught of Wastewater" ProPublica, http://www.propublica.org/article/wastewater-from-gas-drilling-boom-may-threaten-monongahela-river, accesses May 11, 2011.

Sevon, W. D., and Braun, D. D., 1997a, Surficial geology of the Towanda 30 by 60-minute quadrangle, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Open-File Report 97–03, 25 p. [web release].

Sevon, W. D., and Braun, D. D., 1997b, Glacial Deposits of Pennsylvania: Pennsylvania Geological Survey, Map 59.

Soeder, D.J., and Kappel, W.M., 2009, Water resources and natural gas production from the Marcellus Shale: U.S. Geological Survey Fact Sheet 2009–3032, 6 p.

Susquehanna River Basin Commission, "SRBC comments on Susquehanna being named nation's most endangered River In 2011: Call For SRBC to Cease Water Quantity Approvals for Natural Gas Drilling Is Misguided", SRBC Press Release, May 17, 2011 http://www.srbc.net/pubinfo/press/docs/SRBC_Commentary_American_Rivers_5-17-11.pdf, May 19, 2011.

Swistock, B., 2007, A Quick Guide to Groundwater in Pennsylvania, Pennsylvania State College of Agricultural Science Publication No. UH813, 12 pp.

Taylor, L. E., 1984, Groundwater resources of the Upper Susquehanna River basin, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 58, 136 p.

Taylor, L. E., Werkheiser, W. H., and Kriz, M. L., 1983, Groundwater resources of the West Branch Susquehanna River basin, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 56, 143 p.

Williams, J. H., Taylor, L. E., and Low, D. J., 1998, Hydrogeology and groundwater quality of the glaciated valleys of Bradford, Tioga, and Potter Counties, Pennsylvania, Pennsylvania Geological Survey Water Resources Report 68, 89 p.

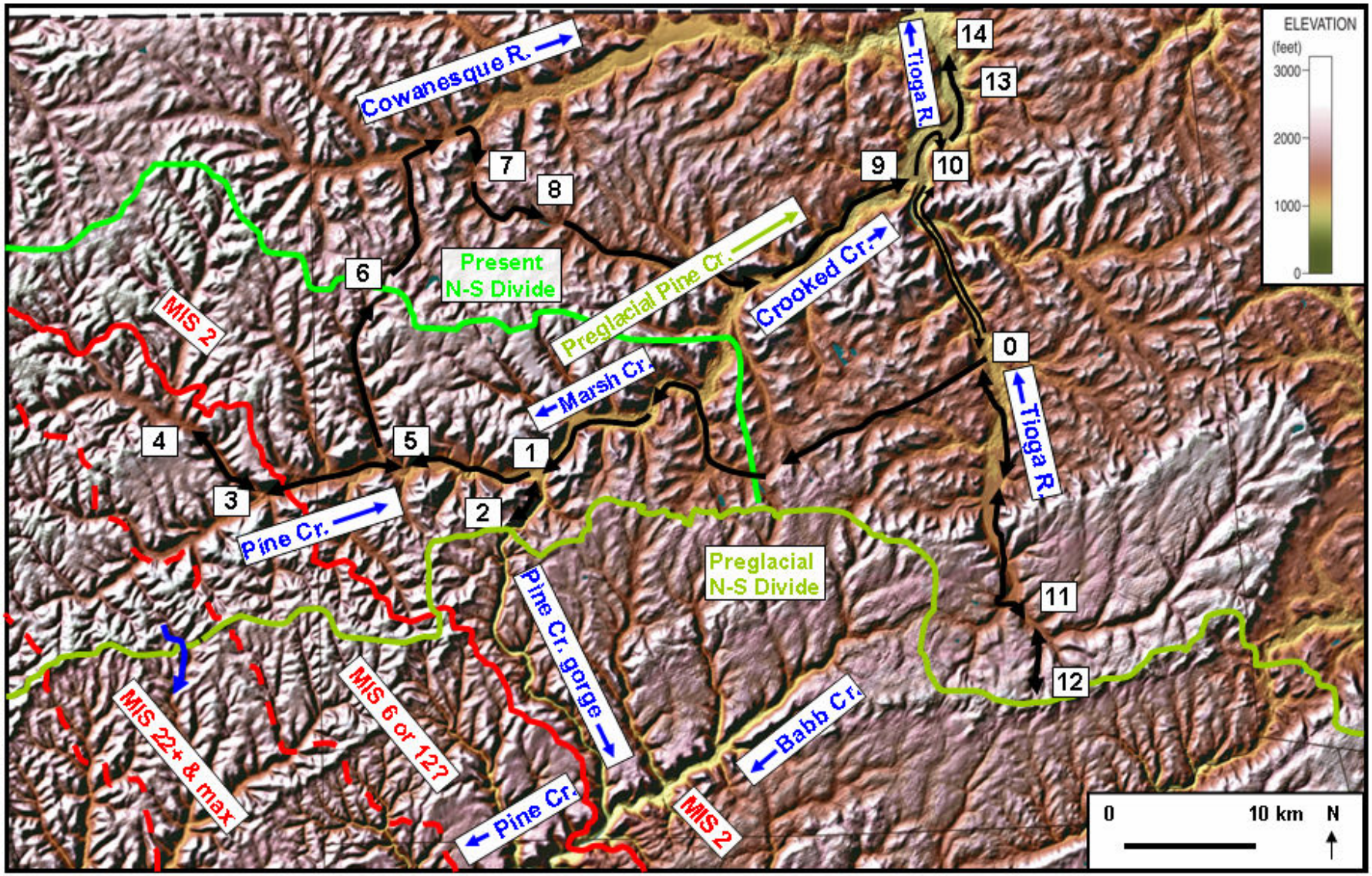
Wood, J.R., Allan, J.R., Huntoon, J.E., Pennington, W.D., Harrison, W.B., Taylor, E., and Tester, C.J., 1996, Horizontal well taps bypassed Dundee oil in Crystal field, Michigan: Oil & Gas Journal, October, p. 60-63.

United States Department of Energy, Strategic Center for Natural Gas and Oil National Energy Technology Laboratory, 2007, DOE's Unconventional Gas Research Programs 1976-1995: An Archive of Important Results, www.netl.doe.gov/kmd/cds/disk7/disk2/Final%20Report.pdf, 10 May 2011

United States Energy Information Agency (a), Shale gas plays, lower 48 states, March 10, 2010, http://www.eia.doe.gov/oil_gas/rpd/shale_gas.pdf, May 17, 2011.

United States Energy Information Agency (b), Natural Gas Consumption by End Use, Area: Pennsylvania, Period: Annual. April 28, 2011, http://www.eia.doe.gov/dnav/ng/ng_cons_sum_dcu_SPA_a.htm, May 17, 2011.

United States Environmental Protection Agency, Region III, Letter to Pennsylvania Department of Environmental Protection, 5/7/2011, http://www.dailykos.com/story/2011/03/10/954784/-De-facto-deregulation-of-shale-gas-drilling-in-PA, May 11, 2011.



Digital terrain map of north central PA showing FOP field trip route (black arrows)and stops (1- 10 Saturday & 11-14 Sunday), glacial limits as red lines (age in MIS #), 1950 ft. sluiceway for Glacial Lake Galeton (blue arrow), and other drainage features.

Saturday field trip by bus starting from the Comfort Inn (41° 48' 28.28", 77° 05' 19.81")

Leave Comfort Inn and descend hill. Turn right and then right again onto US 6 west bound and immediately pass under US 15. Follow US 6 to and through Wellsboro.

At the Comfort Inn, one is 300 ft. below the surface of Glacial Lake Mansfield (Fig. P4). After about 5 miles one crosses a shallow valley on the right that is the entrance to the 1560 ft elevation sluiceway for Glacial Lake Mansfield (no road goes through the sluice). Two miles beyond Mansfield, where PA SR 287 joins US 6, you have now entered the broad, swamp floored southwest draining Marsh Creek valley. That valley is the 10 mile long reversed gradient portion of the northeast draining preglacial Pine Creek. The present north-south drainage divide (Marsh Creek-Crooked Creek) lies about 2.5 miles to the north (right) of the US 6 & PA SR 287 intersection (Fig. P4).

About 11 miles west of Wellsboro, cross the bridge over Marsh Creek and immediately pull over onto the right shoulder of the road for Stop 1.

Stop 1. Entrance to the Pine Creek gorge (stay on bus) (41° 44' 53.14", 77° 25' 43.73")

We have just crossed Marsh Creek whose channel here is aimed directly south towards the entrance to the Pine Creek gorge (Fig. T1). On the south side of US 6 is the entrance to the parking lot for the "put-in" for raft trips down the Pine Creek gorge. Ahead and to the left is the northeast trending (directly at us) preglacial to present valley of Pine Creek. The Pine Creek channel turns a right angle to enter the gorge (actually a T intersection with Marsh Creek).

While Pine Creek now drains south to the West Branch Susquehanna River, in preglacial times it drained northeast along what is now Marsh Creek and Crooked Creek (Fig. P4) to join the Tioga River at Stop 9 (Fig. T8). Ice blockage of the downstream portion of Pine Creek in each glacial epoch permitted progressive incision of the col that is now Pine Creek gorge until the Pine Creek headwaters were permanently diverted to the West Branch Susquehanna River. Northeast of the head of Pine Creek Gorge is the broad, gentle gradient, southwest draining Marsh Creek valley, a reversed underfit stream in a 10 miles (16 km) segment of the northeast draining preglacial Pine Creek (Fig. P3, Fig. P4 line of arrows pointing northeast). Bedrock at the divide at the head of Marsh Creek (Fig 4, line of circles) is at about 1150 feet, the bedrock at the mouth of the gorge is at 1120 feet, so there has been at least 30 feet of erosion of the floor of the preglacial Pine Creek valley to form the reversed course Marsh Creek valley. That erosion is probably more like 140 feet, assuming a reasonable northeast gradient of preglacial Pine Creek. The bedrock floor of Marsh Creek valley has a broad U-shaped cross-section that suggests glacial erosion by Wisconsinan and possibly older glacial advances. That the diversion of Pine Creek is of pre-Wisconsinan age is indicated also by the incision of Pine Creek into the broad floor of the Pine Creek valley upstream of the gorge (Stop 5), and the presence of yellow-red oxidized pre-Wisconsinan glacial gravel deposits near the floor of the Pine Creek valley south of late Wisconsinan terminus in the gorge (Denny, 1956).

Continue on US 6 and immediately turn left onto Colton Road at the sign for Colton Point State Park. Cross Pine Creek and ascend the mountain side. In about miles 3.5 miles park on right at the Colton Point overlook.

Stop 2. The Colton Point overlook into the Pine Creek gorge (41° 42' 31.81", 77° 27' 52.83")

At the Colton Point overlook the view ahead and to the left is towards the deepest part of the glacially diverted course of Pine Creek at the preglacial drainage divide between originally northeast draining Pine Creek and a south draining tributary of the West Branch Susquehanna River. It has been argued that the gorge is a Wisconsinan age diversion due to the relative narrow, deep "youthful" shape of the gorge (Coates and Kirkland, 1974). Most other workers starting with Fuller

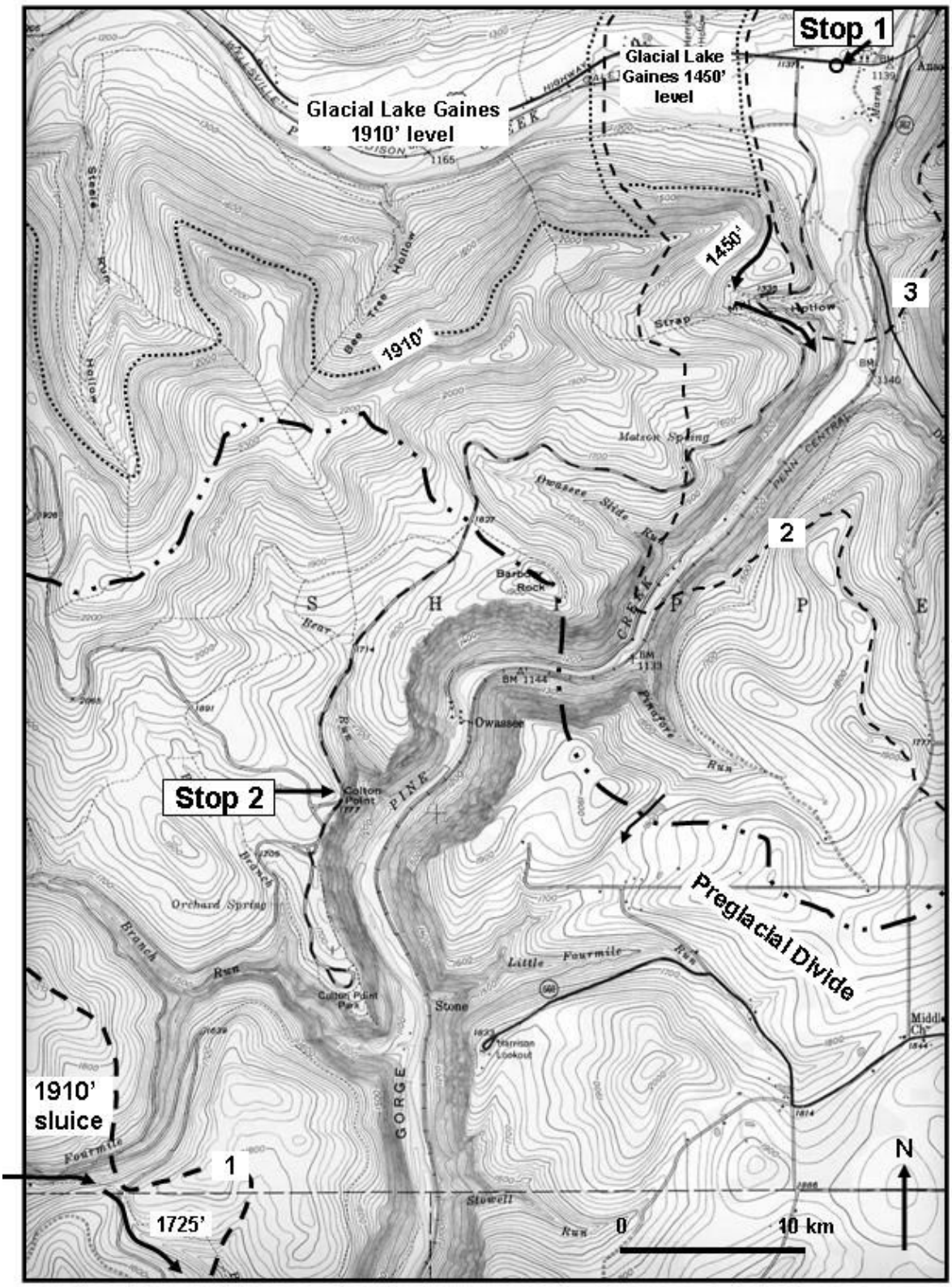


Figure T1. Stop 1. View south to entrance to Pine Creek gorge where Pine Creek turns a right angle to enter the north end of the gorge. **Stop 2.** Colton Point overlook is near the deepest part of the Pine Creek gorge where it cuts the preglacial divide between the original northeast trending Pine Creek and a south trending tributary to the West Branch Susquehanna River. Dashed line ice margin 1 is where ice must be to open the 1910' sluice. Dashed line ice margin 2 is the last ice front position to hold in 1910' level of Glacial Lake Gaines. Dashed line ice margin 3 is the short-lived ice front position that held in the 1450' level of Glacial Lake Gaines.

and Alden (1903) have been convinced that the gorge was initiated in the early Pleistocene and eroded to near the present depth by the mid-Pleistocene (Denny, 1956, Muller, 1957, Crowl, 1981).

The 25 mile (40 km) long Glacial Lake Gaines occupied the headwaters of Pine Creek, a region of uniformly high interfluves (Braun, 1989a). Once each glacial advance traveled southwest of the present gorge, Glacial Lake Gaines would drain to the southeast through a sluiceway 6 miles southwest of Galeton (Stop 4) at an elevation of 1960 feet (598 m). Once each glacier retreated to Gaines (Stop 6), a lower lake outlet opened at 1910 feet (582 m) just west of the Pine Creek gorge (Stop 3), and that started drainage along the southwest flank of the gorge (Fig. T1). The gorge itself was transverse to ice flow and south of the preglacial divide in the terminal zone of the late Wisconsinan glacier. So the gorge was full of ice and glacial sediment when the Glacial Lake Gaines water started flowing along the southwest flank of the gorge. The 1910 feet sluiceway was still as much as 750 feet (230 m) above the bedrock floor of the present Pine Creek gorge. The 1910 feet level would be maintained as the glacier receded northeasterly from the 1910 feet sluice to near the entrance to the Pine Creek gorge (from ice margin 1 to ice margin 2, Fig. T1).

As ice receded eastward down the preglacial Pine Creek valley and receded northward within the gorge, the 1910 feet level Glacial Lake Gaines should have started spilling over a ridgeline on west flank of the entrance to the gorge (between ice margin 2 & 1 on Fig. T1). That initial spillage would start removing ice and sediment infill of the gorge. Glacial Lake Gaines would only drop 50 feet with the opening of this relatively narrow sluice and would not have produced “catastrophic” flooding along the south flank of the gorge. The next lower distinct outlet for Glacial Lake Gaines was a channel cut in the west flank of the entrance to Pine Creek gorge at an elevation of 1450 feet (442 m) (Ice margin 3, Fig. T1). The 460 feet (140 m) drop in lake level occurred over an unknown length of time but would have been expected to be relatively rapid, even catastrophic. As of yet though, features expectable from catastrophic flooding have not been observed along those streams. It is possible that younger late Wisconsinan outwash from other West Branch Susquehanna River tributaries reworked and buried such evidence, especially in the broad West Branch Susquehanna valley where it enters the Ridge and Valley Physiographic Province.

Turn around at Colton Point overlook, return to US 6, turn left onto US 6 west. Take US 6 to Galeton.). Go partway through Galeton, turn left onto West St., cross bridge and immediately turn left into entrance to the town park that is at the confluence of the two branches of Pine Creek.

Stop 3. Galeton, at the late Wisconsinan terminus and 640 feet below the surface of Glacial Lake Gaines (41° 44’ 02.28”, 77° 38’ 49.02”)

At this site we are just one mile (1.5 km) from the late Wisconsinan terminus and 640 feet (195 m) below the 1960 feet surface elevation of ice dammed Glacial Lake Gaines (dotted line, Fig. T2). We have observed exposures of varves in slumps at several valley floor sites in the area (arcuate lines with arrows pointed in the direction of movement, Fig. T2). Varve deposits are far more extensive and thicker in the northwest trending branch of the Pine Creek drainage that lies to the northwest of the area shown on Fig. T2. The late Wisconsinan terminus lay along the northeast side of that drainage and a number of tributaries brought in sediment from the terminus. Varves have not been observed more than 100 feet above valley floor, probably because the clayey material slid off the steep mountainside as or shortly after it was deposited. We have observed severe deformation in the varve deposits remaining, sometimes with well developed recumbent or even isoclinal folds.

The glacial terminus at this site has been drawn as a straight line across the valley, indicating that the ice front was essentially vertical where it “stood” in the lake. This is in contrast to the low ice profile gradient lobes drawn by Crowl and Sevon (1980) that project along the three valleys that join at Galeton (dashed lines, Fig. T2). But these low gradient lobes would have projected out underneath the lake, a physically impossible situation. To support the lobate ice margin Crowl and Sevon mapped two patches of late Wisconsinan till and one patch of late Wisconsinan ice-contact stratified drift within the lobate area. Braun (unpublished mapping) in the mid 1980’s interpreted the two till patches to be colluvium accumulated at the toe of the steep mountain side. The ice-contact stratified drift patch has a terrace landform with a planar top surface sloping downstream parallel to the present Pine Creek floodplain were interpreted by Braun to be an alluvial terrace. Braun agreed

with Crowl and Sevon that there is till on the slope southeast of Galeton but thought much of it was "flowtill" from the glacial front immediately east of that till patch. There has been little actual exposure of the material to verify its precise genesis. Braun traced the glacial terminus across the mountain tops both north and south of the Pine Creek valley and he ended up at either end of the straight line shown crossing the valley. There may have been some sort of calving bay reentrant in the ice front in the Pine Creek valley given that the water depth was on the order of 600 feet. But no arcuate moraine features have been found that are concave to the east (or west for that matter). Hopefully the new Lidar imagery will show something that will clarify the situation.

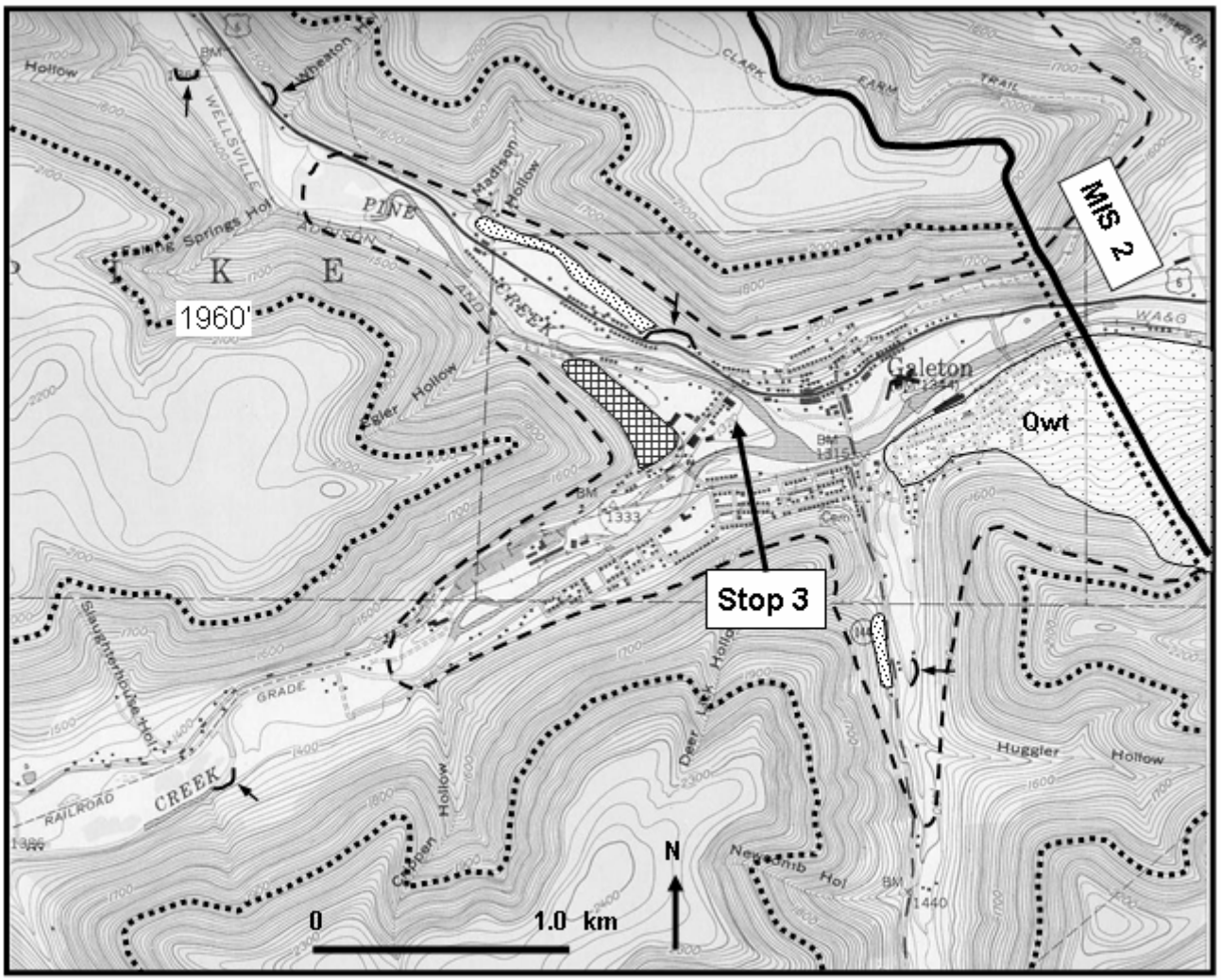


Figure T2. Stop 3 at Galeton, just outside the late Wisconsinan (MIS 2) terminus (heavy line) and 630 ft. under the surface of 1960 ft. surface elevation of Glacial Lake Gaines. Dotted line is the shoreline of the proglacial lake. The lake's outlet is off the map to the southwest. Slumps in varves are denoted by arcuate lines with an arrow pointing in the direction of motion and at the headwall of the slump. Dashed line is Crowl and Sevon's (1980) late Wisconsinan terminus. Grid area is Crowl and Sevon's ice contact stratified drift area here interpreted as an alluvial terrace. Heavy stipple areas are Crowl and Sevon's till areas here interpreted to be colluvium. The lightly stippled till area (Qwt) under the late Wisconsinan terminus is probably "flowtill" in front of the ice front.

Return to US 6 and turn left, continuing west. In about 2.5 miles, turn right into gravel pit beside the road.

Stop 4. Pre-Wisconsinan gravel overlain by late Wisconsinan gravel
(41° 45' 34.09", 77° 41' 03.96")

This site is a gravel pit in a fan that has built out onto the floor of the Pine Creek valley from the mouth of Ansley Hollow, a steep, narrow tributary valley (Fig. T3). Crowl and Sevon (1980) mapped the late Wisconsinan terminus north of the head of Ansley Hollow (dashed line, Fig. T3) and mapped the site as an alluvial fan. Just upstream of the site they drew the terminus as a low gradient sublacustrine ice lobe on the floor of Glacial Lake Gaines (dotted line at lake surface elevation). Sevon and Braun (1997a) traced the late Wisconsinan terminus (solid line, Fig. T3) into the head of Ansley Hollow but retained the alluvial fan designation for the site due to a lack of exposure at that time (the field work was done in the mid-1980's). When the late Wisconsinan

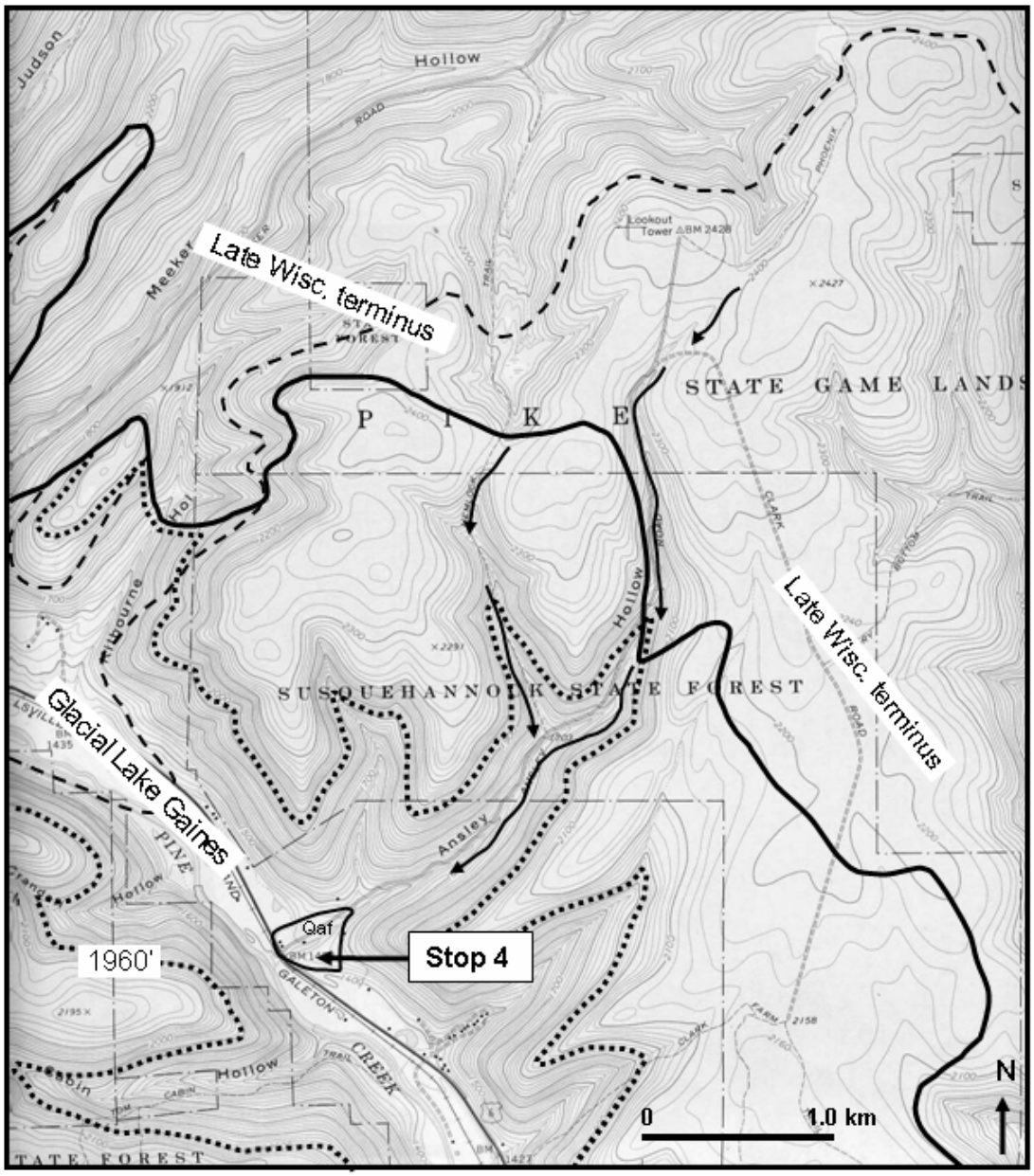


Figure T3. Stop 4. Gravel pit exposing brown Late Wisconsinan gravel overlying yellow-red weathered pre-Wisconsinan gravel in what should have been a sub-lacustrine fan. The dotted line is the shoreline of the 1960 ft. surface elevation of the highest phase of Glacial Lake Gaines. The solid line is the late Wisconsinan (MIS 2) terminus from Sevon and Braun (1997a) and later unpublished work by Braun. The dashed line is the late Wisconsinan terminus from Crowl and Sevon (1980). The arrows are meltwater flow paths to the fan. On the steep, narrow valleys the meltwater probably continued as gravity or debris flows under lake level down to the Pine Creek valley floor.

Figure T4a. Stop 4. Brown Late Wisconsinan gravel over yellow-red oxidized pre-Wisconsinan gravel in what was probably a subaqueous fan. The brown gravel may also partly be a paraglacial deposit from up-valley erosion immediately after ice recession.

FigT4b. Stop 4. Closeup of gravels showing sharpness of contact and clastic imbrication. Closely examine the outcrop and see if there is evidence for the deposit being a series of subaqueous debris flows.

glacier was at the terminus, the fan surface would have been about 500 feet below the surface of Glacial Lake Gaines (dotted line, Fig. T3). So the fan should be a sublacustrine one unless it was built in para-glacial to Holocene times rather than when the glacier was at its terminus at the head of Ansley Hollow.

This is the only site along the glacial terminus zone in north-central Pennsylvania (Potter - Tioga Counties) that I am confident that there are pre-Wisconsinan glacial deposits overlain by late Wisconsinan glacial deposits. The pit was not active and poorly exposed when I was mapping and doing paleomagnetic sampling of the varves in the Pine Creek valley in the mid-1980's. The pit has been only occasionally active in the last couple of decades due to the relatively poor quality of the weathered gravel that forms the most of the deposit. I have only been able to spend a few minutes at the pit preparing for this trip so additional observations from the group would be appreciated.

The brown (7.5 YR 4/2 - 4/4) late Wisconsinan glacial and paraglacial gravel has a sharp contact with underlying more oxidized reddish yellow (7.5 YR 6/6 - 7/6) presumably pre-Wisconsinan gravel (Fig. T4a & T4b). Immediately below the contact is an oxidized bed of silty sand that may represent truncated glacial lake sediments that overlie the older gravels. There does not appear to be a soil profile preserved in the upper part of the reddish yellow gravel. The odd thing about the site is that there are no late Wisconsinan varves either under or over the late Wisconsinan gravel. Both a mile up and down valley of here are thick late Wisconsinan varve deposits. So the lack of varves here is a bit inexplicable.

Turn left onto US 6 east, go back through Galeton and, in a few miles, Gaines. About one-half mile after Gaines turn right into a roadside rest and picnic area.

Stop 5 & Lunch. The inner bedrock gorge of the Pine Creek valley upstream of the gorge (41° 44' 53.46", 77° 32' 46.74")

At this stop one is standing on the top of a 30 feet high bedrock cliff that is being undercut by Pine Creek. This is one side of a 30-50 feet deep, 400-700 feet wide inner gorge cut into the 2000-3000 ft wide overall floor of the Pine Creek valley. The overall floor is covered by 10's to more than a 100 feet thick layer of glacial deposits that Pine Creek has cut through before cutting into the bedrock. This inner valley starts about two miles (3 km) upstream of the entrance to Pine Creek gorge (Stop 1), continues for about seven miles (11 km) upstream while getting gradually shallower, and ends one mile upstream of here at Gaines. Upstream of Gaines the valley floor is entirely mantled by glacial deposits with a broad floodplain like that at Galeton (4.5 miles upstream at Stop 4). This inner gorge or valley is interpreted to be a knickpoint migrating upstream from Pine Creek gorge at Stop 1. The presence of such a knickpoint suggests that the Pine Creek gorge was cut to near its present level in pre-Wisconsinan times, was deepened some in Wisconsinan times, and then a knickpoint migrated upstream from near the gorge entrance in late to post glacial times.

Turn left onto US 6 west and return to Gaines. Turn right onto PA SR 349 and drive up the south trending Long Run valley. In about 6 miles you will be in the deepest part of the 1730 ft. elevation sluiceway for Glacial Lake Cowanesque. Pull over onto the right shoulder of the road.

Stop 6. The Sabinesville Sluiceway, outlet for the two highest phases of Glacial Lake Cowanesque (stay on the bus) (41° 50' 34.46", 77° 33' 03.15")

This 300-400 feet deep sluiceway was cut in bedrock by meltwater flow from the two most headward and highest levels of Glacial Lake Cowanesque (the 1950 and 1730 feet levels) (Fig. T5, H5). During several kilometers of northeasterly ice retreat in the main Cowanesque valley, meltwater from the 1950 ft. level sluice cascaded down into a local proglacial lake in the valley leading to the north side of the 1730 ft. sluice. The floor of 1730 ft sluice was probably somewhat higher than present when the late Wisconsinan glacier first retreated past the site and was eroded down a bit by the first and second phase discharges from the late Wisconsinan Glacial Lake Cowanesque. The floor of the sluice was probably also eroded down a bit by subglacial meltwater flow since the sluice is near parallel to overall ice flow direction and ice surface gradient. The relatively short duration of the glacial lake flows, even with a significant flood during the drop from 1950 feet level to the 1730 feet level, during a single glacial advance and retreat argues against this sluiceway being cut by a single glacial event.

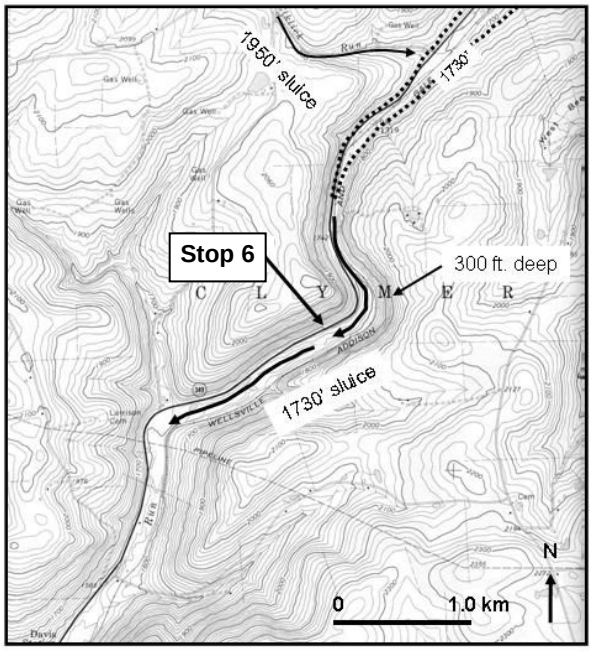


Figure T5 Stop 6. Sabinesville sluiceway, outlet for both the 1950 ft surface elevation first phase of Glacial Lake Cowanesque and the 1730 ft. surface elevation second phase of the lake. The floor of the sluice was probably also eroded down a bit by subglacial meltwater flow since the sluice is near parallel to overall ice flow direction and ice surface gradient.

Continue ahead on PA SR 349, go through Sabinesville and enter Westfield. At the T intersection turn right onto PA SR 49 east. Upon leaving Westfield you will cross the Cownesque River.

You are now heading downstream in the east draining Cowanesque River valley. Glacial Lake Cowanesque's water surface was 400 ft above the present land surface. Thick glaciolacustrine deposits underlie the flanks and floor of the valley. Slump benches on the hillsides are common and particularly prominent north of the village of Cowanesque.

Go through the village of Cowanesque and then turn right onto PA SR 249 and cross the Cowanesque river again.

To the right are high bluffs undercut by the river that expose till and ice-contact stratified drift overlain by varves.

Ascend the Jemison Creek valley and in about 3 miles turn left onto gravel Cooper Rd. Gravel pit visible ahead and to the left. Turn left into the gravel pit.

Stop 7. Local ice-contact delta in the entrance to the 1500 feet sluiceway of Glacial Lake Cowanesque (41° 53' 56.30", 77° 28' 25.60")

This gravel pit is near the entrance to the 1500 feet elevation sluiceway for the 3rd phase of Glacial Lake Cowanesgue (Fig. T6). The gravel deposit has a gently sloping top starting at an elevation of 1600 feet, gently descends to about 1560 feet, and is now incised by two gullies, one on each side of the road. The pit exposes drab-colored pebble to cobble gravels dominated by tabular clasts of the local bedrock. Clast imbrication and stratification indicate flow to the south towards the entrance to the sluiceway. I have tentatively interpreted the deposit to represent a kame or ice-contact delta. If it is a delta, the lake level from here southward to the sluiceway was 100 feet higher than the present bottom of the sluiceway. The trend of the sluiceway is transverse to ice flow, so probably some remnant ice and glacial deposits still partly filled the sluiceway and temporarily ponded water at the 1600 feet level.

I have only spent a few minutes at the pit because I found it when I was running the road log for this trip. The pit was not there in the 1980's when we were working in the area and the soil map (Rayburn and Braker, 1981) indicated a till soil at the site. The soil mapping and the ledge-like landform led us (Sevon and Braun, 1997a) to map it as a bedrock feature. Let's see what additional observations and interpretations the group can come up with as what type of glacial deposit feature this is.

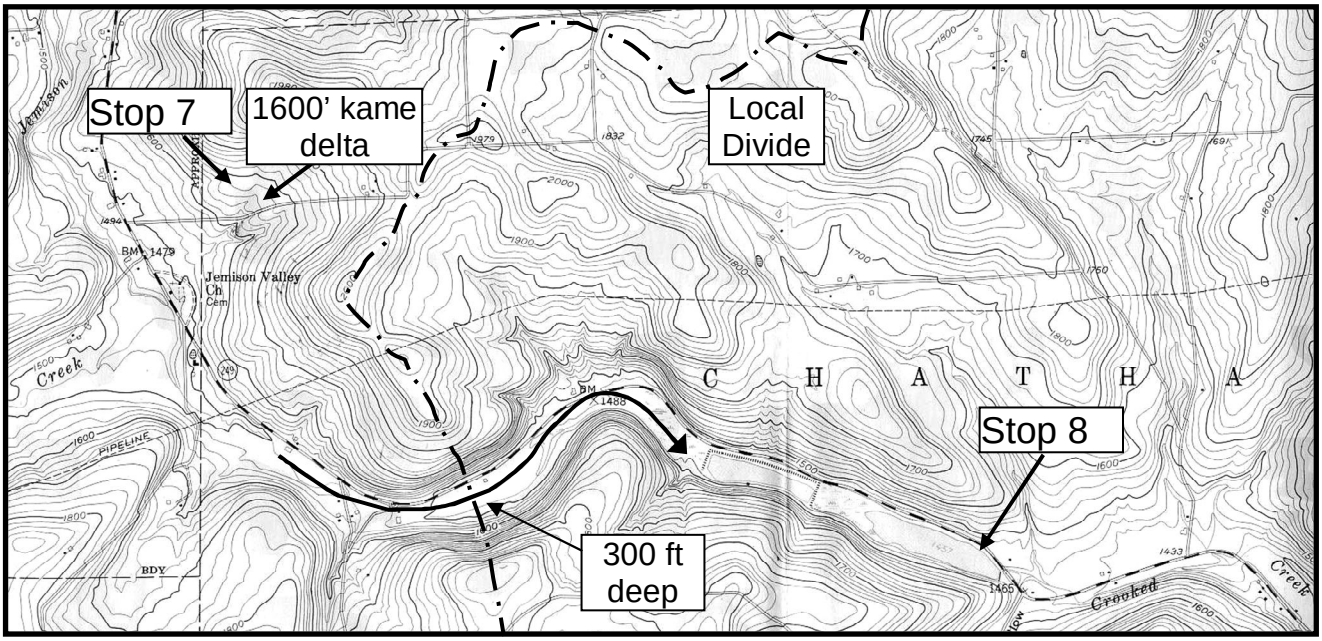


Figure T6. Stop 7. Ice contact delta near entrance to sluiceway. **Stop 8.** Photo stop of the sluiceway for 1500' level third phase of Glacial Lake Cowanesque.

Turn around and return to PA SR 249. Turn left onto PA SR 249 and continue up the valley and into the 300-400 ft. deep cut of the 1500 ft level sluiceway of Glacial Lake Cowanesque. Continue about 2 miles through the deepest part of the sluiceway until you reach a series of ponds on the floor of the sluiceway. Pull over on the right side of the road.

Stop 8. Sluiceway for the 3rd or 1500 feet phase of Glacial Lake Cowanesque

(41° 52' 55.05", 77° 25' 24.90")

This is a stop to take photos of the sluiceway for the 1500' level, third phase of Glacial Lake Cowanesque (Fig. T6 and P5). This sluiceway would have functioned as the outlet for the lake until the glacier retreated 27 km to the northeast to open a 1230 ft. elevation sluiceway near the confluence of the Cowanesque and Tioga valleys (Fig. P4). The sluiceway has been almost entirely cut by lake overflow with possibly some direct glacial drainage as the ice retreated from the valley north of the sluiceway entrance (Stop 8). The sluiceway is transverse to ice flow and ice gradient so subglacial flow did not contribute to its erosion. Again, the cutting of a 300-400 feet deep, 2 mile (3 km) long sluiceway suggests multiple glacial events rather than a single one.

Continue ahead on PA SR 249 and descend the southeast trending Crooked Creek valley. In about 9 miles at a T intersection, turn left onto PA SR 287 and continue down the now much wider and underfit Crooked Creek valley.

One mile to the right (southwest) of the T intersection is the present drainage divide between northeast draining Crooked Creek valley (preglacial Pine Creek course) and present Pine Creek draining south through the Pine Creek gorge. In about 5 miles on the right one will come to the upstream end of the Hammond Dam Reservoir. Across the reservoir are two paraglacial to post glacial fans that have built out onto the broad floor of the Crooked Creek valley from steep, deeply incised tributary streams.

Continue ahead on PA SR 287 past the dam spillway cut into bedrock and into the village of Tioga. At the T intersection turn right onto South Main St. Tioga Dam is straight ahead. Bear right and ascend the access road to the Tioga-Hammond Dam lookout site. Road ends in parking lot.

Stop 9. The Tioga-Hammond Dam's glacial meltwater and human deepened sluiceway
(41° 53' 48.69", 77° 08' 29.87")

The Tioga-Hammond Dams themselves (condensed from Wilshusen and Wilson, 1981)

Construction of the dams took place from 1974 to 1979 and cost $200 million. Subsurface conditions were investigated through 482 diamond drill or soil auger holes plus 53 test pits/trenches. So the subsurface glacial deposits are well described at the site.

Tioga Dam has a 280 mi^2 drainage area and Hammond Dam a 122 mi^2 drainage area. The two dams operate together through a connecting channel as a single flood control project (Fig. T8). Crooked Creek valley, the site of Hammond Dam, is a broad glacial valley with a small, meandering stream. The Tioga River valley is narrow between Mansfield and Tioga and carries a large stream. The Tioga valley is too narrow for economical construction of a large spillway, but it nicely accommodates a large outlet works under the embankment. Conversely, the small Crooked Creek stream channel is not well adapted to a large outlet works, but the broad valley is favorable for a large spillway. The connecting channel allows the dams to share a single spillway and primary outlet works, resulting in a project with each site used to its best advantage. There is a small outlet works through Hammond Dam to maintain minimum flow in Crooked Creek below the dam.

Under normal and low-flow conditions water flowing north in Crooked Creek collects in Hammond Lake with a minimum flow being released through the small outlet works. The remainder flows east into the connecting channel and through controlled gates beneath the channel weir crest and mixes with Tioga River water in Tioga Lake. The latter flows north through the main gates in the Tioga Dam outlet works.

During flood periods, flow conditions are reversed. Water from the Tioga River rapidly fills Tioga Lake. Some water is released through the controlled outlet works, with the excess flowing west over the weir crest in the connecting channel to be stored in the broad Crooked Creek valley. When valley storage is exhausted, the excess from both valleys can flow over the large emergency spillway adjacent to northwest side of Hammond Dam.

The thicknesses and distribution of glacial and fluvial material on bedrock are different at the two dam sites, thus requiring different design criteria. Overburden thickness reaches more than 130 feet at the right abutment of the Tioga Dam and approximately 220 feet at the right abutment of the Hammond Dam. The constructed dams comprise materials appropriate to the conditions encountered.

Slump/earthflow from US 15 onto Tioga dam (revised from Wilshusen and Wilson, 1981)

Across the Tioga valley (east) and just below US 15 road level, one of Pennsylvania's largest historic slope failures occurred in May, 1975 (Fig. T7, stippled area). Movement was continuous over a period of several months and was confined to glacial and colluvial material overlying bedrock. The slide area was approximately 500 feet wide and 900 feet long within which the ground moved by earhflow with rotational slumping near the top. The toe of the earthflow reached the top of the dam embankment then under construction.

The slippage surface was at the top of and within a downhill sloping, 4 to 16 feet thick clayey varve unit. The varves were between two tills or, more probably, between in-situ till and overlying colluvium derived from till. Considerable seepage occurred at the varve zone. The varve unit was at an elevation of 1400 feet and was deposited just in back of the last ice margin to hold in 1560 feet surface elevation Glacial Lake Mansfield (Fig. T7, ice margin 2).

The slope failure area was in a semicircular first order tributary hollow filled with up to 145 feet of unconsolidated glacial and colluvial material. The "cirque like" configuration of the hollow eroded into the bedrock concentrated groundwater flow toward the center of the hollow and the center of the slump/earthflow. The slope failure in this innately unstable location was triggered by a combination of factors: construction activities that involved excavation at the toe, drainage changes and

placement of fill materials at the head, and increased precipitation in the spring of the year.

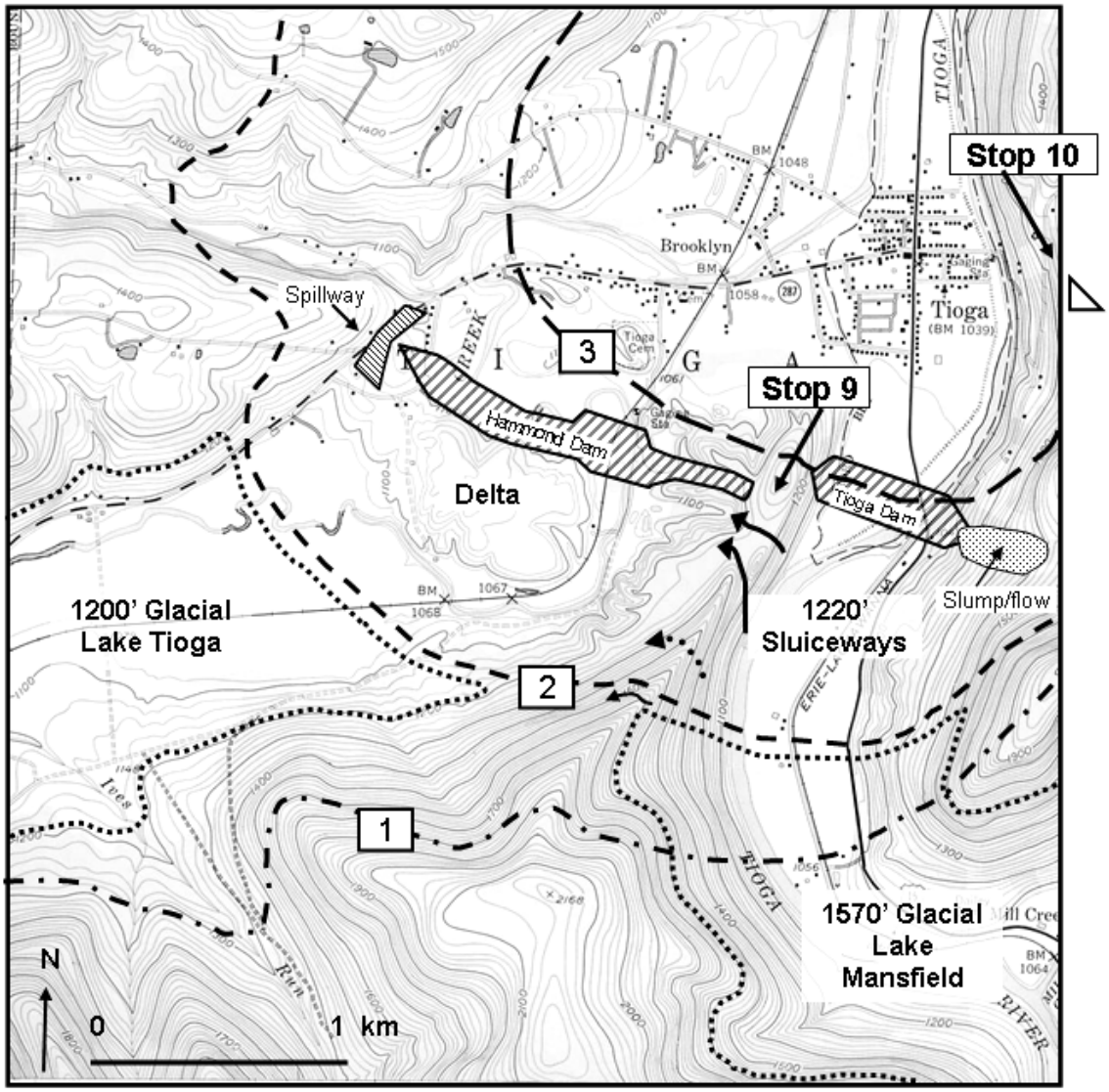


Figure T 7. Stops 9 & 10. Ice margin positions when Glacial Lake Mansfied started draining westward into Glacial Lake Tioga. Ice margin 1 is probably the last or most northerly ice position that held in 1560 feet Glacial Lake Mansfield without any ice marginal or subglacial meltwater drainage westward to 1200 feet Glacial Lake Tioga. Ice margin 2 is where ice marginal drainage westward would have begun (arrow) and there may have been some subglacial drainage (dotted arrow).

The following was done to repair the site: (1) Approximately 1,167,000 yd^3 of slump/flow material above the clay was excavated and stockpiled in the northern part US 15 road-cut north of the failure area; (2) Excavation and disposal of the clayey varve unit off site; (3) Placing of 220,000 yd^3 of rockfill from excavation the southern part of the US 15 road-cut and installing an internal drainage system; (4) Replacement and compaction of the stockpiled slump/flow material: (5) Installation of a surface drainage system and placed rock rip-rap at the toe of the fill; and (6) Installation of instrumentation to monitor compacted fill conditions. This repair was completed in 1978 at a cost of about $3.5 million. No renewed movement has occurred at the site to date. Minor movement has occurred in the fill at the south exit ramp of the PA Visitors Center a few years ago.

The cutting of the sluiceway

At this stop we are at the deep connecting channel between the two earth fill dams (Fig. T7, heavy curved arrows). The US Corps of Engineers cut the channel starting at the bottom of the 1220 feet elevation natural glacial meltwater sluiceway. Ice marginal drainage from the Tioga Valley

crossed the ridgeline to the next valley to the west, the Crooked Creek Valley. There, a plunge pool was cut out of the kame delta as the water continued southwest to the Glacial lake Tioga outlet near the head of Crooked Creek. The present dam and connecting channel system is just a smaller scale version of the previous ice dam and channel system. To the right (west) is the prominent flat topped hill that marks the top of the kame or ice-contact delta that prograded out from the ice front at the 1220 feet sluiceway (Fig. T7, ice margin 3). The Hammond Dam covers the northern ice-contact side of the feature.

This site probably experienced the rapid or even catastrophic failure of the ice dam holding in Glacial Lake Mansfield. At its greatest extent, Glacial Lake Mansfield occupied a 15 mile (24 km) long segment of the Tioga basin south of this site (Fig. P4)(Braun, 1989a). That area is a breached, anticlinal lowland and contained the lake's lowest outlet westward to the Pine Creek gorge at 1,560 feet (476 m). Between the breached anticline and here is a higher elevation synclinal mountain that the Tioga River crosses in a deep, steep-sided valley (Fig. T7, south of ice margin 3). When ice receded to the north side of the synclinal mountain (Fig.T7, ice margin 2) the 1560 feet Glacial Lake Mansfield level would suddenly be able to drop to the 1,200 feet Glacial Lake Tioga level in the Crooked Creek valley, a fall of 340 feet involving tens of cubic kilometers of water. The failure of the ice dam may have started subglacially as the 1,560 feet level waters started working their way through the ice and across the bedrock spur to the 1,220 feet level in Crooked Creek valley (Fig. T7, dotted arrow). But once flow started, piping and total failure of the ice dam should have rapidly resulted. But varves at the 1,400 feet level at a slump/flow site on the east side of the Tioga valley between ice margin 2 & 3 suggest that lowering the Glacial Lake Mansfield level was not complete until the ice front retreated to ice margin 3. The varves there may only be from a very local ice margin lake in the hollow on the side of the mountain. The 23 mile (37 km) long Glacial Lake Tioga probably acted as a flood storage reservoir to dampen the break-out flood crest because evidence of catastrophic flooding has yet to be observed at that lake's outlet. The outlet channel at the divide is now buried by a paraglacial and post glacial alluvial fan from a tributary valley.

Glacial Lake Tioga was unusual in that it had only a single sluiceway at its very upstream head. That outlet drained to the glacial drainage breached Tioga-West Branch Susquehanna divide, Pine Creek Gorge, the so-called Grand Canyon of Pennsylvania (Fig. P3, P4, and T1). This permitted the lake to maintain a single level as ice retreated for 45 miles (72 km) across the region. Usually such north draining ice dammed valleys have a series of proglacial lakes controlled by progressively lower elevation cols in the divide, exposed in sequence as the ice retreated in the downstream direction. That was the situation with Glacial Lake Gaines (Stops 2 and 3) and Glacial Lake Cowanesque (Stops 6 and 8). Glacial Lake Tioga's outlet is also unique in this region in that the sluiceway is in a valley segment where drainage has been reversed by pre-Wisconsinan derangement (Stop 1). The reversed segment provided a low gradient, gravel armored outlet for Glacial Lake Tioga that permitted essentially no outlet incision during the glacial lake's lifetime. Glacial Lake Tioga finally drained, probably “catastrophically”, when ice receded to the Chemung Valley 22 miles (35 km) northeast of here.

Glacial deposits at the site

Under the Tioga Dam the bedrock floor of the valley is nearly flat and the glacial stratigraphy is a simple layer cake across the entire valley (U.S Army Engineer Corps, 1972, Plates 14-18 – 14-22). A basal till layer, 10-40 feet thick, is overlain by 40-70 feet of silty fine sand and sandy fine silt lake sediments that are in turn overlain by alluvium, 10-25 feet thick. There are no coarse grained glaciofluvial deposits under the Tioga Dam site.

Under the broader Hammond Dam site the bedrock floor is again quite flat but the glacial stratigraphy is more complicated and there are considerable amounts of coarse grained glaciofluvial deposits (Fig. T8) (U.S Army Engineer Corps, 1972, Plates 14-24 – 14-31). From the Crooked Creek channel eastward to near the plunge pool by the connecting channel the basal till thickens to 50 feet and the lake sediments abruptly thin to zero and then reappear 550 feet horizontally to thicken to 50 feet before thinning to zero again against the bedrock valley side (right side of Fig. T8). The lake sediments are overlain and cut out by the sandy, bouldery gravel that is more than 100 feet

thick

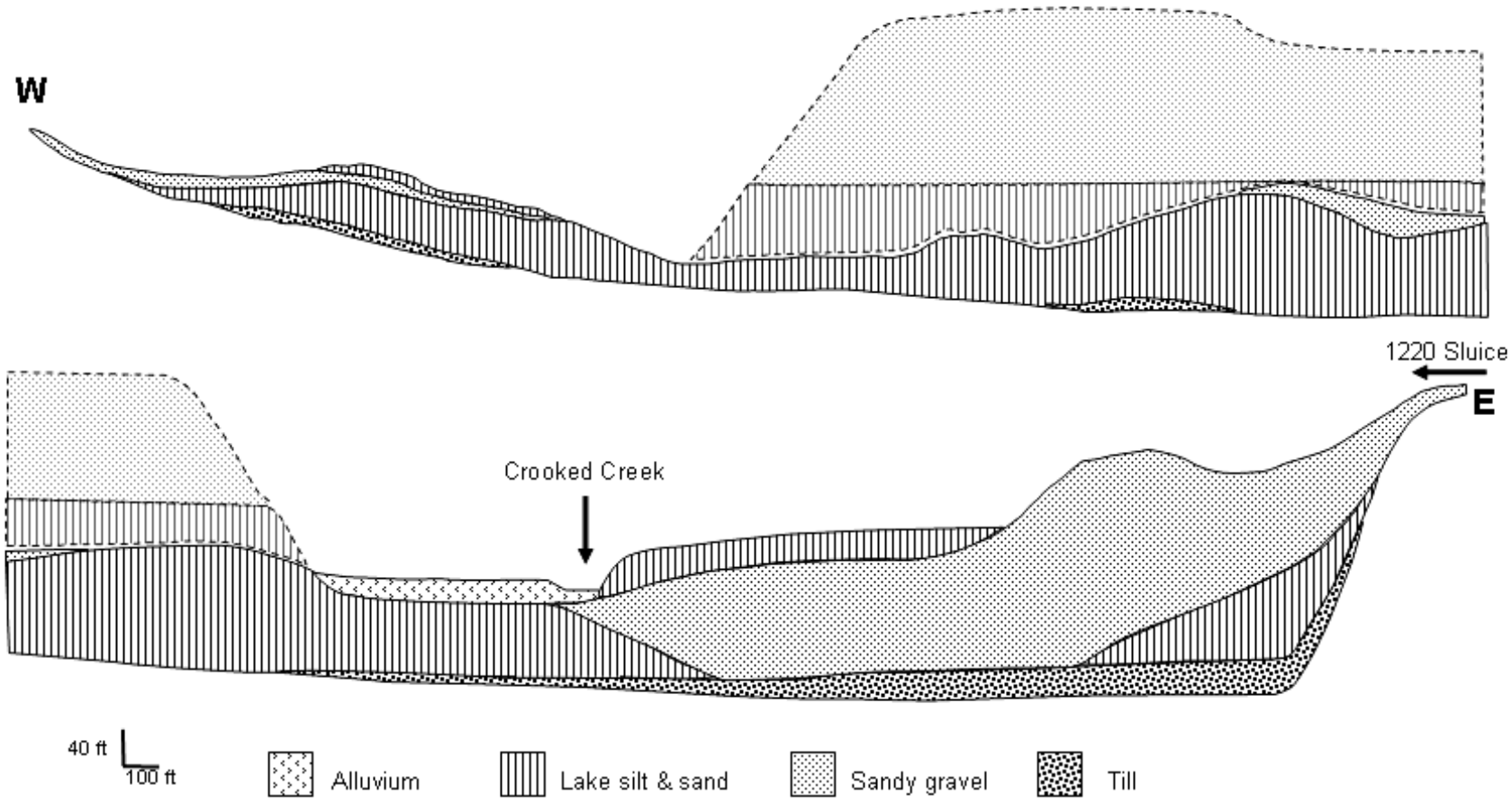


Figure T8. Simplified composite cross-section of glacial deposits under the centerline of the Hammond Dam (solid line outlines) (Engineer Corps, Soil Profile H-2) and of the delta south of the Hammond Dam (dashed line outlines) (Engineer Corps, Soil Profile H7-H9). Borehole sites and thin lenses have been removed Thin sand at top middle of centerline cross-section is probably Holocene slope wash from the higher delta surface to the south of the centerline (see Fig. T7 for topographic setting) .

the from Crooked Creek to near the east valley wall where the gravels quickly thin to 10 feet in thickness in the 1,220 feet sluiceway. Eastward from Crooked Creek to about half way to the valley wall, a second silty sand lake sediment unit overlies the gravel. To the south of the plunge pool area the base of the gravel is seen to gradually cut out the underlying lake sediment and then, under the present Crooked Creek channel, start intertonguing with the lower 40-50 of the lake sediments that lie west of the Crooked Creek channel (shown better in Soil Profile H-4 that is 500-1000 feet south and near parallel to Soil Profile H-2 shown in Figure T8). From the Crooked Creek channel westward to the west valley wall a discontinuous basal till, 0-20 feet thick, is overlain with sandy silt and silty, fine sand, lake sediments, 40-150 feet thick, that are in turn overlain by silty, sandy gravel with cobbles and boulders, 20-90 feet thick. The bedding in the gravel dips steeply southward and thickens towards the south edge of the flat topped hill in the center of the valley. Near the west edge of the valley the lake sediments thin markedly as the gravel thickens near the valley margin.

The stratigraphy from Crooked Creek to the east side of the valley is interpreted to record the initial deposition of Glacial Lake Tioga sediments; the erosion of those sediments by meltwater from Glacial Lake Mansfield coming through the 1,220 feet elevation sluice; and the waning of that flow with the deposition of the gravels and the lower part of the lake sediment unit under the center of the valley. The final recession of the ice north of ice margin 1 on Figure T7 opened up the site for continued deposition of Glacial Lake Tioga sediments.

From Crooked Creek to the west side of the Hammond valley the stratigraphic section is interpreted to record the initial deposition of Glacial Lake Tioga sediments and the deposition of additional lake sediments from the entry of Glacial Lake Mansfield discharges. The upper lake sediments and the overlying gravel record progradation of an ice-contact delta south of ice margin 1 (Figure T7) once the ice front stabilized at ice margin 1 with the draining of Glacial Lake Mansfield. This interpretation requires that there be three lake sediment sequences stacked on top of each other in the middle of the valley where there are lake sediments up to 150 feet thick. The top of the overlying gravels form the 1,180 feet elevation flat-topped hill in the center of the valley. But the 1,180 feet hilltop is 20-40 feet below the estimated 1,200-1,220 feet surface level of Glacial Lake Tioga. Other deltas a few miles to the north of here have 1,220 top surfaces (Stop 13). The top gravels here are probably a delta whose top was eroded by post glacial Crooked Creek before it incised to its present level, though a sub-lacustrine fan origin for part of the deposit cannot be ruled out. The resulting flat-topped hill across most of the valley has been termed a “valley-choker kame” (MaClintock and Apfel, 1944). Coates (1966) used the term “valley-choker moraine” for similar features composed of either till or sand and gravel, though it would be more appropriate to use the term valley-choker kame for the ones composed of sand and gravel.

Return to T intersection in Tioga and continue straight ahead (north) on PA SR 287. In about 1.5 miles bear right and ascend the entrance ramp to 4-lane US 15 south. Continue ascending the mountain ahead and then pull over on the right into the PA visitor center and rest area.

Stop 10. Overlook of the Tioga River - Crooked Creek valleys and the Tioga-Hammond dams
(41° 54’ 01.28”, 77° 07’ 34.90”)

Weather permitting, this stop is a fine overview of the Tioga River-Crooked Creek confluence area and the Tioga-Hammond dams. The site is just off the east edge of Figure T7. At this site at about 20,000 BP, one would have been standing at ice margin 3 (Fig. T7) looking at the ice front curving across the Tioga River-Crooked Creek confluence and watching meltwater pour westward in the 1,220 feet elevation sluiceway channels.

Just after leaving the Visitor Center, one will travel over the repaired 1975 slump/flow site. *Continue ahead on US 15 south. One will drive along above the Tioga Dam and Reservoir on the right, cross an arm of the reservoir on a high bridge, and continue south up the Tioga River valley.*

After crossing the bridge over the reservoir, in about 1.3 miles, to the left and 400 ft above the south bound lanes, is a 1,580 ft top surface elevation hanging delta graded to the 1,560+ ft Glacial Lake Mansfield (KD on Fig. P4). Large sandstone blocks (>3 meters across) from a shattered bedrock

ledge above the delta have been transported across the 500 foot wide, near horizontal delta top and down the delta foreset face. Under present interglacial conditions, the boulders are undergoing weathering and disintegration in place. The only process reasonably able to transport large boulders across such a gently sloping and "under drained site" is gelifluction.

Cross the Tioga River, Mansfield ahead and to the left. Bear right onto the ramp for US 6 east. Turn left onto US 6, descend to just before the bridge over the Tioga River. Turn left onto Lambs Creek Rd. and, almost immediately, turn left into Gateway Drive, the Comfort Inn access road. End of Saturday field trip.

Sunday Field Trip by Car Caravan from the Comfort Inn (41° 48' 28.28", 77° 05' 19.81")

Please car pool in the morning, we will be returning to the hotel to pick up cars before heading north for the last two stops.

Leave Comfort Inn and descend hill. Turn right and then right again onto US 6 west bound, pass under US 15, and immediately turn left onto ramp for US 15 south.

One will ascend the west flank of the Tioga valley with a view to the left (east) of Mansfield and on the skyline a line of windmills on the synclinal ridge to the southeast. At the highest point on US 15, one would still be 200 feet below the surface of Glacial Lake Mansfield. Only the highest hilltops in the area would have projected above the lake surface (Braun, 2009b). Continuing south one will obliquely cross the Tioga River valley, run along the west flank of the valley, enter the deeper valley cut in the synclinal mountain, and then obliquely cross the Tioga valley a second time before reaching the exit for Blossburg.

Bear right onto the exit ramp for Blossburg and then turn left to pass under US 15. Turn left onto Bloss Mountain Rd (old US 15) and then in a short distance turn right onto Taber St. The road will go through an industrial site (foundary). At a T intersection, turn left onto South Williamson Rd. Just after crossing the Tioga River (often yellow-red from acid mine drainage), turn right onto Gulick St. at the T intersection. Follow Gulick St. for about one-half mile and turn left into road for trailer park, bridge over Tioga River is just ahead on Gulick St. Bear left on gravel drive, ascend slope, and bear right to drive above the trailer park. Park at south end of site and walk across slumped area to reach the Tioga River bank.

Stop 11. Large scale slumps sliding on varves buried by till (41° 39' 49.59", 77° 03' 13.90")

The Tioga River is actively undercutting the toe of the slope at this site and perpetuating movement of a series of large, arcuate slumps that extend 1,800 feet (5 km) upslope (260 feet vertically) and 2,500 feet (0.7 km) along slope (Fig. T9). This is one of the sites where the varves are buried by 50 feet or more of glacial till (exposed in headwall and intermediate scarps), are at or near the base of the glacial deposit section, and produce deep failure surfaces with particularly large slump areas. The basal position of the varves indicates that they were deposited as the advancing glaciation first dammed Glacial Lake Blossburg and then were over-ridden by the glacier. The 1,690 feet surface elevation of Glacial Lake Blossburg was 190 feet above the present ground surface. Varves from ice recession have not been yet observed overlying in-situ till in the intermediate slump fissures. But the exposures are poor and there is a mantle of till derived colluivium ("colluviated till") that overlies and tends to "hide" those varves. Near surface varves were observed at elevations of about 1,500 feet in boreholes for new US 15 (future I-99) on the west side of Blossburg (between Blossburg exit and the bridge over the Tioga River).

Return to Garlick St and turn left, immediately crossing the Tioga River and Stop 11 on the left. Garlick St. becomes Ogdensburg Rd. In about 0.4 mile the "main road" curves left to cross the Tioga River again. Continue straight ahead on Ogdensburg Rd. In about 0.3 mile turn right onto graveled Taylor Run Rd.

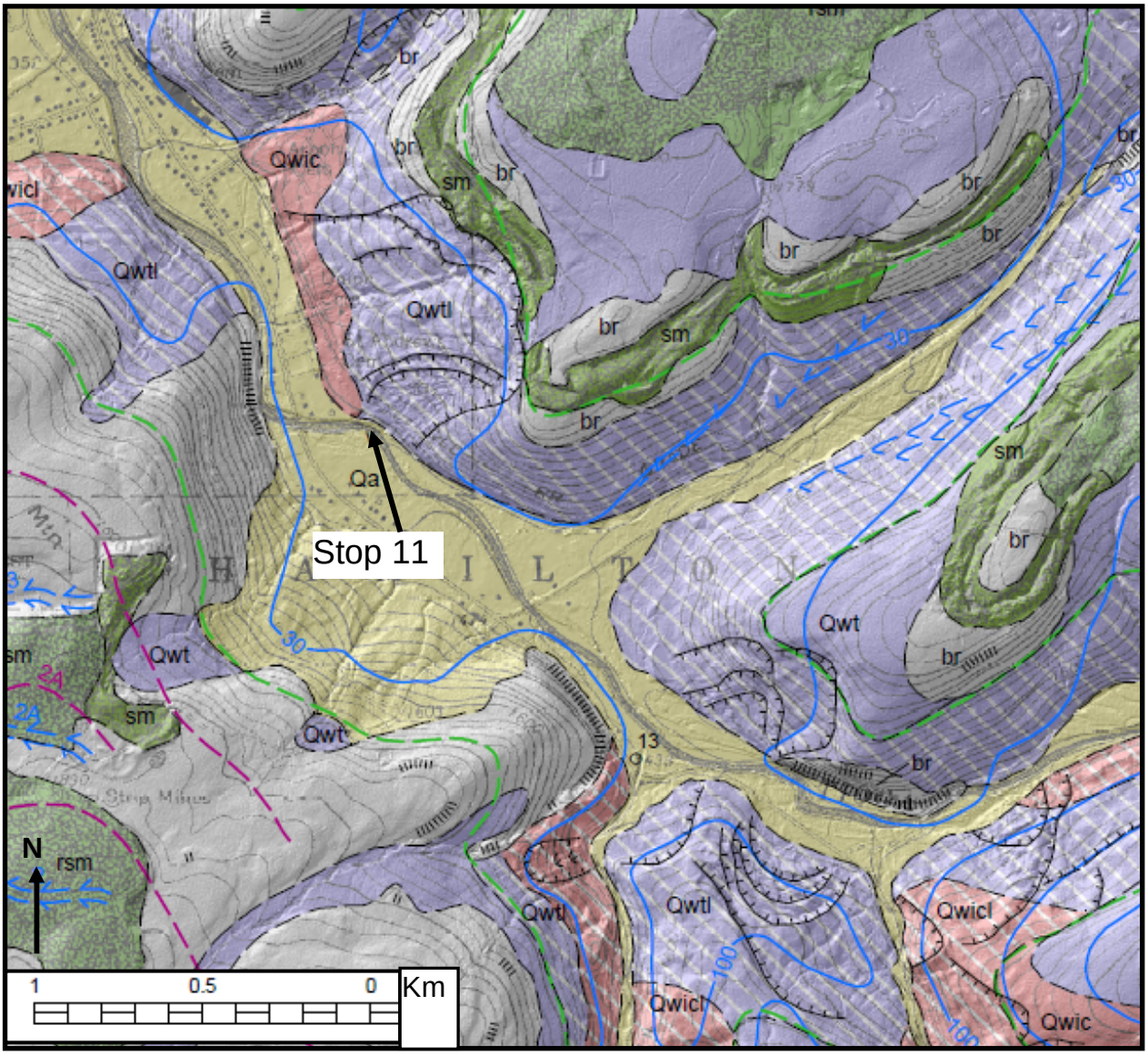


Figure T9. Stop 11. Slumps in varves overlain by till (Qwtl). Slump headwalls marked by arcuate lines with tick marks pointing in the direction of movement. See the surficial deposit legend at end of guidebook for definitions of the deposit labels. Area shown is in the southeastern part of the 7.5' Blossburg Surficial Geology quadrangle (Braun, 2010a)

A short distance up the gravel road a large slump block is on the right (See Stop 12 map) but it is not very apparent from the road. One must walk the slope. There one can walk up a narrow fissure on the north side of the slump to the headwall fissure and then back down a fissure on the south side of the slump. A bit farther up the road on the right are gravel pits in a valley side kame.

In about one mile Taylor Run Rd. becomes a State Forest Land road that is not maintained in the winter and has minimal maintenance at other times. Continue ahead for another two miles to the top of the mountain. Pull over and park along the right side of the road.

Stop 12. Mountain top recessional moraine (41° 36' 56.66", 77° 02' 42.15")

The new Lidar-derived three-foot-scale imagery showed there is one strip of subtle till moraine knob and kettle topography on the top of the broad synclinal upland near the central western edge of the Blossburg quadrangle (Qwtm on figure T10)(Braun, 2010). Walking the site confirmed there are a

number of shallow closed depressions at the scale of a few feet deep and tens of feet across holding vernal pools separated by knobs of similar scale. The area is mostly covered by dense laurel thickets and is well disguised on even the “leaf off” aerial photography.

Also the new Lidar imagery showed there are three 500- to 1,000-foot wide belts of subtle till moraine knob and kettle topography on the top of the broad synclinal upland in the southwest corner of the Liberty quadrangle (immediately south of the Blossburg quadrangle)(Braun, 2010). Walking the site again confirmed there are a number of shallow closed depressions at the scale of a few feet deep and tens of feet across holding vernal pools separated by knobs of similar scale. The southwestern most belt is the late Wisconsinan terminal moraine. The other two moraine belts are parallel to and northeast of the terminal moraine. Each moraine is separated from the other by a 1,000 to 2,500 feet wide ground moraine belt displaying essentially no morainic topography.

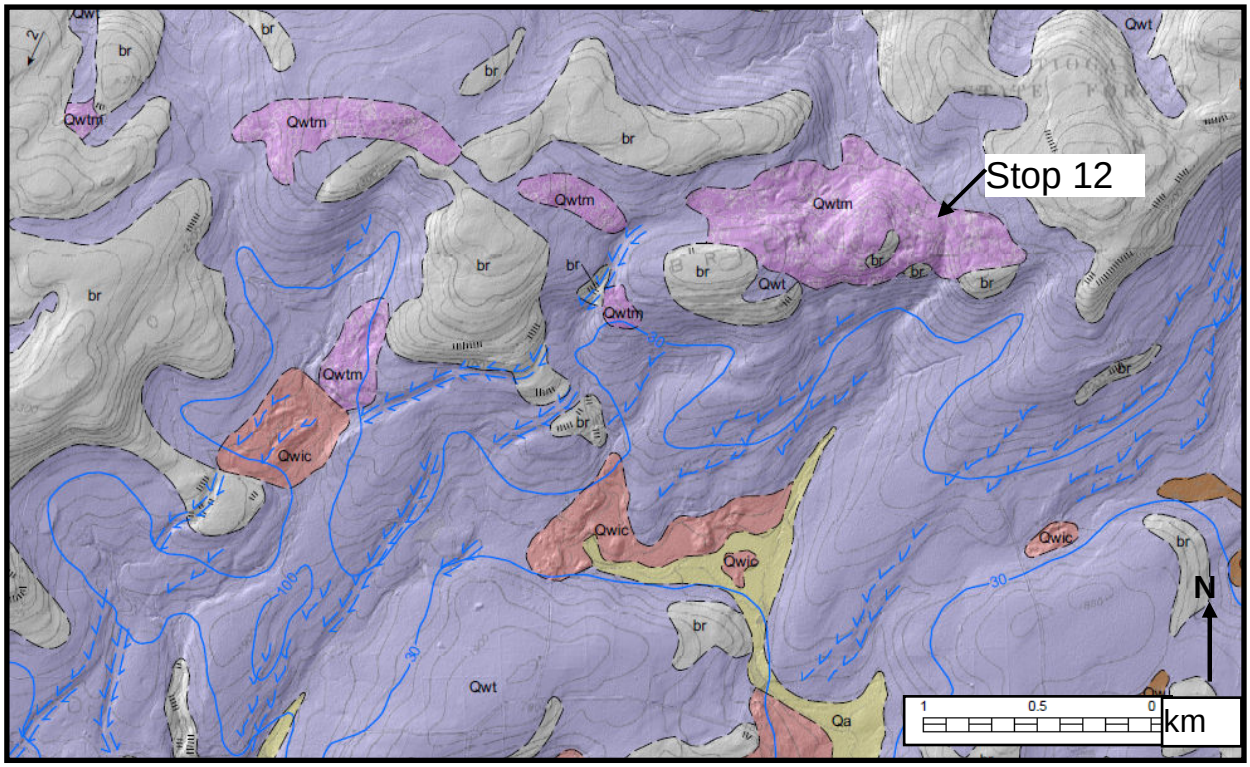


Figure T10. Stop 12. Mountain top moraine (Qwtm) initially observed on the Lidar image. On the south flank of the mountain are a number of ice marginal meltwater channels (barbed dashes). Many channels were cut in till on their northwest sides and were ice walled on their southeast sides. Some channels were cut entirely into till or bedrock. Their southwesterly trend shows the lobation of the ice as it flowed over the mountain top and into the lower hilly area to the south. See the surficial deposit legend at end of guidebook for definitions of the deposit labels. Area shown is in the northeastern part of the 7.5’ Liberty Surficial Geology quadrangle (Braun, 2010b).

Continue ahead to where another roadway enters on the left. Turn around there and retrace route back through Blossburg to US 15. Take north bound ramp of US 15 to head back to the Comfort Inn in Mansfield. Pick up vehicles at the Comfort Inn and return to US 15 taking the north bound ramp. You will be retracing the route from the last stop on Saturday but in the north-bound direction. Go past the PA Visitors Center & Rest Area on the left and in about 1.5 miles, as one descends the mountain, bear right onto the exit ramp to PA SR 287 (old US 15). In about 0.6 mile, bear right onto Bradshaw Rd. In about 0.7 mile turn right into the entrance to the GOH gravel pit, and follow signs to the office.

Stop 13. Ice-contact delta in Glacial Lake Tioga & Lunch (41° 56' 56.64", 77° 05' 54.05")

This site is one of the more active gravel pit operations providing drilling pad material to the gas drilling industry (Fig. T11). Pit faces are changing continuously so one can't predict what we'll be able to see when we arrive. From time to time the topset-foreset contact has been exposed. Usually there are exposures of the foresets and occasionally the bottomsets. The gravel clasts are dominated by subrounded local "drab colored" bedrock fragments. One must search quite a bit for "bright colored" crystalline clasts from Canada or carbonate clasts from New York.

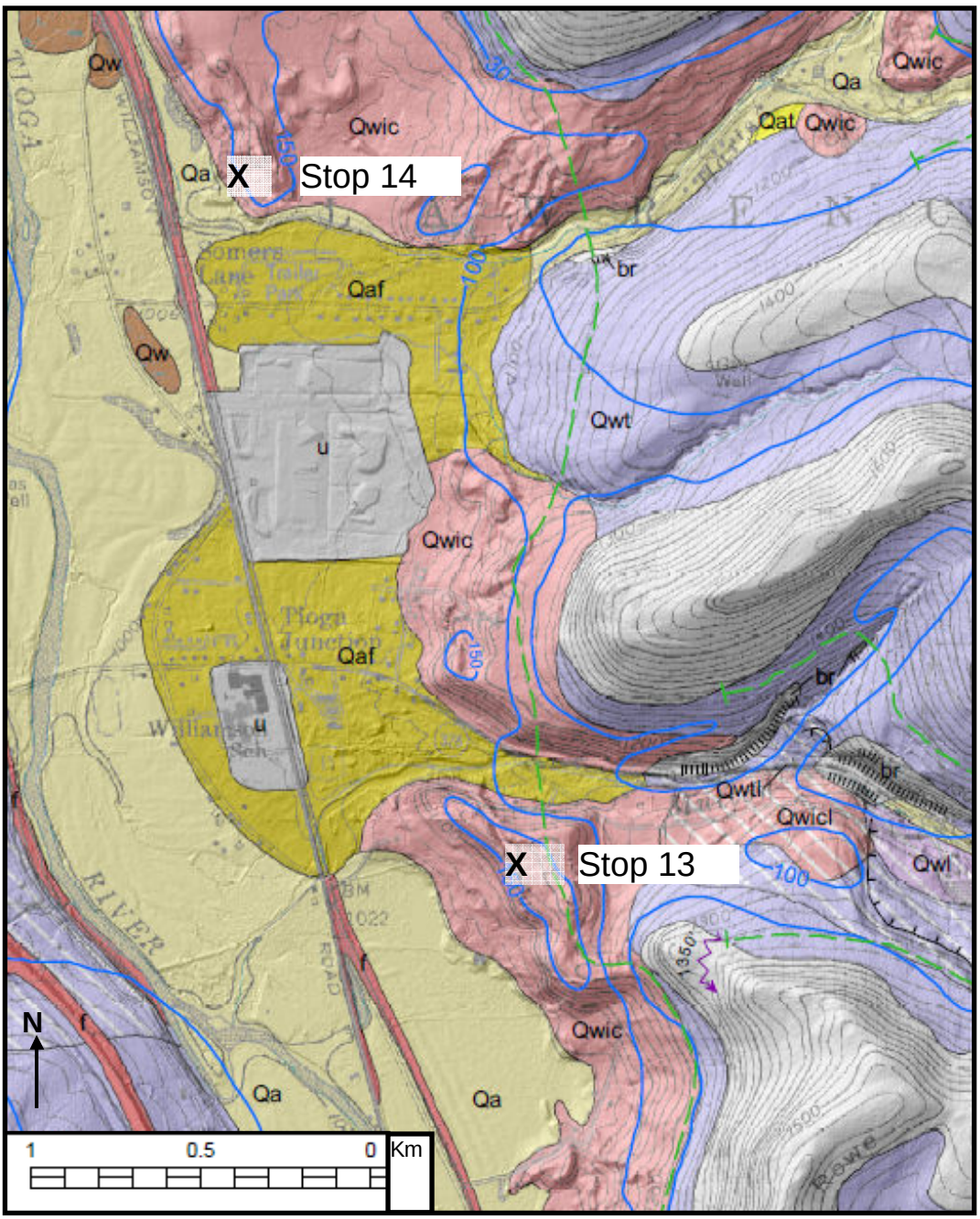


Figure T11. Stops 13 & 14. Deltas built into Glacial Lake Tioga. Stop 14 is the main body of the one delta. Stop 15 is the ice contact face of another delta 2 km north of Stop 14. See the surficial deposit legend at end of guidebook for definitions of the deposit labels. The area shown is in the northwestern part of the 7.5' Jackson Summit Surficial Geology quadrangle (Braun, 2009a).

Denny and Lyford (1963) produced reconnaissance surficial deposit maps of the entire region. They examined all the larger glaciofluvial deposits, focusing on the relative amount of far traveled erratics, particularly carbonates, and how that content influenced the depth of carbonate leaching. On their Plate 3 they noted that the ice-contact glaciofluvial deposits or kames east of Lawrenceville and northeast of Tioga Junction (at and north of Stop 14) both had a significant amount of carbonate and crystalline erratics while the kame deposit southeast of Tioga Junction (Stop 13) had almost no such material. They noted that in general "These deposits (kames) accumulated in association with wasting glacial ice, presumably stagnant, but the location and extent of such stagnant ice is conjectural (p. 19)." They also noted that valley train gravels were washed out ahead of the advancing glacier. The glacier then picked up those gravels and produced till and glaciofluvial deposits enriched in that material in the down ice-flow direction from those valleys.

Glacial Lake Tioga would have been in front of the glacier in the Tioga valley as it retreated across the region and would have remained until the ice retreated another 10 miles to the north to the Chemung valley in New York State. There is a nearly continuous belt of ice-contact stratified drift deposits for 3.5 miles (5.6 km) along the east side of the Tioga valley in this area. In detail, the belt can be separated into a series of south to north segments, each about one-half to one mile apart and marked by a flat topped area at 1,220 feet. The flat topped features between the villages of Mitchell Creek and Tioga junction were sites of sand and gravel pits in the 1980's that showed that the features are large ice-contact or kame deltas with a topset - foreset contact at about the 1,220 or so elevation with the foresets dipping southward. It is reasonable to assume that the other flat topped features to the north are also kame deltas built into Glacial Lake Tioga as the glacier continued its northward retreat. This regular spacing of same elevation kame deltas is strong evidence that the glacier was retreating in an episodic backwasting mode or stagnation zone retreat mode (Koteff, 1974; Koteff and Pessl, 1981). The downwasting retreat mode leaving long tongues of stagnant ice in the Tioga valley was favored by Alden and Fuller (Fuller, 1903b) and Coates (1966b). But that should have left a series of flat topped stratified drift remnants that decline in elevation from north to south due the continuous drainage of the meltwater beside the long, stagnant ice mass rather than a series of equal elevation features.

In the center and west side of the Tioga valley well data shows that there is a continuous belt of glacial lake sediments, locally with some interbeds of coarser deltaic material (Williams and others, 1998, Plate 1A, cross-sections A-A", B-B', C-C'). This indicates that the kame deltas did not prograde across the entire width of the Tioga valley but were restricted to the east side of the valley. That the bulk of the ice-contact stratified drift deposits are on the east side of the north trending valley is expectable due to the overall northeast to southwest orientation of the retreating ice front. When the backwasting ice front occupies any particular position across the Tioga valley, the tributary valleys to the west would be ice free while the tributary valleys to the east would be ice covered and producing sediment laden meltwater both sub-glacially and supra-glacially.

Return to Brashaw Rd., turn right and then bear right onto PA SR 287. In about 1.3 miles turn right into another gravel pit right beside the road.

Stop 14. Esker feeding delta at Stop 13 (41° 58' 13.52", 77° 06' 30.0")

This stop is in an esker on the floor of the Tioga River valley (Fig. T11) that fed the deltas to the south of it (between Stop 13 and 14). It shows the typical ice-contact features of chaotic bedding, abrupt changes in grain size, coarse grained clasts from sand to boulder size, and some faulting of the deposit. Immediately to the south of here the esker has been eroded away so one cannot see how it directly connected to the delta at Stop 13. Today, two large, coalesced para-glacial to post glacial fans from east side tributaries occupy the valley floor between Stop 13 and 14.

End of trip. For those going north turn right onto PA SR 287, go two miles north to Lawrenceville and turn left at traffic light onto PA SR 49. Go about one-half mile and turn right onto ramp for US 15 north. For those going south retrace route to US 15 south.

Explanation Of Surficial Deposit Map Units

f **FILL:**
Rock fragments and/or soil material; typically in road, railroad, or dam embankments; up to several 10's of feet thick.

U **URBAN LAND:**
Cut and fill disturbing more than 50 percent of the ground surface; includes most areas with homes on one-half acre or smaller lots, and commercial/industrial sites.

rsm **RECLAIMED COAL SURFACE MINE:**
Surface mine area now regraded to approximately the original land surface form; partly to completely vegetated.

sm **UNRECLAIMED COAL SURFACE MINE:**
Alternating linear pits and broken rock-waste piles; generally pits are hundreds of feet long and less than a hundred feet deep; nearly all pits are abandoned; some areas have coal remaining at depth.

Qa **ALLUVIUM:**
Stratified silt, sand, and gravel, with some boulders; subrounded to rounded clasts; contains localized lenses of silty or sandy clay; usually is underlain by other unconsolidated material (glacial deposits); 6 feet thick in headward tributary valleys, as much as 15 feet thick in major valleys.

Qat **ALLUVIAL TERRACE:**
Stratified silt, sand, and gravel with some boulders; subrounded to rounded clasts; the deposits form benches running parallel to and a few feet above the present floodplain; usually is underlain by glacial deposits; 6 feet or more thick.

Qaf **ALLUVIAL FAN:**
Stratified silt, sand, and gravel, with some boulders; subrounded to rounded clasts; having a fan shaped landform; usually is underlain glacial deposits; 6 feet or more thick.

Qw **WETLAND:**
Area with standing water for part of each year; usually underlain by peat, clay, silt, sand, or some combination of those materials beneath which are glacial deposits; thickness of peat usually less than 1.5 feet; overall thickness of unconsolidated material is usually greater than 6 feet.

Qwo **WISCONSINAN OUTWASH:**
Stratified sand and gravel typically forming terraces along the flank of a valley; the overall stratification is horizontal with individual strata showing cross-beds, ripples, clast-imbrication, or cut-fill features; thickness typically is 10 to 30 feet.

Qwic **WISCONSINAN ICE-CONTACT STRATIFIED DRIFT:**
Stratified sand and gravel with some boulders; often chaotic stratification; some internal slump structures; gently sloping upper surfaces with a few closed depressions; typically deposited in valley side kames; generally not more than 30 feet thick except along the east flank of the river valley where it exceeds 100 feet in thickness.

Qwt **WISCONSINAN TILL:**
Glacial or resedimented till; texturally a diamicton, a nonsorted or poorly sorted, unconsolidated deposit that contains a wide range of particle sizes, commonly from clay to cobble- or boulder-size, and rounded and/or angular fragments with a clayey, silty, or sandy matrix depending on the local source bedrock; poor to multimodal sorting; unstratified to crudely stratified with a clast fabric; striated cobble and boulder clasts are common; rare clasts are far traveled crystalline erratics; typically occurs as a fairly smooth landform with a bouldery surface and little distinct constructional (knob and kettle) topography on hillslopes; upper 3 feet is often colluviated, displaying a downslope-oriented fabric; thickness is greater than 6 feet, is typically 15 feet , and can be greater than 150 feet in buried to partly in-filled valleys.

Qwl **WISCONSINAN GLACIAL LAKE SEDIMENT:**
Varves, massive to laminated silts with clay drapes, and massive fine sands; coarser sediments occur close to where meltwater flowed from the glacier forming ice-contact deltas or sub-lacustrine fans; varves are alternating layers of silt and clay; each layer usually less than an inch thick; each pair of silt and clay layers (a couplet) represents a single year's summer (silt) and winter (clay) deposition in a proglacial lake; clay usually ¼ to ½ of the volume of the deposit; thickness may exceed 100 feet under the floors of major valleys; hill-slopes underlain by varves typically show a stepped appearance from slumping; such slopes are hazardous to build upon and failure can be aggravated by cut and fill excavations.

Qwol **WISCONSINAN OUTWASH UNDERLAIN BY GLACIAL LAKE SEDIMENT:**
Outwash as described above in Qwo underlain by clay rich proglacial lake sediments; underlying deposits are mostly varves as described above in Qwl; typically the outwash is a few feet to a few 10's of feet thick and is underlain by similar thicknesses of varves.

Qwicl **WISCONSINAN ICE-CONTACT STRATIFIED DRIFT AND GLACIAL LAKE SEDIMENT:**
Ice-contact stratified drift as described above in Qwic underlain by or interbedded with clay rich proglacial lake sediments; underlying deposits are mostly varves as described above in Qwl; typically the ice-contact stratified drift is a few a few 10's of feet to more than 100 feet thick and is underlain by similar thicknesses of varves or coarser grained lake sediment.

Qwtl **WISCONSINAN TILL AND GLACIAL LAKE SEDIMENT:**
Till as described above in Qwt or colluivum derived from till underlain or interbedded with or overlain by clay rich proglacial lake sediments; underlying deposits are mostly varves as described above in Qwl; typically the till is a few feet to more than 100 feet thick and is underlain by similar thicknesses of varves or coarser grained lake sediment.

Qwtm **WISCONSINAN TILL MORAINE:**
Till as described above in Qwt with a knob and kettle land surface topography. Kettles typically have 1 to 10 feet of closure and contain vernal pools.

br **SANDSTONE AND SHALE BEDROCK:**
Bedrock outcrops or clast-rich diamict of residual and colluvial material derived from the directly underlying bedrock of interbedded red and gray sandstone, and shale; clayey silt to sandy silt matrix; clasts are typically matrix-supported with lenses of clast-supported material with or without matrix; tabular clasts generally exhibit a down slope directed orientation within the upper 1.5-3.0 feet of the material; less than 6 feet to bedrock; on greater than 25 percent slopes, typically less than 3 feet thick.

MAP SYMBOLS

Bedrock Ledge Outcrop

30 **Isochore:**
Thickness of surficial deposits in feet, as measured in vertical boreholes or in excavations with vertical faces. Where deposits overlap, contours represent the combined thickness of the stacked surficial units. Shown for thicknesses of 30, 100, and 150 feet.

Glacial Striation:
Arrow shows inferred direction of glacial flow. Site number above the arrow. Point at the head of the arrow marks the location of the striation site.

Esker:
Ridge, often winding, of ice-contact stratified drift (gravel and sand). Typically tens of feet high, tens to hundreds of feet wide, and thousands of feet long. It represents the bed of a meltwater stream that once flowed in a tunnel within or under a glacier. Chevrons point in the direction of transport.

1770' **Glacial Lake Spillway:**
Arrow shows direction of meltwater flow. Number indicates elevation of the floor of the spillway channel.

Lateral Moraine:
A low ridgelike moraine 5 to 15 feet high. The moraines usually trend obliquely across a hillside for several hundred feet and often occur in groups of two to five, near-parallel, somewhat sinuous features.

Slump or Slide:
Headscarp

Headscarp, location approximate

Tick marks on downdropped side of headscarps, indicating direction of movement. Nearly all slump-slide slope failures occur in areas underlain by clayey glacial lake sediments. The clayey sediments are sometimes at the ground surface but more often are buried by other glacial sediments or colluvium. In places, the slope failure occurs where clayey till is sliding on the underlying bedrock.

GLA1

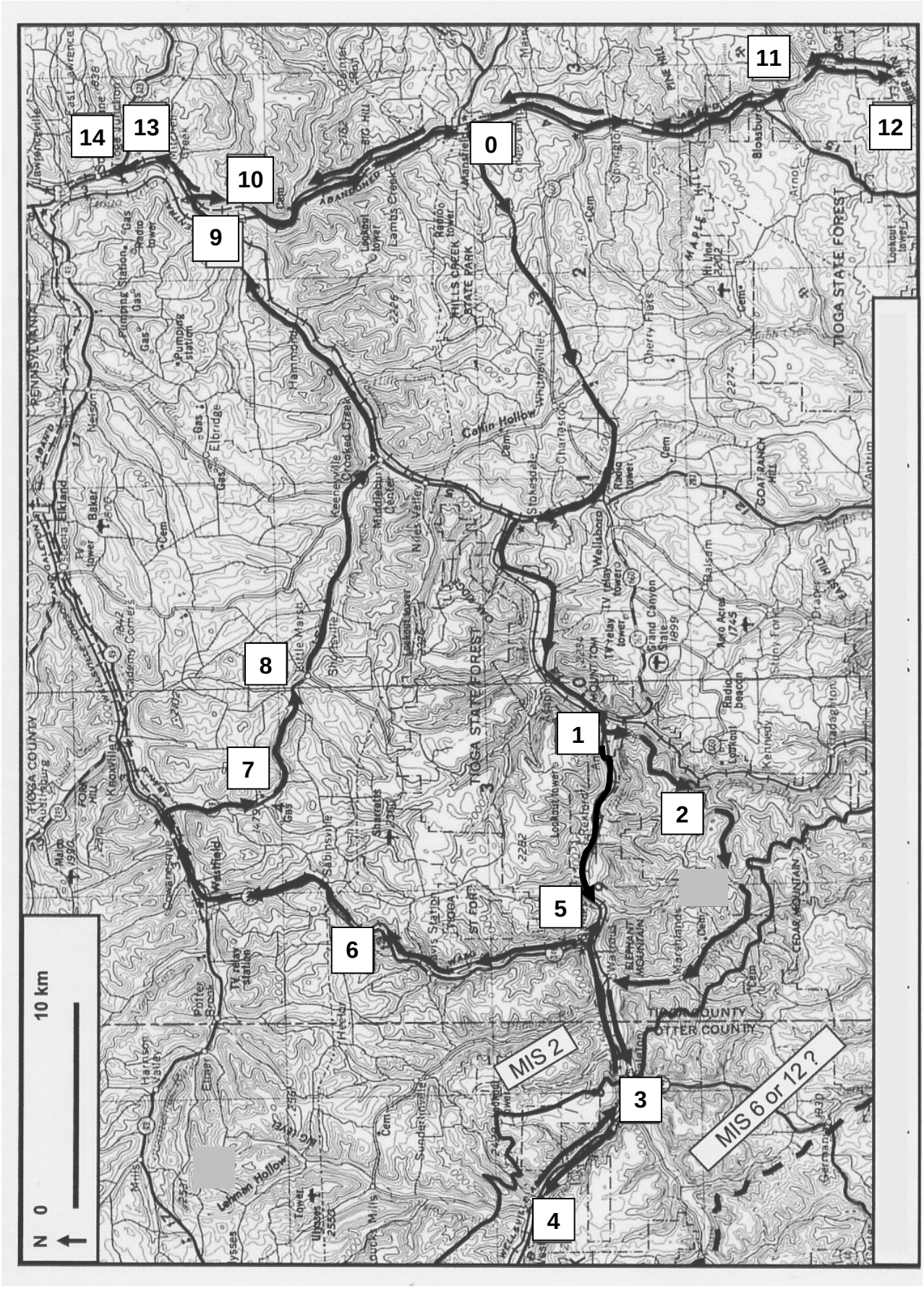
GLA2

Glacial Lake Outline:
Dashed lines mark the outlines of glacial lakes. Hachure indicates estimated point of contact between lake outline and glacier. Arrow connects a higher lake level to a lower lake level. It marks the position of the edge of the glacier when glacial retreat opened up a new, lower outlet for the lakes in that area. Glacial lake abbreviations are on the lake side of the lines (see Table 3).

TK **Till Knob:**
Rounded knob of till on the side of a valley. Typically a few tens of feet high. A wetland is often present up valley of the knob. The stream valley beside the knob has a markedly narrower form than elsewhere in the valley, and often the stream has cut into bedrock on the side of the valley opposite the knob**.**

KK **Knob and Kettle Landform:**
Rounded knobs and shallow depressions, a few feet to a few tens of feet high or deep and tens to hundreds of feet wide. Underlain by glacial till (Qwt) or ice-contact stratified drift (Qwic).

Open water:
Area of open water where the underlying surficial unit is uncertain.



FOP 2011 field trip route, stops 1-10 Saturday, stops 11-14 Sunday.