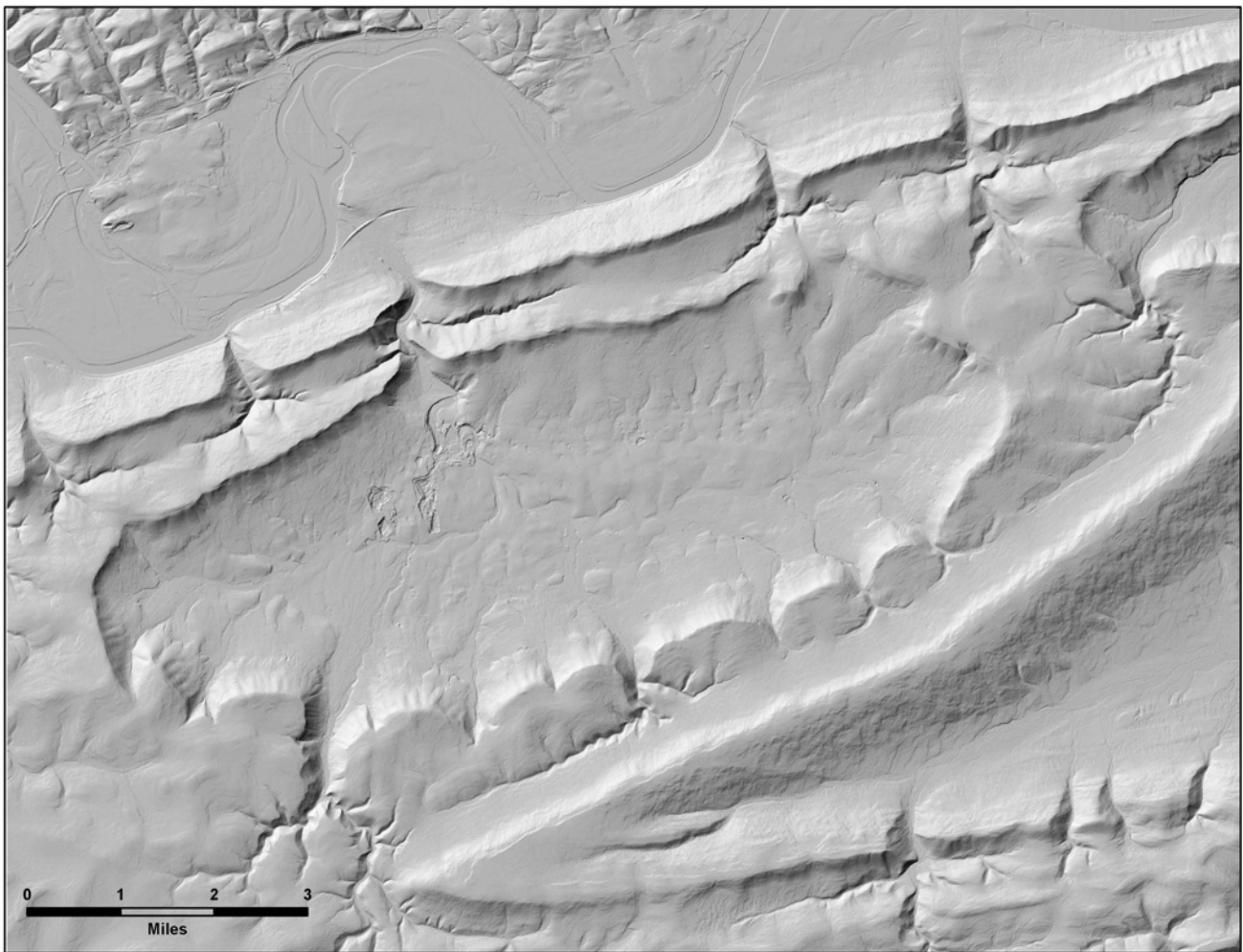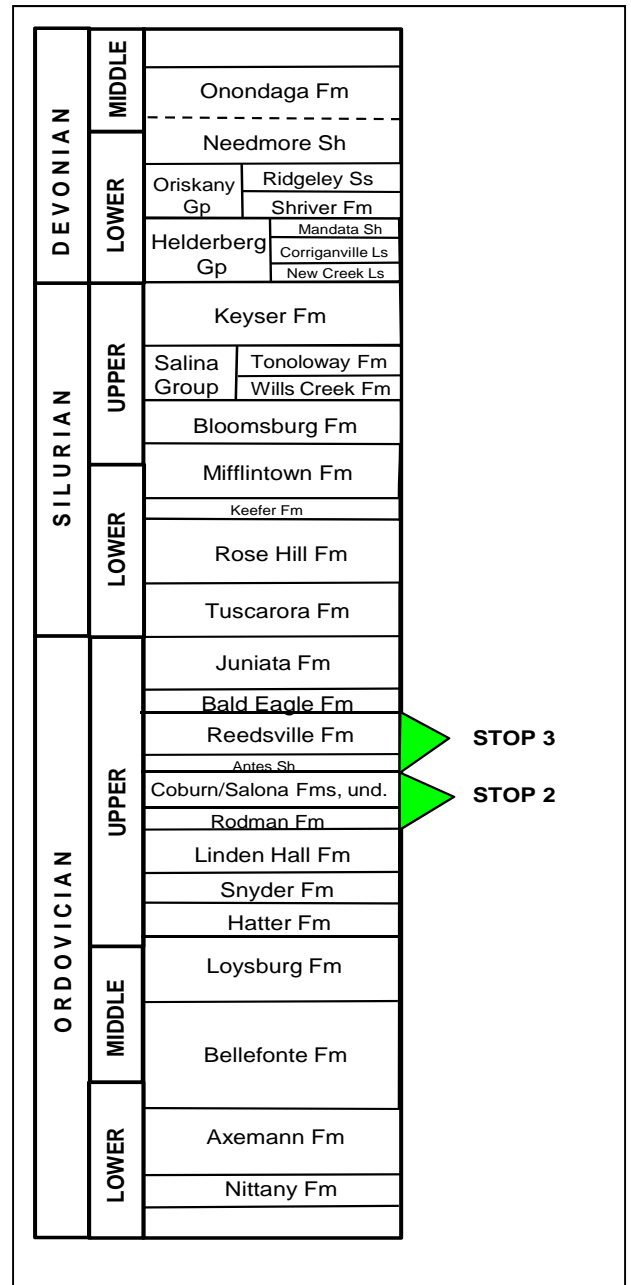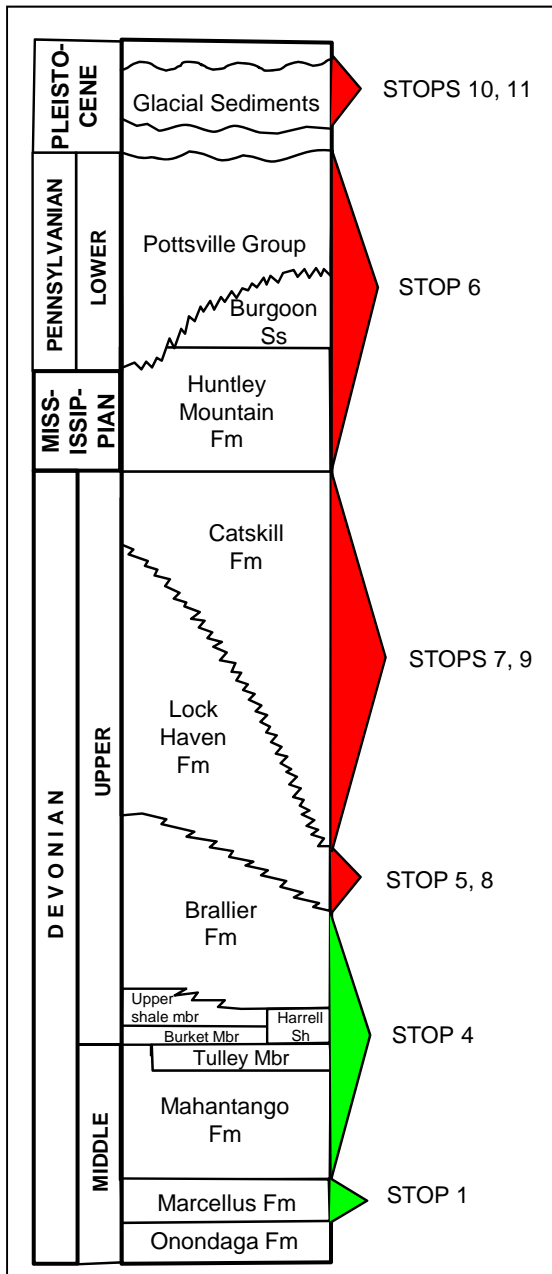# The 78th Field Conference of Pennsylvania Geologists

The Nippenose Valley and the Route 15 Corridor:
A Tale of Two Provinces

September 27 – 28, 2013

Williamsport, PA

Hosts: Pennsylvania Geological Survey
Bloomsburg University of PA

Left column (Devonian–Pleistocene):

PLEISTOCENE		Glacial Sediments	STOPS 10, 11
PENNSYLVANIAN	LOWER	Pottsville Group / Burgoon Ss	STOP 6
MISSISSIPPIAN		Huntley Mountain Fm	
DEVONIAN	UPPER	Catskill Fm	STOPS 7, 9
		Lock Haven Fm	
		Brallier Fm	STOP 5, 8
	MIDDLE	Upper shale mbr / Burket Mbr / Harrell Sh / Tulley Mbr	STOP 4
		Mahantango Fm	
		Marcellus Fm	STOP 1
		Onondaga Fm	

Right column (Ordovician–Devonian):

DEVONIAN	MIDDLE	Onondaga Fm
	LOWER	Needmore Sh
		Oriskany Gp: Ridgeley Ss / Shriver Fm
		Helderberg Gp: Mandata Sh / Corriganville Ls / New Creek Ls
SILURIAN	UPPER	Keyser Fm
		Salina Group: Tonoloway Fm / Wills Creek Fm
		Bloomsburg Fm
	LOWER	Mifflintown Fm
		Keefer Fm
		Rose Hill Fm
		Tuscarora Fm
ORDOVICIAN	UPPER	Juniata Fm
		Bald Eagle Fm
		Reedsville Fm — STOP 3
		Antes Sh
		Coburn/Salona Fms, und. — STOP 2
		Rodman Fm
		Linden Hall Fm
		Snyder Fm
		Hatter Fm
	MIDDLE	Loysburg Fm
		Bellefonte Fm
	LOWER	Axemann Fm
		Nittany Fm

Generalized stratigraphic column for the stops in this year's Field Conference. Column is modified from Berg, T.M., McInerney, M.K., Way, J.H., and MacLachlan, D.B., 1983, Stratigaphic correlation chart of Pennsylvania: Pennsylvania Geological Survey, 4th Series, General Geology Report 76, one sheet.

Guidebook for the
78th ANNUAL FIELD CONFERENCE OF PENNSYLVANIA GEOLOGISTS
September 26-28, 2013

A Tale of Two Provinces:
the Nippenose Valley and Route 15 Corridor

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Hosts: Pennsylvania Geological Survey
             Bloomsburg University of Pennsylvania

Headquarters: Holiday Inn, Williamsport, PA

Cover: Lidar hillshade image of the Nippenose Valley, from Day 1 of the 2013 Field Conference of Pennsylvania Geologists. North White Deer Ridge, composed of the Tuscarora Sandstone, and parallel cuesta ridge, composed of Bald Eagle sandstone, are clearly visible on the south side of the Nippenose Valley in this image. Antes Creek flows through Antes Gap, through both the Bald Eagle and Tuscarora ridges, draining the entire valley to the north.

The Nippenose Valley is a breached, first-order, doubly-plunging anticline. The White Deer Syncline, to the south, is also a first-order fold.

At the northwestern corner of the photo, the beginning of Appalachian Plateau topography is visible. The West Branch of the Susquehanna River flows past Jersey Shore between the Nippenose Valley and the Appalachian Plateau

Group photo of the 2012 Field Conference of Pennsylvania Geologists. Stop 3- Lehigh Gap

Photo by Yuriy Neboga

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INTRODUCTION

The process of selecting an area for the 78th Annual Field Conference started with a handful of geologists standing about 3 feet away from a mounted copy of the state geological map staring at the colored patterns waiting for inspiration. Likely candidates were considered and discussed; pros and cons, been there, done that. Then came the logistics; selecting a motel that could serve 150+ geologists, and most importantly, making sure that there were locations for supplying enough donuts for the morning beverage break.

The theme, "a tale of two provinces," came about after seeing the kink-folded bedrock of the Lock Haven Formation giving way to the near horizontal layering of the Catskill redbeds along US 15 north of Williamsport. Was this the end of the Ridge and Valley and the beginning of the Plateau Province?

This served as the impetus for developing a Conference in the Williamsport region. Previous Field Conferences have explored Tioga County (46th, 1981) and the Lock Haven/Williamsport area of Lycoming and Clinton Counties (60th, 1995). While we may traverse some of the same roads, the focus will cover different themes.

Day 1 begins with examining portions of the Ordovician and Devonian that pertain to major gas plays in Pennsylvania. Exposures of the lower Marcellus (Stop 1) and Lower Reedsville and Antes Formations (Stop 3) provide first-hand view of these important gas-producing zones. The Lockport section (Stop 4) is one of the best exposures in north-central Pennsylvania for the important stratigraphic markers, the Tulley limestone and overlying Burket Member of the Harrell Formation, marking the division between the Middle and Upper Devonian.

Also in Day 1, the Conference will examine the development of the karst landscape in the great amphitheater of Pennsylvania, the Nippenose Valley. Hidden along the pathways of hemlock and hardwoods are sinkholes, caves, and the Nippeno Spring, Pennsylvania's largest (Stop 2).

The final stop of Day 1 will cover a portion of the Pine Creek Trail (Stop 5). The stroll along the scenic trail will trace the structural last gasps of the Ridge and Valley Province observing beds of the Lock Haven Formation go from near-vertically dipping beds to a very abrupt transition to nearly horizontal beds.

The US Route 15 corridor (Day 2) has to be considered one of the most important stratigraphic exposures in Pennsylvania. Rocks ranging in age from the Upper Devonian through the Lower Pennsylvanian are exposed in sometimes lengthy roadside exposures amid some of the most scenic highlands of the High Plateau Province.

One will notice the Lock Haven Formation within the first few miles north of Williamsport bearing remarkably similarities to the section viewed on the previous day at Stop 5.

As one enjoys the scenic drive north, the transition into the characteristic redbeds of the Catskill Formation dominates the roadside viewing. The relatively flat terrain gives way to the long ascent up the flank of the Plateau Province where glacial and periglacial events have directed the beveling and carving of these highlands.

Entering the Blossburg syncline, the stratigraphic sequence changes rather rapidly. Just to the north and south of Blossburg, the contacts between the Upper Devonian Catskill and the Lower Mississippian Huntley Mountain are separated by unconformity (Stop 7). The stratigraphic framework is further challenged by disconformable layering between the Huntley Mountain and overlying Lower Pennsylvanian Pottsville sandstones (Stop 6).

Stop 8 is a lead in to Stop 9 and takes a brief look at the paleontology of the abundantly fossiliferous, Devonian Lock Haven Formation. The gradational marine to non-marine transition of the Lock Haven with the overlying Catskill Formation is best viewed at Stop 9.

Finally, with the Empire State in view, sediments of the Lawrenceville esker provides evidence of the glacial past (Stop 10) and the impact it had on establishing our present-day landscape (Stop 11).

Nippenose Valley, Lycoming County, Pennsylvania

Wayne Welshans

Nippenose Valley is about 10 miles long and four miles wide with a base of limestone completely surrounded by a double row of sandstone mountains. It lies near the Susquehanna River in the Bald Eagle Mountains of Central Pennsylvania and was on the American frontier until after the Revolutionary War. This land was acquired from the Indians by the first treaty of Fort Stanwix in 1768 and the first survey here was in 1769. Its early history inhabitants include: Nippenuce, Wi-daag, John Henry Antes, Peter Pence, John Clark, William Winland, Michael Showers, Jacob Sallada …The second treaty of Fort Stanwix in 1784 cleared the land of Indians and by 1814 approximately 30 families had settled here.

Some 13 small streams enter the valley (five are significant) from the mountains and begin to "sink" upon reaching the limestone at about 870 feet in elevation thus, there was little available water on the valley floor. The springs were located near the base of the mountains and the first settlers built their cabins at these sites. By 1850 the communities of Rauchtown, Jamestown, Oval, Collomsville and Bastress had been established. Schools and churches and mills were built and a patchwork of well kept farms covered the valley floor. It is interesting to note the area's development by way of successive maps

The first map to bear the name "Pennsylvania" was the Thornton Map which was created for William Penn in 1681 to help promote his new colony like had been done 50 years earlier to promote settlement of the Maryland colony. This map showed little but the "Sasquahana" River and mountains. The North and West Branches had not yet been identified, but the map did help promote the availability of cheap land – something that was but a dream to most Europeans of little means. From the beginning boundary disputes were many. This map got the boundary of Pennsylvania 40 miles too far south and the problem was not settled until 1773 by Mason and Dixon. The Connecticut colony claimed the upper third of Pennsylvania and their charter stretched to the Pacific Ocean. Connecticut settlers occupied the land as far west as Muncy. The French claimed the northwestern part of the state and the Virginia colony the southwest.

The **Nicholas Scull** map of 1759 did show the North and West Branches of the Susquehanna connecting near Fort Augusta plus tributaries in our area. Another map by **William Scull** was made in 1770 for Thomas and Richard Penn. Although Nippenose Valley is not identified, close landmarks are: Cawichnowane (Long Island at present Jersey Shore), Pine Creek, Larty's Creek, Great Island (at present Lock Haven) and Nippenosis Creek (Antes Creek today). This was followed by the **Reading Howell** map of 1792 which was presented to Governor Thomas Mifflin. This was the first map to show "Nepanose" Valley complete with 11 sinking streams coming off the mountains, and marked the towns of Jersey Shore, Newberry and Williamsport. The map also shows county and township lines, roads, furnaces, forges, grist and saw mills, as well as Indian paths, minerals and dwelling houses. Some 18 land "purchases" were made with the Indians between 1682 and 1792, and over 110 years they lost 45,000 square miles of land to the state of Pennsylvania.

In 1795 Lycoming County was established. Since 1772 this section of the state was part of Northumberland County which in turn had been part of Berks County since 1752. The first **Lycoming County map** was created in 1795 when Nippenose Valley was still part of Nippenose Township. There were 96 residents living in the township that year. The Great Island Indian Path ran the length of the valley from Elimsport Mountain along route 44 and along what is called Middle Road today, through McElhattan Gap and on to Great Island at present day Lock Haven.

A **land warrant map** for the county was created in 1795. The information for each 300-400 acre parcel includes the name of the warrantee, application number, survey date and warrant date. After the 1868 treaty

the land office opened for business on April 3, 1769. The Surveyor General was John Lukens and the Deputy Surveyor for the valley area was Charles Lukens. William Scull processed applications on the north side of the river. Such applications cost a dollar and the land about 22 cents per acre. Hawkins Boone was the principal guide here. This map shows 52 warrants for Limestone Township, 14 for Bastress and 18 for the "Upper End" which is now in Crawford Township, Clinton County. The valley was divided into about 84 warrants or sections. Most of these large warrants were later cut in half representing the farms of Nippenose Valley today that average 134 acres. 94% of these farms are family owned. The terms "Upper End" and "Lower End" were used early on presumably to indicate relevance to the Susquehanna River. The Upper End is upstream and the lower end downstream. The Lower End was predominantly Catholic and the Lower End Protestant. They did not get along well, and in 1870 there was a petition to divide the township which failed by only six votes.

From 1850 – 1880 large **county wall maps** were produced. H. F. Walling made such a map of Lycoming County in 1861 and Clinton County in 1862, followed by the 1873 Beach Nichols maps. Every property owner was named whether they were on large farms or were town residents; and grist mills, saw mills, roads, churches, hotels, stores, blacksmith shops, streams and sinks were identified. It is interesting to note the many changes that took place over these 12 years. Some 13 churches have been built in Nippenose Valley, the earliest in 1814, and as many schools. This count includes replacement schools with nine existing at one time.

Today there are beautiful **aerial and 3-D maps** available that show the land as never before seen. The valley floor looks like a patchwork quilt as a result of contour farming, interspersed with small villages, roads and streams, all surrounded by a double row of mountains. It would be interesting to bring some of the old settlers back to have a look and comment on 200 years of change.

FORGET THE MARCELLUS - THE UTICA SHALE PLAY IS THE NEXT BIG THING

John A. Harper

INTRODUCTION

There's a "new kid on the block" in the natural gas drilling craze that has hit the Appalachians, and its name is the Utica Shale play. Like the Marcellus Formation, the Utica Shale occurs throughout much of central Appalachian basin, extending across the Appalachian Plateau from New York and Quebec south to Tennessee, and covering approximately 28,500 square miles (mi^2 = 8,690 square kilometers [km^2]) (Figure 1). It typically is a massive, organic-rich, and thermally mature gray to black shale; however, in most places it is less organic-rich (has a lower total organic carbon content) than the Marcellus. It is also much deeper and covers a broader area than the Marcellus (Figure 1). Although some of the Utica Shale is organic-rich, much of it is more properly considered to be non-organic-rich gray shale and siltstone.
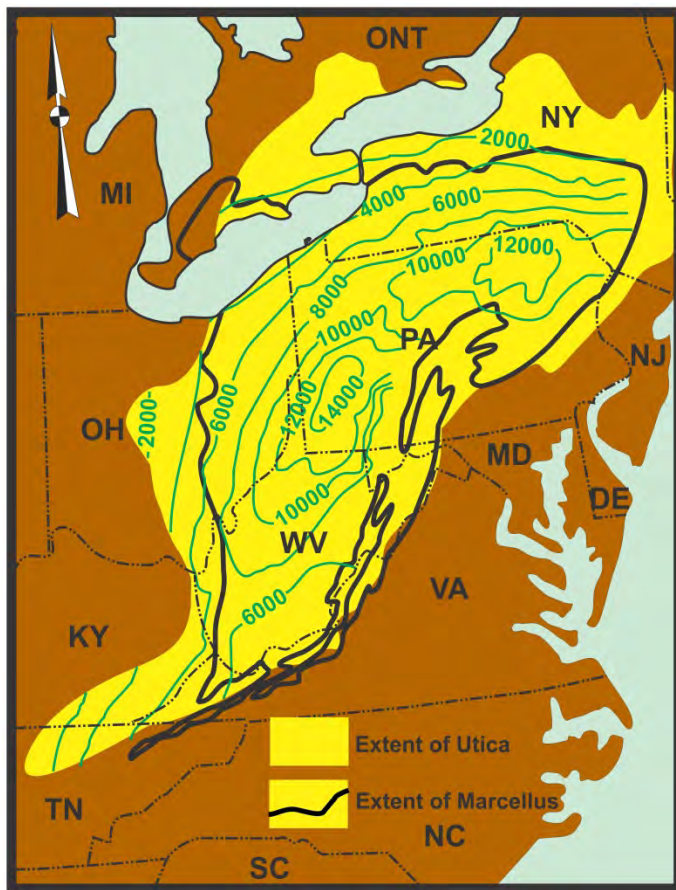

Figure 1. Comparison of the extent of the Utica Shale (yellow) and Marcellus Formation (black line) in eastern North America. The map also shows drilling depth (in feet) to the top of the Utica (green). Modified from Harper (2011).

Bergstrom and Mitchell (1990) called the Utica Shale one of the geographically most widely recognized lower Paleozoic stratigraphic units in the eastern US. It should be pointed out that the name "Utica" as used by drillers in the subsurface of the Appalachian basin is actually a facies name, rather than strictly a formation name. It includes the Point Pleasant Formation of Ohio and western Pennsylvania, the Flat Creek and Indian Castle formations of New York, the Antes Shale of central Pennsylvania, and the unnamed basal black shales and interbedded limestones of the Martinsburg Formation of south-central Pennsylvania, Maryland, Virginia, and West Virginia (Figure 2). The Point Pleasant, like the Flat Creek Shale of Brett and Baird (2002; also Smith and Leone, 2010) in New York, consists of Utica-type dark gray to black shales interbedded with Trenton-type limestones, and seems to be a transitional zone between the end members. Because the Point Pleasant, and perhaps the Flat Creek, are (or will be) important components of the Utica Shale play, Taury Smith, former State Geologist of New York, suggests that the play might be more appropriately called the "Utica Shale and associated organic-rich calcareous shale and interbedded limestone and shale play". In fact, the Utica Shale *per se* typically is not the best reservoir in this play. In Ohio and western Pennsylvania, the underlying Point Pleasant Formation is actually the main target of drillers.
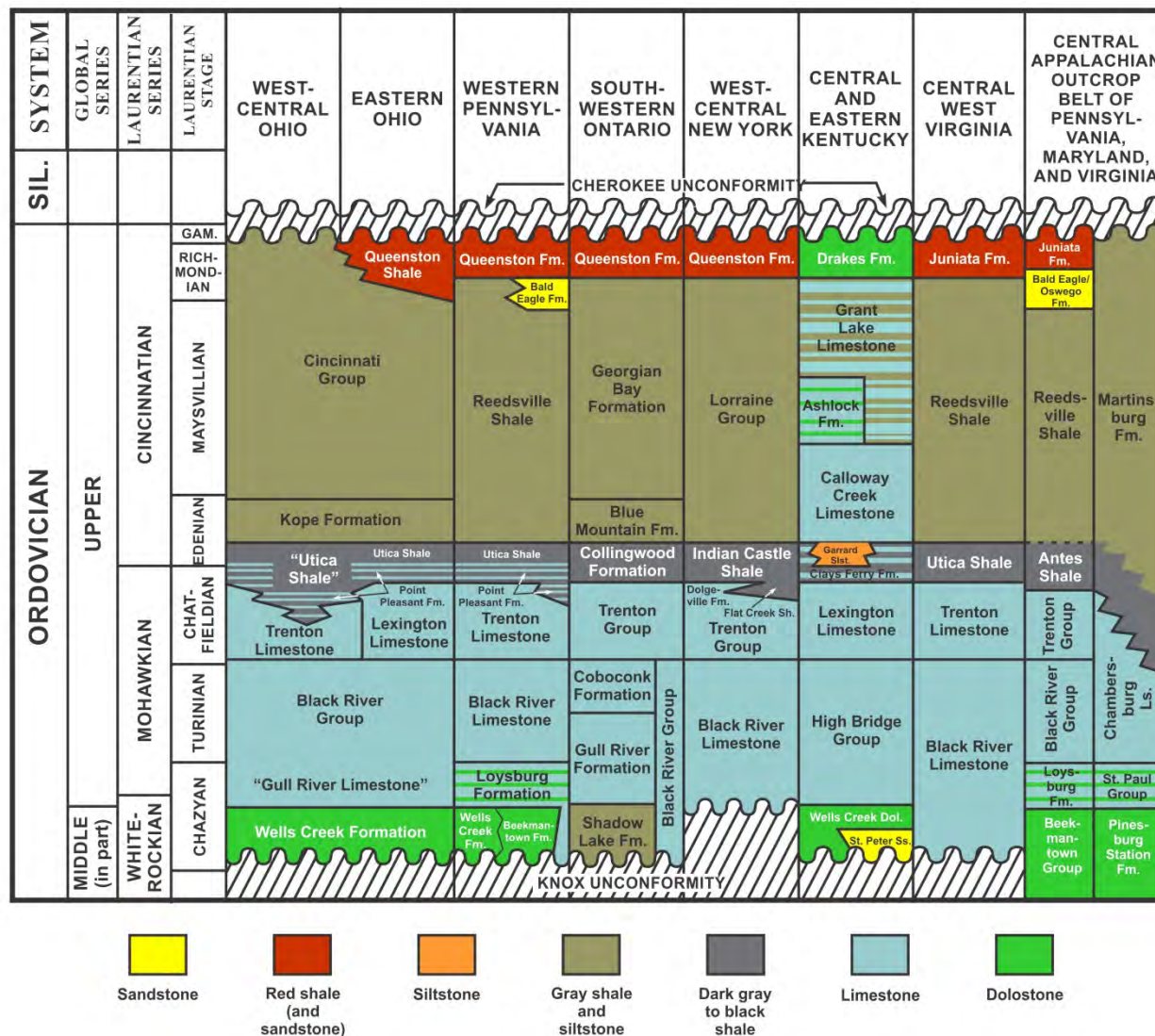
Figure 2. Correlation of Upper Ordovician strata in the Appalachian basin (modified from Harper, 2011).

STRATIGRAPHY

Classic New York Outcrop

Emmons (1842) named the Utica Shale for 75 ft (23 m) of black shale lying between the Lorraine shales (= Reedsville Shale of Pennsylvania) and the Trenton limestones exposed near the town of Utica, New York. Over the years, the term "Utica" came to mean different things to different geologists. Today, the name is used in New York alternately as a group name (i.e. Brett and Baird, 2002) and as a facies name (i.e. Baird and Brett, 2002; Lehmann and others, 1995, used the old Ken Caster term "magnafacies").

Classic stratigraphic studies of the Utica in New York by Prosser and Cummings (1897), Ruedemann (1908; 1912), Kay (1953), and others established that the "Utica" included all the black shale both above and time-correlative with the limestones of the Trenton Group in the Mohawk Valley. More recent work relying on K-bentonites, submarine discontinuities, fossils (especially graptolites), and other markers (Goldman and others, 1994; Mitchell and others, 1994; Baird and Brett, 2002) indicates that the Utica is a

time-transgressive facies in New York, older to the east, younger to the west (Figure 3). Lehmann and others (1995) recognized at least five intervals within the Utica facies that are bounded by unconformities and/or condensed beds, although some of the intervals probably do not belong to the Utica facies, strictly speaking.

At the base of the Utica, the Flat Creek Shale (or Member, depending on the author) lies disconformably on, and is laterally equivalent to carbonates of the middle Trenton (Denley Limestone of New York) (Figure 3). The Flat Creek is an organic-rich, calcareous shale that is time-equivalent to part of the upper Trenton Group and lies on a subaerial disconformity. The upper part of the Flat Creek grades laterally into the alternating thin-bedded, turbiditic limestones and shales of the Dolgeville Formation (Figure 3).

Smith & Leone (2009) divide the upper Utica facies, called the Indian Castle Formation (or Member) into a low-organic-carbon upper Indian Castle and a high-organic-carbon lower Indian Castle. Baird and Brett (2002, p. 213) described the lower Indian Castle beds as "hard, blocky to sheety black shale with bundles of tabular impure limestone beds." These rocks lie on the Dolgeville Formation. In the west, along the New York State Thruway, the Dolgeville and lower Indian Castle are distinctly separated by a regional disconformity called the Thruway disconformity (Figure 3). In western New York where the Utica facies is missing, this disconformity is probably a subaerial erosion surface. To the east, the Dolgeville exists as a distinctive, widespread interval that separates the Flat Creek and Indian Castle formations without apparent disconformity (Brett and Baird, 2002). The Upper Indian Castle beds are "monotonous, fissile black shale to silty black shale broken only by infrequent thin K-bentonites or by widely scattered concretionary limestone or calcareous siltstone beds." (Baird and Brett, 2002, p. 215-216).
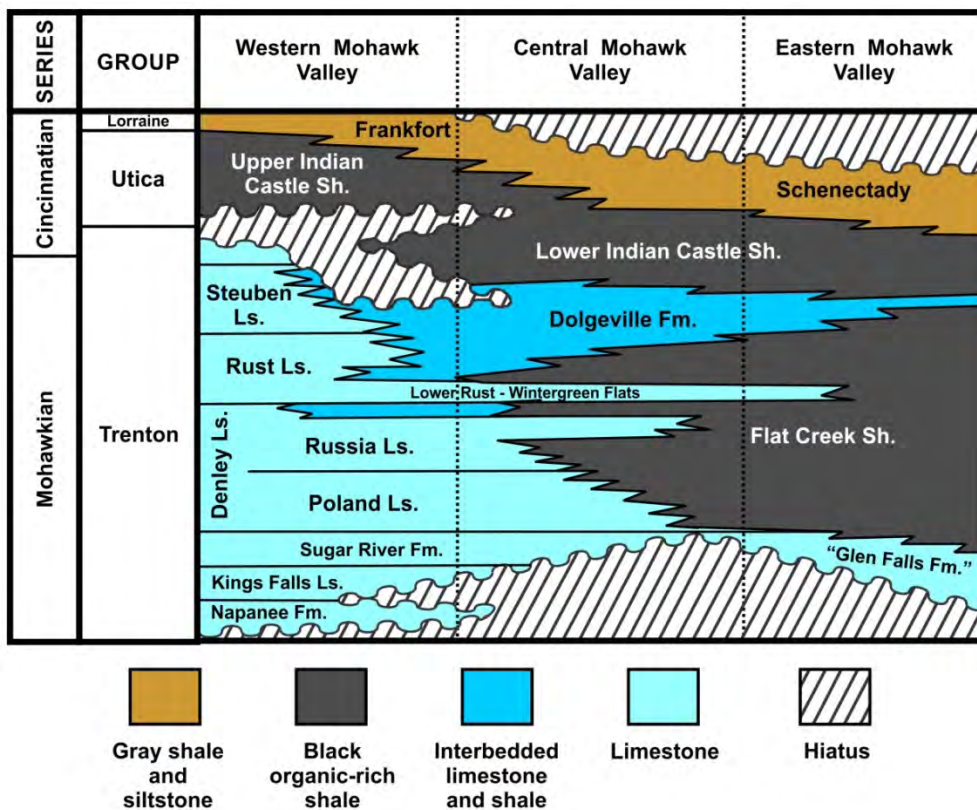
Figure 3. Stratigraphic diagram of Utica facies and associated formations in the Mohawk Valley of New York (based on Smith, 2011). The unconformity within the Indian Castle Formation in the western Mohawk Valley is referred to as the Thruway disconformity.

Subsurface of Ohio, Western Pennsylvania, and West Virginia

In Ohio, the strata representing the Utica facies range in age from late Middle Ordovician to middle Late Ordovician (late Chatfieldian to Edenian). Based on graptolites, Bergstrom and Mitchell (1990) suggest that

the Utica in north-central Ohio ranges from middle Edenian to middle-upper Maysvillian, with the lower contact being disconformable above the Trenton; this indicates that the lower Edenian and possibly the upper Mohawkian are probably missing. Throughout most of Ohio, however, the Utica is essentially early Edenian in age (Figure 2). This also appears to be the case in the plateau in Pennsylvania and West Virginia. It is probable that the formation called "Utica" in the subsurface of Ohio, Pennsylvania, and West Virginia is correlative with the upper Indian Castle Shale of New York.

Also present throughout most of the plateau of Pennsylvania and West Virginia is a formation beneath the Utica that, until recently, had been considered merely part of the Utica Shale that graded downward into the Trenton limestones. This is the Point Pleasant Formation of Ohio, recognized in the subsurface of Pennsylvania during the Trenton/Black River project (Patchen and others, 2006). The Point Pleasant Formation consists of interbedded argillaceous limestone (i.e., Trenton-type carbonates) and dark gray to brown to black calcareous shales (i.e., Utica facies). The number of limestone beds decreases upward. In the subsurface, the boundary typically is considered the last limestone bed recognizable on wireline logs. The Point Pleasant is considered to be stratigraphically equivalent to the upper Trenton in Ohio (Wickstrom and others, 1992), as it is in Pennsylvania. Based on its position within the Upper Ordovician, its composition, and its higher organic content, the Point Pleasant is most likely equivalent to the lower Indian Castle Shale of New York (Figure 4).
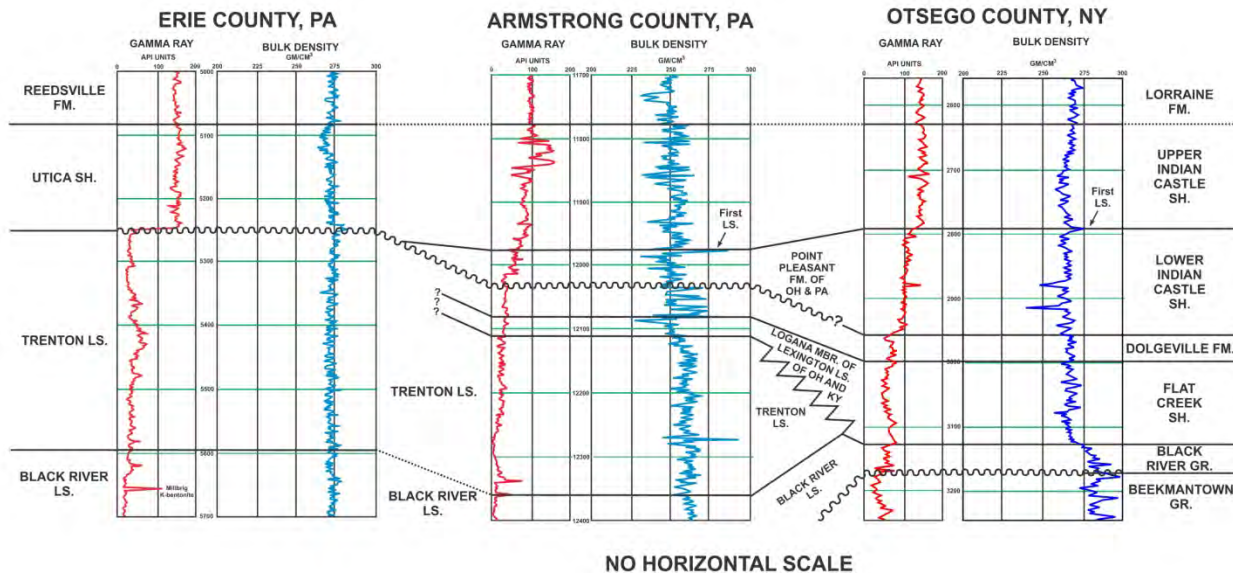
Figure 4. Generalized correlation of Utica facies in Pennsylvania and New York using geophysical logs (gamma ray and density) (modified from Harper, 2011). The typical Utica Shale (= upper Indian Castle Shale of New York) lies directly on Trenton limestones in Erie County (log A), suggesting the presence of the Thruway disconformity known from outcrop in western New York. Proceeding into the basin, log B indicates the presence of a thin section of Point Pleasant Formation, suggesting the Thruway disconformity continues at least as far southeast as Armstrong County. The presence of shale in the Trenton in log B suggests correlation with the Logana Member of the Lexington Limestone, which is considered to be part of the Upper Ordovician source rock sequence in Kentucky and Ohio. This, in turn, suggests that the Logana is a distal correlative of the Flat Creek Shale of New York.
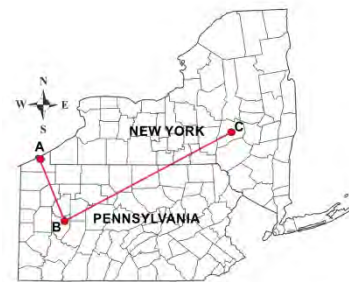
PALEOGEOGRAPHY, PALEOOCEANOGRAPHY,
AND DEPOSITIONAL ENVIRONMENT

During Late Ordovician time, what is now the Appalachian basin comprised a small portion of the Laurentian craton, lying about 25° south of the equator (Figure 5). From Late Precambrian through Middle Ordovician time, this region was part of an eastward-thickening, miogeoclinal basin that accommodated mostly carbonate platform sediments (Thompson, 1999), part of the "Great American Carbonate Bank" (GACB) that extended more than 1,865 mi (3,000 km) along nearly the entire length of what was the southern seaboard of the Laurentian continental mass. The platform prograded eastward through Cambrian and Early Ordovician time, and its eastern terminus seaward was at a continental slope beyond which lay deep ocean basin sediments. Beginning in the Middle Ordovician, the craton margin was uplifted during the Taconic orogeny (Faill, 1999). The platform was progressively submerged. The basin changed from a miogeoclinal carbonate platform to an exogeosynclinal foreland molasse basin (Thompson, 1999). This basin received large amounts of terrigenous sediment from eastern highlands, which were deposited in the foreland basin as the Taconian clastic wedge.
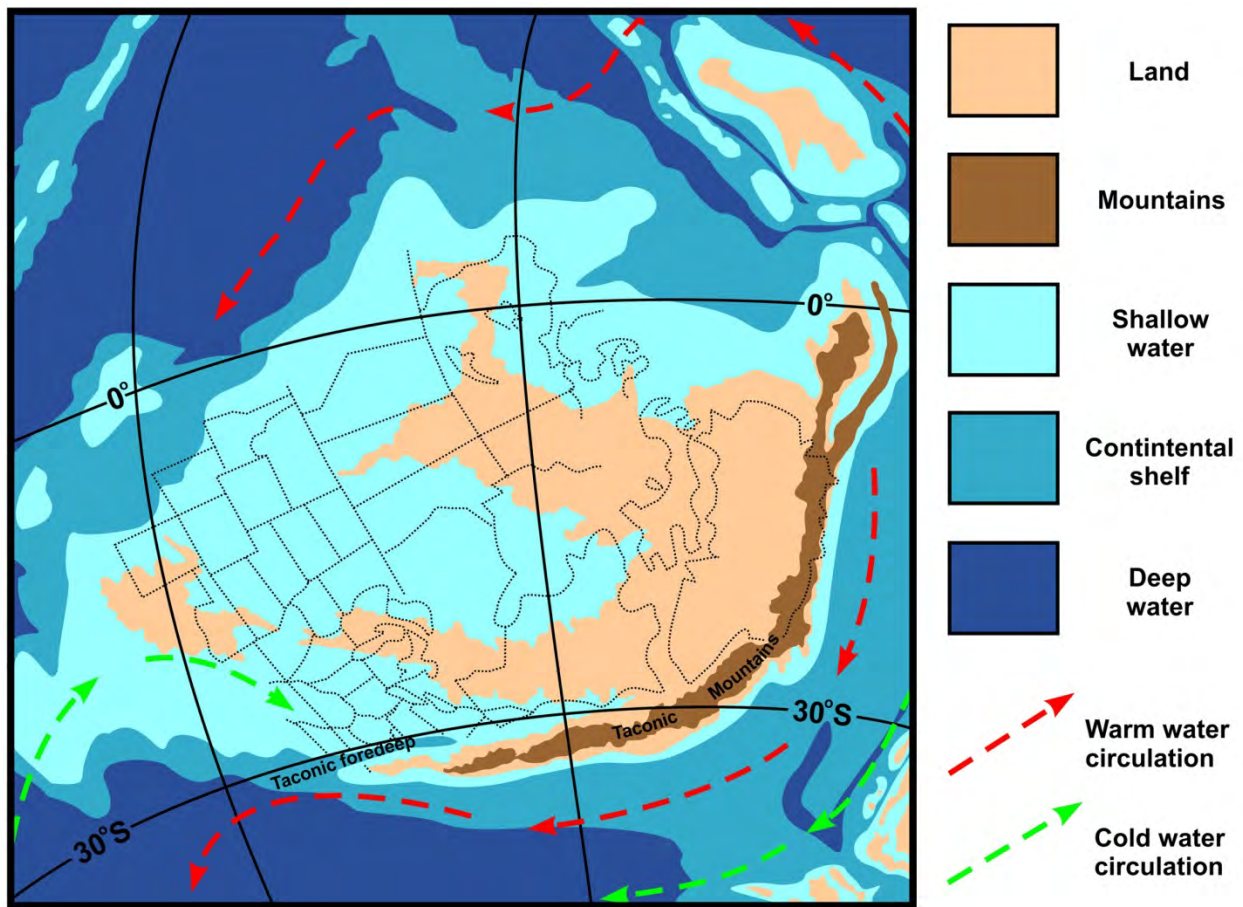
Figure 5. Paleogeographic reconstruction of Laurentia during the Late Ordovician (redrawn from Blakey, 2011).

The distribution of landmasses and oceans suggested for the late Middle and early Late Ordovician (Figure 5) should have affected global seasonal weather patterns essentially the same as we experience today (Wilde, 1991), but with the northern and southern hemispheres reversed. Summer months feature low-pressure air masses developing over large landmasses, resulting in the flow of large amounts of moist air (monsoons). Because landmasses were essentially restricted to the southern hemisphere at this time, monsoons would have occurred only in that hemisphere. Anti-trade monsoonal winds generated along the

coast of Gondwana drove warmer water northward, countering the flow of cooler water from northern midlatitude high-pressure systems. Because Pennsylvania lay on the southwest-facing coast of Laurentia around 25° south latitude, the dominant winds would have been trades similar to Recent trade winds, with some variation due to continental configurations. Warm maritime air would have provided abundant rainfall from the equator to about 35° or 40° south latitude.

The Ordovician stratigraphic record in Pennsylvania reflects the control of tectonics and paleogeography on sedimentation. Lower through Upper Ordovician carbonates indicate prolonged passive-margin carbonate accumulation in semi-arid conditions on the GACB, transitioning to foreland deposition and transitional climates as the Taconic orogeny created flexure and increased downwarping. Subsidence rates increased markedly beginning ca. 460 ma, marking the termination of the GACB (Goldhammer and others, 1987). The appearance of dark brown and black shales marked a significant change in paleogeography during Late Ordovician time (Kolata and others, 2001) as clastic muds began filling the Taconic basin from the island arcs and mountains in the east. These Late Ordovician clastics reveal a history of basin filling by marine and terrestrial clastics on a continental margin subjected to moist air traveling westward in lower latitude.

Patchen and others (2006) referred to a Utica/Point Pleasant sub-basin ("Point Pleasant" basin of Wickstrom and others, 1992) within the carbonate platform area of western Taconic basin that included the narrow Sebree trough (Figure 6). The Sebree trough was a northeast-southwest trending feature from western Kentucky through southeastern Indiana, and into northwestern Ohio, and may have extended across western Pennsylvania. During deposition of the Utica Shale, the intensity of the Taconic orogeny once again increased causing a rapid rise in sea level or increased subsidence of the region, which resulted in the Utica Shale replacing carbonate deposition on the platforms. Consequently, relatively thick deposition of the Point Pleasant and Utica occurred in both the Utica/Point Pleasant sub-basin and the central Appalachian basin area, followed by eventual drowning of the carbonate platforms.
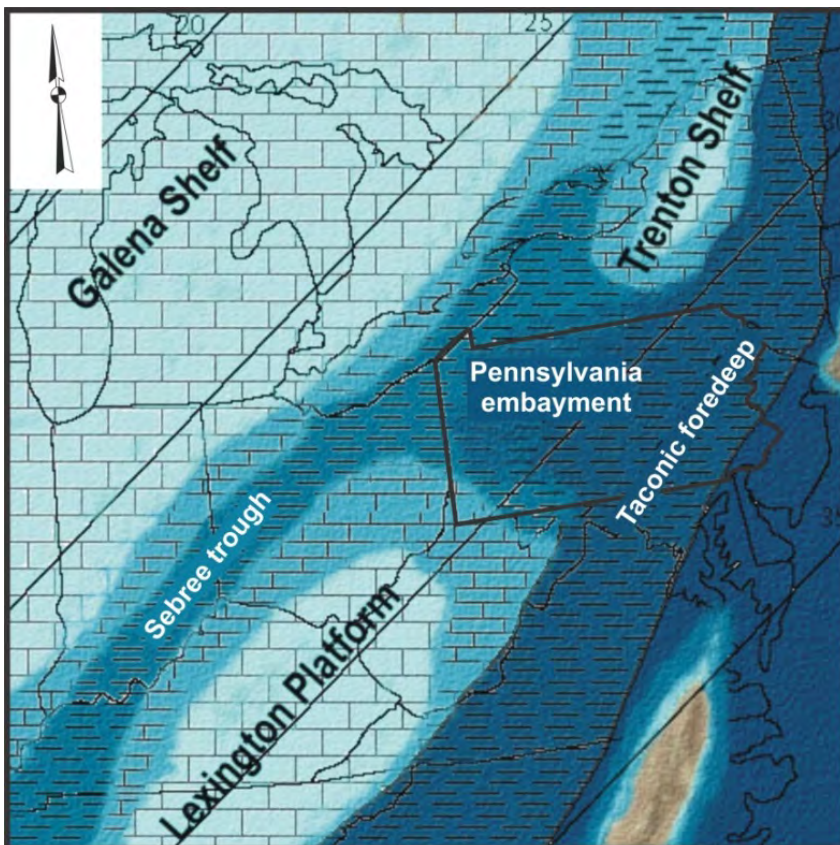
Figure 6. Paleogeographic reconstruction of the eastern United States during the early Late Ordovician (modified from Brett and Caudill, 2004).

Traditionally, the depositional environment for organic-rich black muds has been interpreted as deep, anoxic water at the bottom of an enclosed basin. For example, Lehmann and others (1995) suggested that the shift from limestone deposition on carbonate platforms changed abruptly to deposition of deeper-water organic-rich Utica muds as a result of faulting. Contrary to the deep-water concept, however, Utica mudrocks probably were deposited in relatively shallow water, forming mainly on the western cratonward side of the Appalachian foreland basin where they onlapped unconformities in relatively shallow water, rather than in the deepest part of the basin (Figure 7). Relative

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water depth estimates based on sedimentary structures such as storm beds, grainstones, mineralized discontinuity surfaces, and trace fossil assemblages suggest depths of only 30-160 ft (10-50 m) of water (Jones and others 2011). This concept emphasizes the existence of the Thruway Disconformity, which separates New York's Indian Castle Shale above from the Trenton limestones (Dolgeville Formation) below. This equates to the disconformity that separates the Utica Shale from the Trenton limestones in northwestern Pennsylvania (Figure 4). In addition, an older unconformity separates the basal Utica facies (= Flat Creek Shale) from the underlying Trenton in New York. I am unaware of a specific unconformity separating the lower Antes in central Pennsylvania, and lower Martinsburg in southcentral Pennsylvania, from the lower Trenton/Black River/Chambersburg limestones, but it might exist. It is unlikely that the Ordovician carbonates graded uninterruptedly into the Utica facies.

Jones and others (2011) suggested that distinguishing the various alternative depositional models using local geological data will require careful analysis to discriminate regional and local effects on lithology and accommodation space.
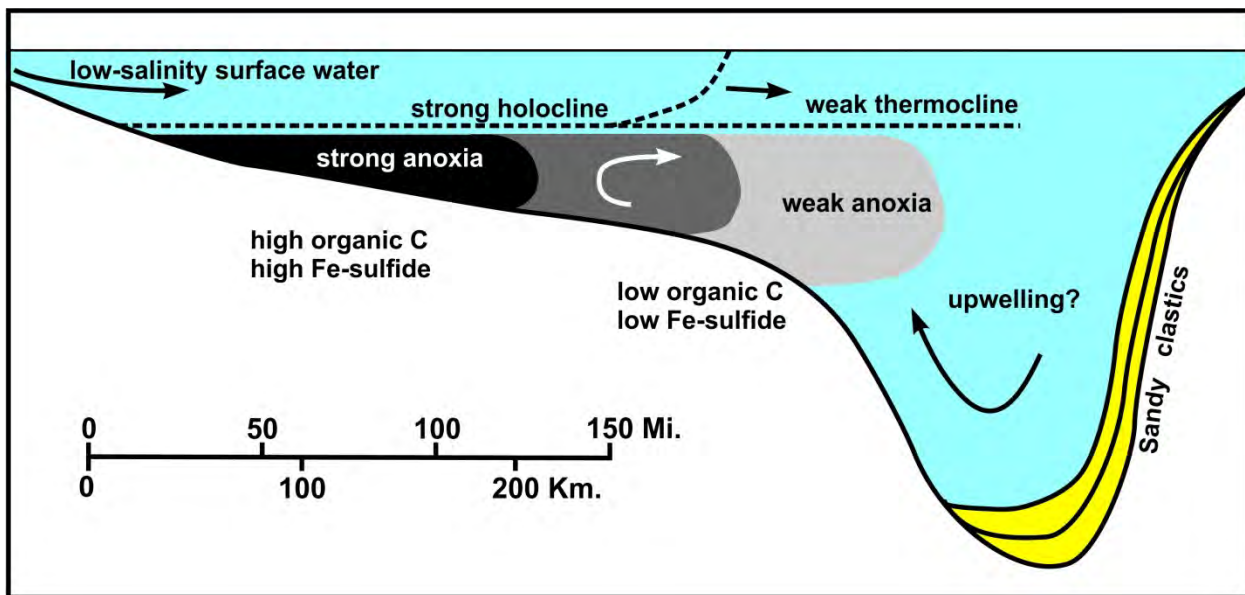
Figure 7. Shallow-water model for deposition of Utica facies rocks (based on Algeo and Wilkinson, 1989, and Smith and Leone, 2010).

STRUCTURE AND TECTONICS

Figure 8 illustrates the geologic structure on top of the Trenton limestones in the Appalachian basin. Because the Utica facies in the form of the Flat Creek, Point Pleasant, lower Indian Castle, or Utica (as used outside of New York) sits directly on Trenton limestone throughout the region, this map equates to a structure map on the base of the Utica facies. It is likely that certain structures not shown, such as the Tyrone-Mt. Union lineament, are present, but may have had little or no impact on the shales. Of course, lack of detailed stratigraphic data throughout most of western Pennsylvania could also account for an apparent lack of structural impact.
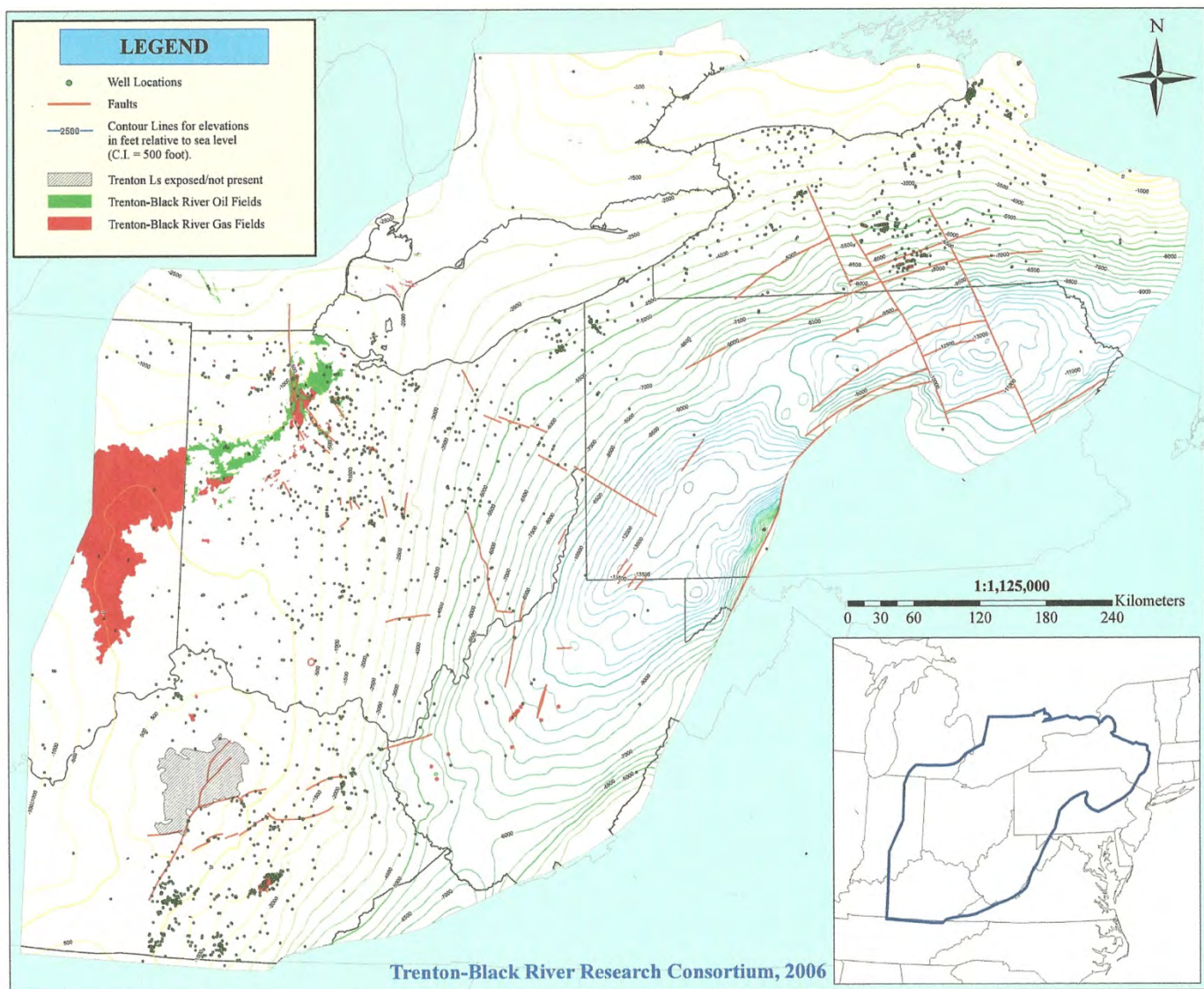
Figure 8. Structure map on top of the Utica Shale in the Appalachian basin (from Patchen and others, 2006).

The Taconic orogeny was a complex series of orogenic episodes spread over the larger part of the Ordovician. Originally considered to have been a single event, the orogeny is now known to consist of at least three major episodes. The first episode occurred during the Early Ordovician in northern New England and the Canadian Maritimes. The second occurred during the Middle Ordovician with eastern Tennessee as the center of the episode. The third occurred during the Late Ordovician with the largest amount of disturbance in the central Appalachians (from Virginia to New York). Flexure of Laurentia during the orogeny created some deep sedimentary basins that accumulated as much as 1,000 ft (300 m) of sediment in some areas, resulting in accumulations such as the Queenston Delta formations (Reedsville/Martinsburg through Queenston/Juniata – see Figure 2).

Disconformities disturbed what otherwise would have been continuous carbonate accumulation, and widespread deformation occurred. The Utica facies units, bounded as they are by disconformities, implies that each resulted when a pulse of tectonic subsidence occurred in the foreland basin, followed by a pulse of siliciclastic sedimentation (Lehmann and others, 1995). Lehmann and others (1995) also documented a shift in the basin axis throughout the late Middle Ordovician and the early Late Ordovician. They found the basin axis migrated westward over 60 mi (100 km), resulting from deformational loading of the continental

margin and progressive foreland flexure, with smaller-scale structural elements, and normal fault-bounded basement blocks, superposed on the large-scale geometry of the Taconic foreland-basin.

MINERAL COMPOSITION

Utica/Point Pleasant mineralogy in wells from New York, Ohio, Quebec, and Pennsylvania is similar, consisting primarily of clay minerals (mostly illite and chlorite), quartz, feldspars (K-feldspars and plagioclase), and carbonates (both calcite and dolomite). Minor, but significant, minerals include pyrite and gypsum. The ternary diagram in Figure 9 shows the distribution of these major components in the Trenton through Reedsville equivalent rocks in Pennsylvania and New York, as well as in the Antes and Reedsville in outcrop from central Pennsylvania. It is to be expected that the Reedsville and equivalent Lorraine are mostly quartz/feldspar- and clay-rich, but the Trenton carbonates actually have a surprisingly high percentage of clay and quartz. The Utica facies rocks are highly variable. Lower Indian Castle (= Point Pleasant) and Flat Creek (= lower Antes and Martinsburg) shales both have fairly high percentages of carbonates, whereas the typical subsurface Utica (= upper Indian Castle) has relatively little carbonate content.

Surprisingly little is known about the distribution of minerals in the organic-rich beds of the Utica facies, outside of clay analyses performed in the 1960s and 1970s (O'Neill and others, 1965; O'Neill and Barnes, 1979). Current study of the Utica by a consortium of state geological surveys of New York, Pennsylvania, Ohio, West Virginia, and Kentucky hopefully will provide more information on this and related topics.
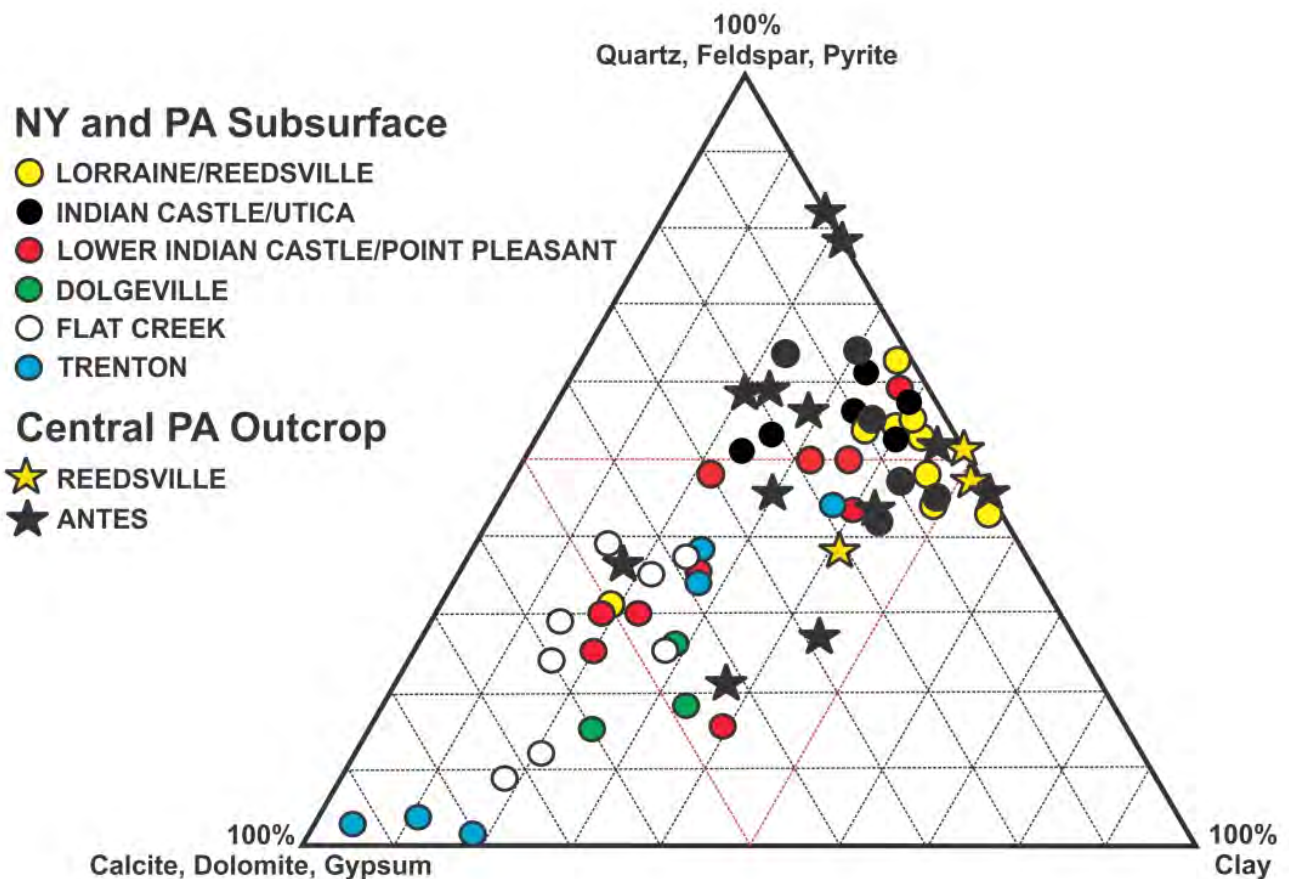

Figure 9. Ternary diagram of samples of Upper Ordovician rocks from New York and Pennsylvania. Data from Nyahay (2011) and files of the Pennsylvania Geological Survey.

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ORGANIC GEOCHEMISTRY

Total Organic Carbon

Total organic carbon (TOC) in the Utica facies generally is low (typically less than 1.8 wt. %) and exhibits geographic variation with ranges typically between 0.0 and 4.3 wt. %, and averages 1.3 wt. % (Laughrey and Baldassare, 1998; Peters and others, 2005). However, Jones and others (2011) suggest locally enhanced TOC values up to 13 wt. % are present in the Flat Creek Shale in New York.

Non-proprietary measurements of TOC in the Utica Shale are sparse. Where they have been reported, however, TOC values over 3 wt. % are common, at least in the lower beds of the Utica facies that lie on disconformities (Smith, 2011). Nyahay and others (2007) and Smith (2011) reported TOC values for core and outcrop samples of the Flat Creek, Dolgeville, and lower Indian Castle in New York generally range from 0.5 to 3. TOC values in the upper Indian Castle generally fall below 0.5 wt. %. TOC values as high as 3.0 wt. % have been reported in eastern New York and 15 wt. % in Ontario and Quebec.

TOC values in Ohio are available as a map (Ohio Geological Survey, 2013) indicating the maximum TOC per well for the total Ordovician organic-rich interval, which includes Utica, Point Pleasant, Lexington, and Logana. (The Lexington and Logana intervals correlate with the Dolgeville Formation and Flat Creek Shale, respectively, of New York.) Because the values are combined for all formations, it is difficult to compare these data with those of the other states. Values shown on the map are, basically, "all over the map" so to speak. The lowest value appears to be 0.6 wt. % for a well in Champaign County in the east-central part of the state, whereas the highest value appears to be 7.6 wt. % for a well in Marion County a few counties to the northeast. In general, the highest maximum TOC values occur in eastern Ohio, within four counties of the Pennsylvania-Ohio border.

Based on data available from the Pennsylvania Geological Survey (Laughrey and others, 2009), Utica TOC values range from 0.17 wt. % and 2.87 wt. %, averaging 1.47 wt. %. Point Pleasant TOC values range from 0.8 wt. % to 3.74 and average 2.25 wt. %. As in New York, the upper Utica facies (upper Indian Castle Shale in New York = Utica Shale in Pennsylvania and Ohio) has a lower TOC value than the lower facies (lower Indian Castle Shale in New York = Point Pleasant Formation in Pennsylvania and Ohio). Industry is aware of this also – most of the wells permitted for the "Utica" drilling in Pennsylvania identify the Point Pleasant as the target formation.

It is important to note, however, that Jarvie and others (2005) found that TOC values measured from both cuttings and conventional core analysis for the same depth interval in a single well were very different, with core samples yielding 2.36 times higher values than cuttings. Since all of the Utica and Point Pleasant TOC values in the Commonwealth were derived from cuttings, the averages for these two formations in Pennsylvania actually could be much higher than reported.

Thermal Maturation

Hydrocarbon production from shales, which are going to have very low permeabilities, requires at least 2 wt. % TOC to begin with, then maturation through burial and catagenesis to crack hydrocarbons from the organic matter. With increasing thermal maturation, the shale will lose carbon and hydrogen due to hydrocarbon generation, so that increasing thermal maturity will decrease the TOC and hydrogen values in the rock. Therefore, even dismissing the conclusions of Jarvie and others (2005) as discussed above, measured TOC values may not accurately reflect the hydrocarbon potential of a rock.

Figure 10 is a thermal maturity map constructed for the Trenton play book (Patchen and others, 2006). It shows that the Utica facies rocks throughout most of Pennsylvania and eastern New York are overmature;

that is, they have very little or no potential for hydrocarbon generation and production is unlikely. Western Pennsylvania and New York, southeastern Ohio, and northern West Virginia are in the mature (dry gas) window, and indeed, western Pennsylvania has been the target of drilling, especially in Beaver, Butler, Lawrence, and Mercer counties. Much of eastern Ohio and down into eastern Kentucky is in the wet gas and oil generating windows. Eastern Ohio is actually the main target for Utica/Point Pleasant drilling given that liquid hydrocarbons are selling at higher prices than typical natural gas (mostly methane).
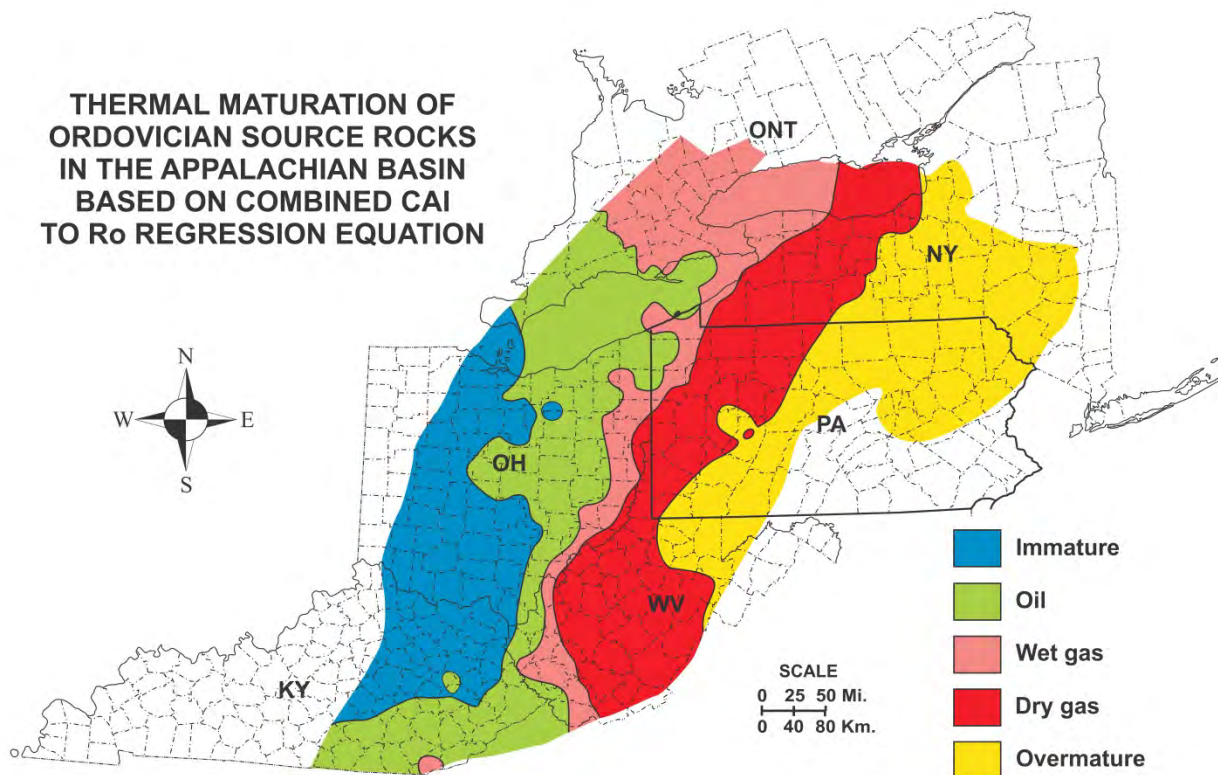
Figure 10. Map of Appalachian states showing thermal maturation of Upper Ordovician source rocks (Utica facies). Redrawn from Patchen and others (2006).

HYDROCARBON POTENTIAL

Historically, the Utica facies has been considered the source rock for more conventional gas reservoirs such as the Lower Silurian Medina Group sandstones and deeper reservoirs of eastern Ohio and northwestern Pennsylvania (Laughrey and Baldassare, 1998; Ryder and others, 1998). For example, Ryder and others (1998) determined that crude oils from the Upper Cambrian and Lower Ordovician Knox Dolomite in central and eastern Ohio were derived from "Utica and Antes shales" (Ryder and others, 1998, p. 433-434). Fresh samples containing residual kerogen and other petroleum residuals reportedly have been ignited and can produce an oily sheen when placed in water. In fact, as far back as the Second Geological Survey of Pennsylvania, Lesley (1892, p. 541) reported that, "The black *Utica slate*, and many darker layers of the Hudson River slate [Reedsville and equivalent Martinsburg] . . . have been so heavily charged with carbon from the decayed bodies of the creatures which filled the sea, that hand specimens will smoke and flame in a blacksmith's fire. This has given them the mineralogical name of *fire slate* (pyroschists)." (Italic in the original) Lesley went on to describe attempts to distill oil from the Utica that resulted in 3 to 5 percent oily and tarry matter with combustible gases and water.

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As with the Marcellus shale, significant gas shows historically were reported in places when drilling through the Utica facies while exploring for hydrocarbons in older carbonate rocks. Yet, as with the Marcellus, no one bothered to test the Utica for production because, after all, it is merely impermeable shale. All of the gas was coming from fractures in the rock. But with the advent of drilling horizontally with long laterals, and the ability to perform multi-stage hydraulic fracturing within those laterals, industry can now produce from these impermeable organic-rich shales where, perhaps, as much as 90 percent of the original hydrocarbon reserves can still be accessed. Test wells completed in the Utica in Quebec that have produced up to one million cubic feet of natural gas per day (1.0 MMcfgpd). Wells drilled to the Utica and Point Pleasant in Pennsylvania (Figure 11) have been reported with successful after-treatment flows of between 2.7 and 10.13 MMcfgpd. Even in Tioga County, where the Utica is considered to be overmature, there has been production from the Utica (commingled with Trenton, so the exact amount coming from the Utica is unknown). This seems to indicate that, although the shale is spent in that area, it is still a viable reservoir.
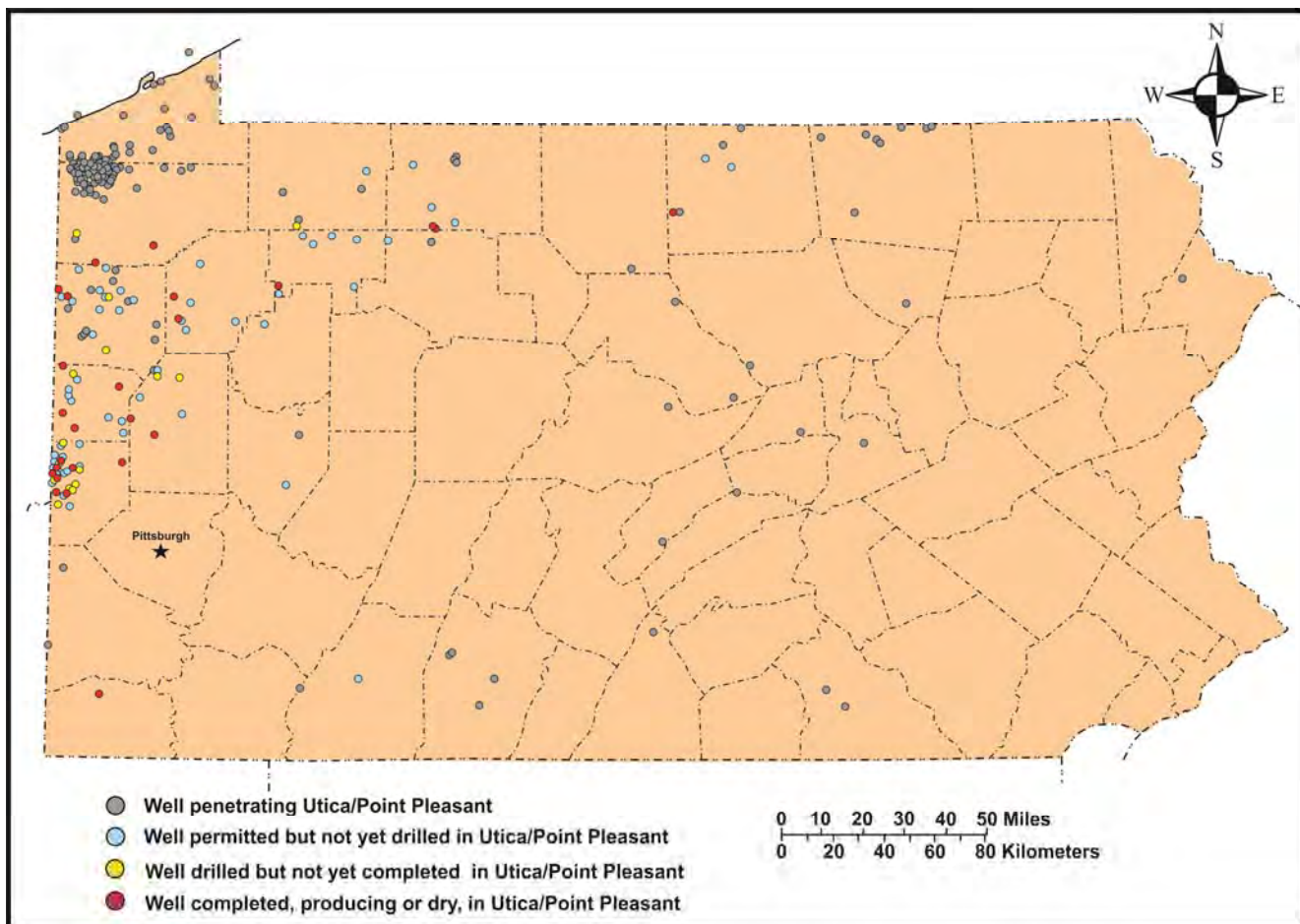
Figure 11. Map of Pennsylvania showing wells that have been permitted, drilled, and/or completed in the Utica Shale and Point Pleasant Formation, as well as non-Utica/Point Pleasant wells that have penetrated those formations. Data, as of July 1, 2013, from the Pennsylvania Geological Survey.

Ohio, falling within the wet gas and oil windows (Figure 10) has been benefiting from Utica and Point Pleasant drilling the way Pennsylvania has been benefiting from Marcellus drilling. Initial productions have been reported between 1.5 MMcfgpd and 9.5 MMcfgpd with 1,425 barrels of oil per day (BOPD) (Wickstrom and others, 2012). Wickstrom and others (2012) estimate that the Utica and Point Pleasant in Ohio could hold as much as 3.75 to 15.7 trillion cubic feet of gas and 1.31 to 5.5 billion barrels of oil!

INTO THE FUTURE

The Utica and Point Pleasant constitute a still-emerging play. There is so much more to be learned; comparatively little is actually known about the play in Pennsylvania because it is only a few years old. In fact, much more needs to be done across the entire basin before anyone will have a strong grasp of the true potential for the Utica facies in the central Appalachian basin.

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THE STRUCTURE OF THE TRANSITION ZONE BETWEEN THE VALLEY-AND-RIDGE AND PLATEAU AROUND THE GREAT BEND IN CENTRAL PENNSYLVANIA

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INTRODUCTION

The Nittany Anticline forms the outermost portion of the Valley-and-Ridge province in central Pennsylvania. The curvature of the anticline and the associated structures, both within the anticline and in the adjoining transition to the plateau have been recognized as one of the great mysteries of central Appalachian structure for many years. Recent mapping in the eastern Nittany Valley (Washington, 2009) and adjoining areas (Washington, 2013) indicates that this curvature represents the interference of four separate deformational events, representing four distinct phases of the Alleghanian orogeny. In addition, the saddle separating the Nittany and Nippenose Valleys can be explained by a pre-Alleghanian deformation that affected at least the eastern Nittany Valley (Washington, 2009).

The topographic expression of the Nittany anticline extends from the West Branch of the Susquehanna River by Muncy on the east to Hollidaysburg at the southwest, though the fold structure can be traced further on both ends. The north/northwest flank of the anticline is marked by the Bald Eagle Mountain ridge (called Brushy Mountain south of the Litte Juniata River) held up by the Late Ordovician to Early Silurian Bald Eagle-Juniata-Tuscarora sequence of sandstones, siltstones, and shales of the Queenstown clastic wedge; much of this ridge is a topographically double ridge because the Juniata Formation is predominantly shale in this area and therefore less competent than the Bald Eagle and Tuscarora sandstones. Within the Nittany, Nippenose, and Mosquito Valleys, the Late Ordovician to early Silurian sandstones have been breached to expose Cambro-Ordovician carbonate shelf strata overlain by Late Ordovician shale (Reedsville and Antes shales, a.k.a. Utica shale). The deformation of the carbonate sequence within the Nittany Valley is quite complex, but is relatively simple within the Nippenose and Mosquito Valleys; therefore, much of the discussion will rely on information garnered from the Nittany Valley.

The area to the north/northwest of the Bald Eagle ridge and south/southeast of the nearly flat-lying Mississippian sandstones that form the lip of the Allegheny plateau is generally viewed as a transition zone even though it is locally structurally continuous with one or the other structural region.   The transition zone can be subdivided into three regions based on deformational patterns: an eastern region, roughly from the village of Woolrich eastward; a central region, roughly from Woolrich to the village of Bald Eagle (just north of Tyrone); and a southwestern region, roughly southwest of Bald Eagle. The field trip will visit representative central region structure at Lockport across the river from Lock Haven and representative eastern region structure along the Pine Creek Rail Trail north of Jersey Shore.

The eastern region is marked by high-amplitude plateau-style deformation, including folding and thrust faults verging both north (forethrusts) and south (back thrusts). The central region is marked by a nearly unbroken monocline connecting the Bald Eagle ridge to the plateau. Except for a couple of forethrusts near the front of the Bald Eagle ridge, back thrusting is dominant within this monocline wherever displacements can be observed, and the transition is disconnected from the Nittany anticline by major backthrusting. The southwestern region is much more like a normal fold-and-thrust belt with forethrusting dominant; we will not visit this region during this field trip.
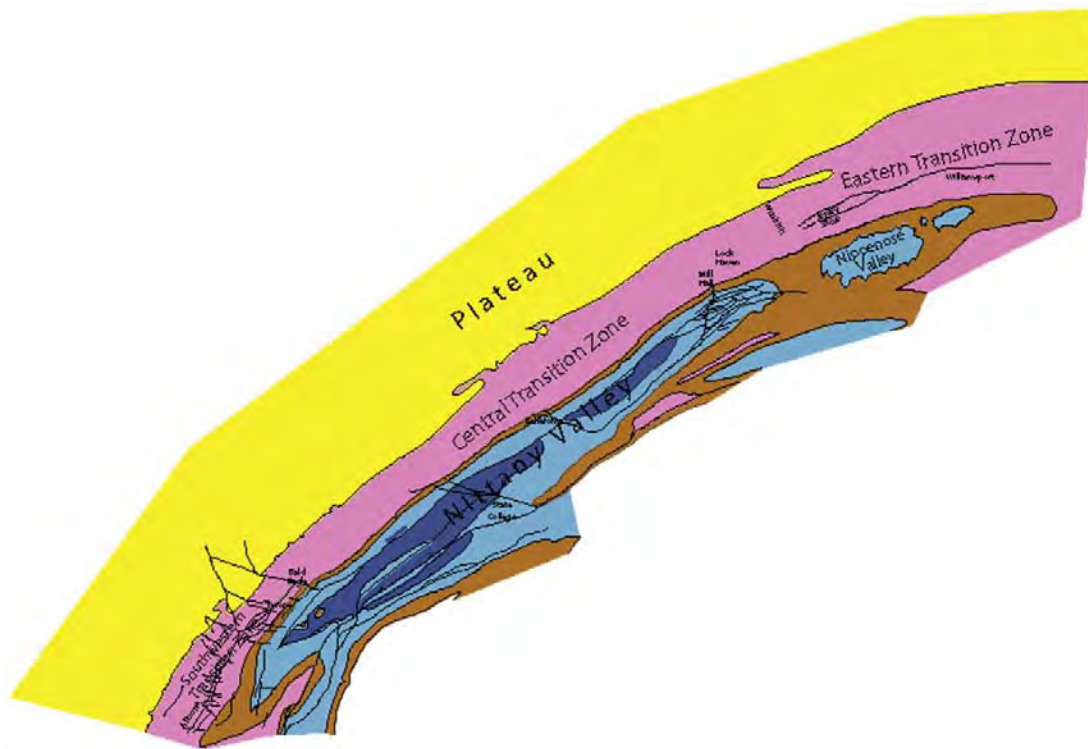
Figure 1. Major structural elements of the outer edge of the Valley-and-Ridge around the bend in central Pennsylvania.

DEFORMATION WITHIN THE ANTICINAL CORE

Deformation within the core of the Nittany anticline is mostly derived from outcrop studies because there have only been two exploration wells drilled within the anticlinal core – the California Co.'s W.E. Snyder #1 (T.D. 5808 ft.) at Oriole in the Nippenose Valley in 1951 (Fettke, 1952, table 1, column 78), and SOCONY Mobil Oil Co.'s J. Franklin Long #1 (T.D. 15,663 ft.) 2.4 km east of Jacksonville in the eastern Nittany Valley in 1963 (Lytle et al., 1965, table 12, column 15). Although seismic data exists, the high dips (locally near vertical to overturned) makes interpretation difficult.

Eastern End (Nippenose & Mosquito Valleys area)

The Nittany anticline is nicely expressed toward its east end in the Nippenose and Mosquito Valleys and along the Hagerman's Run valley, with a steeply dipping northern limb, a relatively flat crest, and a less steeply dipping southern limb (Faill et al., 1977). Nevertheless, two different directions of shortening are recorded by cleavage in this area: a first direction of nearly due north (present coordinates) and a second direction of 20° to 30° west of north. Although the primary anticlinal trend (except at its extreme eastern end) reflects the later direction, the primary cleavage orientation within the eastern Nippenose Valley and the eastern end of the anticline strikes nearly due east reflecting the earlier direction; the later cleavage is weak to non-existent in most outcrops. Cleavage in this area is generally expressed as a sub-centimeter pencilling of the shales and a multi-centimeter-spaced cleavage in the sandstones and is mostly found in the northern limb; cleavage is rarely evident in the carbonates of the eastern end of the anticline. Although both cleavages are rotated by the folding, the stronger E-W-striking cleavage is oblique to the local fold hinge whereas the weaker N65E cleavage is roughly parallel to the local fold hinge. This suggests that the weaker cleavage is formed earlier than the folds but during the same shortening event, whereas the stronger cleavage was formed during an earlier non-coaxial shortening event.

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The W.E. Snyder #1 well drilled at the village of Oriole in the center of the Nippenose Valley (Fettke, 1952) indicates that the entire Cambro-Ordovician carbonate sequence was involved in the thrusting and that the Nittany anticline in this area is a relatively simple ramp anticline. The thrust sheet apparently climbs onto the Late Silurian Salina detachment surface beneath the anticline, so all of the strata exposed within the anticline are passively folded across the ramp anticline. However, the surface structure suggests that the apparently flat-lying footwall is part of a fault-propagation fold, giving rise to the backthrusting that decapitates the anticlinal structure.

The divergence in orientation of the cleavage and the anticlinal trend is explained by invoking plateau-style deformation in the area of the Nittany anticline during the first stage of Alleghanian deformation, with the primary thrusting occurring during the second stage. Cleavage in a plateau area occurs in areas of concentrated shortening, primarily above low-displacement ramps, so the occurrence in the northern limb of penciling and spaced cleavage related to the early deformation suggests that there was minor displacement on the ramp at that time. The change in anticlinal orientation south of Williamsport probably indicates that the initial ramp was only developed east of the Williamsport area, so the later displacement direction controlled the ramp orientation to the west of that bend. This is further supported by the change in orientation of the pencilling in the western Nippenose Valley to that of the second event. Due to the poor exposure of the shale, it is impossible to define the point of change, but there is a week overprinting in the eastern Nippenose Valley.

Eastern Nittany Valley

In the eastern Nittany Valley, the picture becomes considerably more complicated. Although the anticlinal structure appears to continue from Nippenose Valley to the east, detailed mapping (Washington, 2009) shows that the carbonates are highly deformed and the entire anticlinal crest is detached by a major back thrust. The earliest deformation within the carbonates is expressed as a thrust system with west-to-southwest and east-to-northeast transport (fore-trusts and back thrusts) accompanied by cm-scale spaced cleavage in the carbonates (dolostones as well as limestones). This deformation is rotated by the Alleghanian deformations, so this is interpreted to be a pre-Alleghanian event; the timing of this event is not determined, but there are offsets in the Silurian and Devonian strata of the transition zone which may be related to this event. Whatever its origins, the deformation increased the effective thickness of the Cambro-Ordovician sequence, producing the western end of the saddle between the Nippenose and Nittany Valleys. The thrust surfaces associated with this deformation may have controlled the location of the western end of the frontal ramp of the second Alleghanian event.

Although dominant at the eastern end of the anticlinal structure, the first Alleghanian event is poorly expressed in the eastern Nittany Valley. The second Alleghanian event is the primary producer of cleavage on the northern limb of the antiform (I use this term in favor of anticline because the anticlinal structure is cut by multiple back thrusts, thereby segmenting what may have originally developed as an anticline). A third Alleghanian event becomes increasingly dominant southwest of the Mill Hall gap; this event has a structural strike of about N50˚E.

From the Salona area (east of the Mill Hall gap) westward, the frontal limb is clearly detached from the rest of the structure along a major back thrust. The detached strata include the entire post-Beekmantown stratigraphic interval and slivers of upper Beekmantown and older strata either included as part of the back thrust sheets or occurring as horses along the back thrust traces. Numerous duplications occur within this area along additional back thrusts, making measurement of stratigraphic sections suspect.

An additional back thrust decapitates the top of the structure, pushing the strata exposed along the valley axis southeastward onto the southwestern limb. This back thrust can be traced to the east end of the valley where it transects the Bald Eagle sandstone. The northern limb back thrust apparently joins this central back thrust near the east end of the valley. The magnitude of back thrust displacement dies out

rapidly east of Salona, and the back thrust cannot be traced very far into the intervalley saddle. The southwest end of the detachment is linked to a cross-fault that connects to the Bellefonte gap.

In the detached axial portion of the eastern Nittany Valley, multiple thrust surfaces can be mapped east of the village of Lamar; most of these are imbricate thrusts within larger thrust sheets, with the imbricates either dying out or becoming obscure westward. The structurally higher thrust, often considered a northern continuation of the Birmingham thrust but called the Sand Ridge thrust herein, brings Cambrian strata to the surface. The lower sheet contains an equivalent stratigraphic package (as seen in the J. Franklin Long #1 well in Marion township), though only the younger strata are exposed except at the southern end. The stepwise younging of the basal strata, both in the thrust sheet and along the top of the underlying thrust sheet, northeastward from the cross fault suggests that these thrust surfaces were originally north-verging, but have been reactivated as northwest-verging thrust surfaces. A similar scenario is suggested for the Birmingham thrust (*sensu stricto*) in the southwestern portion of the Nittany Valley. This interpretation also explains the localized occurrence of overthrust Siluro-Devonian carbonates near Mill Hall and around Milesburg.

The decapitated valley axial portion between Bellefonte and Salona appears to have been thrust obliquely eastward during the last phase of deformation. Motion would have been obliquely updip with a right-lateral strike-slip sense on the underlying backthrust. A cross-fault extending south from the Salona area is actually a thrust accommodating this oblique motion.

Western Nittany Valley

South of the Skytop cross-fault, the Nittany Valley looks very much like the western portion of the eastern Nittany Valley. Cambrian strata are brought to the surface on two imbricates of the northwest-verging Birmingham thrust. A major back thrust separated the northwestern limb from the antiformal core, with southeast-verging imbrications being well-developed with the back-thrust sheet. Small sympathetic backthrusts related to this displacement can be seen in a couple of the new roadcuts along the I-99 corridor just east of the Skytop gap. As with the eastern Nittany Valley, a second major backthrust separates the Birmingham thrust sheet from the strata to the southeast along the southeast side of the valley; this back thrust was mapped around the south side of the Birmingham window structure, but has not been carried northeastward to State College though detailed mapping shows that this extension exists. The back thrusts can be shown to join just east of the village of Skelp southwest of Birmingham.

Generally, evidence for either of the first two phases of Alleghanian deformation is either not developed or thoroughly obscured by the last two phases in this area. If they exist, more work will be necessary to tease out the earlier deformational signatures.

The last phase of Alleghanian deformation increasingly overprints the third phase toward the southwest end of the Nittany antiform. Back thrusts and accompanying folding dominate the obvious structures associated with the last phase, but transverse faulting and new and reactivated forethrusts are also evident. Displacement directions are generally N60-70˚W and S60-70˚E, following the primary central Appalachian fold-and-thrust structural trend that dominates from this area south to Roanoke, Virginia.

The increase southward in shortening in the plateau is accompanied by a great deal of transverse faulting in the Nittany Valley. Three major cross-faults offset the northwestern limb - at Bellefonte, Skytop, and Bald Eagle; the first two connect to displacement zones that cross much of the anticinal core, whereas the last connects to a backthrust that also cuts across the anticline core. The eastern imbricate in the Birmingham thrust sheet between this last backthrust and the Skytop cross-fault (at State College) is broken by small displacement right-lateral transverse faults with approximately 500 m spacing. This suggests a late-stage counterclockwise torsion within the Valley-and-Ridge.

THE TRANSITION ZONE

The transition zone in the Williamsport area is dominated by imbricate thrusting above the Salina detachment. This structural style dominates everything east of the village of Woolrich, but progressively dies out westward, being recognizable for only 3 km west of Woolrich. From just west of Woolrich to the village of Bald Eagle (just north of Tyrone), the transition zone is dominated by a massive monocline that flattens northwestward into the plateau; only minor back thrusts are evident in most of this area, though there are a couple of small thrusts and associated folds found near the edge of the Valley-and-Ridge. Southwest of the village of Bald Eagle, the transition zone structures start transitioning into the continuation of the main central Appalachian folding extending the Valley-and-Ridge structure west of the Nittany anticlinal axis.

Eastern Zone

The imbrication of the eastern transition zone is dominated on the south edge by thrusts that place Siluro-Devonian carbonates and immediately overlying strata onto stratigraphically much younger strata (Fail et al., 1977, Faill and Wells, 1977; Hoskins, 1976). There is no evidence of any involvement of pre-Salina strata in the structure, and the short wavelengths of the folds argues strongly against any ramp development below the Salina. The transfer of the primary basal displacement from the lower Cambrian to the Salina results in a much thinner stratigraphic package, and the lateral compression of that package obviously resulted in the imbrication of that thinner package.

The southern edge of the transition zone is dominated by emergent thrusts, the northern of which, the Jersey Shore thrust, is traceable across much of this zone (Hoskins, 1976; Faill et al., 1977). North of the Jersey Shore thrust, most of the primary thrust ramps are blind, producing dramatic fault-propagation folds and backthrusts. The folds that will be seen on the Pine Creek Rail Trail are excellent examples of these fault propagation folds, with shallowly dipping trailing limbs and steep to vertical frontal limbs. Along the Rail Trail, there are three such folds, two of which will be seen; the third fold turns the strata vertical right up to the edge of the plateau (i.e., just north of the cross-roads labeled Torbert on the Jersey Shore 7.5 min quadrangle). The northern of these folds, which is also the largest, can be traced eastward to US Rte. 15 just south of the Hepburnville interchange. North of this fold, the folding is much less dramatic, reflecting structures much more like the plateau to the north. Backthrusting within this folded zone and to the north is probably related to the bed length problems associated with fault propagation folding (see Mitra, 1990). The displacement on these folds dies out westward, disappearing west of Woolrich.

It should be noted, however, that there are two folding and displacement directions represented within these structures - a primary northward (c. 0°) displacement with concomitant east-west fold trends and cleavage strikes, and a later northnorthwestward (c. 330°) displacement with concomitant ENE folds trends and cleavage strikes (Engelder, 1979; Engelder and Geiser, 1979; Geiser and Engleder, 1983). The east-trending folds (stage 1) dominate in the east, whereas the ENE trending folds (stage 2) dominate in the west. However, local features for each can be found across the entire area. Some penciled shales in the transition zone contain diamond-shaped pencils reflecting the two directions of shortening. In other places, phase I cleavage is found within phase 2 folds.

Even more striking, NE-striking and NNE striking features also can be found locally, indicating that the area has been influenced by the later deformations that dominate to the southwest. Structures associated with the last phase commonly show displacements consistent with the aforementioned torsion within the thrust stack. The incomplete separation of the hangingwall from the footwall beneath the fault propagation folds probably limited the lateral movement on these surfaces. However, the curl on those folds at their western end (see Taylor, 1977) suggests that they experienced transverse movement during the later, non-coaxial phases of Alleghanian deformation.

Central Zone

The structure within the central transition zone is deceivingly simple. The strata form a great monocline with an average 40°NW that appears to include all of the strata from Late Ordovician to Mississippian. Only adjacent to Central Mountain High School by Mill Hall and in a narrow strip from Milesburg northeast to the village of Curtin is there direct evidence of thrusting atop a Salina decollement. In the former area, Siluro-Devonian carbonates can be seen thrust over Marcellus in a north-verging thrust, but this is likely just the westernmost expression of the Jersey Shore thrust system. In the later area, the S-D carbonates are folded and thrust westward onto the trailing edge of a larger sequence of Devonian strata. Other than these cases, all of the significant thrusting and associated folding indicates southeastward verging thrusts, i.e. back thrusts. Even these are sparse, leaving a nearly uniform monoclinal fold containing the entire transition zone.

One interesting complication is that the apparent stratigraphic distance between the Tuscarora atop the northwestern ridge of Bald Eagle Mountain and the late Devonian strata varies significantly along strike. Where that distance is least, the topographic expression of the S-D carbonates and even the first overlying sandstone disappears, suggesting that the ridge is thrust westward onto the monoclinal Siluro-Devonian strata. This fault has only been mapped along the north side of the ridge in the eastern zone, but the evidence appears to require that it continues nearly the entire length of the central portion as well. If so, this fault would probably be post-back thrusting, making it one of the last structures formed in the system. It is possible that the thrust represents an extreme example of the forelimb out-of-sequence thrusting associated with fault propagation folding creating this monoclinal limb. Small displacement faults exposed in the quarries adjacent to the ridge cutting the steeply dipping uppermost Ordovician carbonates and the back thrusts within those carbonates are consistent with this late-stage out-of-sequence northwestward verging thrust interpretation. Small versions of this style of late-stage thrusting within the frontal limb of a fault-propagation fold are seen at Station 6 on the Pine Creek Rail Trail.

Southwestern Zone

Southwest of a cross-fault by Bald Eagle, the S-D carbonates and later strata become involved in imbricate thrusting again, indicating that a more normal style of thrust transition to the plateau occurs there. Major back thrusts disrupt this imbricate thrust system and adjacent Valley-and-Ridge structures, but these die out southward as new ridges become prominent within the plateau. These ridges eventually become a westward continuation of the Valley-and-Ridge in West Virginia.

The collapse of the plateau to the west of the southwestern transition zone probably accounts for the torsion within the thrust stack and transition zone farther north. The high density of faults in the plateau just north of the transition at Bald Eagle (Glass, 1971; Faill, 1981) is also consistent with this interpretation.

EFFECTS OF FAULT PROPAGATION FOLDING

Fault propagation folds form as a ductile bead ahead of a thrust ramp that is propagating at a slower rate than displacement is occurring on the ramp. The relative rate of displacement and propagation controls the fold angle and dip on the leading limb. The height of the fold reflects the amount of displacement that has occurred on the ramp. Many fault propagation folds have been mistaken for fault-bend folds, but the lack of angular correspondence with either Mode 1 or Mode 2 fold geometries (Suppe, 1983) gives them away.
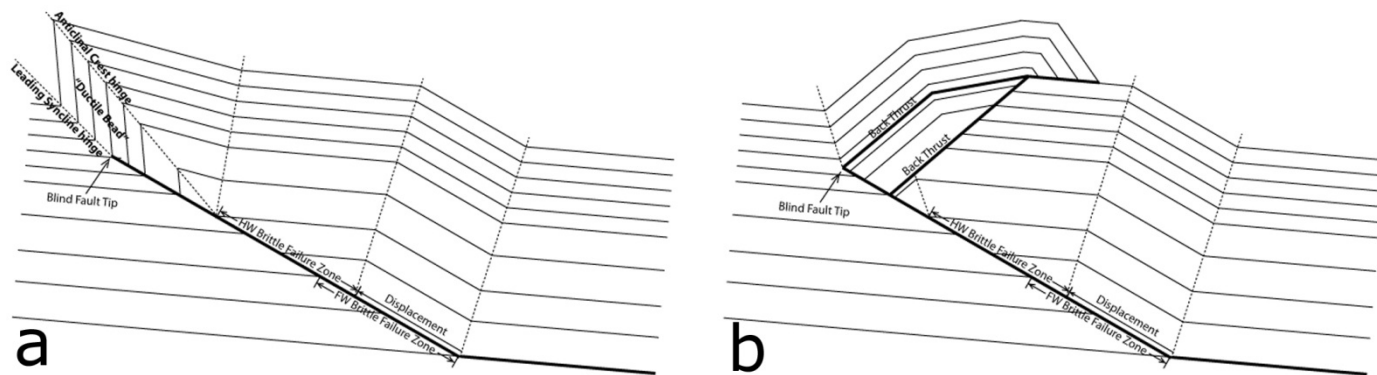
Figure 2. Fault propagation fold geometries that do not require back thrusts (a) and that do require back thrusts (b). Most of the dramatic fault propagation folds seen in the transition zone are of the first variety, but the main portion of the Nittany anticline and most of the plateau folds are the latter variety.

As the ramp propagates forward, the trailing limb and anticlinal crest remain passive, whereas the leading synclinal hinge migrates forward in concert with the progression of the fault tip (see Figure 2a). As Mitra (1990) pointed out, fault propagation folds produce a structure that does not accommodate the bed length of the overlying strata, so there is a kinematic requirement for additional shortening of this strata. Though Mitra (1990) suggested that this could be accommodated through forethrusting of the overlying strata, the common association of fault propagation folds and back thrusts in central Pennsylvania suggests that much of the excess bed length is accommodated by peeling off the top of the fold through the development of back thrusts (Figure 2b).

Along the Pine Creek Rail Trail, the backthrusting at Station 1a and the concomitant folding on the south side of Station 1b is an expression of the backthrusting associated with many fault propagation folds. The lack of similar backthrusting associated with the two larger folds to the north can be explained by the achieving of the critical condition of a vertical leading limb so that a divergence of dip on the anticlinal crest axial plan and the leading synclinal axial plane can accommodate the small remaining bed length disparity.

The lower leading limb dips in the Nittany anticlinal structure, including the monoclinal transition zone, would require much more bed length accommodation, thereby producing the major back thrust along the frontal limb of the anticline and even the decapitating back thrust within the anticlinal core. The transecting of the anticlinal core and frontal limb created a nascent triangle zone, suggesting that other triangle zone development can be explained by the development of low-angle fault-propagation folds forcing the development of back thrusts.

DISCUSSION

Deformation in the central Appalachians of Pennsylvania proceeded through a series of non-coaxial phases of shortening representing four distinct displacement vectors rotating progressively from nearly due north to westnorthwest. The structures of the outermost Valley-and-Ridge and adjoining transition zone to the plateau reflect these changes, with abrupt changes in structural style in the transition zone marking the boundaries between areas dominated by each of the last three phases of deformation.

The northeastern end of the bend is dominated by phase 1 and 2 structures, containing thrusts and folds that are developed above a detachment horizon that appears to be the Salina slat. The Valley-and-Ridge folding in the eastern Nittany anticline is a relatively simple ramp anticline containing the entire Cambro-Ordovician carbonate section thrust onto the Salina detachment. The adjoining portion of the transition zone is dominated by emergent thrusts, which give way northward to dramatic fault propagation folds above blind ramps. These fault propagation folds abruptly give way northward to low-angle folds that decrease in amplitude as they pass into the plateau to the north. These low amplitude folds are cut by numerous small forethrusts and back thrusts, suggesting that they are fundamentally fault-propagation folds.

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The western end of this deformation in the Valley-and-Ridge occurs within the eastern Nittany Valley and beneath the saddle between the Nittany and Nippenose Valleys, and in the transition zone just to the north of that area. This is an area where the Cambro-Ordovician carbonate section is cut by pre-Alleghanian faults that lie nearly perpendicular to the Alleghanian structural trend. It is expected that these faults served as a transverse ramp zone for the transfer of the Nippensone Valley ramp displacement directly onto the Salina surface, thereby allowing the continuation of the plateau deformation to occur to the west during this early phase of the Alleghanian orogeny.

The central Nittany Valley developed after the eastern end (Nippensoe Valley and eastward). Failure of the Cambrian through Silurian strata involved development of a fault-propagation fold beneath the Nittany anticline with a great monoclinal forelimb. The excess limb length created by this geometry requires accommodation, commonly by backthrusting. The back thrusts within the frontal limb and cutting across the Nittany anticline have been mapped and create a triangle zone geometry. The northeastern extension of this deformation appears to have been accommodated by transverse motion on the fault surfaces of the previous phases, so there was little transfer of this motion into the plateau to the north, though there is some extension of the deformation in the plateau to the northwest.

The last phase of Alleghanian deformation is not well expressed in the northeastern end of the great bend. This phase involved the imbrication of the Cambro-Ordovician strata west of the Valley-and-Ridge (as seen in central Pennsylvania) with a westnorthwest displacment direction. The torsion associated with this last motion was accommodated where possible by slippage on the earlier faults, but it created additional backthrusting in the southwestern Nittany Valley and an area of high density minor faulting in the plateau to the west.

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A NOTE ABOUT FORELAND STRUCTURES

Paul Washington

The foreland edge of a fold and thrust belt along with the adjoining deformed plateau (a.k.a. foreland basin) contain a distinct suite of structures that are often more subtle and less well developed than those found in more interior parts of an orogenic belt. It is worth discussing these structures separately before we proceed to talk about their relations to regional structural patterns.

Folds

In the plateau and foreland portions of the fold and thrust belt, folds exist only as drapes over other structures, generally thrust ramps though ductile collapse zones also exist very locally. Ramp-bend folds are asymmetric, with the steeper limb (actually, the limb having the greater angle with the footwall bedding) being the leading limb developed above the hangingwall cutoff. Because there is generally a forelandward progression of thrusting, it is common for ramp-bend folds to be rotated by subsequent underlying ramp-bend folding. In many places, there will be small (outcrop-scale) ramps, termed wedges by Cloos (1961), within these larger structures, with each ramp creating its own ramp-bend fold in the overlying strata. The amplitude of a ramp-bend fold reflects its stratigraphic throw, and the width reflects its lateral displacement (Suppe, 1983). Fault-propagation folds form a significant subset of ramp-bend folds in which the ramp propagates at the same time as slip is occurring on the ramp (Mitra, 1990). These folds generally have steeper forelimb dips and resolution of the subsurface structure from their shape is somewhat less obvious.

The Valley-and-Ridge folds are ramp bend folds (both standard ramp-bend and fault-propagation) developed above imbricate ramps cutting across the Cambro-Ordovician carbonate shelf sequence. The thickness of the Cambro-Ordovician section creates the very large folds characteristic of this province. The folds we will see in the transition zone are ramp bend folds developed above ramps rising off of the Salina decollement and cutting up through the Siluro-Devonian carbonates and overlying clastics. Because the stratigraphic section is much thinner and horizontal displacements diminish dramatically as we cross into the plateau, these folds are generally much smaller. Recumbent folds and footwall synclines and are rare, but can occasionally be seen (or mapped) in the outer portions of the fold-and-thrust belt.

The folds of the plateau are also ramp-bend-folds, but the ramps are generally blind thrust structures (i.e., they do not cut to the surface). As we go further north into the foreland, the amplitude and asymmetry of these folds diminishes; the loss of asymmetry may be due to the interference of doubly verging thrust systems, creating equivalent fold shapes on both limbs of the larger overlying structure.

Faults

Most faults seen in the outer portions of the Valley-and-Ridge and plateau are thrust faults. These thrust consist of layer-parallel portions (a.k.a. flats, decollements, detachments) connected by ramps that cut across the layering. It is the misfit created by the movement of the stepped hangingwall up the stepped footwall that produces the ramp-bend folds.

Thrust systems generally form by progressive breaking of ramps in a forelandward progression; this progression is called "in sequence." In a classic in-sequence thrust succession, no thrust transects an earlier formed thrust; rather, later thrusts use the same detachments used by earlier thrusts and carry any overthrust sheets along in a relatively passive piggyback manner. Any thrust that cuts across earlier structures is considered "out of sequence." Out-of-sequence thrusts are quite common in this area and become dominant in the Nittany Valley and adjoining transition zone to the southwest.

The most common out-of-sequence thrust in this area is the back thrust, a thrust that verges opposite the regional transport direction. In this area, in-sequence thrusts verge northward and back thrusts verge southward. Many of these small back thrusts can be linked to room problems created by fault-propagation

folding, but significant back thrusts whose origins are not as clear will also be seen. To the west of this area, these back thrusts create a triangle zone, a thrust system in which earlier in-sequence thrusts are transected by major back thrusts, making the advancing thrust wedge into an insertion wedge driving under the adjacent plateau. A much less common out-of-sequence thrust structure is one formed by foreland-verging imbrication of the existing faulted material. This appears to occur within the Nittany Valley, beng the mostly likely explanation for the Birmingham thrust system, including the Birmingham window itself.

Cross-faults, a.k.a. tear faults, are strike-slip faults that offset the dominant contractional fault and fold structures. Although there are places where these faults are obviously lateral edges to thrust sheets, i.e. essentially vertical lateral ramps, that does not appear to be the general case in this area. Rather, these faults appear to be transfer zones between thrust faults that experienced reactivated slip, usually oblique to the original slip, during later oblique transport of the thrust mass. This same oblique reactivation creates local normal motion on some fracture surfaces. Essentially, the northern edges of the fold-and-thrust belt experienced torsion during the subsequent northwestward and west-northwestward displacement that dominates the Valley-and-Ridge to the southwest.

Cleavage and Joints

Closely spaced to continuous cleavage of the sort seen within the interior parts of the orogen are rare within the foreland fringe of a fold-and-thrust belt. Nevertheless, we use the term to designate a penetrative spaced fracturing (a.k.a. fracture cleavage) that parallels the structural strike. Within shales, spaced cleavage often produces a distinctive penciling paralleling the intersection of the spaced cleavage and bedding (Engelder and Geiser, 1979). Similar patterns can be seen in siltstones and even sandstones, though the cleavage spacing tends to increase as the clay percentages decrease.

Cleavage is caused by a domainal removal and concomitant redistribution of material in response to shortening. The greater the shortening, the more closely spaced the surfaces will tend to be, so cleavage spacing tends to increase outward onto the plateau. Eventually, the spacing becomes so great that they are seen only as a consistent fracture orientation parallel to structural strike – Engelder's (1979) strike-parallel fracture set. The presence of crinkle folding in pelite films in jointed sandstones, and the presence of obvious concentration zones in thin-section along incipient fracture traces demonstrates that these are indeed shortening fabrics that open as joints during exhumation. Cleavage never represents the entire layer-parallel shortening that the rock has experienced. Rather, significant (probably >10%) bulk shortening usually accompanies the first appearance of cleavage and increases as the intensity of the cleavage increases.

Within the outer Valley-and-Ridge and transition zone, it can be shown that the most intense cleavage is associated with ductile shortening around ramps. The strata above a hangingwall ramp are commonly cleaved … and then the cleavage is rotated as the ramp-bend fold forms. This should be expected since the failure and growth of the ramp is never simple, with the ramp fracture nucleating within more brittle layers as the stress is concentrated by the ductile shortening of the adjoining layers. The cleavage, therefore, should represent the ductile shortening of those more ductile layers.

It should be noted that diverging shortening directions will often result in overprinted cleavages. However, if the cleavage formed during ramp formation, changing displacement directions on the ramp will not generally result in the formation of a new cleavage except where there is some obstruction to that new displacement direction.

Cross-strike joints have been used extensively to map paleostress field directions in the plateau of northern Pennsylvania and western New York (see Engelder, 1979; Geiser and Engelder, 1979; Younes and Engelder, 1999). These joints are generally much more planar than the strike-parallel joints, and they are more likely to be decorated with plumose structures and other hackle. Different populations of these joints are generally distinguished by their orientations.

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VIEW OF A SILURIAN AGED OCEAN IN CASTENEA, PA

Dana Heston, Donald Hoskins, Sarah Bradley

This large exposure of the Mifflintown Formation is shown in Figure 1. It consists of the McKenzie Member of the Mifflintown Formation, evident by the interbedded fossiliferous limestone and medium to dark gray shale beds. The contact with the overlying Bloomsburg Formation was discovered on the same property. The exposure is dipping 43º to the north with a joint set dipping 88º to the east.

Figure 1 Mifflintown Formation exposure in Castenea, Pa

Figure 2 Evidence of a fault was found at approximately the midpoint of this exposure. This picture shows the offset ripple marks, and slickensides that indicate movement.

Heston, D., Hoskins, D.M., and Bradley, S. (2013)View of a Silurian aged ocean in Castenea, PA, *in*, Fleeger, G.M., Schmid, K., and Anthony, R., eds., A Tale of Two Provinces: the Nippenose Valley and Route 15 Corridor, 78th Annual Field Conference of Pennsylvania Geologists, Williamsport, PA, pp. 35 – 36.

Figure 3 Small to medium cross bedded ripple marks

Different layers of the exposure contain different ripple mark patterns, indicative of changes in sea level and flow direction. The choppy appearance of the ripple marks on the bottom bedding plane of the exposure are consistent with a lower sea level; while, the top bedding plane shows that sea level is increasing by the flow direction in the larger ripple marks. The cross-pattern of the ripple marks shown in figure 2 shows a change in the flow direction.

Several species of fossils were discovered in the study area. Trilobites are present in the limestone bedding. Evidence of worm burrows, ostracods, cephlapods, and worm burrows have also been identified on site.

Figure 4 Orthocone or rusophychus fossils in the McKenzie Member of the Mifflintown Formation.

Figure 5: Burrows in the McKenzie limestone

UPPER DEVONIAN TERRESTRIAL-MARINE TRANSITION ALONG RT. 15, TIOGA COUNTY, PENNSYLVANIA

Chris Oest

INTRODUCTION

Road cuts along US Route 15 expose the Lock Haven and Catskill Formations (Figure 1). The contact between these formations is gradational, representing the gradual transition from marine to terrestrial depositional environments. Driving north along Route 15 from Trout Run, the transition becomes apparent as alluvial plain facies of the Catskill Formation interfinger with prodelta facies of the Lock Haven formation.

TECTONIC AND PALEOGEOGRAPHIC SETTING

During the Acadian Orogeny, pulses of clastic sedimentation associated with collisions between Laurentia and island arcs resulted in the progradation of a clastic wedge into an epicontinental sea (Mintz, Driese, & White, 2010). This basin formed on the western flank of the orogen in response to crustal loading due to uplift along this collision margin (Ettensohn, 1985). Paleocurrent, isopach, petrographic, and lithofacies data confirm that the Acadian highlands not only created the basin in which the Catskill wedge was deposited, but was also the source of the sediment which ultimately filled the basin (Sevon, 1985).
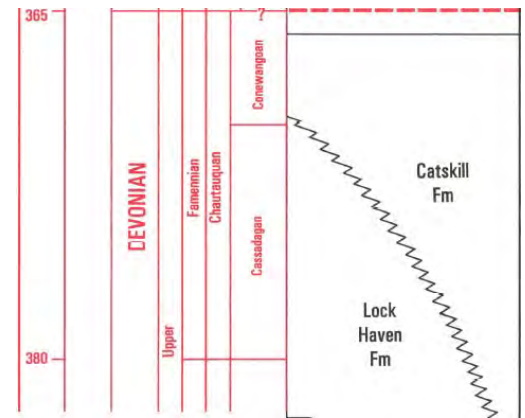
Figure 1. Selected Tioga County Stratigraphy. Modified from Berg, McInerney, Way, & MacLachlan 1993.

Paleomagnetic data indicates that an epicontinental sea received sediments shed from the Acadian highlands which straddled the equator (Figure 2) in the Middle to Late Devonian (Kent, 1985). Lithologic data such as calcareous paleosols, mudcracks, and rain drop impressions Indicate a seasonally wet and dry conditions. Additionally, fossil data such as Dipnoan (lungfish) aestivation burrows and paleobotanic data (Woodrow, 1985) indicate seasonal fluctuations in precipitation, which supports an equatorial depositional setting for the Catskill clastic wedge.

Figure 2: Paleogeography of the Catskill Delta.
A: epicontinental sea   B: Acadian Highlands
C: Alluvial plain, modified from Woodrow, 1985.

Figure 3. Road cut exposing the upper Lock Haven Formation. Located just to the east of Tioga along Route 15 along the northbound lane.

DEPOSITIONAL ENVIRONMENTS

Lock Haven Formation

The Lock Haven Formation (Figure 3) consists of light olive-green interbedded shale, siltstone and sandstones with ripples and flaser bedding (Faill & Wells, 1977). The unit is commonly fossiliferous, containing mainly brachiopods and bivalves. However, bryozoans and crinoid colum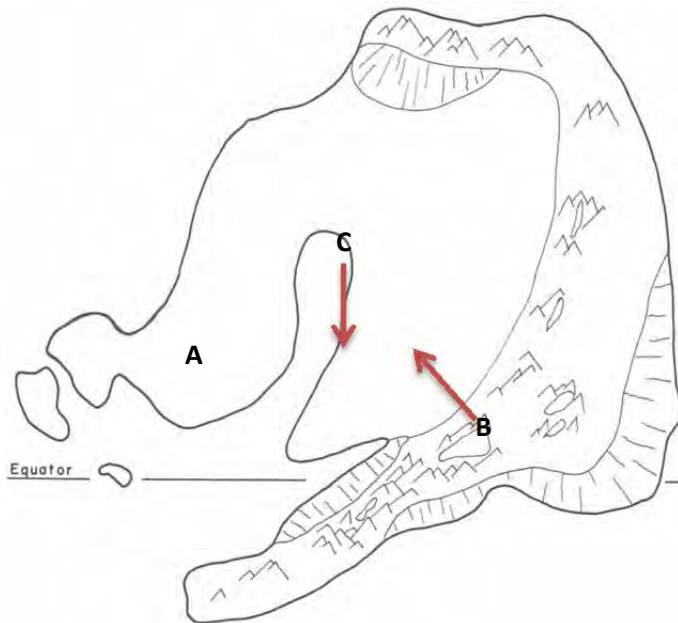nals are also present. Thin shell hash lenses, possibly representing storm lag deposits, are also present. Exposures along Route 15 match descriptions of Lundegard et al.'s "delta front facies" of the lower Lock Haven Formation[1] (1985).

Catskill Formation

The Catskill Formation consists of red to red-gray mudstones, siltstones and sandstones arranged in fining upward sequences. The overall lack of marine fossils, fining upward sequences and evidence of pedogenesis imply a fluvio-deltaic depositional setting for the Catskill Formation. Delta plain, alluvial plain and braided stream facies are found in vertical succession, representing the progradation of the Catskill delta through time. Here, delta plain and alluvial plain facies will be discussed.

Delta Plain: Delta plain facies of the Catskill formation consist of fining upward cycles of fine- to coarse-grained gray-green or grayish red channel sandstones overlain by red siltstones and mudstones (Sevon, 1985). Fine-grained overbank deposits dominate delta plain facies. Pedogenesis is common in interfluve areas, which indicates relatively long term landscape stability between depositional events. Green layers and drab halo root traces indicate accumulation and in-situ decomposition of organic matter. Angular to subangular blocky soil ped structures indicate the possible presence of Bt horizons[2] (horizon of translocated clays) (Retallack, 1988). Soil profiles are truncated by overlying channel sandstones (Figure 4).

Figure 4. A) Gray-green basal channel sandstone cutting into overbank deposits. B) Drab halo root traces and green reduced layers resulting from decay of organic matter. C) Angular blocky peds D) Truncation of potential Bt horizon by channel

With subsidence and sediment supply in an ideal balance, minimal erosion results in the preservation of earlier deposits (Sevon, 1985) allowing for moderate soil development.

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[1] Lundegard et al. refer to the lower Lock Haven formation as the lower "Chemung" Formation. Faill and Wells (1977) proposed renaming the unit between the Braillier and Catskill Formation from Chemung to Lock Haven in Pennsylvania.
[2] The presence of a Bt horizon must be confirmed through geochemical and micromorphological analysis of the horizon in question to prove clay translocation. This is outside the scope of this paper.

Figure 5. Lingula fossil

Soil development is impeded by flooding and influx of sediments. Water conditions were likely brackish as small *Lingula* fossils were found in red beds of this facies (Figure 5).

Alluvial Plain: Alluvial plain facies of the Catskill formation consist of fining upward cycles of gray (Figure 6A) or red medium grained or larger sandstones overlain by red siltstones and claystones (Sevon, 1985).

This facies is dominated by channel sandstones as a result of streams avulsing and meandering across a broad alluvial plain (Sevon, 1985 and Rahmanian, 1979). Evidence for rapid channel avulsion can be inferrred from cross cutting channel bodies (Figure 6B).
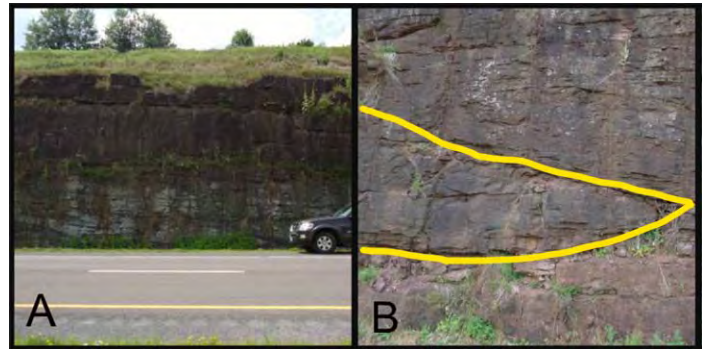
Figure 6. A) Basal gray channel sandstones. B) Cross cutting channel bodies (outlined in yellow).

Figure 7. Pedogenically modified oxbow. Hammer for scale at center of picture.
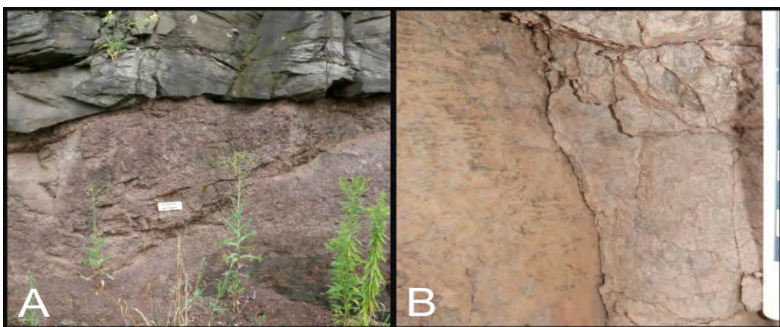
Figure 8: A) paleosol with vertic features B) Slicks on underside of subhorizontal fracture planes resulting from expanding clay

Pedogenically modified channel forms indicate meander cut offs and the formation of oxbow lakes (Figure 7).

Rare well developed paleovertisols are observed in this facies along Route 15 (Figure 8). Vertic features (listric fracture planes resulting from shrinking and swelling of expanding clays) indicate seasonal wetting and drying of this alluvial plain. Seasonal wet-dry periods are supported by Woodrow's (1985) proposal of a rain shadow effect that the Acadian highlands had on the alluvial plain on the western margin of the highlands. Airmasses would have been forced to rise on the eastern margin of the highlands resulting in orographic precipitation east of the alluvial plain and dominantly dry conditions to the west of the highlands.

DISCUSSION

Driving north along Route 15 from Willaimsport to Tioga, one travels up section from the Upper Devonian Lock Haven and Catskill formations, into Mississippian and Pennsylvanian rocks then back down section ultimately back to the Lock Haven exposed in a road cut across from the Route 15 rest stop on the southbound side of the road.

Traveling this route exemplifies Walther's Law, which states that laterally adjacent depositional environments occur in vertical stratigraphic succession. Figure 9 is a block diagram showing a possible reconstruction of "Catskill" paleoenvironments. To the east are the Acadian highlands with alluvial fan (9B) deposits prograding onto the alluvial plain (C). The alluvial plain consists of meandering streams and associated facies. In the central part of the diagram, the alluvial plain grades into the delta plain (9D) where terrestrial and marginal marine facies interfinger. The western edge of the block diagram shows shallow shelf to deep water depositional environments (9E and 9F respectively).
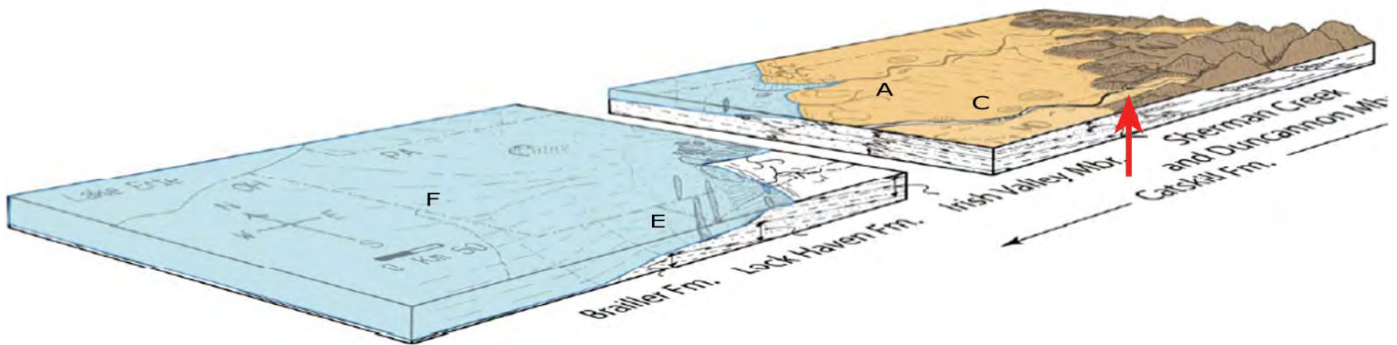
Figure 9: Block diagram showing lateral relationships of various facies. (Modified from Slingerland, Patzkowski, & Peterson, 2009). A) Approximate location of Williamsport B) Alluvial fans C) Alluvial Plain D) Delta plain E) Shallow shelf F) Deep marine

The resulting stratigraphic sequence (Figure 10) from the progradation of the Catskill clastic wedge into the Acadian foreland basin is an overall coarsening upward sequence consisting of packages of fining upward sequences. Distal basin muds are overlain by shallow shelf and shoreline siltstones and sandstones. As the clastic wedge continues to prograde into the basin, shelf and shoreline deposits are overlain by meandering fluvial sandstones and overbank mudstones. Paleosols are common in interfluve areas of this facies. Finally, the dominantly fine grained meandering fluvial facies is overlain by courser grained braided stream and alluvial fan deposits.
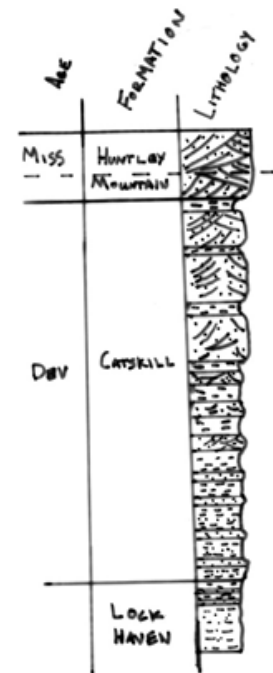
Figure 10.   Illustrative stratigraphic section of the Catskill formation along Route 15 in the Mansfield-Tioga area. Modified from (Dyson, 1967) and (Rahmanian, 1979)

Paleosols examined for this paper were non-calcareous. While this contradicts Woodrow's assertion of an equatorial, seasonally wet-dry climate, the absence of calcareous paleosols may not be a function of climate. Jenny (1941) cites 5 factors which influence soil formation: climate, biologic activity (flora and fauna), relief, parent material and time. Gile et al. (1966) describe 4 stages of carbonate accumulation morphologies. These morphologies indicate the length of time that carbonate has been accumulating, with thin pebble coatings developing in 5,000 to 10,000 years to nodules and duricrusts (pedogenic limestone beds) requiring up to 50,000 years to develop. The absence of any carbonate in the paleosols along Route 15 could indicate that rapid channel avulsion or high sedimentation rates in this part of the basin resulted in insufficient time for carbonate soil development. Another possibility is that this part of the basin was outside of the area affected by the rain shadow created by the highlands and was therefore not dry enough for carbonate accumulation in soils. Vertic features in the soil horizons indicate ample precipitation to translocate swelling clays.

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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., 1978; Braun, 1994, 1999, 2011a; 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-Illinioan 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 (Figure 1). The trace of each successive glacial advance's maximum limit is remarkably similar (essentially parallel) to that of the previous advance (Figure 1).
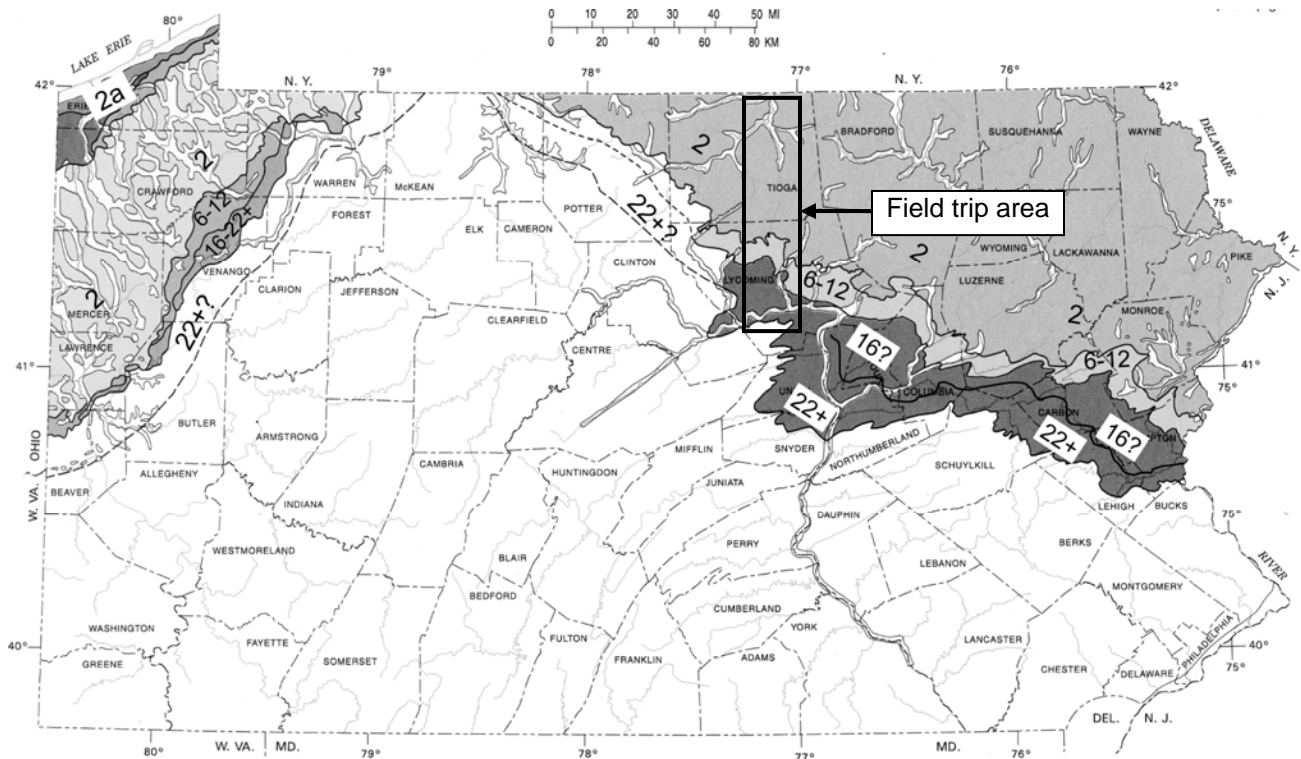
Figure 1. Map of the extent and age of glacial deposits in Pennsylvania. Ages given by Marine Isotope Stage (MIS) number: (Braun, 2011a, Figure 1, modified from Sevon and Braun, 1997b).

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) (Figure 2). The triangular shaped unglaciated area between the two lobes in Pennsylvania and New York is called the Salamanca Reentrant (Figure 2). In northwestern Pennsylvania the relatively low relief (300 to 1000 feet), 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; Marine, 1997). 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 (1000 to 1500 feet) 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, 1895, 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, 1903, Fuller and Alden, 1903; Braun, 1988), and north of the Late Pleistocene limit as ice receded from Pennsylvania (Willard, 1932, Coates, 1966; Gardner *et al.*, 1993; Braun, 1989a, 1997, 2002, 2011a). Part of the northeasterly drainage was diverted south to form the 'Grand Canyon' of Pennsylvania, a 750 feet deep bedrock gorge (Fuller and Alden, 1903; Crowl, 1981; Braun 2011b).

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 Pennsylvania and caused periglacial climate conditions there (Braun, 1989b,1994, 2011a). 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, 2006, Potter, 2013).   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, 2011b).

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 2 is a more recent compendium of ice flow direction (Crowl and Sevon, 1980, 1999).

Fuller and Alden (1903) 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 grouping 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 S62°W direction (Fuller and Alden, 1903). The regional S46°W direction was from thick ice flowing all the way to the terminus about 20 miles southwest of the site. The more westerly S62°W 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.
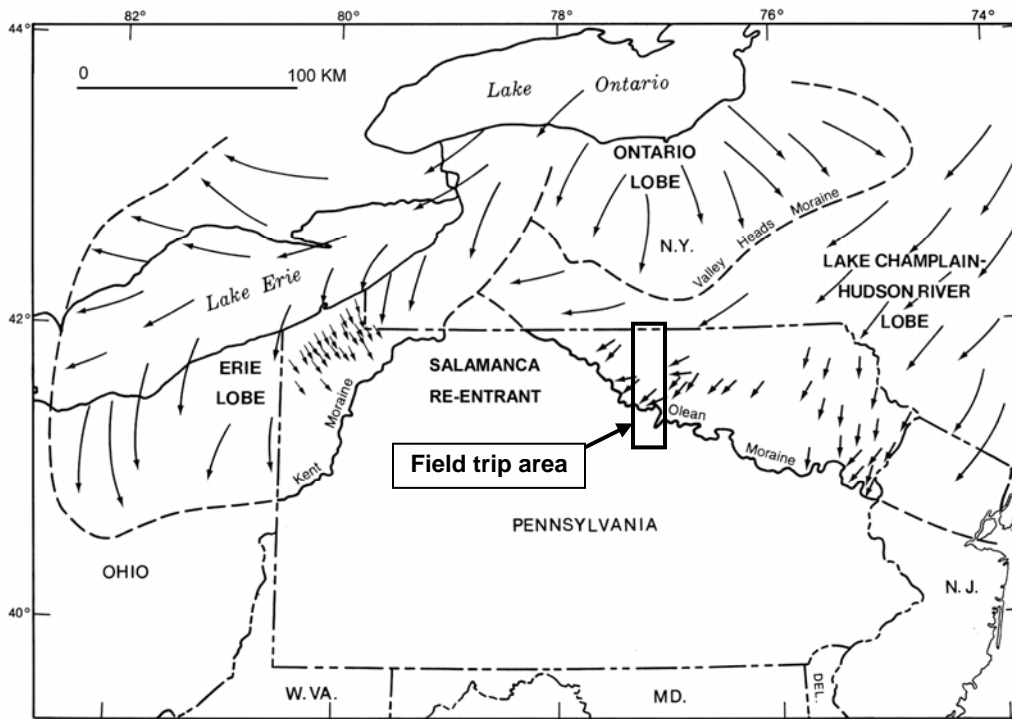
Figure 2. Map showing the ice flow directions into northeastern and northwestern Pennsylvania to either side of the Salamanca Re-entrant (from Crowl and Sevon, 1999, Figure 15-2).

GLACIAL BOUNDARIES IN THE WILLIAMSPORT AREA

A belt of thick glacial deposits runs just northeast of Williamsport, Montoursville, and Muncy. That line of deposits is thought to represent the Illinoian or older glacial boundary (Figure 3, MIS 6 or 12?) (Braun, 2011a; Braun *et al.,* 2008). Only five miles or so northeast of and near parallel to that border lies

the late Wisconsinan boundary (Figure 3, MIS 2)(Braun, 2011a; Braun *et al.,* 2008). The late Wisconsin glacier terminated on the Appalachian Plateau except where an ice lobe protruded down the Muncy valley reentrant in the Plateau. A few miles to the southeast and southwest of Williamsport there is a discontinuous belt of thick glacial deposits that is thought to represent a pre-Illinoian glacial terminus (Figure 3, MIS 16?) (Braun, 2011a; Braun *et al.,* 2008). That terminus should have wrapped around the east end of Bald Eagle Mountain and dammed the West Branch Susquehanna River. About ten miles to the southwest of Williamsport in the West Branch Susquehanna valley is the oldest glacial boundary. Glacial lake sediments there have a reversed polarity magnetic direction (Ramage, J. M., Gardner, T. W., and Sasowasky, I.D., 1998) and are thought to represent an early Pleistocene glacial advance (Figure 3, MIS 22+)(Braun, 2011a; Braun *et al.,* 2008). That glacier dammed the West Branch Susquehanna River from when it advanced over Williamsport until when it retreated northeast of Williamsport. The early Pleistocene terminus (Figure 3, MIS 22+) crosses Bald Eagle Mountain to the east of the Nippenose valley and then to the south the terminus runs along the down plunge ends of the anticlinal ridges.
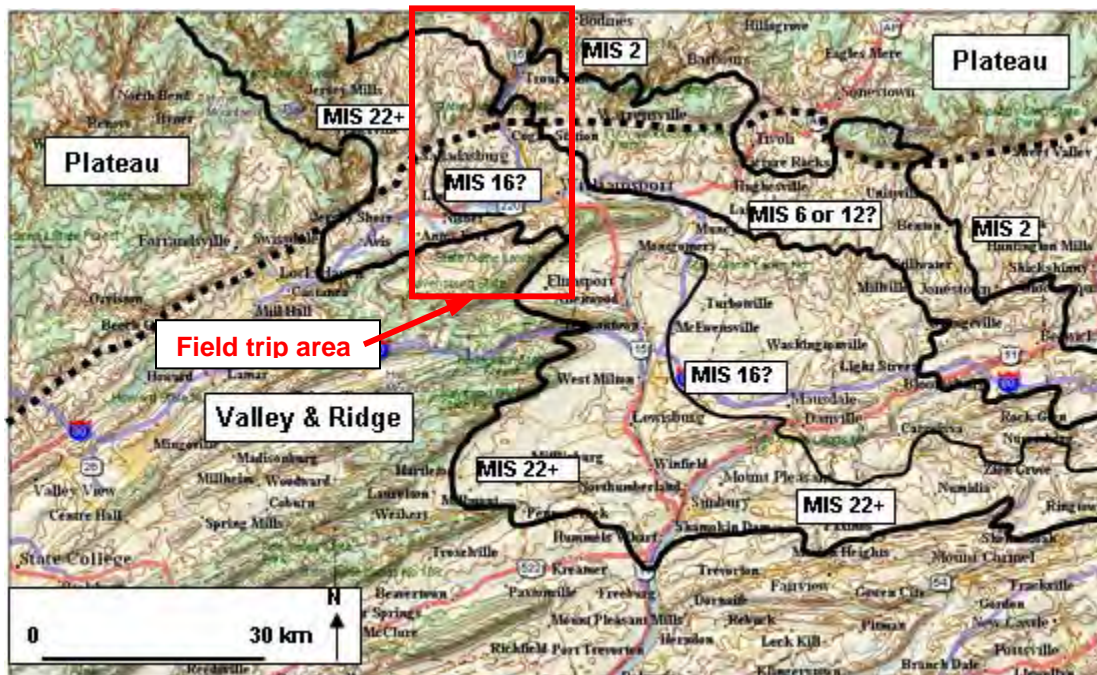
Figure 3. Central Pennsylvania glacial borders. Black lines are the glacial borders of the various glacial advances into Pennsylvania labeled with probable age given in terms of Marine Oxygen Isotope Stage numbers (MIS #). MIS 2 = Late Wisconsinan (25 Ka calendar yrs.); MIS 6 = Late Illinoian (150 Ka); MIS 12 = pre-Illinoian-B (450 Ka); MIS 16 = pre-Illinoian-D (650 Ka); MIS 22+ = pre-Illinoian-G (850+ Ka). Material northeast of the MIS 16? line has normal magnetic polarity while material southwest of that line has reversed magnetic polarity. Dotted line is the physiographic province boundary.

PROGLACIAL LAKE LESLEY IN THE WEST BRANCH SUSQUEHANNA, NIPPENOSE, AND NITTANY VALLEYS

Williamsport is at the east end of Glacial Lake Lesley where the lake had its greatest extent and depth (about 150 m or 500 ft). The lake was spilling over the divide to the Juniata River in the Bald Eagle valley 112 km or 70 miles to our west (Williams, 1895; Leverett, 1934; Ramage, Gardner, and Sasowsky, 1998). The bedrock floor of that outlet was at an elevation of about 1020 feet (borehole data in Ramage, Gardner, and Sasowsky, 1998). With a 1020 water surface elevation, Glacial Lake Lesley would have flooded the entire Bald Eagle strike valley and lower elevation parts of the Nippenose and Nittany breached anticline valleys (Figure 4). As noted previously, the lake sediments exhibit reversed magnetic direction, so first Lake Lesley was Early Pleistocene in age (Figure 3, MIS 22+). A second shorter lived Lake Lesley

should have been formed by the Middle Pleistocene glacial advance (Figure 3, MIS 16?). As each glacier receded to Williamsport, the lake started spilling around the plunging anticlinal nose of Bald Eagle Mountain due to the northeast-southwest orientation of the terminus (Figure 3) that would have permitted drainage to the east around nose or east end of Bald Eagle Mountain. The Illinoian or older glacial border is trending southeast-northwest just to the north of Bald Eagle Mountain (Figure 3, MIS 6 or 12?) and that glaciation probably did not dam the West Branch Susquehanna River or if it did so, it would only have formed a temporary low level lake before the river breached the relatively thin ice dam.
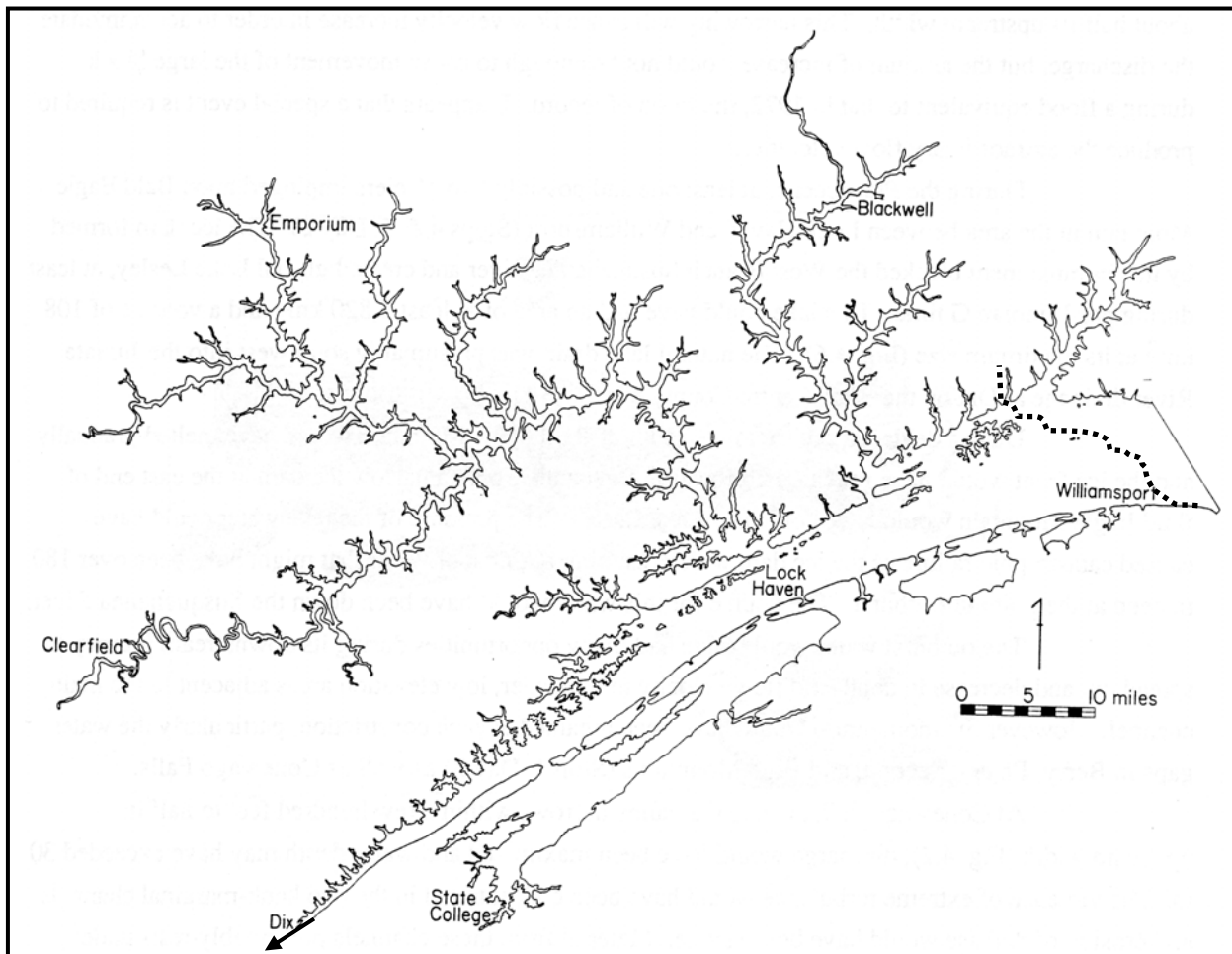
Figure 4. The maximum extent of Glacial Lake Lesley, drawn along the 1100 feet contour. Its outlet at Dix (arrow) drained southwest into Juniata River drainage (from Gardner et.al, 1993, Fig. 4.6). Dotted line is Braun's estimate of where the ice front was when water started spilling eastward around the plunging anticlinal nose of Bald Eagle Mountain.

EARLY, MIDDLE, AND LATE PLEISTOCENE DEPOSITS AT THE ANTES FORT "FAN"

Williams, who named Glacial Lake Lesley in 1895, called the landform at Antes Fort a fan-cone (Figure 5). He thought that it formed from a torrential discharge of glacial origin through the Antes Creek gap in Bald Eagle Mountain from the Nippenose valley on the south side of the mountain (Williams, 1917, 1920). Leverett (1934) used both fan-cone and alluvial fan to describe the landform. He noted that in the railroad cut 27 ft of gravelly material was on top of laminated clay that extended down to shale. He also thought that the 660 ft (200 m) elevation of the top of the fan marked the maximum level of the glacial lake. Andreus (1993) did a more detailed description of the railroad cut (Figure 6) and found that the gravel was only a few feet thick veneer over tens of feet of glacial lake sands and varves, and, at the west end, glacial till under the varves. The apex of the fan-like landform is actually a bedrock hill capped by glacial

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sediments. The gravelly veneer is alluvium deposited by Antes Creek after the lake drained and before the River and the Creek incised into the glacial deposits.

The early Pleistocene till and varves are essentially unweathered at this site thanks to burial by overlying sands and gravelly alluvium. The varves are reddish from clay derived from the thick redbed sequence in the Devonian Catskill Formation. They retain a strong reversed magnetic polarity signature and place the material in the early Pleistocene (MIS 22+) (Gardner et.al., 1993; Gardner, Sasowsky, and Schmidt, 1994). It has argued elsewhere (Braun 1989b, 1994, 2011a; Ramage, Gardner, Sasowksy, 1998) that these sediments were deposited in MIS 22 since that was the first of the major cold events of the last million years and that this landscape has been eroding relatively rapidly in the Pleistocene.

On the mountain toe-slope immediately south of the fan-like landform are boulder colluvium deposits from periglacial gelifluction activity in mid and late Pleistocene times. The boulder colluvium extends from 1000 to 560 feet elevation, well below the 1020 feet surface elevation of Glacial Lake Lesley and must post date the last impoundment of the lake in the mid-Pleistocene (MIS 16 ? advance). The ubiquitous boulder colluvium deposits in eastern, central, and southern Pennsylvania have been highly stable (not being eroded or deposited) under the present climate and have many morphological features that indicate a gelifluction origin (Clark and Ciolkosz, 1988; Clark *et. al.*; 1992; Braun, 1994, Potter, 2013, Day 2, Stop 2).
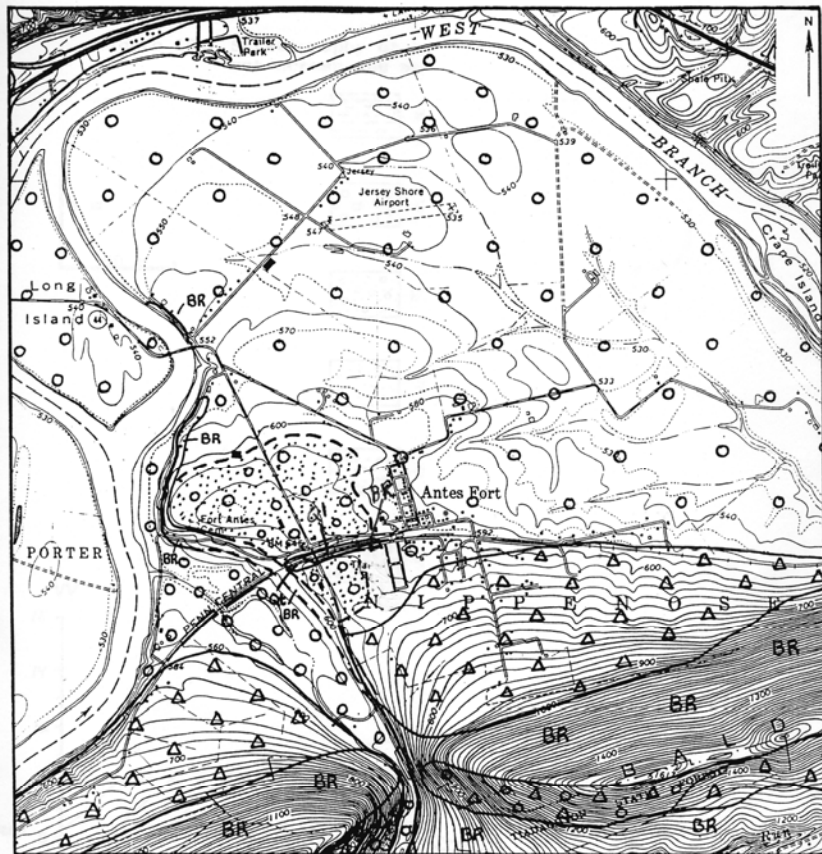
Figure 5. Topography and surficial deposits of the Antes fan-like feature at Antes Fort (Andreus, 1993, figure 9)

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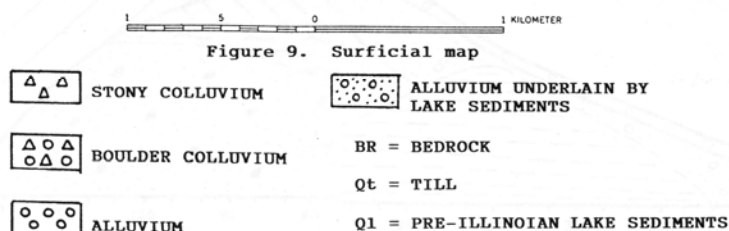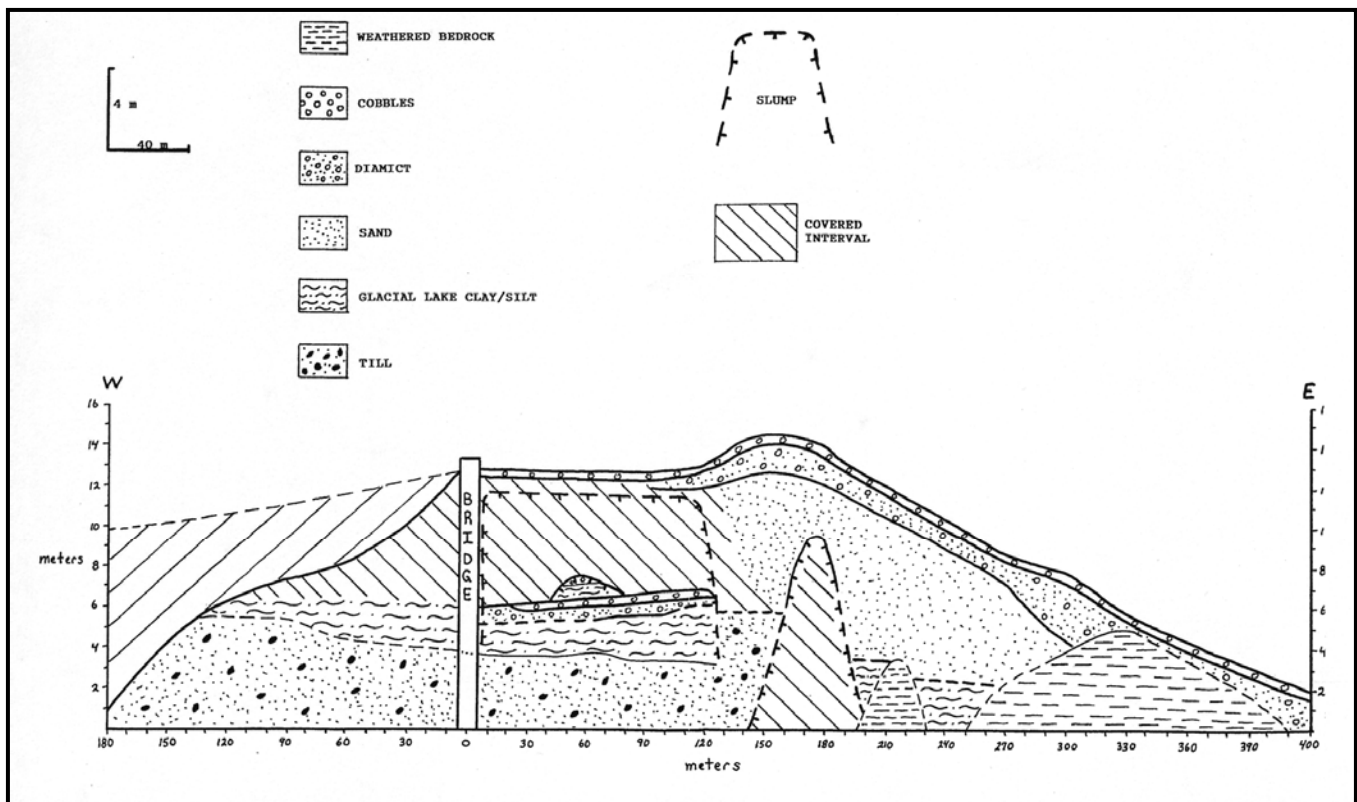
Figure 6. Stratigraphic section of the north face of the railroad cut at Antes Fort (Andreus, 1993, Figure 10).

     Immediately north of the fan-like landform are a series of late Pleistocene outwash terraces on the inside of a large incised meander loop of the West Branch Susquehanna River. At this site, Peltier (1949, Table 41) had noted a 1000 feet wide Binghamton terrace about 25 feet above the river and a 3000 feet wide Olean terrace, about 45 feet above the river. The Olean terrace represented outwash from the late Wisconsinan terminus and the Binghamton terrace represented conditions when the late Wisconsinan terminus had retreated to the Pennsylvania - New York border. The West Branch Susquehanna was receiving no meltwater by the time the glacier had retreated to New York so the Binghamton terrace represented a temporary stabilization of the post glacial down-cutting of river. Peltier did not note that at the village of Antes Fort there is an Illinoian or older terrace remnant 60-70 feet above the river but he noted such remnants at that elevation elsewhere along the West Branch Susquehanna. Braun (1995) corroborated Pelter's terrace mapping at the site.

THE WIND GAP THROUGH BALD EAGLE MOUNTAIN

     US Route 15 crosses Bald Eagle Mountain in a 215 m or 700 ft deep wind gap whose bottom elevation is presently 367 m or 1205 feet (Figure 7). The floor of the center of the wind gap is covered by boulder colluvium but bedrock is exposed starting at least at 340 m or 1120 ft in a gully on the north side of the floor of the wind gap. Immediately north of the wind gap and Bald Eagle Mountain, the West Branch Susquehanna River is at an elevation 152 m or 500 feet, 188 m or 620 ft or so below the bedrock floor of the wind gap. The nearness in elevation of the wind gap floor and the highest level of Glacial Lake Lesley has led some to informally suggest that the gap was an outlet for the lake. But the cutting of the wind gap started at the 1900 ft level in the top of the mountain, well above lake level. Also the 700 feet of incision into the Tuscarora sandstone could not have occurred in the few thousand year life of the glacial lakes.

     The cutting of the wind gap is from stream capture around a plunging anticlinal nose, probably sometime in the Pliocene (Braun, *et. al*, 2008).. It is a type of stream capture peculiar to fold belts where a larger trunk hung up on resistant bedrock is captured by one of its own tributaries working headword around

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the nose of the plunging anticline on much weaker bedrock. As the entire landscape was eroding down, the West Branch Susquehanna would have found itself incising into the Tuscarora sandstone at the crest of the plunging anticline. That would have slowed its incision and greatly increased its gradient at the site. This would have kept the river upstream of the gap from incising at the rate of the river downstream of the gap, it would have been "hung up". A strike valley on much weaker shale and limestone would have existed around the plunging nose of the aniticline, as is does today. An east bank or down plunge direction tributary to the river on the south side of the plunging anticlinal nose would be able to readily work headword around the nose of anticline. Eventually the tributary would capture the river on the north side of the anticline, bringing the river around the nose of the anticlinal ridge, its present position. The capture would have been accelerated by headword spring sapping from subterranean flow through the limestone core of the strike valley. The process is starting again today at the axis of the anticline where the river now has an abruptly steepened gradient.
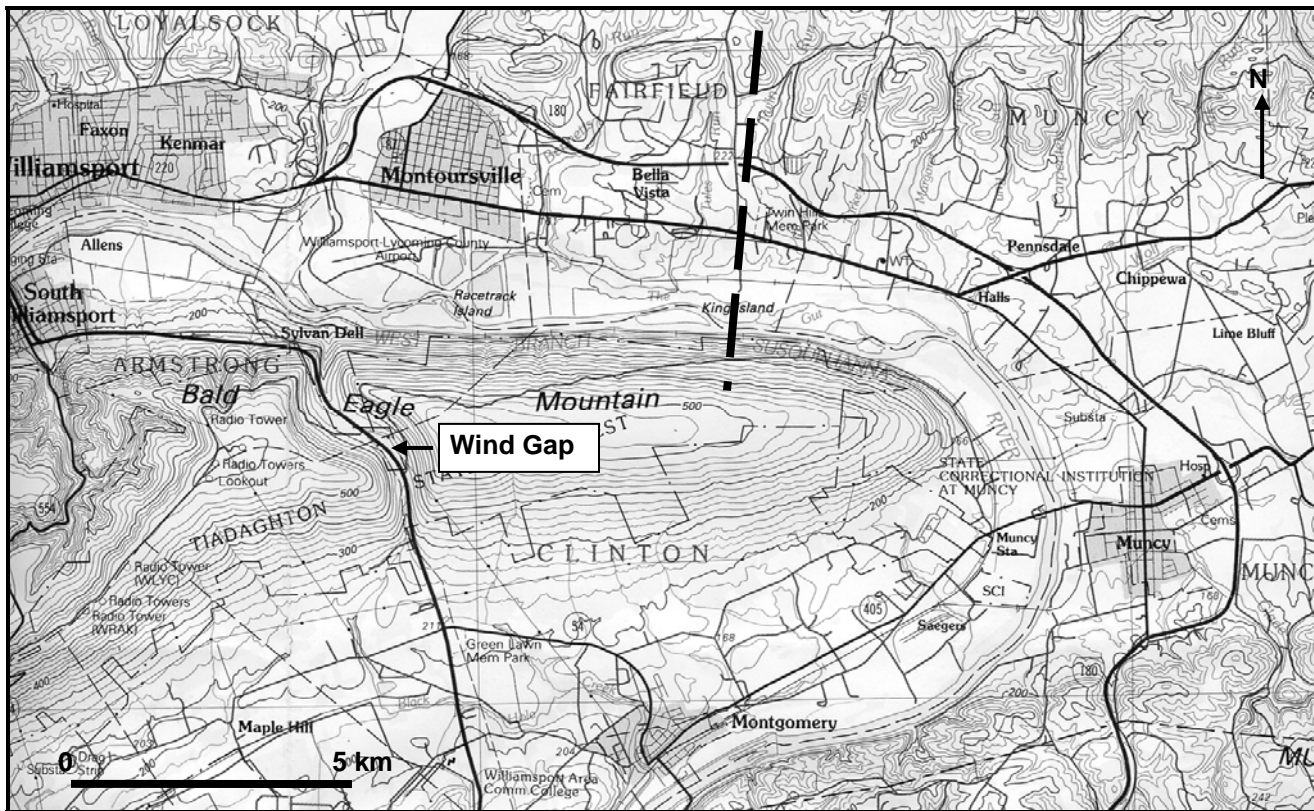
Figure 7. Topographic map of the Williamsport area showing the wind gap through Bald Eagle Mountain. Contour interval = 20 m. Heavy dashed line is probable local divide in the strike valley that was breached by headward erosion of a tributary to the river to capture the river to its present location.

THE LARGE-DEEP PROGLACIAL LAKES IN THE TIOGA RIVER BASIN

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 (Stop 6). 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

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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, valley wide 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 Figure 10 E) in a tributary valley to the Tioga valley in Mansfield area.

Coates (1966) 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, 2011b) emphasized that there were a number of large proglacial lakes, each with several levels, in north-central Pennsylvania (Figure 8). Sevon and Braun (1997a) did reconnaissance mapping (in the middle 1980's) the glacial deposits in the region. The field maps were done at 1:24,000 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 Fuller and Alden (1903), and Denny and Lyford (1963). In those smaller valleys, outcrops of clayey to sandy glacial lake sediments at the toes 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 equally positioned 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 (Figure 8, arrows labeled 1730, 1500, 1230), 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

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south had a marked change in geology and elevation at the site of the Pine Creek gorge (Figure 8, PCG) (Braun, 2011b). 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 over the divide to its opposing south draining valley than any

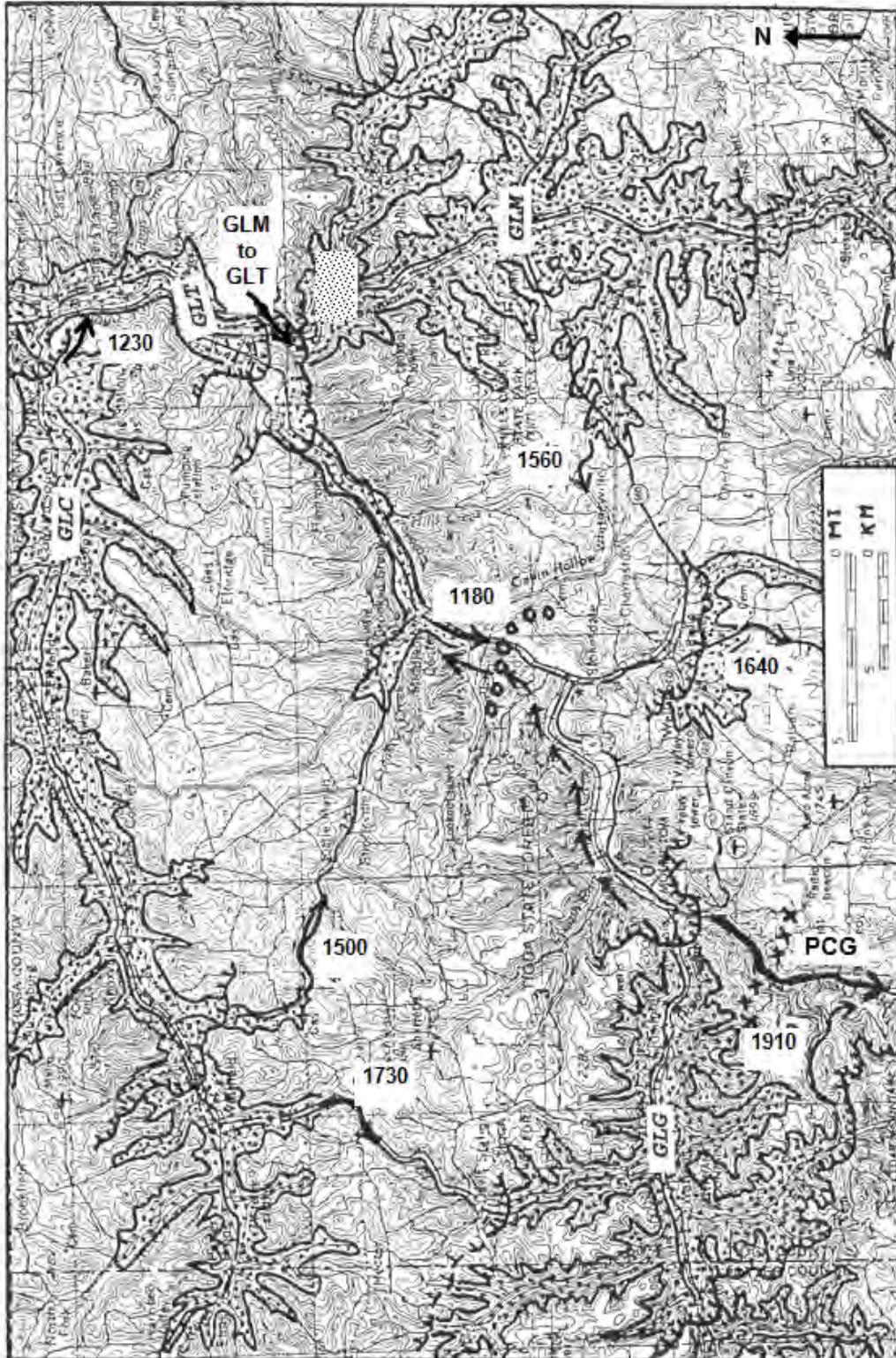Figure 8. Topo map showing the extent of the various proglacial lakes in the region. Sluiceway floor elevations in feet. GLC - Glacial Lake Cowanesque, GLG - Glacial Lake Gaines, GLM - Glacial Lake Mansfield, GLT - Glacial Lake Tioga, PCG - Pine Creek Gorge, .x.x.x.x Preglacial divide at Pine Creek gorge, o o o o Present Marsh Creek - Crooked Creek divide (modified Fig. 14, Gardiner *et al.,* 1993).

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 northeast in the rolling hills lowland that would open up as the ice continued to recede. The Pine Creek Gorge was initiated in the Early Pleistocene, deepened to near its present elevation in the Middle Pleistocene, and cut to its present level in the Late Pleistocene (Fuller and Alden, 1903; Muller, 1957; Crowl, 1981; Braun, 2011b).

The headwater part of the north draining Tioga River valley was occupied by Glacial Lake Mansfield (Figure 8, GLM). That lake had only a single sluiceway between the sluiceway near the head of the drainage and the Crooked Creek (preglacial Pine Creek) – Tioga River confluence due to changes in geology going downstream (Figure 8, arrow labeled 1560). 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 dam site (Stop 11, Figure 6-1). The 1560 feet 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. That 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 stage (Figure 8, 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 (Figure 8, arrow labeled 1180). 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)(Figure 8, line of circles). In the early Pleistocene stage 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 to the Pine Creek gorge and would have been at least 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.

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

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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 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

Fuller and Alden (1903) noted and had a photograph of glacial clays (varves) 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. More detailed 7.5' quadrangle mapping by Braun in the region from 2009 to present (maps available as

open file reports on the PAGS website) has delineated many more slumps and outlined areas where varves should expected to be present. Anywhere within the areas that were covered by the proglacial lakes in the region (Figure 8) 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 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 6).

PROBLEMS FOR MARCELLUS SHALE GAS EXTRACTION 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 gas lines for export from the area. Those plastic gas lines 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 gas lines 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 gas lines 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 gas line 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. The drill pads are being made into "bathtubs" to prevent such contamination, but the large number of tank trucks supplying the frac jobs will from time to time crash a cause localized contamination problems.

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BEDROCK GEOLOGY ALONG PART OF THE US 15 CORRIDOR, EASTERN TIOGA COUNTY, PENNSYLVANIA

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INTRODUCTION

This description of the bedrock geology of part of Tioga County, PA is based on mapping of the Mansfield (McLaurin, 2010) and Blossburg (McLaurin and Dodge, 2012) 7.5-minute quadrangles which was supported by the USGS National Cooperative Geologic Mapping Program. The study area is located within the Glaciated Low Plateau and High Plateau sections of the Appalachian Plateau Physiographic Province (Berg et al., 1989; Briggs, 1999; Sevon, 2000) in the eastern portion of Tioga County, Pennsylvania. This area is along the US 15 / I-99 corridor (Figure 1) and stratigraphic units range from Devonian to Pennsylvanian in age, encompassing marine rocks at the base of the section within the Upper Devonian Lock Haven Formation and extending to the mostly nonmarine, fluvial-deltaic coal measures within the Pennsylvanian Pottsville and Allegheny Formations (Figure 2). It includes part of the Blossburg-Antrim coal field, a subdivision of the North-Central coal fields (Dodge and Edmunds, 2010). Two broad fold structures trend northeast-southwest through the area with the Blossburg syncline occupying the higher elevations near Blossburg and the Wellsboro anticline traversing through the Mansfield area.

PREVIOUS STUDIES

Taylor (1835) first described the general economic geology of the coal field around Blossburg with more extensive geologic studies undertaken by the First Geological Survey of Pennsylvania (1836–58) (Rogers, 1858). Rogers (1858) described the prominent anticlinal feature in the Mansfield area as simply "Anticlinal Axis No. 3" and recognized synclinal "mountains or table-lands" with sandstone escarpments, conglomerate, and coal characteristically forming the "third coal-basin" at Blossburg (Figure 3). The stratigraphic nomenclature utilized by Rogers (1858) for the northeastern district of Pennsylvania, covering the Mansfield-Blossburg area, included in ascending order (oldest to youngest) the Vergent shales, Ponent series, Vespertine series, Umbral red shales, Seral conglomerate, and Coal measures. These units are roughly equivalent to the modern geologic formations recognized in north-central Pennsylvania that include the Lock Haven, Catskill, Huntley Mountain and Burgoon, Mauch Chunk, Pottsville and Allegheny, respectively. Later mapping by Sherwood (1878) as part of the Second Geological Survey of Pennsylvania (1874–95) used a mixed terminology that included the Chemung, Catskill, and Pocono Sandstone for the lower units, but adhered to Roger's (1858) use of Umbral, Seral, and Coal measures for the uppermost stratigraphic units. In addition, Sherwood (1878) described the structural characteristics, defining both the "Wellsborough anticlinal" and "Blossburg Mountain synclinal". (Figure 4). He also provided a detailed description of the coal measures and associated units of the Blossburg coal basin. Subsequent mapping by Fuller (1903) of the Elkland and Tioga 15' quadrangles (Figure 5) 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 use of the term Chemung for the lowermost "Vergent shales" was questioned by Frakes (1963), who noted it relied on biostratigraphic criteria instead of strict lithostratigraphic designation, which is necessary to define formations according to the North.
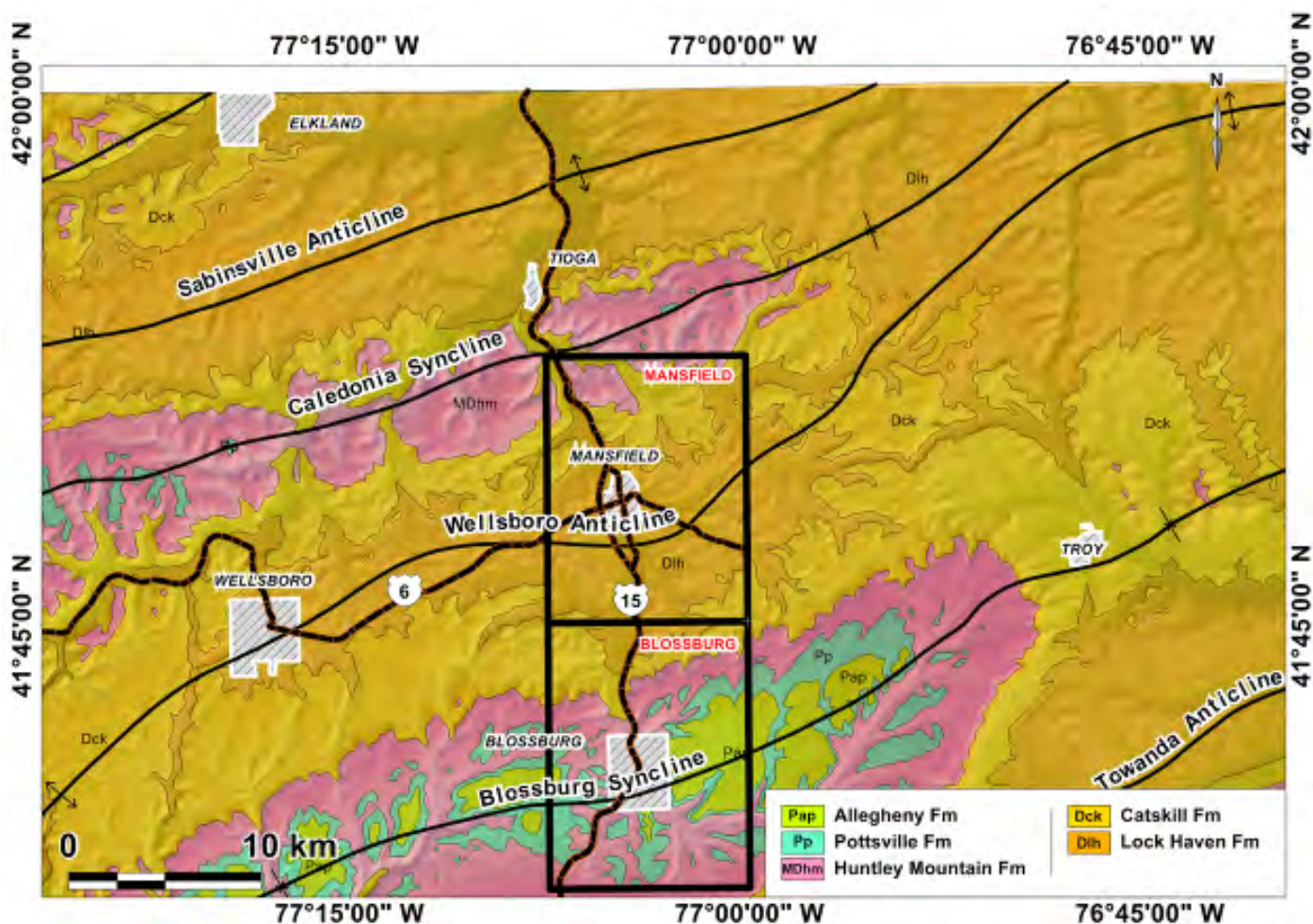
Figure 1. Geology of part of Tioga County with the locations of the Blossburg and Mansfield quadrangles highlighted. Geology taken from Berg et al (1980). Fold axes from Faill (2011).

American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 2005). Thus, he encouraged restriction of the term "Chemung" to a localized time-stratigraphic unit. Faill and et al. (1977) established the Lock Haven Formation to encompass this largely marine stratigraphic interval between the Brallier and Catskill Formations. Further revisions of the stratigraphy were proposed by Berg and Edmunds (1979), who defined the Huntley Mountain Formation to cover the stratigraphy in the transition between the Catskill Formation and Burgoon Sandstone. The use of Huntley Mountain Formation is restricted to the northern part of Pennsylvania and replaces the Pocono Group and Oswayo Formation previously established in this area, as shown on the 1960 "Geologic Map of Pennsylvania" (Gray et al., 1960). The Burgoon Sandstone was previously mapped in the Blossburg area (Berg et al., 1980), but our work suggests that the sandstone and overlying red shale intervals below the Pottsville quartz arenite are more characteristic of the Huntley Mountain Formation than the Mauch Chunk Formation and Burgoon Sandstone. Furthermore, the red shale is considered to correlate lithostratigraphically with the "Patton shales" as described by Colton (1963) and not the Mauch Chunk as previously mapped by Gray et al. (1960). Thus, we believe that the Pottsville Formation disconformably overlies the Huntley Mountain Formation and the Burgoon Sandstone is absent due to erosion (Berg and Edmunds, 1979; Edmunds, 1981). The contact between the coal-bearing Pennsylvanian Pottsville and Allegheny Formations in north-central Pennsylvania is defined as the base of the Cannel coal, which is now considered equivalent to the Brookville (Clarion no. 1) coal of the Main Bituminous coal field of western Pennsylvania. (See Dodge and Edmunds, 2010, for a general discussion of bituminous coal nomenclature and correlations in Pennsylvania.) Accurate mapping of this boundary by fieldwork is not possible due to poor exposures; thus the position of this coal marker is based on correlation

of drill-hole records, identification of coals in abandoned surface mines, and projection by intervals from other key beds.

Figure 2. Composite stratigraphic column for the Tioga county area of the northern tier of Pennsylvania. Modified from Berg et al. (1993).
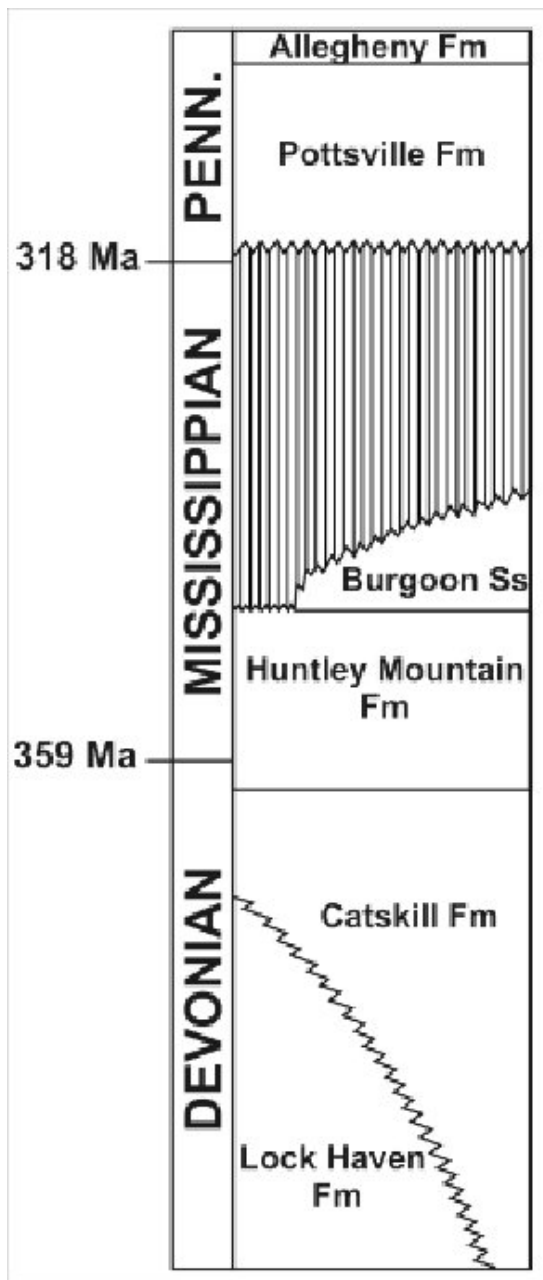
STRATIGRAPHY

The stratigraphic nomenclature applied in this study is consistent with that used for compilation of the geologic map of Pennsylvania (Berg et al., 1980). The stratigraphic units that were mapped include in ascending order (oldest to youngest) the Upper Devonian Lock Haven and Catskill Formations, the Upper Devonian-Mississippian Huntley Mountain Formation, and the Pennsylvanian Pottsville and Allegheny Formations.

Limited field exposures of these stratigraphic units have increased the importance of acquiring and using subsurface data for correlations and geologic mapping. In particular, in the Blossburg-Antrim coal field, exploratory core drilling and drill-hole records from public and private sources was instrumental in the mapping process. Recent Marcellus Shale drilling is generating new, high-quality datasets of geophysical logs and mudlogs, often run from the surface to total depth that can provide much greater resolution and understanding of the bedrock geology. Figure 6 is an example of a gamma ray log from a Seneca Resources well drilled approximately 400 m southwest of field conference Stop 7.

Upper Devonian

Lock Haven Formation

The Lock Haven Formation is the oldest stratigraphic unit exposed within the Mansfield and Blossburg quadrangles. It primarily occupies broad valleys, particularly along the trace of the Wellsboro anticline around Mansfield. Approximately 130 m (430 ft) to 175 m (574 ft) of the uppermost part of the unit is exposed. The total thickness of the Lock Haven Formation in the type section and the Williamsport area is 1181 m (3875 ft) (Faill et al., 1977). Exposures of the Lock Haven within the area are limited to road cuts along new US Route 15 and parallel secondary roads and rare outcroppings along creeks. The Lock Haven Formation is characterized by interbedded gray-green shale and thin, flaggy, very fine to fine-grained, gray to gray-green sandstone that is commonly calcareous (Figure 7A). The sandstones are massive to laminated and may contain abundant brachiopod fossils and bioturbated zones (Figure 7B). The sandstone units are typically less than 1 m (3 ft) in thickness, but can locally exceed 2 m (7 ft). Shale intervals are up to 8 m (26 ft) thick, but when interbedded with sandstone are 50 cm (1.6 ft) to 1 m (3 ft) thick. Sherwood (1878) describes hematitic sandstone that was mined for iron ore in the vicinity of Mansfield from the upper parts of the Lock Haven Formation.
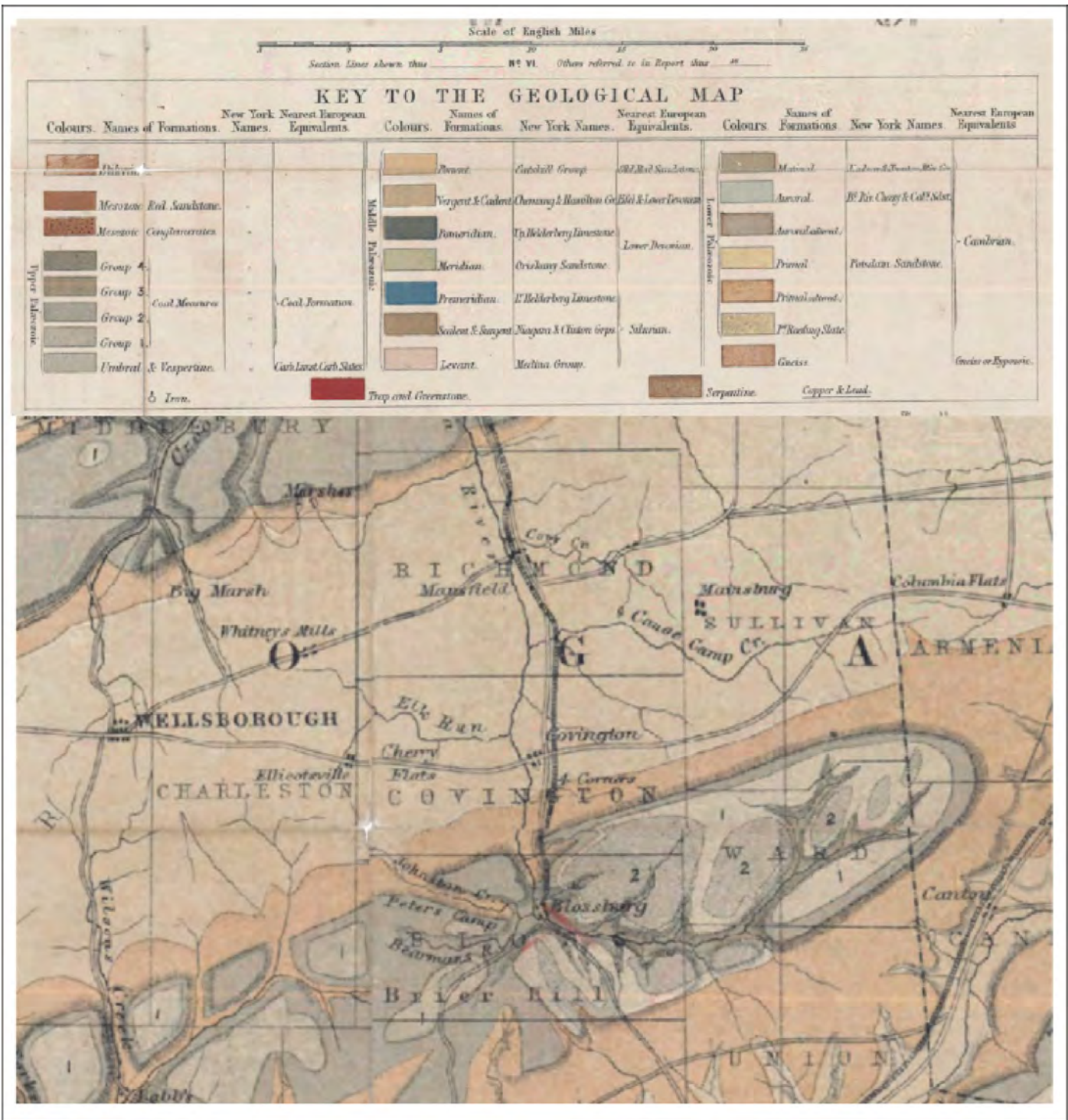
Figure 3. Geologic map of the Tioga County area from the 1858 Geologic Map of Pennsylvania (Rogers, 1858)

Catskill Formation

Road cuts on the east side of US 15N between Blossburg and Mansfield provide the best exposures of the middle to upper parts of the Catskill Formation. The lower 50 to 80 m (160 to 260 ft) of the formation is poorly exposed, and the contact with the underlying Lock Haven Formation is not observed. This transition is present further north in the Jackson Summit and Tioga quadrangles and is the focus of field conference Stop 9. The total stratigraphic thickness of the Catskill Formation in the area thins northward
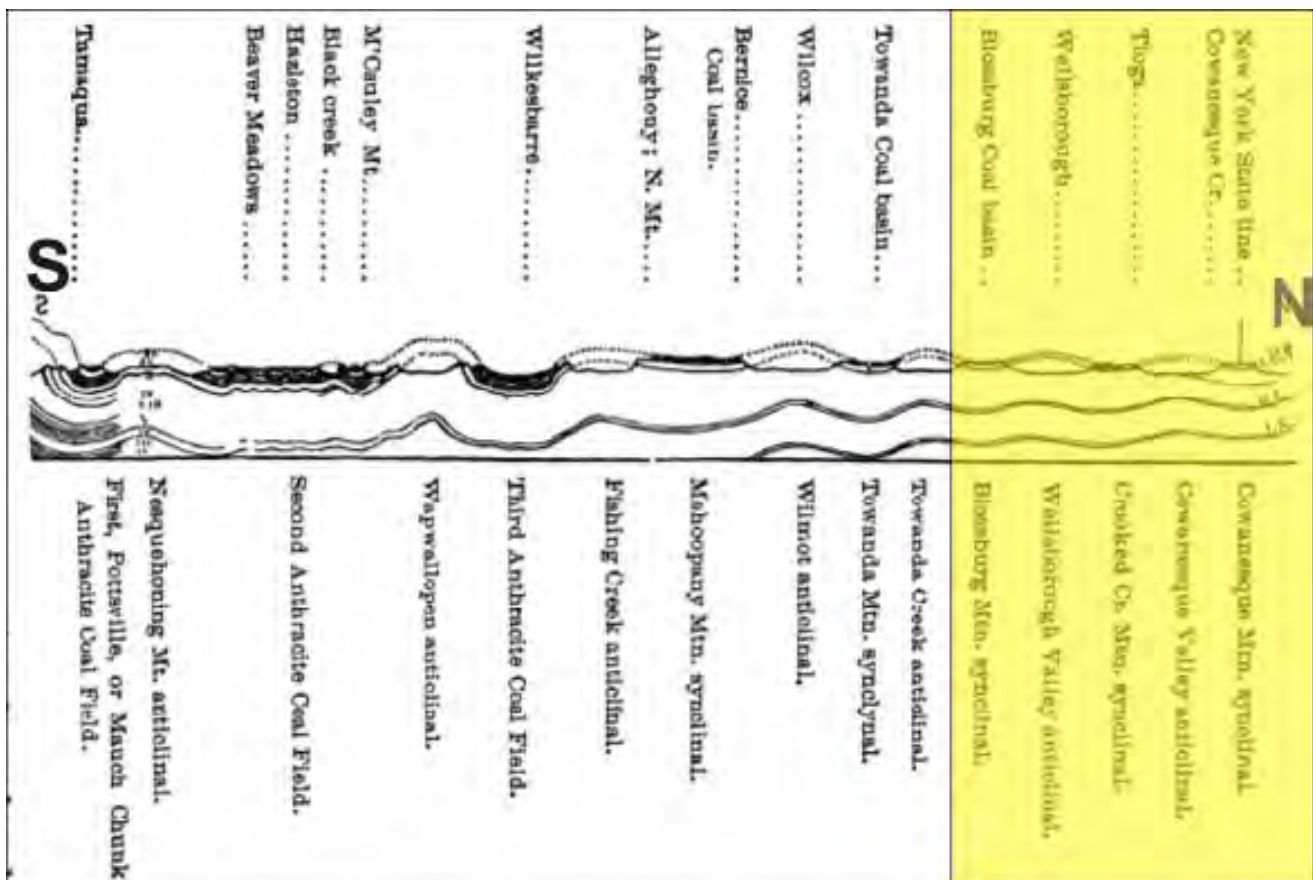
Figure 4. Schematic cross section across Appalachian Plateau folds. From Sherwood (1878). Yellow shading covers area from Blossburg quadrangle, northward to the Pennsylvania-New York state line.

from approximately 312 m (1020 ft) in the Blossburg quadrangle to 260 m (850 ft) measured in the Mansfield quadrangle (McLaurin, 2010). By comparison, the Catskill appears to gradually thicken southward to the Williamsport area, where it is up to 610 m (2000 ft) thick (Faill et al., 1977).

The lowermost Catskill Formation is placed at the first red mudstone or shale that marks the beginning of an interval of approximately 80 m (260 ft) of alternating red Catskill-type sandstone and shale with gray-green Lock Haven-type sandstone and shale (Figure 8A). This transitional zone, observed at field conference Stop 9, 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 et al., 1977). Within this zone, there are five of these red/gray-green couplets that reach a maximum thickness of 24 m (80 ft). The sandstone within this interval is fine grained with rip-up clasts and cross-lamination (Figure 8B). Interbedded with the sandstone are heterolithic layers containing 1- (0.4-in.-) to 2-cm- (0.8-in.-) thick sandstone and 3- (1-in.-) to 5-cm- (2-in.-) 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 sandstones that are arranged in fining-upward cycles, culminating in finer-grained interbedded sandstones and shales. This part of the Catskill Formation is exposed in a road cut along new US Route 15N, just north of Blossburg and approximately 130 m (430 ft) thick (Figure 9A). Sandstone units vary in thickness from less than 1 m (3 ft) to greater than 6 m (20 ft). Interbedded sandstone and shale units are up to 24 m (79 ft) thick (Figure 9B). This interval within the Catskill is similar in appearance to the Sherman Creek Member, although it is considerably thinner than has been mapped in other areas (Faill et al., 1977). The uppermost part of the Catskill Formation is characterized by gray-red, fine- to medium-grained, cross-bedded sandstone (Figure 10A). These sandstones
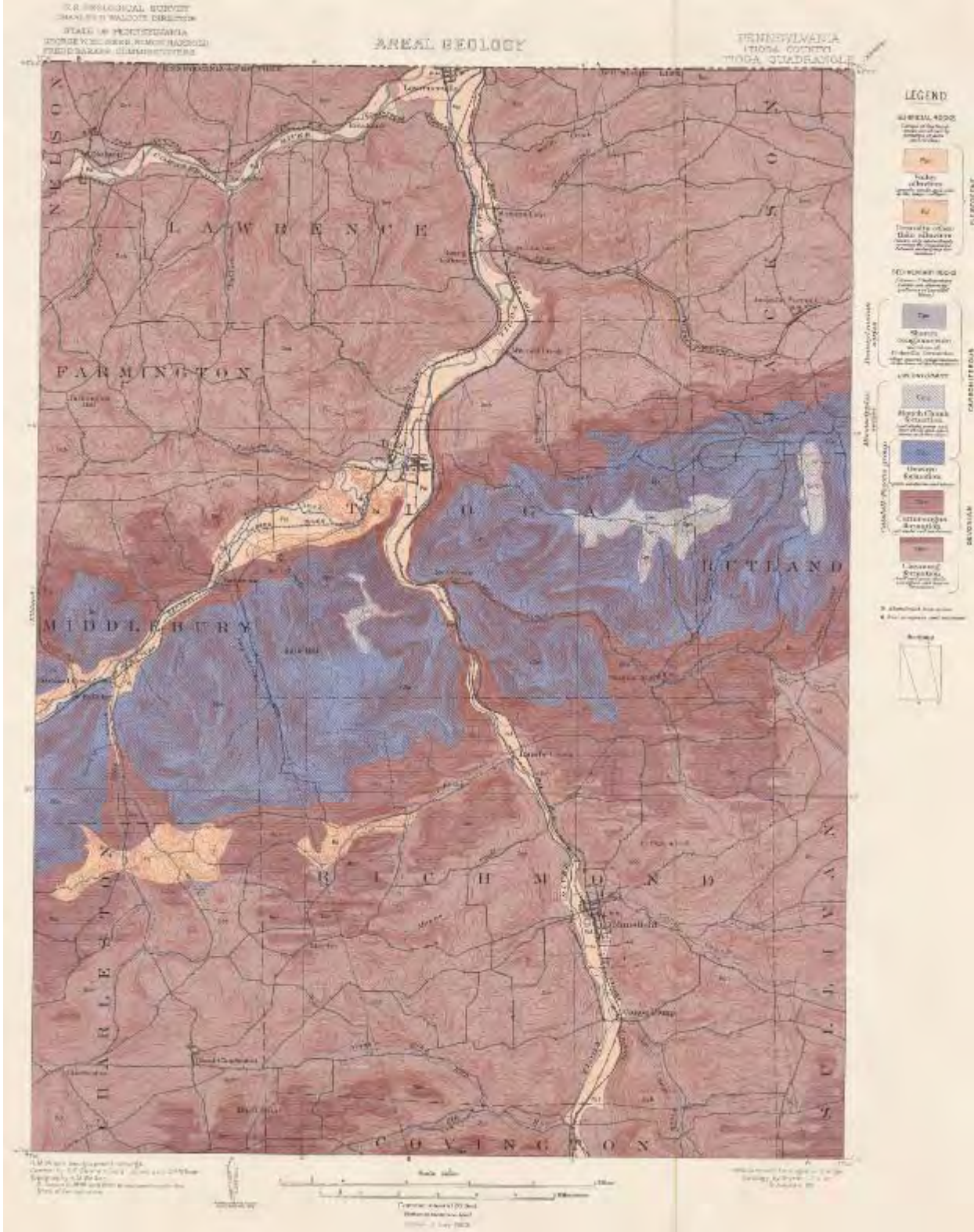
Figure 5. Geologic map of the Elkland -Tioga area from Fuller (1903).

are locally amalgamated into larger sheet sandstones that exceed 10 m (33 ft) in thickness (Figure 10B). Interbedded red, clayey silt and shale units intercalated with thin, very fine grained sandstones can exceed 10 m (33 ft) in thickness as well.
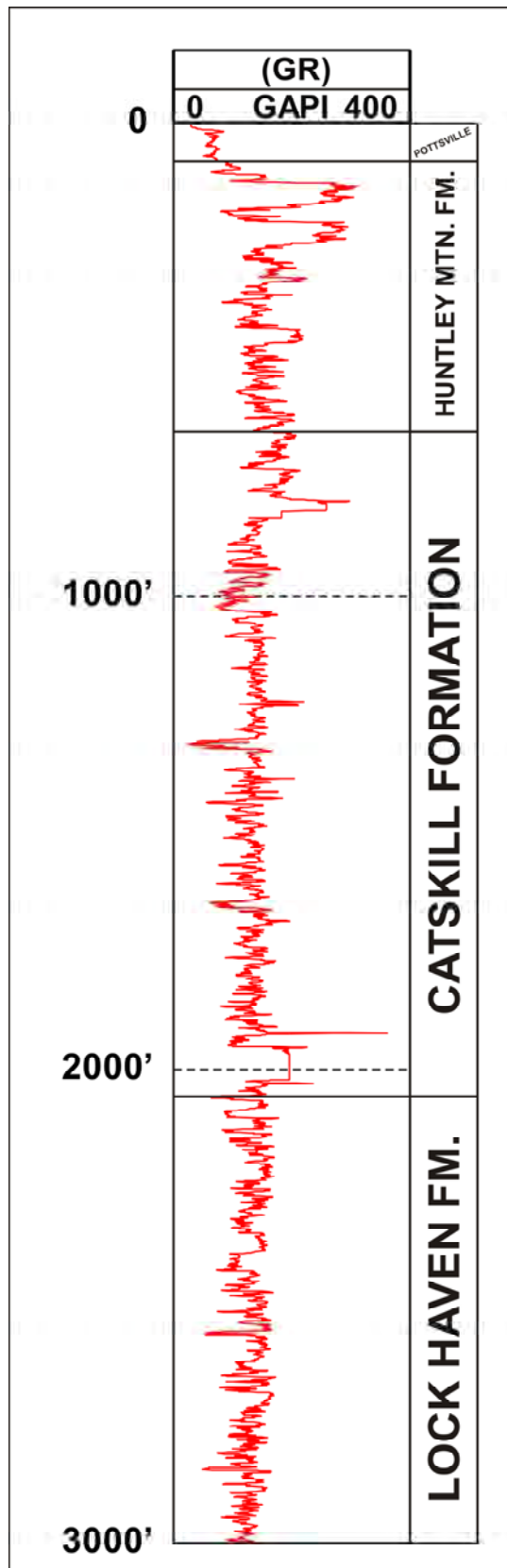
Figure 6. Representative gamma ray log of the Lock Haven - Pottsville Formation interval from the Seneca Resources DCNR 595 70v well. Printed with permission of Seneca Resources Corporation.
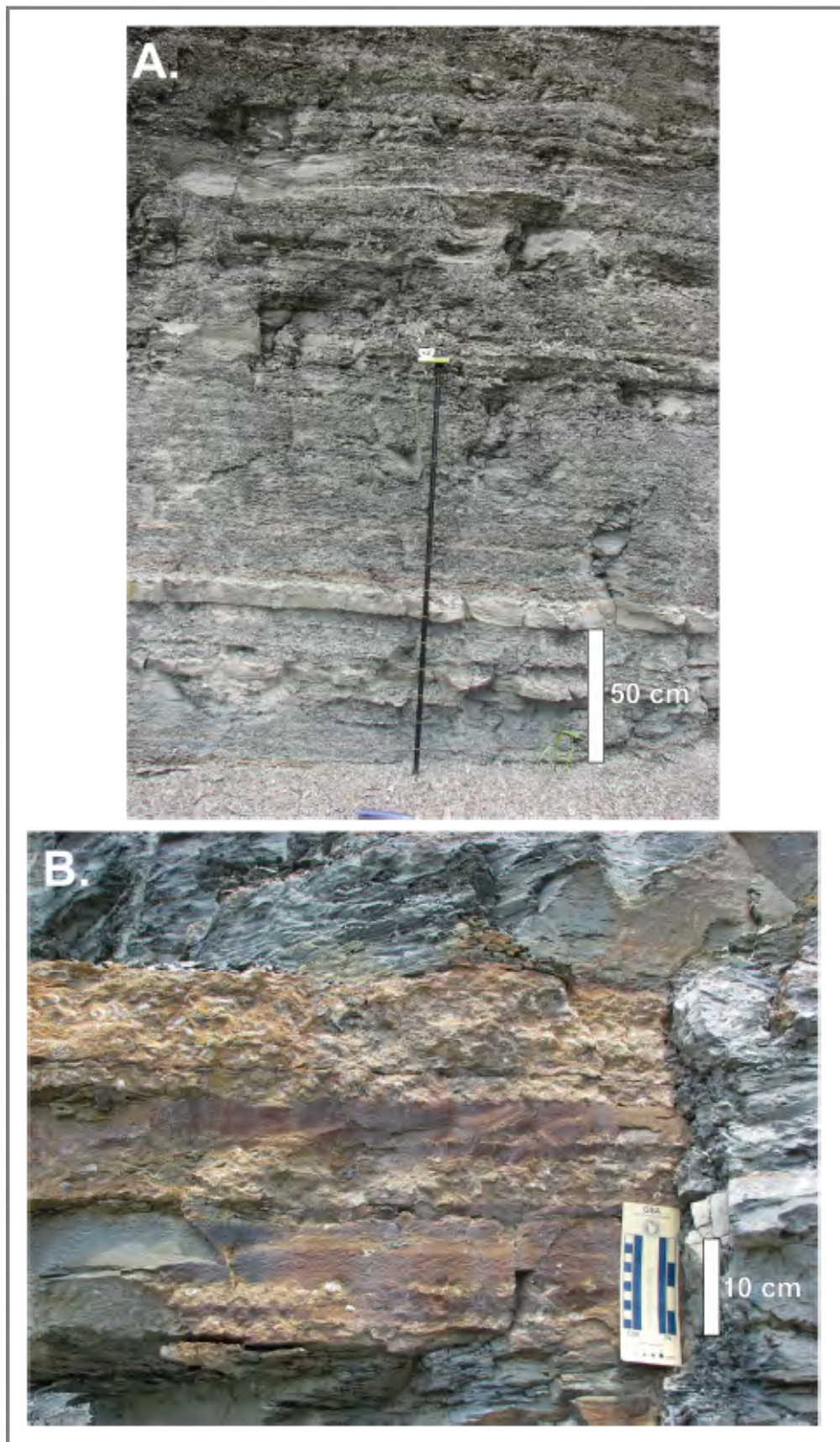
Figure 7. A. Interbedded sandstone and shale of the upper part of the Lock Haven Formation. B. Zones of brachiopod shell-lag within the Lock Haven Formation.

Figure 8. A. Contact between the Lock Haven and Catskill formations. B. Cross-laminated sandstone within the lowermost Catskill Formation.

Figure 9. A. Middle portion of the Catskill Formation along US 15, just north of Blossburg. B. Shale zones within the middle part of the Catskill Formation.

Figure 10. A. Overview of the upper part of the Catskill Formation, just below the contact with the Huntley Mountain Formation. B. Cross-bedded sandstone in the upper part of the Catskill Formation.

Figure 11. A. Sheet sandstone bodies with intercalated finer-grained gray shale. B. Channel-bodies within the Huntley Mountain Formation.

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.

Upper Devonian–Mississippian

Huntley Mountain Formation

The Huntley Mountain Formation forms steeper slopes compared to the underlying Catskill and Lock Haven Formations and is much sandier. It is approximately 160 m (525 ft) to 175 m (569 ft) thick. The contact between the Catskill and overlying Huntley Mountain was defined 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 11A and 11B). The sandstone occurs as amalgamated sheets that are up to 10 m (33 ft) thick, and average 5 to 6 m (16 to 20 ft). Gray shale to siltstone intervals are not readily observed in outcrop, but are commonly present in road cuts where they rarely exceed 3 m (10 ft) in thickness. There is a succession of green and red shale approximately 60 to 80 m (200 to 260 ft) above the base of the Huntley Mountain Formation that is 5 m (16 ft) thick. This shale unit may be equivalent to the "Mount Pleasant" red shale of Ebright (1952). Sandstones contain a high density of erosional surfaces, and where shale intervals are observed, they commonly cannot be traced more than a few meters (yards) laterally before they are cut out by overlying channel sandstones. The lowermost parts of the Huntley Mountain Formation contain finer-grained zones that are composed of interbedded very fine grained sandstone and gray siltstone to shale. A distinctive, uppermost interval of Huntley Mountain is 15 to 20 m (49 to 66 ft) thick and includes interbedded red shale and gray to buff sandstone that occurs as fine- to medium-grained and very fine to fine-grained beds, 50 to 80 cm (1.6 to 2.6 ft) thick (Figure 12).

Figure 12. Red shale in the upper part of the Huntley Mountain Formation, just below the Pottsville Formation.

Pebbly zones are present in the sandstone and consist of up to 1-cm- (0.4-in.-) diameter red chert pebbles. This red shale and sandstone interval in the uppermost Huntley Mountain Formation is considered

correlative to the informal Patton red beds as described by Colton (1968) and is truncated by overlying qurtz arenite of the Pottsville Formation.

Pennsylvanian

The Pottsville and Allegheny Formations of Pennsylvanian age are present in the Blossburg quadrangle. They commonly underlie the upland areas within the Blossburg syncline (Blossburg-Antrim coal field) and have been covered in many places by glacial deposits of variable thickness (Duane Braun, personal communication). The lack of exposures and poor quality of outcrops within these formations make it impossible to identify the Cannel coal (Brookville coal equivalent) and other key beds from fieldwork alone. Therefore, the contact between the Pottsville and Allegheny Formations was determined from subsurface data, using exploratory drill-hole records and other sources for correlation. Although eight or nine coal seams have been identified within the quadrangle (Rogers, 1858; Sherwood, 1878; Dodge and Edmunds, 2010), none were observed in outcrop. The coal nomenclature of this report is consistent with that commonly used in the North-Central coal fields (Dodge and Edmunds, 2010).

Pottsville Formation

The Pottsville Formation is characterized by buff, medium- to coarse-grained sandstones (quartz arenites), 2 to 3 m (7 to 10 ft) thick (Figure 13). The sandstones may contain medium conglomeratic interbeds in places, up to 30-cm (1 ft) thick and having quartz clasts up to 1 cm (0.4 in.) in diameter. The sandstone units are commonly trough cross-bedded in sets 3 to 20 cm (1 to 8 in.) thick and may also contain interbeds of gray shale up to 10 cm (4 in.) thick. Thicker fine-grained units are typically, gray-black, paper-thin shale, ranging in thickness from 1 to 6 m (3 to 20 ft), and contain interbedded buff-gray, fine- to medium-grained cross-bedded sandstone. Dodge (1995) has observed rare, restricted-marine, invertebrate fossils (*Lingula* sp.) in core from a Pennsylvania Geological Survey drill hole located about 1.6 km (1 mi) east of the Blossburg quadrangle near Fall Brook. The fossils occur within medium-dark-gray siltstone, about 6 m (20 ft) above the Bloss coal. The thickness of the Pottsville Formation is approximately 48 m (157 ft).

The coal seams present in the study area within the Pottsville Formation include in ascending order (oldest to youngest) the

Figure 13. Conglomeratic sandstone within the lower part of the Pottsville Formation.

Kidney, Bear Creek, and Bloss (Figure 14). The Kidney and Bear Creek coals are highly lenticular, generally ranging in thickness from 0 to <0.8 m (0 to <2.5 ft), and were rarely mined in the past. The Kidney and Bear Creek occur about 18 to 21 m (60 to 70 ft) and 6 to 9 m (20 to 30 ft), respectively, below
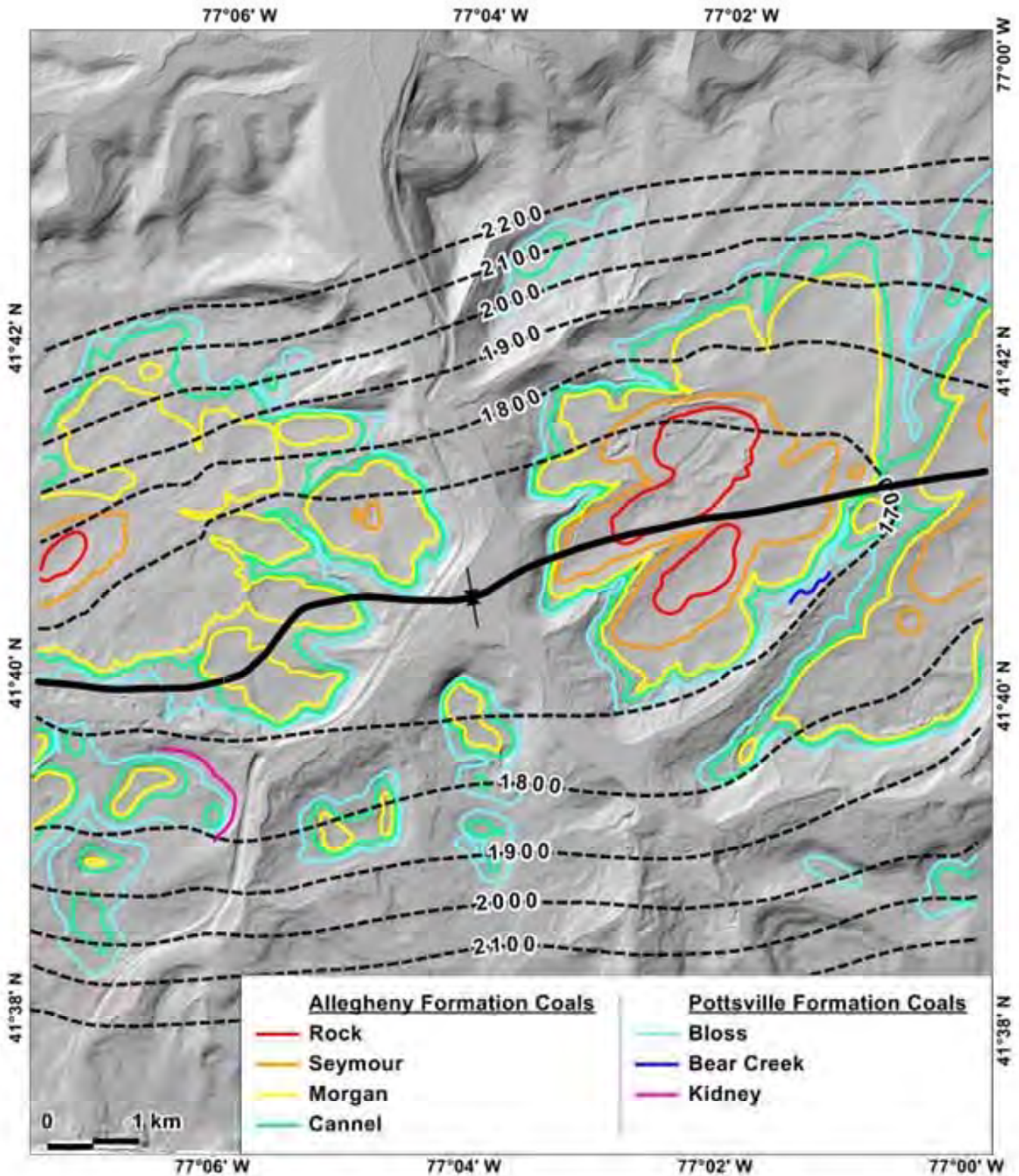

Figure 14. Structure-contour map of the base of the Bloss coal with the locations of coal seams within the Pottsville and Allegheny Formation. Contours in feet relative to mean sea level. Modified from McLaurin and Dodge (2012).

the Bloss. The Bloss coal was formerly the most important economic seam of the area, where it was extensively deep and strip mined. The Bloss generally ranges in thickness from about 0.8 to 1.5 m (2.5 to 5 ft), but it locally splits into two benches that may be separated from one another by as much as 1.5 to 3 m (5 to 10 ft). Each bench varies in thickness from about 0.6 to 1.1 m (2 to 3.5 ft), but where one bench is thicker (i.e., toward the upper limit of its thickness range) the other is thinner. The Bloss coal is the most important key bed within the coal measures for correlation and was used for structure contouring the coal basin (Blossburg syncline). The Bloss coal is now considered equivalent to an Upper Mercer coal of the Main Bituminous coal field.

Allegheny Formation

The Allegheny Formation contains cyclic sequences of sandstone, siltstone, shale, claystone (paleosol), and coal. Conglomerate occurs in places as well. Several coals are present that were economically important in the past. Dodge (1995) has observed dark-gray shale overlying the Seymour coal and containing marine invertebrate fossils and siderite concretions. The marine fossils have been identified in talus from abandoned strip-mine highwalls just beyond the eastern edge of the quadrangle and include *Chonetes* sp., *Dunbarella* sp., and *Lingula* sp. This marine shale represents the Columbiana marine zone that over lies the Lower Kittanning coal of the Main Bituminous coal field. (See Dodge, 1995; Dodge and Edmunds, 2010.) As mapped, the Allegheny Formation is confined primarily to the upland areas in the middle of the quadrangle, toward the deepest parts of the Blossburg syncline. The estimated maximum preserved thickness of the Allegheny Formation is about 72 m (236 ft).

The minable coals of the Allegheny Formation in the Blossburg quadrangle include in ascending order (oldest to youngest) the Cannel, Morgan, Seymour, and Rock (Figure 13). The Cannel coal generally ranges in thickness from 0.3 to 0.6 m (1 to 2 ft), though it is locally absent. It has been surface mined in places in the past. The Cannel lies about 9 to 12 m (30 to 40 ft) above the Bloss coal. As previously stated, the Cannel coal is now considered equivalent to the Brookville coal of the Main Bituminous coal field. The Morgan coal varies in thickness from about 0.5 to 0.9 m (1.5 to 3 ft), though it is thin or absent in places. It has formerly been both surface mined and deep mined locally. The Morgan coal is situated about 15 to 21 m (50 to 70 ft) above the Bloss. It is equivalent to the Upper Clarion coal of western Pennsylvania. The Seymour coal is about 0.6 to 0.9 m (2 to 3 ft) thick and has been mined by both surface and underground methods in the past. It occurs about 40 to 46 m (130 to 150 ft) above the Bloss coal and is areally restricted to the higher elevations of the uplands within the deeper parts of the Blossburg syncline. It is equivalent to the Lower Kittanning coal of the Main Bituminous coal field. The Rock coal is about 0.6 to 0.8 m (2 to 2.5 ft) thick. It was previously strip mined and is nearly all mined out. It is situated around 52 to 55 m (170 to 180 ft) above the Bloss. The extent of the Rock coal is limited to several hilltops near the axis of the syncline. The Rock is equivalent to the Middle Kittanning coal of the Main Bituminous coal field.

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. The prominent structural features within the area are the Blossburg syncline to the south and the Wellsboro anticline to the north. North of the Wellsboro anticline is the Caledonia syncline and south of the Blossburg syncline is the Towanda anticline (Faill, 2011).

Wellsboro Anticline

Mapping of the axis along the Wellsboro anticline proved problematic due to poor outcropping of the Lock Haven Formation in the Mansfield area. Fuller (1903) illustrates the fold axis orientation, of the Wellsboro anticline, as variable, trending 90° - 94° in the western part of the quadrangle, approximately 74° in the central part of the quad where it crosses just north of Canoe Camp. East of Mansfield the axis is

shown as swinging to a 44° trend. Field mapping within the Mansfield quadrangle suggests that the fold axis is approximately 1 km south of where it is mapped by Fuller (1903) as it enters the western part of the study area. Its trend is consistent at approximately 72°. No sense of fold plunge could be ascertained from field examination. The transition to a more northeasterly orientation as indicated by Fuller (1903) in the eastern areas could not be verified due to the lack of suitable outcrop. Examination of large scale SRTM (shuttle radar topography mission) data does not support a significant deviation of the fold axis and thus an orientation of 72° is maintained into the Roseville quadrangle to the east.

Blossburg Syncline

Mapping of the axis of the Blossburg syncline proved more successful due to an extensive dataset of drill hole and mine data from coal mining activities. Using the top of the Bloss coal seam within the Blossburg quadrangle indicates that the fold axis is oriented along a trend of 76°–256°. The axis is projected to pass through the village of Arnot, the southern part of Blossburg, and just north of Morris Run. The structure-contour mapping indicates that the Blossburg syncline is a doubly plunging syncline. In the western half of the quadrangle, the syncline plunges to the east at about 5.1 m/km (27 ft/mi). In the east half of the quadrangle, the fold plunges westward from as little as 2.1 m/km (11 ft/mi) near Blossburg to about 15 m/km (78 ft/mi) to the north of Morris Run. The structurally lowest part of the fold axis trends through the Borough of Blossburg and crosses U.S. Highway 15 near the mouth of Boone Run. Strike and dip measurements show that the strata, on average, dip from 4° to 6°. Field measurements and data from drill holes and underground coal-mine maps indicate that the north limb of the syncline dips slightly steeper than the south limb. This observation is consistent with that of Sherwood (1878) and Rogers (1858), who noted that the anticlinal limbs show steeper southerly dips and more shallow northerly dips. Along the north limb of the syncline, beds dip south-southeast at approximately 53 m/km (280 ft/mi), whereas along the south limb, beds dip north to north-northwest at about 44 m/km (230 ft/mi). The fold height (structural relief), measured from the bottom of the Blossburg syncline to the top of the Wellsboro anticline in the Mansfield quadrangle, is about 442 m (1450 ft). The wavelength of the Blossburg syncline is approximately 25 km (15-1/2 mi), measured between the axes of the flanking folds—the Wellsboro and Towanda anticlines.

In addition to the Blossburg syncline, a secondary feature was identified in the northwestern part of the quadrangle, just west of Covington. Here, it was observed that the dips of the strata within the Lock Haven and Catskill Formations show a deviation from the regional trend. Instead of dipping to the southeast, the dip directions range from northwest to southwest. Although outcrops within the Catskill Formation are lacking, a fold trend can be approximated for this limited area. This feature, herein termed the Covington anticline, is limited to the area west of the Tioga River. Exposures on the east side of the river do not show the deviations in dip direction and instead have consistently southeast dips. The axis of the Covington anticline can be inferred to trend northeast–southwest to the south of Covington before turning to a more northwesterly trend. Dip values are similar to most of the other dips and range from 4° to 6°. Further work is required to more adequately define this feature over a broader area.

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COAL GEOLOGY AND MINING HISTORY OF THE BLOSSBURG–ANTRIM COALFIELD, SOUTHEASTERN TIOGA COUNTY, PENNSYLVANIA

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INTRODUCTION

Recent development of the Marcellus Shale gas play in north-central Pennsylvania has led to renewed interest by geologists and others in the bedrock geology of the region. As a result, the Pennsylvania Geological Survey, Department of Conservation and Natural Resources, has undertaken several new studies, involving geologic mapping and drilling, for parts of Tioga and Bradford Counties (e.g., McLaurin, 2011; McLaurin and Dodge, 2012; Behr and Hand, 2013; and Dodge and Markowski, in press). A focus of this endeavor has been the examination of the once-famous Blossburg–Antrim coalfield in southeastern Tioga County. Investigations in the Blossburg–Antrim coalfield have created a greater understanding and appreciation of the coal geology and mining history, which have important regulatory implications for Marcellus drilling under the Pennsylvania Oil and Gas Act of 1984 (Act 224), with Amendments (e.g., special drilling and casing requirements for workable coal seams, mine voids, and groundwater protection), and for future additional mine permitting, mine reclamation, and acid-mine-drainage remediation by state government, in cooperation with private partnerships.

The historical importance of the Blossburg–Antrim coalfield cannot be overstated, as it was the source of the famous Bloss (Blossburg) coal, which contributed greatly to the economic development and vitality of north-central Pennsylvania for nearly a century (Figure 1). The Bloss was especially noteworthy for its production, quality, and versatility. Between 1815 and 2012, the estimated total production of coal from the Blossburg–Antrim coalfield is about 72 million net tons, at least three-quarters of which probably came from the Bloss seam alone. Nevertheless, this part of the story cannot be separated from past mining practices during an era of relatively little or no government oversight (i.e., nineteenth through mid-twentieth centuries) that resulted in a legacy of unreclaimed mine lands and acid-mine drainage, creating significant mine hazards and water pollution that have only been addressed to a large extent in recent decades. At present, there is no active coal mining in Tioga County on a commercial scale (only some minor incidental mining associated with expansion of the Phoenix Resources Landfill near Antrim).

This purpose of this paper is to provide an overview of current knowledge of the geology and past-mining activities of this fascinating and underappreciated coalfield.

LOCATION

The Blossburg–Antrim coalfield of southeastern Tioga County is named for two of the principal former mining communities situated toward the east (Blossburg) and west (Antrim) ends of the field. It is the largest of four small bituminous coalfields comprising the North-Central coalfields (Figure 2). The other three fields include the Little Pine Creek or English Center (northwestern Lycoming County), Ralston (northeastern Lycoming County), and Barclay (southwestern Bradford County). The North-Central coalfields are confined to a series of small detached structural basins or closed synclines.
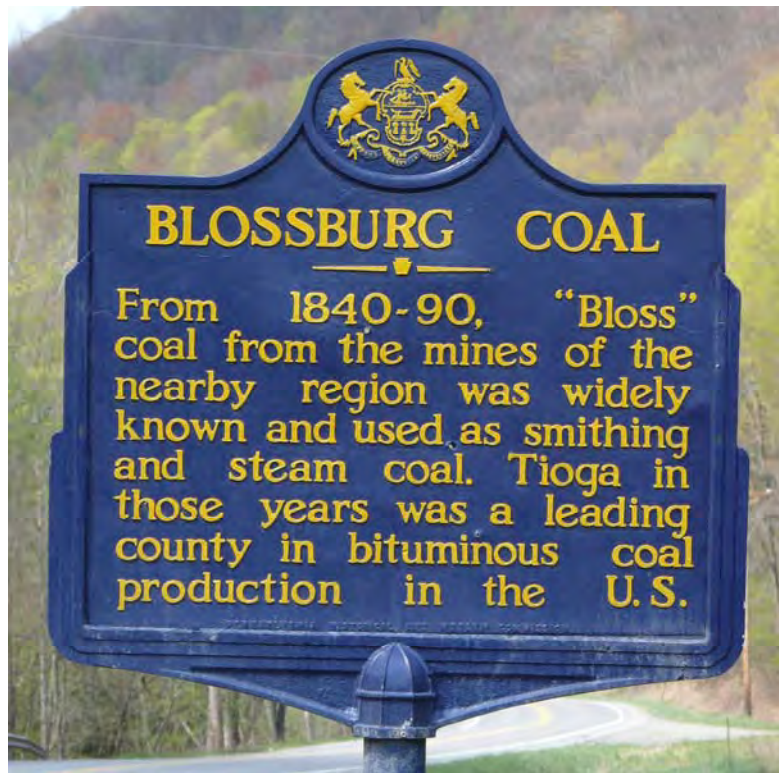
Figure 1. Marker explaining historical significance of Bloss (Blossburg) coal. Erected by Pennsylvania Historical and Museum Commission near Antrim, Pennsylvania.

The Blossburg–Antrim coalfield is about 23 mi long by 2-1/2 to 4 mi wide and covers an area of approximately 60 mi^2. It trends N60–80°E/S60–80°W. The coalfield is confined to upland areas and upper valley slopes within portions of (from east to west) the Gleason, Blossburg, Cherry Flats, Antrim, and Morris 7.5-minute topographic quadrangles and covers parts of (from east to west) Ward, Hamilton, Covington, Bloss, Duncan, Delmar, and Morris Townships. It is bounded to the north by a dissected ridge that is variously named (from east to west) Armenia Mountain, Pine Hill, Maple Hill, Goat Ranch Hill, and East Hill, and to the south by a similar ridge differently named (from east to west) Armenia Mountain, Huckleberry Mountain, Brier Mountain, Bloss Mountain, Hickory Ridge, and Tannery Hill. The upper reach of the Tioga River trends parallel to the coalfield along its south side toward the east end, and Babb Creek runs parallel along the south side near Morris at the west end of the field. Several streams transect the coalfield, including (from east to west) the Tioga River at Blossburg, Babb Creek near Landrus, Wilson Creek near Antrim, and Stony Fork.

PHYSIOGRAPHY

The Blossburg–Antrim coalfield is mostly in the Glaciated High Plateau Section of the Appalachian Plateaus Physiographic Province. The section is characterized by broad to narrow, rounded to flat, elongated uplands and shallow valleys, which here follow the structural axis of the basin (Sevon, 2000). The topographic form of the section is chiefly the result of fluvial and
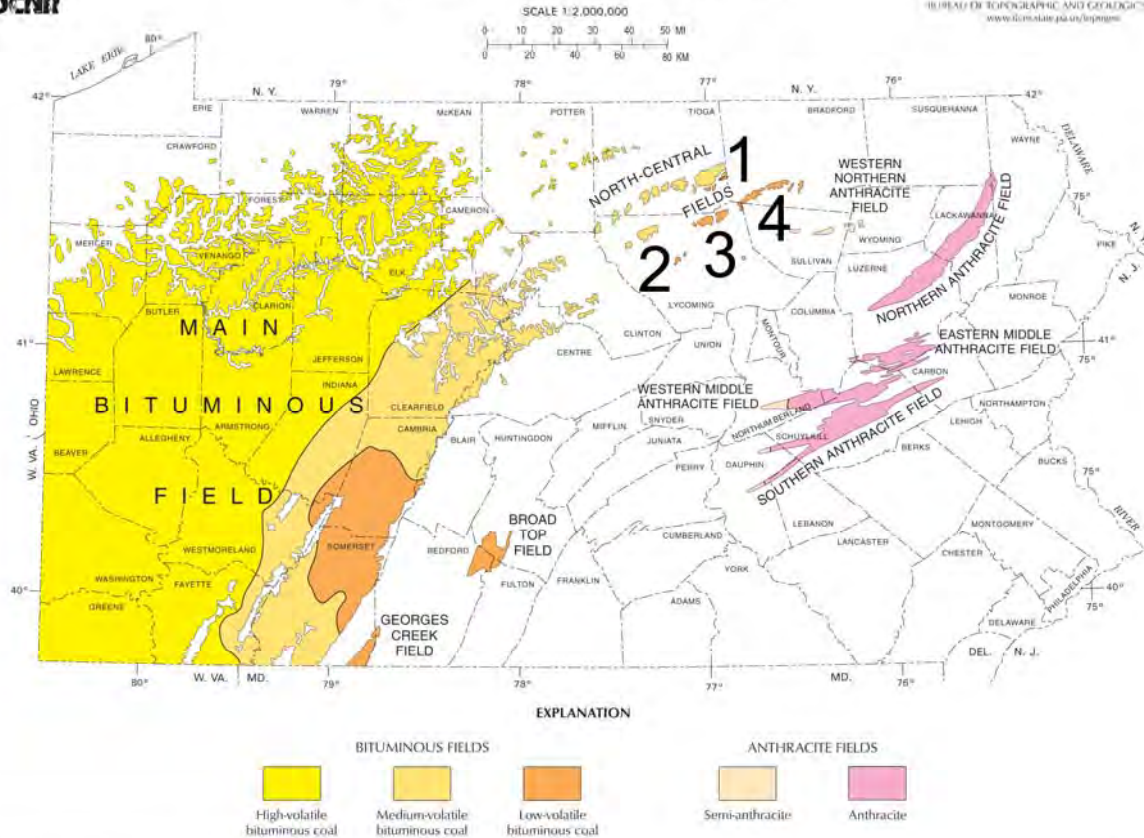
Figure 2. Map showing distribution of coals in Pennsylvania. Bituminous coalfields comprising North-Central coalfields are as follows: 1–Blossburg–Antrim, 2–Little Pine Creek (English Center), 3–Ralston, 4–Barclay. (Modified from Pennsylvania Geological Survey, 2000.)

glacial erosion, coupled with glacial deposition. Depth of unconsolidated (glacial) cover to bedrock may be as much as 60 ft or more locally. Local relief exceeds 950 ft in places.

Toward its west end, the Blossburg–Antrim field lies within the Deep Valleys Section of the Appalachian Plateaus Physiographic Province. The section is distinguished by very deep, angular valleys, with some broad to narrow uplands (Sevon, 2000). The boundary between the two physiographic sections in the coalfield occurs in Babb Creek valley near Landrus and Wilson Creek valley near Antrim. The topographic form of the Deep Valleys Section is the result of fluvial erosion and periglacial mass wasting. Thin glacial deposits occur locally in the upland areas. Local relief may reach 1000 ft in a few places. The Late Wisconsinan glacial border is only about 3 mi west of the terminus of the coalfield (Berg and others, 1980; Sevon and Braun, 1997).

COAL GEOLOGY

Previous Investigations

In 1832, R.C Taylor (1789–1851), a highly regarded English geologist living in the United States, was employed by the Tioga Improvement Company to conduct a detailed geologic survey of the coal and iron-ore beds around Blossburg and to determine the best route for a railroad from Blossburg to the New York State Line. His survey report, one of the first of its kind in North America, was published the following year and included eight measured sections (partly leveled) and detailed text that contained a description of the thickness and quality of the coals and ores examined and an estimate of the coal resources

for the coal basin (Taylor, 1833). He later summarized some of these finding in the Transactions of the Geological Society of Pennsylvania (Taylor, 1835), with the assistance of Clemson (1835) for several analyses of coal and iron ore.

During the field season of 1839, J. T. Hodge (1816–71), assistant geologist of the First Geological Survey of Pennsylvania (1836–58), and a volunteer assistant examined the Blossburg–Antrim coalfield or "Third Coal-Basin" of Rogers (1840, p. 166–174; 1858a, p. 519–524). Sections were measured or obtained from mining companies operating in the area, and rocks were correlated. The most detailed descriptions of the geology were made in the vicinity of the active mining district around Blossburg.

In 1841, while touring the United States, Sir Charles Lyell (1797–1875), the preeminent English geologist, visited Blossburg as the guest of Dr. Lewis Saynisch, president of the Arbon Coal Company (Meginness and others, 1897). Lyell was impressed by his visit to the mines and by the geology he saw. He later recounted that "it was the first time I had seen the true 'Coal' in America, and I was much struck with its surprising analogy in mineral and fossil characters to that of Europe—the same white grits or sandstones as are used for building near Edinburgh and Newcastle—similar black shales, often bituminous, with the leaves of ferns spread out as in a herbarium, the species being for the most part identical with British plant fossils—seams of good bituminous coal, some a few inches, others several yards in thickness—beds and nodules of clay iron-stone; and the whole series resting on a coarse grit and conglomerate, containing quartz pebbles, very like our Millstone Grit, and often called by the American as well as the English miners the 'Farewell Rock,' because when they have reached it in their borings, they take leave of all valuable fuel. Beneath this grit are those red and grey sandstones already alluded to as corresponding in mineral character, fossils, and position, with our 'Old Red'" (Lyell, 1845, p. 49–50). Based on his observations of the similarities of the coal measures at Blossburg and in South Wales, Lyell was the first to suggest some causal connection between the two in time and space.

Rogers (1858b) released the first geologic map to include the outcrop pattern of the Blossburg–Antrim coalfield in some detail.

Macfarlane (1875) published a detailed account on the history of development and mining of the Blossburg coal region. He also included measured sections and coal descriptions. Most of the local names of the principal coals discussed in his publication are still in use today.

Sherwood (1878), a member of the Second Geological Survey of Pennsylvania (1874–95), authored a geologic map of Tioga County that included a more detailed depiction of the Blossburg–Antrim coalfield. He also defined the "Blossburg Mountain synclinal," in which the coalfield is situated, describing its structural characteristics and comparing its geometry with that of other folds in the region. Platt (1878), another member of the Second Survey, presented the first comprehensive description of the geology of the coalfield and published numerous measured sections of the coal-bearing rocks. He also discussed in detail the coal geology of each active mining operation, focusing especially on the Bloss coal. Furthermore, he presented many coal analyses and described the character, quality, and uses of the Bloss and other identified coals, including several that were not actively mined at the time.

Thereafter, for almost a century, little new work was published on the geology of the coalfield. Most investigations were directed toward data collection for coals—measuring sections and sampling and analyzing coals from underground- and surface-mine operations (e.g., U.S. Bureau of Mines, 1939; Sponseller, 1973; Skema and others, 1977).

As part of the Pennsylvania Operation Scarlift Program, a feasibility study was made to determine the mine drainage abatement measures needed to restore abandoned mine lands in the Tioga River watershed near Blossburg and Morris Run (Gannett Fleming Corddry and Carpenter, 1972). Included in the report were two maps showing the locations of numerous mine adits and surface-mined areas.

Also as part of the Operation Scarlift, Boyer Kantz and Associates (1976) undertook a mine drainage abatement project for the Babb Creek watershed, which included a review of the coal geology and nomenclature, compilation of outlines of underground mines, and structure-contour mapping of a portion of the coal basin.

Buss (1986) investigated the hydrogeologic and hydrologic system of the abandoned Arnot no. 2 underground-mine complex at Arnot in conjunction with his "connector-well" project. He also prepared a structure-contour map of the area and published the records of boreholes drilled for his study.

In the past several years, the Pennsylvania Geological Survey has initiated new areal bedrock geologic mapping in the coalfield and adjacent areas, which has provided detailed information on the geology and structure between Morris Run and Antrim (i.e., McLaurin and Dodge, 2012; Dodge and Markowski, in press).

Stratigraphy

Nomenclature and Correlation

Eight principal coal horizons have long been recognized in the Blossburg-Antrim coalfield. The corresponding local coal names used today date from at least the 1870s, and several from as far back as the 1830s (Taylor, 1833; Macfarlane, 1875; Platt, 1878). In ascending stratigraphic order (oldest to youngest), the coals were designated the Kidney, Bear Creek, Bloss, Cannel, Morgan, Monkey, Seymour (less often called Cushing), and Rock. Several of the coals were also given letter designations early on, including A, for Bear Creek; B, for Bloss; C for Morgan; D for Seymour, and E, for Rock (Macfarlane, 1875). Later workers sometimes used C, for Cannel (or its equivalent), and C', for Morgan (e.g., Boyer Kantz and Associates, 1976). Letter designations never caught on very much for use in the Blossburg-Antrim coalfield and do not equate with the lettered coal names used in the Main Bituminous coalfield of western Pennsylvania. Early correlations across the basin, based on measured sections and limited drill-hole data, tended to be least reliable for the Cannel, Morgan, and Monkey coals (e.g., Platt, 1878), which are now known to be less persistent than previously believed. Complicating matters is the local occurrence of one or more thin, lenticular coals (splits or riders) anywhere between the Bloss and Monkey coal horizons that may contribute to misidentifications. Where the Bloss coal splits in places, the lower bench is sometimes miscorrelated with the Bear Creek.

Principal coals of the Blossburg-Antrim field have been successfully correlated with those of the Main Bituminous coalfield (in terms of relative stratigraphic position, but with no implication of seam continuity), based in part on the recognition of the Columbiana marine shale overlying the Seymour coal, which equates to the Lower Kittanning coal (Dodge, 1995; Edmunds, 1996; Dodge and Edmunds, 2010). As follows, the equivalent Main Bituminous coal name is given in parentheses after each Blossburg-Antrim coal name: Kidney (Quakertown); Bear Creek (Lower Mercer), Bloss (Upper Mercer); Cannel (Brookville, or Clarion no. 1); Morgan (Upper Clarion, or Clarion no. 3), Monkey (Lower Kittanning leader); Seymour (Lower Kittanning); and Rock (Middle Kittanning).

The mappable bedrock geologic units of the Blossburg-Antrim coal measures, recognized by the Pennsylvanian Geological Survey, are the late Early to Middle Pennsylvanian Pottsville Formation and Middle Pennsylvanian Allegheny Formation. The contact between the two formations is defined here as the base of the Cannel coal, which is equivalent to the Brookville coal of western Pennsylvania (McLaurin and Dodge, 2012).

Pottsville Formation

Lesley (1876) introduced the name Pottsville as a synonym for the older term "Seral Conglomerate" of the First Geological Survey of Pennsylvania, recognizing the excellent development and exposure of the clean, coarse sandstones and conglomerates near the town of Pottsville, Schuylkill County, Pennsylvania, in

the Southern Anthracite coalfield. In the Blossburg–Antrim coalfield, the Pottsville Formation consists of a complex, heterolithic, mostly nonmarine sequence of predominantly light- to medium-gray sandstone and extraformational conglomerate (quartz arenite), with much subordinate medium-gray siltstone, medium-dark-gray to dark-gray shale, olive-gray claystone (paleosol), and coal. The sandstones are generally medium to very coarse grained and conglomeratic in places, clean, siliceous, well cemented, thin to thick bedded, and trough cross-bedded. They contain scattered fossil plant impressions and compressions. The formation is more commonly conglomeratic in the lower part. The conglomerates occur as very thin to medium interbeds and lenses and contain clasts (extraformational) of milky-white quartz up to large and locally very large pebbles. Dodge (1995) reported rare, restricted-marine invertebrate fossils (*Lingula* sp.) about 20 ft above the Bloss coal (upper Pottsville) within medium-dark-gray siltstone in core from a Pennsylvania Geological Survey drill hole near the former mining village of Fall Brook, about 4 mi east of Blossburg. The lower contact of the Pottsville Formation is sharp and erosional. A major regional unconformity is present at the base. The Pottsville is present along the upper valley slopes and upland areas within the coalfield. The thickness of the Pottsville is approximately 150 ft.

The pre-Pottsville regional unconformity is believed to have begun no later than the Early Bashkirian (Early Morrowan) or early Early Pennsylvanian (Englund, 1979; Edmunds and others, 1999). The erosional surface has been interpreted as the result of eustatic sea-level decline associated with the world-wide mid-Carboniferous erosional hiatus (Saunders and Ramsbottom, 1986; Blake and Beuthin, 2008) or tectonic uplift due to a peripheral bulge (Ettensohn and Chestnut, 1989). W. E. Edmunds (written commun., 2010) provided evidence that both mechanisms may be involved in Pennsylvania, resulting in two closely timed sequential unconformities occurring in the Late Serpukhovian (Late Chesterian), for eustacy, and Early Bashkirian (Early Morrowan), for structural arching.

Three coals of historical economic importance are present within the Pottsville Formation and include in ascending order (oldest to youngest) the Kidney, Bear Creek, and Bloss. The Kidney coal was named for the kidney-shaped iron ore (siderite concretions) that commonly occurred in the overlying shale. Bear Creek coal was named for its occurrence along Bear Creek, east of Blossburg. The Bloss coal derives its name from Aaron Bloss, an early settler who first worked the coal (Taylor, 1835).

The Kidney and Bear Creek coals are lenticular and generally range in thickness from 0 to about 0.2.5 ft. They were rarely mined in the past on a commercial scale. The Kidney was once mined about 2.5 mi southwest of Blossburg. The Bear Creek was best developed and formerly mined in places along the Tioga River valley between Bear Creek and Morris Run. The Kidney and Bear Creek occur about 50 to 75 ft and 15 to 30 ft, respectively, below the Bloss. The stratigraphic position of the Bloss coal is about 35 to 55 ft below the top of the Pottsville. The Bloss coal was formerly the most important economic seam of the basin, where it was extensively deep and strip mined. The Bloss ranges in thickness from about 2.5 to 5 ft, but locally forms a coal complex where it has split into two principal benches that may be separated from one another by as much as 5 to 10 ft. Each bench varies in thickness from about 2 to 3.5 ft but where one bench is thicker (i.e., toward the upper limit of its thickness range) the other is thinner and may contain more partings. The thinner bench was generally uneconomical to deep mine along with the thicker bench, owing to bed thickness, seam impurities, or distance between the two benches. The Bloss coal is the most important key bed within the coal measures of the Blossburg-Antrim coalfield and has been used for correlation and structure contouring of the coal basin (McLaurin and Dodge, 2012; Dodge and Markowski, in press). As mentioned previously, the Bloss coal complex is now considered by the Pennsylvania Geological Survey as equivalent to the Upper Mercer coal complex of the Main Bituminous coalfield.

Allegheny Formation

The Allegheny Formation was named for exposures of the lower productive coal measures along the Allegheny River in the western part of the state and replaced earlier usage of the term "Allegheny River Series" of the First and Second Geological Surveys of Pennsylvania (see Platt, 1875). In the Blossburg–

Antrim coalfield, the Allegheny Formation is a poorly exposed and highly heterolithic unit composed of cyclic sequences of light- to medium-gray sandstone to conglomeratic sandstone (quartz arenite) and subordinate medium-gray siltstone, medium-dark-gray to dark-gray shale, olive-gray claystone or underclay (paleosol), and coal. The major sandstone units are fine to very coarse grained and locally conglomeratic, siliceous, thin to thick bedded, and cross-bedded. They contain scattered fossil plant impressions. Milky-white quartz clasts in the conglomerate range up to medium and locally large pebbles. Some finer grained units contain scattered fossil plant compressions and impressions. Dark-gray shale containing marine invertebrate fossils and siderite concretions overlies the Seymour coal (Dodge, 1995; Dodge and Edmunds, 2010). Identified marine and restricted-marine fossils include *Chonetes* sp., *Dunbarella* sp., *Lingula* sp., and nuculid bivalves. The marine shale represents the Columbiana marine zone that also overlies the Lower Kittanning coal of the Main Bituminous coalfield. The lower contact of the formation is sharp and corresponds to the base of the Cannel coal (equivalent to the Brookville coal of the Main Bituminous coalfield). The Allegheny Formation is confined to the upland areas toward the deepest parts of the Blossburg syncline. The estimated maximum preserved thickness of the formation is about 235 ft , in an area to the east of Blossburg.

Five coals of historical economic importance are present within the Allegheny Formation in the coalfield and include in ascending order (oldest to youngest) the Cannel, Morgan, Monkey, Seymour, and Rock. The Cannel coal was named for its duller luster and cannel-like appearance; it is not a true cannel coal. Above this coal is the Morgan, which was informally known at one time as the "Dirty vein" because of the shale partings and other impurities it contained in places. The Monkey coal was given its name by miners for its general thinness. It is typically associated with an overlying thick, cross-bedded sequence of interbedded sandstone and conglomeratic sandstone that formed a conspicuous topographic feature in parts of the basin called the "Monkey Ledge." The Seymour coal was named in honor of former New York Governor Horatio Seymour (1810-86), who owned a large tract of land where the coal was first mined. The Rock coal received its name for the thick sequence of coarse-grained sandstone to conglomeratic sandstone that overlies it. (See Macfarlane, 1875.)

The Cannel coal generally ranges in thickness from about 1 to 2 ft, though it is absent in places. It has been surface mined in the past in the vicinity of Morris Run. The Cannel lies about 30 to 40 ft above the Bloss coal. The Morgan coal varies in thickness from about 1.5 to 3 ft, though it is thin or absent locally. It has been both underground and surface mined in the past in the area around Morris Run. The Morgan coal is situated around 50 to 70 ft above the Bloss coal. Some workers in the past regarded the Cannel and Morgan as a coal complex in which the two coals were separated from each other by no more than several feet of shale or claystone (see Dodge and Edmunds, 2010). However, recent stratigraphic-correlation work and geologic mapping by McLaurin and Dodge (2012) and Dodge and Markowski (in press), coupled with a review by this author of the original definitions and correlations (i.e., original intent) of the two coals as presented in Platt (1878), show that the Cannel and Morgan are discreet coals that are not associated with each other. The Monkey coal is highly lenticular, ranging in thickness from 0 to 4 ft, though it is generally less than 2 ft thick. It is best developed to the east of Antrim, where it was formerly underground and surface mined. (Here, it had previously been miscorrelated as the Seymour coal.) The Monkey coal occurs about 80 to 110 ft above the Bloss coal. The Seymour coal is about 2 to 4 ft thick. It has been mined by both underground and surface methods in the past, particularly in the vicinity of Fall Brook, Morris Run, and Arnot. The Seymour is around 130 to 150 ft above the Bloss coal and areally restricted to the higher elevations of the uplands in the deeper parts of the Blossburg syncline. The Rock coal is the highest preserved coal in the basin and is about 2 to 2.5 ft thick. It was previously surface mined and is nearly all mined out. The areal extent of the Rock is limited to several hilltops near the axis of the syncline to the northwest of Morris Run.

Depositional Environments and Paleoclimate

The lower part of the Pottsville Formation was deposited as sediments by gravelly or sandy braided-stream systems crossing an alluvial plain (Meckel, 1964; Dodge, 1992). Regional changes in sediment load/gradient and stream competency were primarily in response to changes in base level of the fluvial system brought on by glacial-eustatic fluctuations of sea level (Heckel, 1994; 1995). Upper Pottsville and lower Allegheny rocks (i.e., Bloss coal and above) represent depositional environments chiefly characterized by high-sinuosity meandering-stream systems in a proximal-alluvial-plain or coastal-plain/paralic setting. Peat accumulated in the coal swamps that developed in response to changes in groundwater level and accommodation space mostly during times of near-maximum to maximum transgressions (highstands) that were controlled by glacial eustasy (Heckel, 1994; 1995). The rapidity and duration of the transgressions were primary controls on the development and persistence of the coal swamps. Based on the known occurrence of marine fossils, it appears that the strongest (most penetrating) Pennsylvanian transgressions into Tioga County were associated with the Bloss and Seymour coal cycles (the Seymour being the most significant). Pottsville and lower Allegheny sediments for this area were derived mostly from a cratonal source to the north during the initial suturing between Gondwana and Laurussia to form Pangea (Meckel, 1964; Edmunds and others, 1999; Blakey, 2008).

Paleoclimate reconstructions of the Carboniferous by Cecil (1992) and Cecil and Eble (1992) suggest that the lower Pottsville Formation was humid, mostly ever-wet tropical, with a transition to more wet-dry (seasonal) tropical conditions toward the top of the Pottsville and upward through the Allegheny Formation. During the deposition of the Pottsville and Allegheny, the paleolatitude was about 5 to 10 degrees south of the equator (Edmunds and others, 1979; Blakey, 2008).

Structure

The Blossburg–Antrim coalfield lies within the Blossburg–Antrim structural coal basin, an overall inwardly doubly plunging or canoe-shaped segment of the Blossburg syncline. The Blossburg syncline is one of a series of NE-SW-trending first-order folds found throughout Tioga County and vicinity. The fold axis of the Blossburg syncline undulates within the coal basin, creating several closed structural depressions that have their minimums near (from east to west) Blossburg, Landrus, and Antrim. Within the coalfield, the fold plunges at about 10 to 40 ft/mi. At the east and west end of the basin, plunges may increase to 100 to 150 ft/mile. The Blossburg syncline is slightly asymmetrical, having a steeper dipping north limb (McLaurin and Dodge, 2012). This is consistent with observations made by earlier workers, who noted that the limbs of anticlines throughout the region showed steeper southerly dips and shallower northerly dips (Rogers, 1858a; Sherwood, 1878). Along the north limb of the syncline, beds dip south-southeast at about 280 to 400 ft/mi, whereas along the south limb, the beds dip north to north-northwest at approximately 230 to 340 ft/mi. The structural relief (fold height), measured from the bottom of the Blossburg syncline to the top of the Wellsboro anticline to the north, is about 1450 ft (McLaurin and Dodge, 2012). The wavelength of the Blossburg syncline varies from approximately 16 mi toward the east end of the basin (measured between the axes of the flanking Wellsboro and Towanda anticlines) to about 12 mi near the west end (measured between the axes of the Wellsboro and Slate Run anticlines).

No demonstrable tectonic faults have been observed within the Blossburg–Antrim coal basin. Numerous underground mine maps examined by the author show no offsets in the coal. However, a small fault, trending NE–SW and having a 3-ft throw, was encountered in three adits on the Bloss coal along the north side of Bear Creek near Blossburg (Rogers, 1858a, p. 521). Boyer Kantz and Associates (1976) reported a fault trending NW–SE through two adjacent underground mines west of Wilson Creek near Antrim. Vertical displacements are said to be negligible to as much as 10 to 20 ft locally. However, the structure contours shown on the mine-map compilations indicate little or no offset along the presumed fault. Furthermore, the purported fault is oriented obliquely (i.e., a cross-fault) to the fold axis of the Blossburg syncline, or grain of the regional structure, which is highly unusual in this geologic setting. Thus, it appears

that offsets at the second location are isolated and may not be part of a continuous feature. In the absence of any other known occurrences, these reported features are considered very local and are attributed to paleoslumping.

COAL QUALITY

For more than a century, coals of the Blossburg–Antrim basin have been sampled and analyzed for various coal-quality parameters by several state and federal agencies. Three of the most basic parameters determined include total sulfur content, total ash content, and heat value. The total sulfur content of coal, on an as-received basis, is classified by the U.S. Geological Survey (USGS) as low (1 weight percent or less), medium (more than 1 but less than 3 weight percent), and high (3 weight percent or more). The total ash content of coal, on an as-received basis, is classified by the USGS as low (less than 8 weight percent), medium (8 to 15 weight percent), and high (more than 15 weight percent). Heat value, a measure of energy content, is commonly reported as British thermal units (Btu) per pound, on an as-received basis. Coal-quality data for the basin can be found in Platt (1878), U.S. Bureau of Mines (1928), U.S. Bureau of Mines (1939), Sponseller (1973), Skema and others (1977), U.S. Geological Survey (1994), and Dodge and Edmunds (2010). Some of the older analyses should be interpreted with caution, as they were often made on "representative" samples from run of mine rather than standard channel samples. "Representative" samples were subject to hand picking and mechanical sorting to remove binders and other impurities, and do not necessarily reflect the quality of the entire seam or a composite of the coal benches mined. For a given coal of the same area, run-of-mine analyses may be lower in sulfur and ash content than comparable analyses from benched channel samples. Nevertheless, samples and analyses for run of mine generally do represent the product as it was shipped to market.

The coals of the Blossburg–Antrim coalfield are classified as medium-volatile bituminous rank, which is defined as coals having from 69 to less than 78 weight percent fixed carbon on a dry, ash-free basis. As a whole, the Blossburg–Antrim coals tend to be low to medium in sulfur and medium in ash. Heat values, on an as-received basis, range from about 12,600 to 13,900 Btu per pound. Some aspects about the coal quality of the individual coals are as follows:

1. The quality of the Kidney coal is unknown, but based on the physical description of the coal and associated roof rocks of black shale, probably pyritic, it is likely that the coal is medium in sulfur and medium to high in ash.
2. The Bear Creek coal is generally low to medium in sulfur and medium in ash.
3. The Bloss is the premier coal of the basin in terms of persistence, thickness, and quality. It is generally low in sulfur and low to medium in ash. Sulfur content is typically less than 0.7 percent and ash less than 10 percent for the old underground mines at Fall Brook, Morris Run, Blossburg, Arnot, and Antrim, where operations extracted the best quality coal first. Newer analyses from mines tend to show a low-to-medium sulfur content and medium ash content; sulfur may vary from 0.8 to greater than 2.5 weight percent, and ash content may run from around 8 to 16 percent. The Bloss is also an excellent coking coal, with yields of about 77.5 to 80.5 percent. The coal was well suited for a variety of uses, including blacksmithing, salt production, puddling, heating in roller mills, and especially generating steam in locomotive and stationary boilers (Platt, 1878).
4. The quality of the Cannel coal is not known. However, the physical description of the coal suggests that it is most likely low in sulfur and medium in ash.
5. The Morgan coal ("Dirty" vein) is variable in quality, ranging from low to medium in sulfur and low to medium in ash.
6. The Monkey coal tends to be low to medium in sulfur and medium in ash.

7. The Seymour (Cushing) coal is generally medium to high in sulfur and medium in ash. It has the highest average sulfur content for any of the coals in the basin. Older analyses tend to be lower in sulfur and ash than the newer ones, either because the better quality coal was mined first or because sampling was less representative of the whole seam.
8. The little information available for the quality of the Rock coal suggests that it is medium in sulfur and medium in ash (Dodge and Edmunds, 2010).

MINING HISTORY

The Blossburg–Antrim coalfield was one of the oldest productive coal-mining districts in Pennsylvania. For a time, the demand and popularity of the famous Bloss coal rivaled that of other notable bituminous coals, including the Sharon of Mercer County and the Pittsburgh of Allegheny County and vicinity. Much of the following discussion is based on Meginness and others (1897).

Coal was apparently first discovered in Tioga County in 1792 at "Peters' Camp," now the present borough of Blossburg. Two noted Indian scouts, the Patterson brothers, were said to have made the discovery while escorting German and English immigrants from Williamsport to "Genesee Country" in southern New York (Meginness and others, 1897). Coal was first mined for shipment in Tioga County by David Clemons (1767–1833) around 1815. Clemons discovered coal on the lands of Aaron Bloss (1775–1843) along Bear Creek near "Peters' Camp" and made arrangements with Bloss to develop it. The coal mine became known as "Clemons' opening" and the coal was named "Clemons' coal or vein." Because of poor road conditions and limited local markets, Clemons shipped his coal by wagon or sleigh northward to Painted Post, New York. Soon after Clemons started mining, Aaron Bloss discovered another coal on Bear Creek about 11 feet below "Clemons' coal," which became known as "Bloss' vein." Bloss did not mine the coal himself but rather made financial arrangements with others to develop the seam. It is believed that the "Clemons coal" was an upper split of the Bloss coal that was thickest near its outcrop and pinched out some distance into the hillside. Aaron Bloss, of course, will always be remembered for discovering the main part of the celebrated Bloss coal. Another coal was later discovered and worked about 15 to 20 feet below the Bloss along Bear Creek, and it was called the Bear Creek coal.

As early as 1807, Bloss was also aware of the presence of other minerals on his lands, including iron ore (siderite). Around 1827, Judge John H. Knapp, a businessman from Elmira, New York, saw promise in further developing the mineral resources around Blossburg and in creating a center there for mining and manufacturing. As such, he purchased over 200 acres of land from Bloss. Through his investments, he opened a coal mine on Coal Run (near Blossburg) and established an iron works nearby. Thereafter, Knapp and other like-minded businessmen formed the Tioga Navigation Company to develop the means to transport coal and iron to markets in the state of New York. In 1832, the Tioga Improvement Company hired Richard C. Taylor to make a thorough geologic survey of the coal, iron ore, and other minerals around Blossburg and to determine the best route for a railroad between Blossburg and the Pennsylvania–New York border. Taylor's (1833) report on his findings provided the detailed information and facts needed to make investors confident in going forward with their plans. Because of failing health, Knapp sold his interests, which were eventually purchased by Dr. Lewis Saynisch and his business partners in 1834. Saynisch continued to acquire additional tracts of land near Blossburg. By 1838, Saynisch and his partners formed the Arbon Coal Company to mine coal from the old "Clemons' opening" and further created the Arbon Land Company to promote the rapid building of a railroad from Lawrenceville to Blossburg. By September 1840, the Corning and Blossburg Railroad was completed, securing reliable transportation to markets and ushering in a new era of development for the coalfield.

The Arbon Coal Company operated the Blossburg mines until 1845 when ownership was acquired by John Ward and Company, which leased the operations to William M. Mallory and Company until 1857 and thereafter to John Magee until 1859. With the mines playing out and with the desire to stop leasing and to acquire his own mining company, Magee ended his lease and founded the Fall Brook Coal Company,

which formally began operations in 1860. Thereafter, the Blossburg mines ceased shipping coal, but mining continued for a number of years for local use. Between 1840 and 1859, a total of 533,745 gross tons of coal were mined around Blossburg (Meginness and others, 1897).

During the second half of the nineteenth century, three major mining companies dominated coal production in the Blossburg–Antrim basin, mostly from the Bloss seam. In 1853, the Tioga Improvement Company began to develop the coal lands up the stream valley of Morris Run, about 2-1/2 mi east of Blossburg. With the completion of a railroad from Blossburg to the new village of Morris Run that same year, the company began shipping Bloss coal from its new mines. From 1853 to 1862, the Morris Run mines produced 323,174 gross tons of coal (Meginness and others, 1897). In 1862, the mines were sold to the Salt Company of Onondaga, Syracuse, New York, which operated them until 1864. At that time, the mines were acquired by the newly formed Morris Run Coal Company. Mining operations expanded greatly over the next few years, and the company was considered to be a first-class operation. The Morris Run mines were noteworthy for the exceptional quality and thickness of its Bloss coal. (See Platt, 1878.) The company reorganized as the Morris Run Coal Mining Company in 1877.

In 1856, Duncan Magee, son of John Magee and superintendent of the Blossburg mines, entered into an agreement with C. L.Ward of Towanda, Pennsylvania, to explore his lands for coal in Ward and Union Townships with an option to purchase them. The explorations proved successful, and a deal was struck to acquire about 6000 acres of land. Magee's assistants determined that the best outlet for the Bloss coal was through the stream valley of Fall Brook. With this in mind, the mining village of Fall Brook was started in 1859, and the Fall Brook Railroad was completed the same year from Blossburg to Fall Brook, a distance of about 4 miles. The mines of the Fall Brook Coal Company formally opened in 1860. After this time, mine production here grew rapidly. Between 1866 and 1868, the Fall Brook Coal Company also explored potential coal lands along the valley of Wilson Creek, about 12-1/2 mi west of Blossburg, and found them very promising. The mining village of Antrim was founded in 1868. The Fall Brook Coal Company incorporated the Lawrenceville and Wellsboro Railroad in 1867 to provide the means to ship coal from its Antrim mines. In 1872, the railroad was completed to Antrim, and the mines formally began operations (Figure 3).

Figure 3. Stereograph of Antrim no. 1 drift, Fall Brook Coal Company, at Antrim, Pennsylvania. Photograph circa 1872. Published by Gates and Company, Watkins Glen, N.Y.

With the discovery of large deposits of coal along Johnson Creek, about 3 mi southwest of Blossburg, a group of capitalists formed the Blossburg Coal and Mining Company in 1866 and purchased several thousand acres of land in Bloss Township. Later that year, test adits were driven into the hillside, and the village of Arnot (originally named Draketown) was established. A railroad was also completed from Blossburg to Arnot in the summer of 1866. In 1868, the Blossburg Coal and Mining Company purchased the Tioga Railroad Company, which had leased the Blossburg and Corning Railroad line (a reorganization of the Corning and Blossburg Railroad), thus securing shipment of its coal to market. Thereafter, output from the Arnot mines increased greatly. Mining was initially on both the Seymour and Bloss coals, but the former played out in a few years. In 1880, the company began producing coke of excellent quality but eventually could not compete with the Connellsville area, Fayette County, and other large coke-producing centers. In 1882, the Blossburg Coal and Mining Company completed the Arnot and Pine Creek Railroad between Arnot and Hoytville, a distance of about 12 mi. Along this route, the company founded the village of Landrus that same year next to Babb Creek at the site of its newly constructed sawmill. (All of the mining companies had sawmills to provide wood for mine props and general construction.) In 1888, the Blossburg Coal Company (this became the formal name of the company in 1885) opened its Bear Run mine on the Bloss coal to the northeast of Landrus. This operation was the first fully electric coal mine for all its machinery in the United States and purported the world.

Prior to 1861, annual coal production in the Blossburg–Antrim coalfield never reached 100,000 net tons. Mine output grew dramatically during the 1860s, following the organization of the Fall Brook Coal Company and Blossburg Coal and Mining Company, as well as improvements made to the Morris Run Coal Company. Total annual coal production grew from 135,000 net tons in 1861 to 840,000 net tons in 1870. Annual production exceeded 1 million net tons for the first time in 1873, totaling 1,109,984 net tons. Total output declined somewhat in the second half of the 1870s but surpassed 1 million net tons again in 1880. With one exception, annual coal production exceeded 1 million net tons every year during the 1880s, and output reached its zenith for the coalfield in 1886 with a total of 1,384,800 net tons. By the mid-1890s, coal production in Tioga County declined relative to the previous decade and was affected by labor unrest, the national economy, and competition from the Clearfield County mines, which could operate more efficiently and at lower cost. (See Eavenson, 1942; Schanz, 1957.)

Grievances by miners over issues such as housing, union formation, and wages led to a number of strikes in the late nineteenth century, including 1865, 1873, 1879–80, 1890, 1894, and 1899–1900. Conditions did not always improve upon their return to work. In 1899, Mother (Mary Harris) Jones (1837–1930), the well-known union activist, visited Arnot during a bitter strike and succeeded in organizing the wives of striking miners to prevent replacement workers from taking their husbands' jobs.

As the twentieth century began, the three major mining companies continued to dominant coal production in the coalfield. They expanded existing operations and recently opened up new mines, such as the Fall Brook Coal Company's Anna S mine, west of Wilson Creek, and Blossburg Coal Company's Maple Hill mine, north of Lick Creek. During the 1900s, annual production of the coalfield generally ranged from about 700,000 to slightly more than 1 million net tons. The last time total output ever exceeded 1 million net tons was 1910, with a total production of 1,037,417 net tons. However, for the most part, total production continued to exceed 3/4 million net tons per year between 1911 and 1918. Total production fell below 1/2 million net tons in 1921, for the first time since 1866. The coal industry in Tioga County struggled under weak demand and declining reserves. Operators, such as the Morris Run Coal Company, began to deep mine other seams, such as the Morgan and Seymour coals. Coal output from the mid-1920s to the late 1930s never reached 200,000 net tons per year. Production did increase during World War II to as much as 358,801 net tons in 1942 but fell back below 200,000 net tons after the war ended. (See Schanz, 1957; Pennsylvania Department of Environmental Protection, 2000.)

Surface-mine operations commenced in 1941 and grew dramatically following the end of World War II. The last underground mine ceased operations in 1962. Annual coal production generally remained below 200,000 net tons between 1946 and 1956, decreasing to 55,289 net tons in 1954. In the early 1950s, Jones and Brague Mining Company became the successor to the Morris Run Coal Mining Company. Jones and Brague was largely a surface-mine operator thereafter and became the dominant mining company in the coalfield from the late 1950s to early 1980s. Starting in 1957, production surged upward again, reaching a maximum of 873,239 net tons in 1971. Total annual output generally exceeded half a million net tons through the remainder of the decade. Beginning around 1980, Antrim Mining Company, Inc. was established and began a number of operations in Tioga County. Coal production totaled 496,447 net tons in 1980 and steadily declined for the most part afterwards until all mining ceased in 1990. (See Pennsylvania Department of Environmental Protection, 2000.) No commercial-scale coal mining has occurred in Tioga County since then. Lack of interest in future development and mining of coal results from the small remaining reserves, weak market conditions, and highly acidic (toxic) overburden requiring specialized treatment and handling under current state regulations.

ACKNOWLEDGEMENTS

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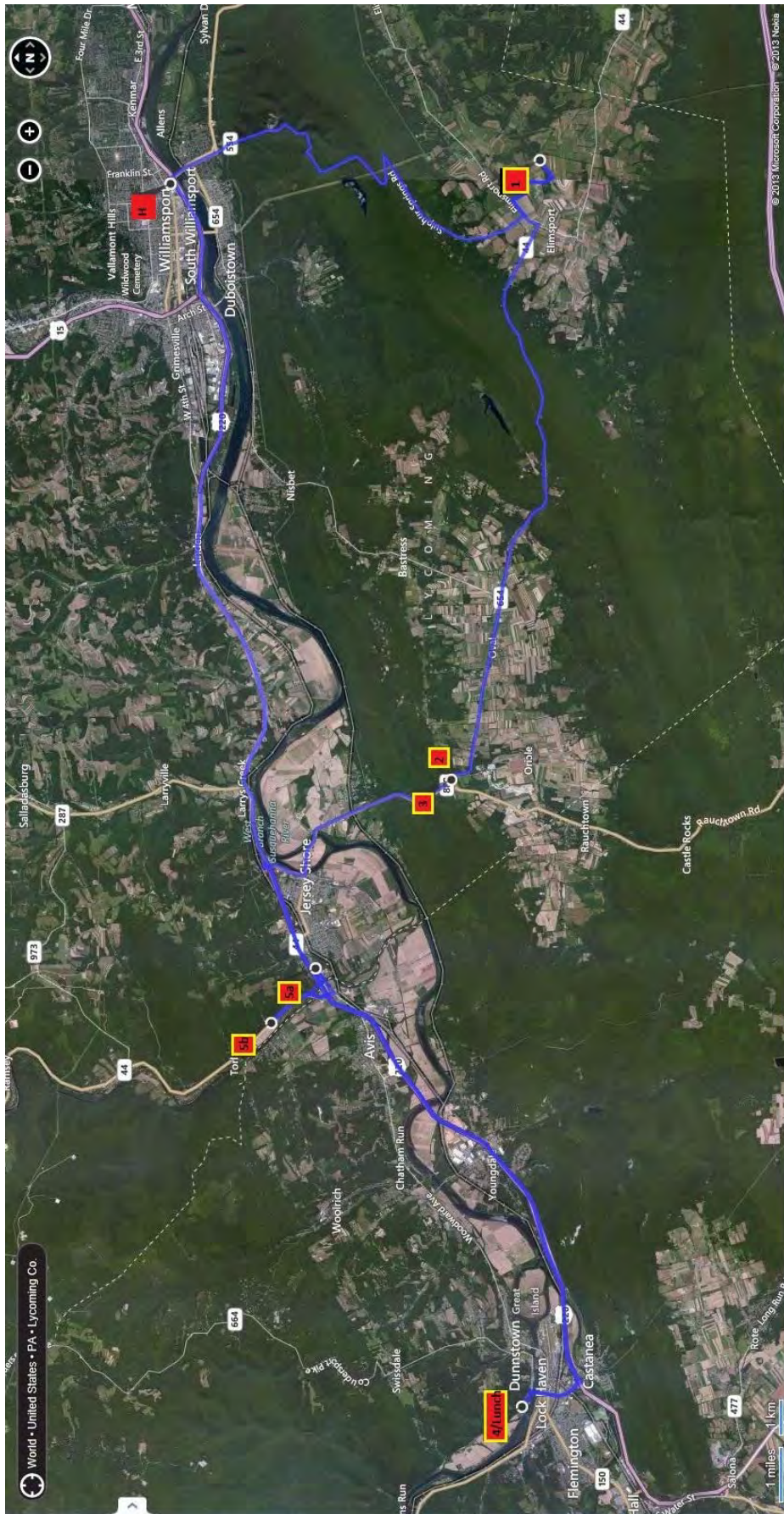
Road Log- Day 1

Int	Cum	Description
0.0	0.0	Leave parking lot of Holiday Inn, left onto Court Street.
0.1	0.1	Stop sign. Right onto Church Street.
0.1	0.2	Traffic light. Continue straight on 15 south.
0.2	0.4	Crossing Susquehanna River, West Branch.
0.3	0.7	Traffic light, Southern Avenue. Stay in right lane.
0.1	0.8	Traffic light, Central Avenue. Stay straight.
0.4	1.2	Continue straight ahead on SR 554 towards Elimsport.
0.5	1.7	Red sandstone of the Juniata Formation on the left.
0.4	2.1	Small quarry to the left in sandstones of the Bald Eagle Formation.
0.3	2.4	Hagermans Run Aggregate Facility Plant #12. Mining of the Bald Eagle sandstone.
0.3	2.7	Bald Eagle sandstone outcropping on the right.
0.9	3.6	Hair-pin turn. Bald Eagle sandstone outcropping on the right.
2.1	5.7	Crest of North White Deer Ridge.
0.2	5.9	Splendid view on the right of the White Deer Valley and South White Deer Ridge; the synclinal valley is comprised of primarily Silurian and Devonian age rocks.
0.5	6.4	Another excellent view of the White Deer Valley and South White Deer Ridge in the background to the left.
3.4	9.8	Stop sign. Continue right onto SR 554.
0.3	10.1	Continue straight on SR 44/554 south.
0.1	10.2	Enter village of Elimsport.
0.7	10.9	Stop sign. Turn left staying on SR 44 south.
0.4	11.3	Turn left onto Pikes Peak Road.
1.3	12.6	Turn right into shale pit. **Stop 1 Marcellus shale, jointing, volcanic ash and calcareous concretions near Elimsport, PA**. Leader- Donald Hoskins See page 103.
	12.6	Exit Stop 1. Turn left onto Pikes Peak Road.
1.3	13.9	Stop sign. Turn right onto SR 44 north.
0.3	14.2	Enter village of Elimsport.
0.3	14.5	Stop sign. Turn right, staying on SR 44 north.
0.8	15.3	Turn left staying on SR 44 north.
0.1	15.4	View of synclinal fold ahead and to the left, comprising South White Deer Ridge and North White Deer Ridge.
3.9	19.3	Crest of North White Deer Ridge.
0.2	19.5	Scenic view of cuestas (hogbacks) of the Nippenose Valley. The topographic features are comprised of sandstones of the Bald Eagle Formation.
1.7	21.2	Passing through the cuestas of the Bald Eagle Formation.
0.6	21.8	Entering the Nippenose Valley. Some question as to where the valley begins - at the base of the cuestas or the beginning of the carbonates.
0.1	21.9	Goose farm on the right.
0.5	22.4	Entering the Village of Collomsville.
0.7	23.1	Intersection with SR 654.
0.1	23.2	To the left is a splendid view of the cuestas along the south side of the valley.

0.4	23.6	Nippenose Valley Elementary School on the left.
0.5	24.1	Middle Road on the left. Yes, it does run down the middle of the valley.
0.2	24.3	Entering the village of Oval.
0.2	24.5	Limestone Township Municipal Building to the right.
0.1	24.6	Reclaimed limestone quarry in vegetated area on right.
0.1	24.7	Note that the right side of the Nippenose Valley does not have the cuesta features as observed on the south side of the valley. Bedding is more steeply dipping.
1.0	25.7	Lime kiln on the right.
0.9	26.6	Turn left onto Quarry Road.
0.1	26.7	**Stop 2**. Park on the right side of the road and disembark.
		Nippeno Spring and Sinkholes. Leader- Bill Kochanov See page 109.
0.1	26.8	Leave Stop 2. At stop sign, turn left onto SR 44 north.
0.2	27.0	Cross Antes Creek. Ordovician Coburn Formation outcrops along the north side of the stream bank.
0.3	27.3	SR 880 on left, stay straight on SR 44; cross Antes Creek.
0.1	27.4	Park in pull off area on right. **Stop 3**.
		Reedsville Formation and Antes Shale at Antes Gap, Lycoming County.
		Leader- John A. Harper See page 117.
0.1	27.5	Leave Stop 3. Continue north on SR 44. Entering Antes Gap.
0.6	28.1	Old paint mill on left.
1.1	29.2	View of Allegheny Front to left.
0.1	29.3	Entering village of Antes Fort.
0.1	29.4	Village of Old Fort to the left.
0.4	29.8	Riding the flood plain of the West Branch of the Susquehanna River.
0.4	30.2	Crossing the West Branch.
0.7	30.9	Crossing West Branch and entering Jersey Shore.
0.1	31.0	Traffic light. Turn right, continuing on SR 44.
0.6	31.6	Bear left (left lane). Stay on SR 44 and SR 220 south.
0.4	32.0	Merge onto SR 220 south.
2.8	34.8	Cross West Branch of the Susquehanna River.
0.1	34.9	SR 44 Exit.
1.7	36.6	SR 150/Avis, Exit 118.
1.2	37.8	Entering Wayne Township; crossing the West Branch of the Susquehanna; flood plain on the right.
0.6	38.4	Smelly Landfill I. Wayne Township Landfill. The landfill is permitted for municipal and residual waste disposal (Hershey and Pollok, 1995).
0.6	39.0	McElhattan/Woolrich, Exit 116.
0.7	39.7	Bald Eagle Ridge on the left.
3.5	43.2	Entering Castanea Township.
0.4	43.6	Mifflintown Formation exposed in abandoned shale pit on the left.
1.5	45.1	Exit Castanea/Lock Haven, SR 120W, Exit 111.
0.3	45.4	Merge onto SR 120 (Jay Street).
0.6	46.0	Traffic light. Straight on SR 120. Entering Borough of Lock Haven.
0.1	46.1	Traffic light. Continue straight across bridge over West Branch of the Susquehanna

		River.
0.2	46.3	Turn left onto River Drive.
0.4	46.7	Turn left into River View Park. Lunch. **Stop 4**. **Lockport .** Leader- Dr. Thomas C. Wynn See page 129
	46.7	Leave Stop 4. Return to SR 120, east (right) across bridge and continue through Lock Haven.
1.7	48.4	Turn left onto US 220 north.
0.5	48.9	Lock Haven Flood control project to right. Designed to protect the city of Lock Haven against the 200-year flood level through a series of levees, flood-walls, and closure structures (Way and Yowell, 1995).
1.2	50.1	Cross West Branch of the Susquehanna River.
0.2	50.3	Mifflintown Formation exposed in abandoned shale pit on the right.
1.0	51.3	Smelly landfill II on right.
0.6	51.9	Cross west Branch of the Susquehanna River.
0.4	52.3	Exit 118. SR 150/Avis.
2.2	54.5	Exit 120. SR 44N Pine Creek. For future visits, use the alternative directions.
0.6	55.1	Entering Lycoming County, crossing the West Branch of the Susquehanna.
1.2	56.3	Exit right at Thomas Street, SR 150.
0.2	56.5	Right onto Thomas Street, SR 150.
0.2	56.7	Right onto Railroad Street, SR 150 west.
0.6	57.3	Disembark buses at access to Pine Creek Rail Trail. From here we walk.

****

Alternate Directions

2.2	54.5	Exit 120. SR 44 N/Pine Creek
0.4	54.8	Stop sign. Left onto SR 44 N
0.2	55.0	Cross West Branch of Susquehanna River
0.4	55.4	Right onto Minnie Drive. Park on left. From here it is a scramble up the steep hill to the Rail Trail. Vertical beds will be facing you just a bit to the left.
0.7	56.1	To get to the gift shop, continue on 44 N. Creekside Country Gifts and Furnishings on the right.

****

		Stop 5. Pine Creek Rail Trail by Jersey Shore. Leader- Paul Washington See page 133. Buses will be parked at Creekside Country Gifts and Furnishings (see alternative directions at mileage 54.5). The store will be open.
	57.3	Leave parking area. Left onto SR 44. Will use the same ending mileage at the point we got off the busses just for convenience.
1.1	58.4	Cross the West Branch of the Susquehanna River.
0.3	58.7	Turn left onto US 220 north, merge.
0.2	58.9	Crossing Pine Creek and entering Lycoming County.
1.7	60.6	Prepare to stay in the right lane!
0.3	60.9	Stay in the right lane and continue straight on US 220 north.
1.7	62.6	Bald Eagle Mountain on the horizon to the right.
0.7	63.3	Sheetz on the right.
0.3	63.6	Harvest Moon Plaza and Restaurant on left. Foot-long hot dogs with sauerkraut!

0.8	64.4	Henry's Bar-B-Q on the right.
3.9	68.3	Crossing Lycoming Creek.
2.1	70.4	Bear right and exit onto the ramp for US 15. Once on the ramp, stay in the left lane.
0.1	70.5	Traffic light. Turn left onto Market Street.
0.1	70.6	Turn left onto Church Street.
0.1	70.7	Turn left onto Court Street.
0.1	70.8	Turn right into parking lot of Holiday Inn.

End of Field Trip. Day 1.

STOP 1: Marcellus shale, jointing, volcanic ash and calcareous concretions near Elimsport

Leader: Don Hoskins

Here, a portion of the lower part of the Marcellus black shale is well exposed *(Figure 1-1)*. Regionally, the outcrop is located nearly along the axis of a broad Appalachian syncline defined by two nearby, prominent and ridge-forming, eastward-plunging Tuscarora anticlines.

Over nearly the extent of the exposure, the layers dip to the east at a relatively low angle of 7 degrees. At the extreme western end the layers dip shallowly to the west defining a small anticline. The outcrop approximately emulates the horizontal position of layers presently encountered at depth by drillers in northern and western areas of Pennsylvania.

Figure 1-1. Marcellus black shale in Finck quarry near Elimsport

The Elimsport exposure includes a rare outcrop of one of the Tioga A-G ash layers often referred to as "bentonites". The U. S Geological Survey   (Roen & Hosterman) recommends that volcanic layers lacking smectite, that is present in type bentonites, should be referred to as "ash layers". The ash layer at Elimsport is here in the form of a weakly cemented micaceous rock referred to as tuff *(Figure 1-2)*.

Comparing this ash layer to the now unavailable exposure in the nearby Zeigler pit near Lewisburg (Way, Smith and Roden, 1986) the layer exposed here may be the "F-G" layers. An "H" layer was also present at the Zeigler pit. Is it here?

Figure 1-2. Volcanic ash layer and overlying blocky black shale. The base of the ash layer begins at the junction of the blue handle with the metal part of the rock hammer.

Hoskins, D.(2013) Stop 1– Marcellus shale, jointing, volcanic ash and calcareous concretions near Elimsport, *in*, Fleeger, G.M., Schmid, K., and Anthony, R., eds., A Tale of Two Provinces: the Nippenose Valley and Route 15 Corridor, 78th Annual Field Conference of Pennsylvania Geologists, Williamsport, PA, pp. 103 – 108.

The Elimsport exposure also provides more commonly observed geologic features present in nearly all outcrops of the Marcellus. These include numerous prominent joints, including the orthogonal joint directions that are of significant interest to the drillers of natural gas *(Figure 1-3)*. At least two additional joint directions are also present *(Figure 1-4)*. The weathered characteristic of the exposure makes it difficult to determine the full extent of joint direction. Some joint surfaces have patterns that may relate to the original hydrodynamic formation.

Figure 1-3. Excavation of Finck quarry at Elimsport by Tera Tek of Salt Lake for static tests of blocks of Marcellus. The excavation parallels the exhumed J1 joint visible to the left. The pit's refuse pile was obtained from this excavation, now filled. *Photo by Terry Engelder, PSU.*

Figure 1-4. Volcanic ash layer and underlying blocky black claystone. Note orthogonal and other joint directions in underlying layers.

Rare fossil debris occurs. A small brachiopod mold was found, and in the refuse pile at the east end, styliolinids and microscopic globules, some of which are replaced by pyrite *(figure 1-5)*.
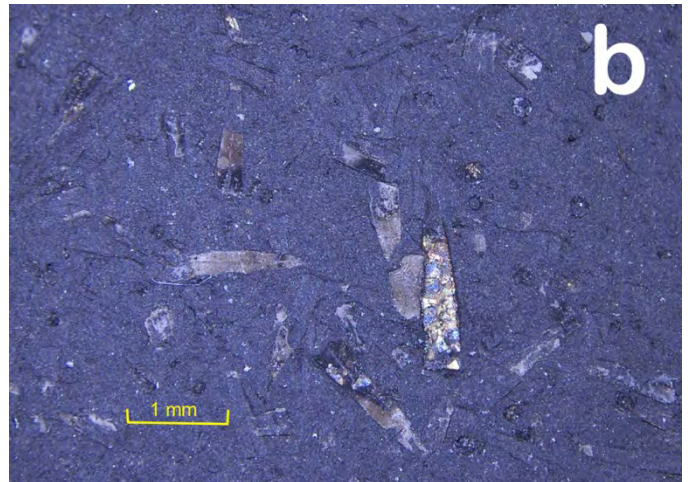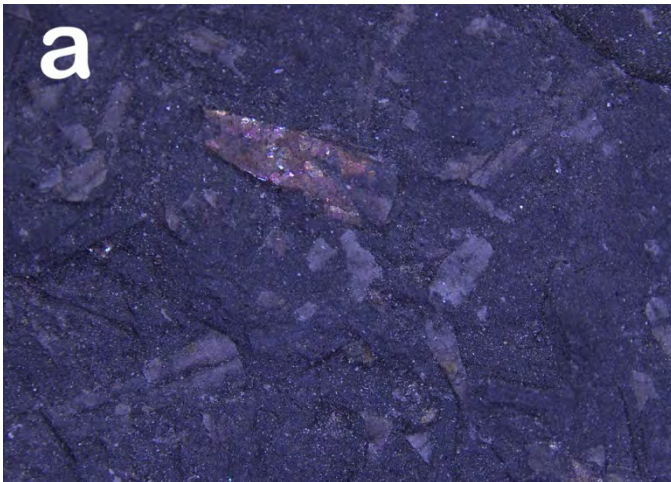

Figure 1-5. Styliolinid fossil debris with pyrite replacements. Styliolinids are tube shaped, generally with a central linear depression likely resulting from being crushed when buried *(a)* and generally are less than 0.2 cm in length *(b)*.

Additionally, at least one dark gray carbonate concretion outcrops in place in the exposure *(figure 1-6)*. Numerous other concretions are present in a refuse pile and other parts of the pit *(figures 1-7 & 1-8)*. The concretions do not exhibit the joint patterns seen in the black shale.

Figure 1-6. Large carbonate concretion interbedded with Marcellus black shale. Note lack of jointing in the concretion that is present in underlying and overlying shale layers.

Figure 1-7. Large concretion at eastern end of pit.

Figure 1-8. Small concretions found in spoil and rock fall from outcrop.

Lacking at this exposure are structural features that occur in many of the Appalachian Folded Belt outcrops of the Marcellus such as thin zones of cleavage duplexes and chevron folding interpreted as evidence of pre- or early Alleghanian deformation that occurred prior to the formation of the folded structures. One carbonate concretion retrieved from a spoil pile exhibits slickenlines in secondary calcite that show interlayer slip movement directions *(figure 1-9)*.

Figure 1-9. Slickenlines in secondary calcite coating a carbonate concretion. Slickenlines indicate horizontal interlayer movement prior to Alleghanian folding.
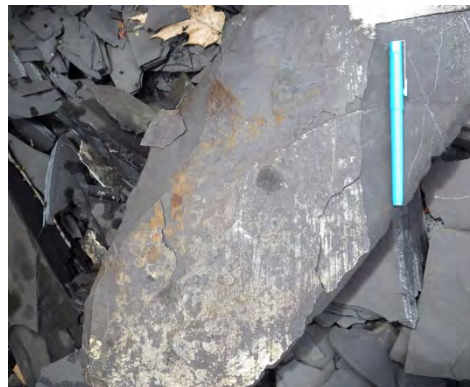
The Elimsport outcrop is frequently visited for academic purposes, notably those by PSU. Terry Engelder, Dept of Geosciences, The Pennsylvania State University has described this outcrop in a field trip guide prepared for the Pittsburgh Society of Petroleum Geologists (Engelder & Gold, 2008). Appendix 1 is the field trip description of the Finck quarry from their guidebook.

APPENDIX I

Dipping Union Springs Member of the Marcellus with J_2 joints propagating around concretions, *after* Engelder, T. and Gold, D.P., Stop 4 in Guidebook for the Pittsburgh Association of Petroleum Geologists Field Trip, September 12-13, 2008, pp 53-54 *(see references cited).*

Elimsport (Finck quarry along Pikes Peak Road off of Route 44 east northeast of Elimsport) [Google Earth UTM coordinates: 18 T 332747 m E – 4556008 m N]

Jumping over to the south flank of the Nittany Anticlinorium moves us into the transition between the Allegheny Front where gas shale is prospective to a region of the Valley and Ridge is overmature *(see Table 1 below).* In fact, industry dogma at the time of preparation of this field guide is that vigorous leasing of the Marcellus should remain north of an E-W line marked by Rt 118 in Lucern, Columbia, and Lycoming Counties.

Rt 118 is a virtual extension of the Allegheny Front east of the Susquehanna River.

The Finck quarry at Elimsport exposes the Union Springs Member of the Marcellus somewhere above the top bentonite in the Marcellus *(Figure 1-10).* The organic content of the shale (TOC > 6%) reveals that this portion of the Marcellus is in the hot bottom section as observed on gamma-ray logs.

J_1 is well developed in this portion of the Marcellus and exhibits the characteristics of natural hydraulic fracturing by passing around some large concretions within the Union Springs. In general, the surfaces are not as planar as see in outcrops of black shale in the Finger Lakes District, NY. Like the J_1 joints in the Snook quarry at Antis Fort, these joints form normal to bedding and when bedding is rotated to horizontal the J_1 joints return to vertical, again a sign of a prefolding origin. Also, like the Snook quarry, the J_1 joints have a relatively weak cluster. In the Finck quarry the vector mean strike for the J_1 set of 061°

While clustering is weak at both Stops 2 and 3, it is significant to note that the orientation of J_1 on both sides of the Nittany Anticlinorium is counter clockwise from the strike of J_1 joints in the Finger Lakes District of NY.

Table 1. Samples sent to Humble Labs for TOC and Rock-Eval measurements.

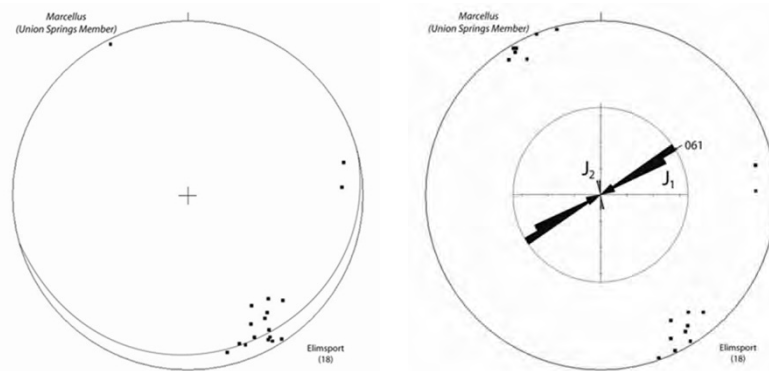|	TOC	S1	S2	S3	T_{max}	R_o (calc)
Matrix	6.05	0.04	0.22	0.56	593	3.51

Figure 1-10. Examples of joint development in the Union Springs Member of the Marcellus at the Delmar Finck Quarry along Pikes Peak Road off of Route 44 east northeast of Elimsport. Bedding is 075/10. Joints plotted in present coordinates (right) and rotated to their postion with horizontal bedding using a fold axis plunging 05 toward 075 with a rotation of 10º (left).

REFERENCES CITED

Engelder, Terry and David P. Gold, 2008, Structural Geology of the Marcellus and other Devonian gas shales: geological conundrums involving joints, layer-parallel shortening, and the contemporary tectonic stress field: Guidebook for the Pittsburgh Association of Petroleum Geologists (PAPG) Field trip, September 12-13

Engelder, Terry, Rudy Slingerland, Michael Arthur, Gary Lash, Daniel Kohl, and David P. 2011, An introduction to structures and stratigraphy in the proximal portion of the Middle Devonian Marcellus and Burket/Geneseo black shales in the Central Appalachian Valley and Ridge: The Geological Society of America, Field Trip 20, 2011

Engelder, Terry, Rudy Slingerland and Alfred Lacazette, 2013, The stratigraphy and structural geology of the Catskill Delta Complex of the Central Appalachian Mountains: Gas-bearing rocks between the Middle Devonian Marcellus black shale and the Mississippian Pocono Formation; Privately printed by Terry Engelder for the 2013 AAPG ACE, Pittsburgh, PA

Roen, John B. and John W. Hosterman, 1982, Misuse of the term "bentonite" for ash beds of Devonian age in the Appalachian basin: Geological Society of America Bulletin, vol. 3, no. 9, p. 921-925

Way, John H, Robert C. Smith II, and Mary Roden: 1986, Detailed Correlation across 175 miles of the Valley and Ridge of Pennsylvania using 7 ash beds in the Tioga zone *in* Guidebook , 51st Annual Field Conference of Pennsylvania Geologists, Selected geology of Bedford and Huntingdon Counties p. 55-72

Stop 2: Nippeno Spring and Sinkholes,

William E. Kochanov, Pennsylvania Geological Survey

This stop (Figure 2-1) is characteristic of the karst landscape prevalent throughout the Nippenose Valley. Sinkholes, disappearing streams (swallets), and caves are common features. We will view several outstanding examples of sinkholes and the largest spring in Pennsylvania, the Nippeno Spring.

As one observes the sinkholes you want to take note of:

1. The angularity of the bedrock exposed along the sidewalls of the sinkholes.
2. The lack of a thick residual soil developed atop the limestone.
3. The minimal amount of rock lying at the bottom of the sinkholes.
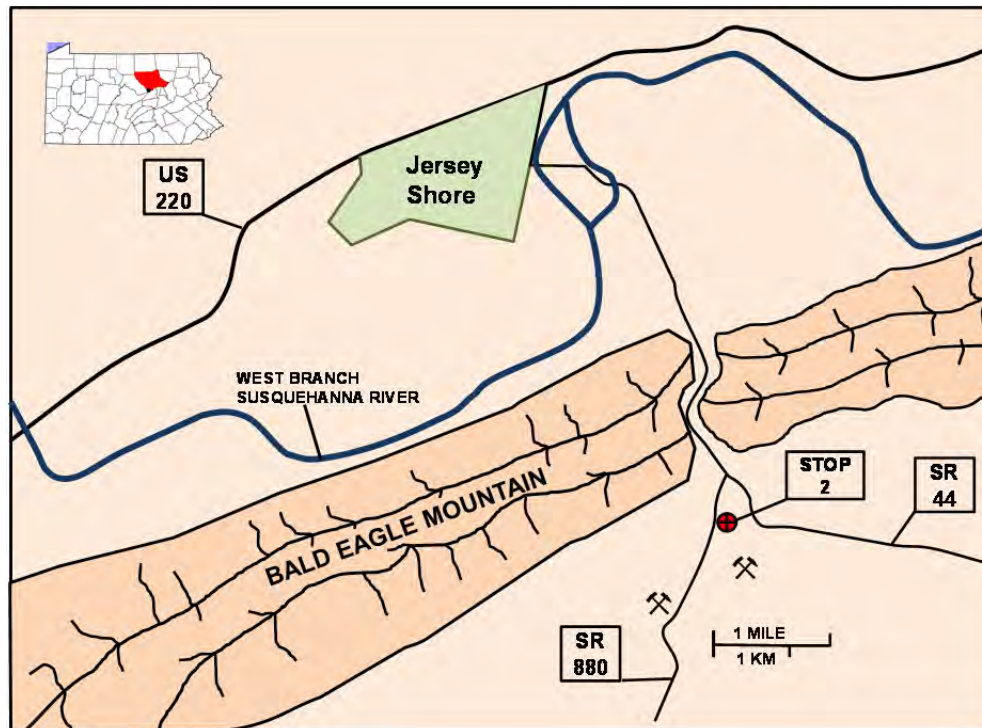4. The sinkholes occur in a preferred alignment.

Figure 2-1. Location Map

Logistics

We will depart from the busses and follow the marked path westward across the field into the woods and meet at the sinkhole designated as SH1 for a brief discussion. You will then be let loose to view the sinkholes, cross the bridge and end up at the Nippeno Spring. We will then return to the busses. We will spend approximately 1.5 hours at this stop.

Refer to the sketch map as a general guide (Figure 2-2).

Cautionary notes: We have permission to go down into sinkhole SH2. The steep slopes that form the sidewalls of the sinkhole tend to reside in a state of perpetual dampness and when coupled with loose rock and vegetation, **may not** provide the best environment for good footing. The odds are fair that one will end up on their rear at some point on the way down or encounter poison ivy along the rim on the way up. Just

Kochanov, W.(2013) Stop 2- Nippeno Spring and sinkholes, *in*, Fleeger, G.M., Schmid, K., and Anthony, R., eds., A Tale of Two Provinces: the Nippenose Valley and Route 15 Corridor, 78th Annual Field Conference of Pennsylvania Geologists, Williamsport, PA, pp. 109 – 115.

telling you to be aware of your steps if you decide to go down into the sinkhole and to be careful of what vegetation you grab onto on the way out.

The marked pathways around the sinks provide good vantage points for observations. Stick to the marked trails and try to travel with group. Do **not** go south from SH1 (the highwall of the Hanson Quarry is nearby) and do not go wandering off to the west (over the hill) from sinkhole SH1. There is nothing down there but spiders and uncharted sinkhole shafts covered with leaf litter.

There are natural "bridges" separating the sinkholes. The one separating sinkholes SH3 and SH4 tends to have a moderate downhill component before it flattens out in the middle. Footing can be tricky if the surface is wet so take your time; there are no hand rails and it is a good 40+ foot drop on either side. The bridge separating SH2 and SH3 is only accessible from the south end. It is not as steep as the other and is easier to navigate.
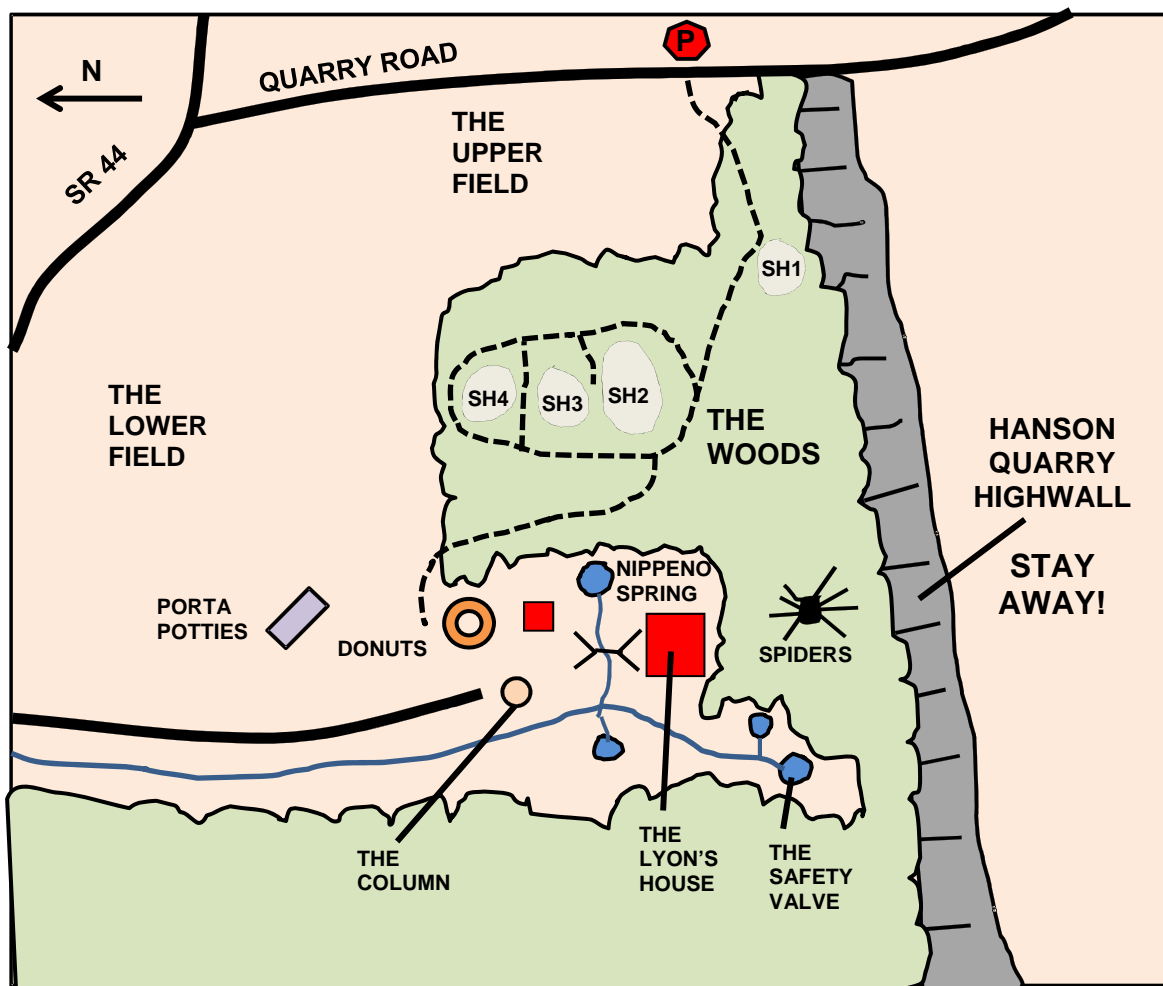

Figure 2-2. Sketch map for Stop 2. Not to scale

Geologic Setting

The Nippenose Valley is a breached anticline. From the Valley rim to its floor, the stratigraphic succession ranges from the more durable sandstones of the Tuscarora, Juniata, and Bald Eagle Formations, down the colluviated slopes to less resistant shales of the Reedsville and Antes Formations, until coming into contact with the predominantly soluble limestone beds of the Coburn through the Bellefonte Formations at its core (see stratigraphic column on inside of the front cover for reference).

The structural imprinting of stress during various times in the geologic past has determined the pathways for surface and ground water in the form of fractures.

Karst Features

The Reading Howell map of 1792 and the revised 1811 map are the first to show streams draining into sinkholes within the Nippenose Valley (Figure 2-3). Present-day lidar imagery has increased the resolution revealing many more of these insurgencies (Figure 2-4).

The lidar imagery also shows that swallets, as well as sinkholes, are more abundant along the northern half of the Valley as opposed to the south. It may be that the amount of alluvial cover on the southern half has masked the underlying karst surface. However, surficial mapping by Faill and others (1977) do not recognize thick colluvial deposits within the Valley. The mapped stoney colluvium is generally less than a meter thick and its location does not coincide with the locations of the swallets and sinkhole swarm.

It is interesting to note on the lidar image that the drainageway that runs east-west down the approximate center of the anticlinal valley (Figure 2-4) has configurations that have characteristics combining straight and meandering channels.

This pattern of stream development can be related to a form of sinkhole piracy whereby sinkholes occurring within the channel of a stream, begins the meandering process by directing the flow from side to side. These sinkhole deflections can be accentuated during periods of flood causing overbank waters to connect with sinkholes that may be present in the floodplain. Repetitive cycles of flooding and waning "connect the dots" eventually establishing a pattern of anastomosing and straight, segmented channels (Kochanov, 2005).

Sinkhole 1 (SH1)

This sinkhole (Figure 2-2) stands apart from the local group. From the northern and southern vantage points this near vertical shaft opens to a depth of approximately 20 meters. The elongate outline of the sinkhole coincides with the dominant joint orientation for the general area at N15-20E. Cave openings are also present at the north and south ends. Cave-roof breakdown has occurred and is representative of an intermediate phase of shaft development.

Recent sinkhole activity has been taking place over the past few years south of SH1 (over the hill down with the spiders). These sinkholes are at a closer proximity and elevation to the Safety Valve Spring which may make them more influenced by local hydrologic changes. There was significant flooding from Tropical Storm Agnes in 1972 (W. Welshans, pers. comm., 2013) and it may be that some of the surface and subsurface plumbing had been altered as a result leaving the local karstic drains more susceptible to fluctuations in water levels and more prone to sinkhole occurrences.

Sinkhole 2 (SH2)

Sinkhole SH2 (Figure 2-2) is much larger than the other sinkholes and is likely the result of the coalescing of several sinkholes. This sinkhole is somewhat different from the other sinkholes in that it has a large block on its floor at the north end. It is apparent that this block detached from the sidewall. This may represent a more advanced stage of sinkhole development, i.e., post shaft. In this scenario, once the shaft has breached the surface, the sinkhole expands in diameter as the sidewalls continually spall off. Eventually the sinkhole expands towards nearby sinkholes narrowing the separating bridges until these septa are removed over time and the sinkholes coalesce. After the sinkholes have coalesced, erosional and weathering would continue to work along discontinuities to destabilize the existing sidewalls and promote rock fall.

SH2 also has a cave opening at its north end hidden by the large block. If one goes down into the sinkhole, one can see into the next sinkhole (SH3) through the cave opening.

Figure 2-3. The "Nepanose" Valley shown on the Reading Howell map of 1811. The map shows stream segments starting at the base of the ridges and ending mid valley in circular-shaped sinkholes. Image is from http://www.mapsofpa.com/antiquemaps31.htm
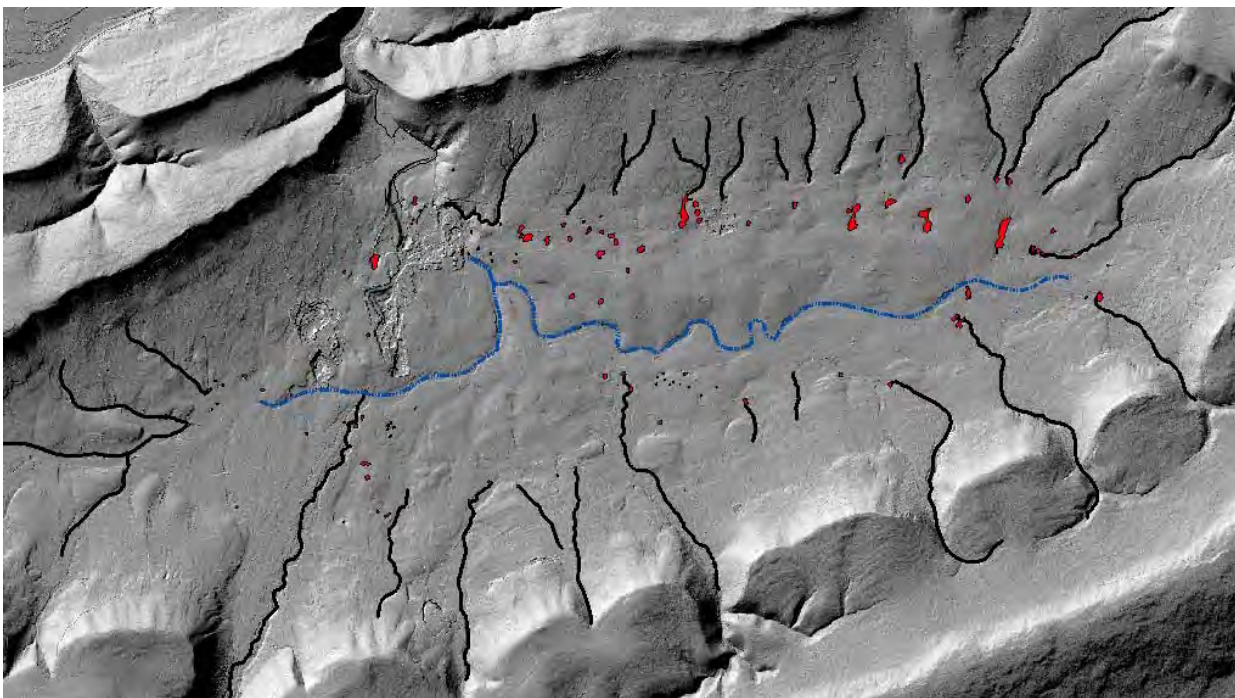

Figure 2-4. Lidar image of the Nippenose Valley showing valley-wall streams outlined in black, central drainage in blue, and sinkholes in red. Note the increase of larger sink structures on the north side of the Valley. North is up, scale 1:50,000.

View from the Bridge

The view southward from the bridge between sinkholes SH3 and SH4 provides an excellent vantage point to see a portion of the breached limb of an anticlinal fold (Figure 2-5). The moss-coated beds of limestone draw one's eyes towards the central vee of the breached fold.

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Figure 2-5. Breached fold of sinkhole SH3. Also visible is the bridge (B) separating SH2 and SH3, and the cave connecting to sinkhole SH2 (C).

Sinkhole 3 (SH3)

Peering down into this sinkhole one notes that the individual beds exposed along the sidewalls show more angularity as compared to sinkholes SH2 and SH3. It may be that this sinkhole (called the wishing well) is the youngest of the series of sinkholes SH2-4.

Sinkhole Formation

Shaft formation is outlined in Hess (2005) and serves as an appropriate model in this instance (Figure 2-4). The process progresses once certain conditions already exist, i.e., the presence of a fracture zone and a cave with a stream flowing through it (Figure 2-6a). Stoping, the upward propogation of a void as a result of rock failure and the subsequent breakdown of the ceiling rock, takes over as the primary mechanism in the evolution of the shaft (Figure 2-6b). The stoping process can be best seen in sinkhole SH1. Once the thicker beds have collapsed, it would be much easier for the unsupported, more thin-bedded Coburn/Salona beds to fail due to gravity.

A   **B**   **C**
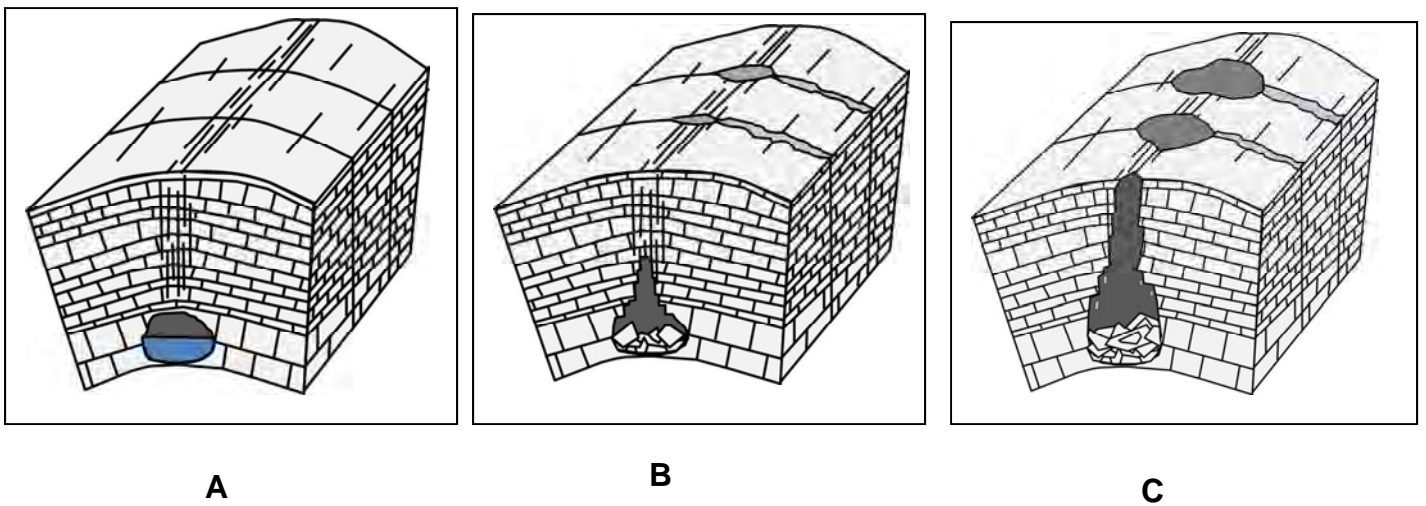
Figure2-6. Shaft development. Modified from Hess (2005).

This repeated breaking and collapse process continues until the failed roofs breach the land surface (Figure 2-6c). As the collapse pipes breach the land surface, the shafts then serve as a localized drain (swallet) for more chemically aggressive surface waters and enhance the dissolution of the broken limestone bedrock at the site of a sinkhole.

All of the sinkholes show some connection to the other through an established cave system that trends along the regional fracture patterns. It is likely that the folding of the Nippenose anticline as well as the movement of the nearby St. James Fault Complex (Faill and others, 1977) provided the regional impetus for creating the dominant fracture systems in this part of the Valley. Subsequent fracturing through the "unroofing" of the Valley, would have provided additional flowpaths for groundwater and the development of caves and further the mechanical development of sinkholes within the Nippenose Valley at this location.

The Nippeno Spring

Note: At the request of Mr. Lyon, there are to be no more than 15 individuals on the bridge at any one time while viewing the spring.

The Nippeno Spring (also the Big Spring, the Enchanted Spring, Figure 2-7) is reported to be Pennsylvania's largest with a median rate of 18,000g/gpm (Flippo, 1974). Additional discharge from the nearby Old Safety Valve Spring to the south and adjacent smaller unnamed springs contribute to creating the headwaters of Antes Creek. The discharge from these smaller springs responds to more seasonal peaks, such as winter thaw, and storm events in comparison to the Nippeno Spring where discharge is relatively constant.

Figure 2-7. The Nippeno Spring.

The story of the Nippeno Spring starts as the surface water draining into the closed valley, comes into contact with the carbonate bedrock. From there it enters the sinkhole drains and goes subsurface with the majority of these swallets occurring within mapped portions of the Linden Hall Formation (Faill and others, 1977). This subsurface flow travels down gradient westward and resurfaces at the Nippeno Spring; the primary discharge point for the entire Nippenose Valley (Figure 2-7). A tracer dye was injected into the Sawmill Run sinkhole approximately 6 km to the east of the Nippeno Spring (Loges, 1997) with breakthrough occurring approximately 6 to 7 hours later.

It has been reported that the discharge at the Nippeno Spring does not match the recharge volumes. Based on a water budget study (Loges, 1995) and groundwater chemistry (Ulmer and Sasowsky, 2001) it has been suggested that recharge to the Nippeno Spring may be contributed from sources outside of the Nippenose Valley;). Nearby Sugar Valley to the south has a 150 m higher base level elevation than the Nippenose Valley (Loges, 1995) and is connected through the same regional anticlinal structure. Groundwater flow from regions of higher to lower hydrostatic pressure provides some plausibility to the argument.

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STOP 3: Reedsville Formation and Antes Shale at Antes Gap, Lycoming County

Leader: John A. Harper

Introduction

Antes Gap, a very picturesque water gap formed where Antes Creek carved an entrenched, meandering channel through Bald Eagle Mountain, is a classic locality that exposes Upper Ordovician shales and siltstones of the Reedsville Formation and Antes Shale (Figure 1). It is here that Kay (1944) named the Antes Shale, and here that Hoskins and others (1983) designated Lycoming County's official state fossil collecting locality.

Figure 2 is a generalized structural and stratigraphic cross section of this part of Lycoming County showing the major formations exposed in Bald Eagle Mountain and the Nippenose Valley. Although the rocks that crop out along Antes Creek in Antes Gap range from the lower sandstone beds of the Bald Eagle Formation to the carbonates of the Salona Formation, at Stop 3 we will be looking only at the Reedsville Formation and Antes Shale, which dip 31° NE and strike N58°E (Gross, 1955).
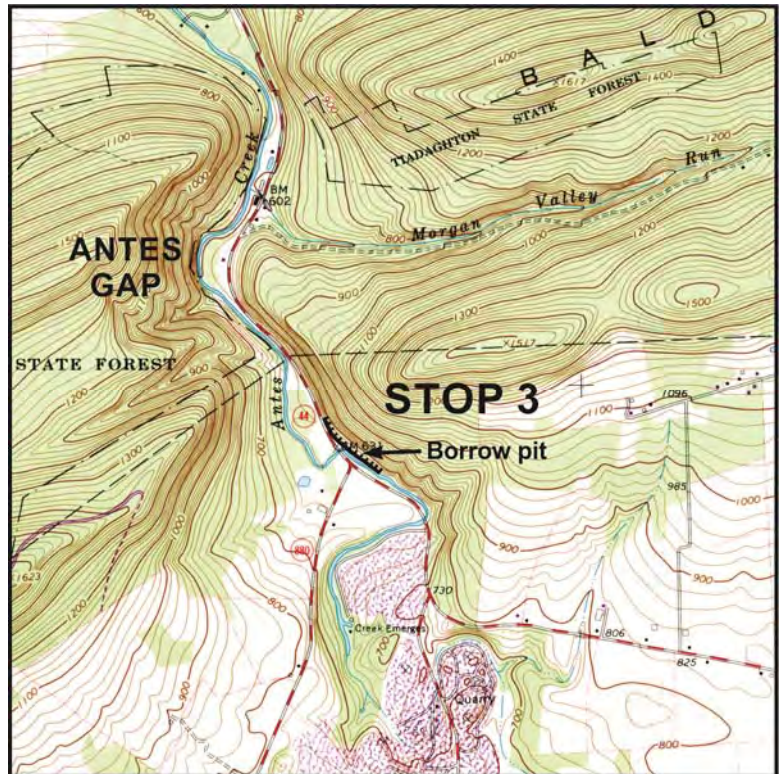

Figure 3-1. Portion of the Linden 7.5-minute topographic map showing the location of stop 3 at Antes Gap, Lycoming County
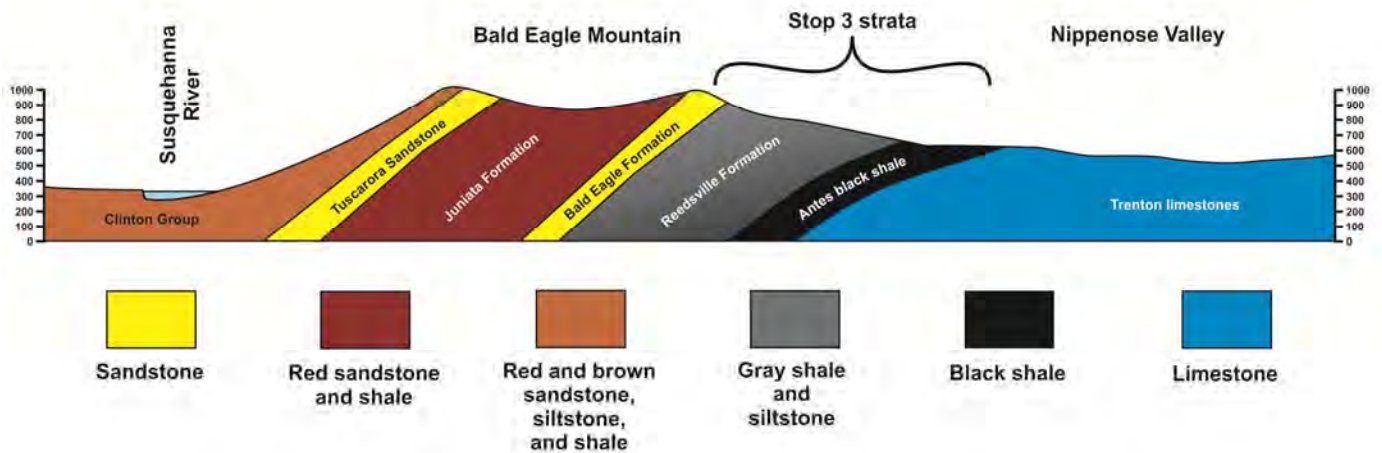

Figure 3-2. Generalized structural and stratigraphic cross section of Bald Eagle Mountain and Nippenose Valley in the vicinity of Stop 3 (redrawn from Sherwood and Platt, 1880).

Late Ordovician Finer-Grained Clastics

   The Reedsville Formation and Antes Shale at Antes Gap represent just one small exposure of a sequence of generally fine-grained, often calcareous clastic rocks that carry an indisputable Late Ordovician fauna. Tracing the Reedsville/Antes lithologies around the Ordovician outcrop is quite challenging because they are easily deformed tectonically and rarely crop out. These shales weather easily into clayey loam with a few shale chips (Pierce, 1966; Faill and Wells, 1977), and on the slopes of mountains they tend to be covered by colluvium. Tracing them around the Appalachian basin, especially in the subsurface, is not quite as challenging, despite the inherent confusion surrounding "state-line faults", because they are easily recognizable in drill cuttings and on geophysical logs.   See, for example, Harper (this guidebook, fig. 2) for the correlation of the Reedsville and Antes through adjacent Appalachian states, and Figure 3 for the correlation of these rocks within Pennsylvania.

   Gross (1955) studied this locality and documented the lithology and paleontology of the Reedsville Formation and Antes Shale. It should be noted that he disagreed with Kay's (1944) consideration of the Antes as a separate formation. Instead, Gross (1955) described the Reedsville Shale as divisible into an upper "Antes Creek Shale and Siltstone Member" and a lower "Antes Black Shale Member". Figure 4 illustrates his measured section at the Antes Gap exposure from the Bald Eagle Formation back in the gap to the last exposure of Antes black shale in an abandoned borrow pit across Antes Creek from the intersection of PA 44 and PA 880 (see Figure 1).

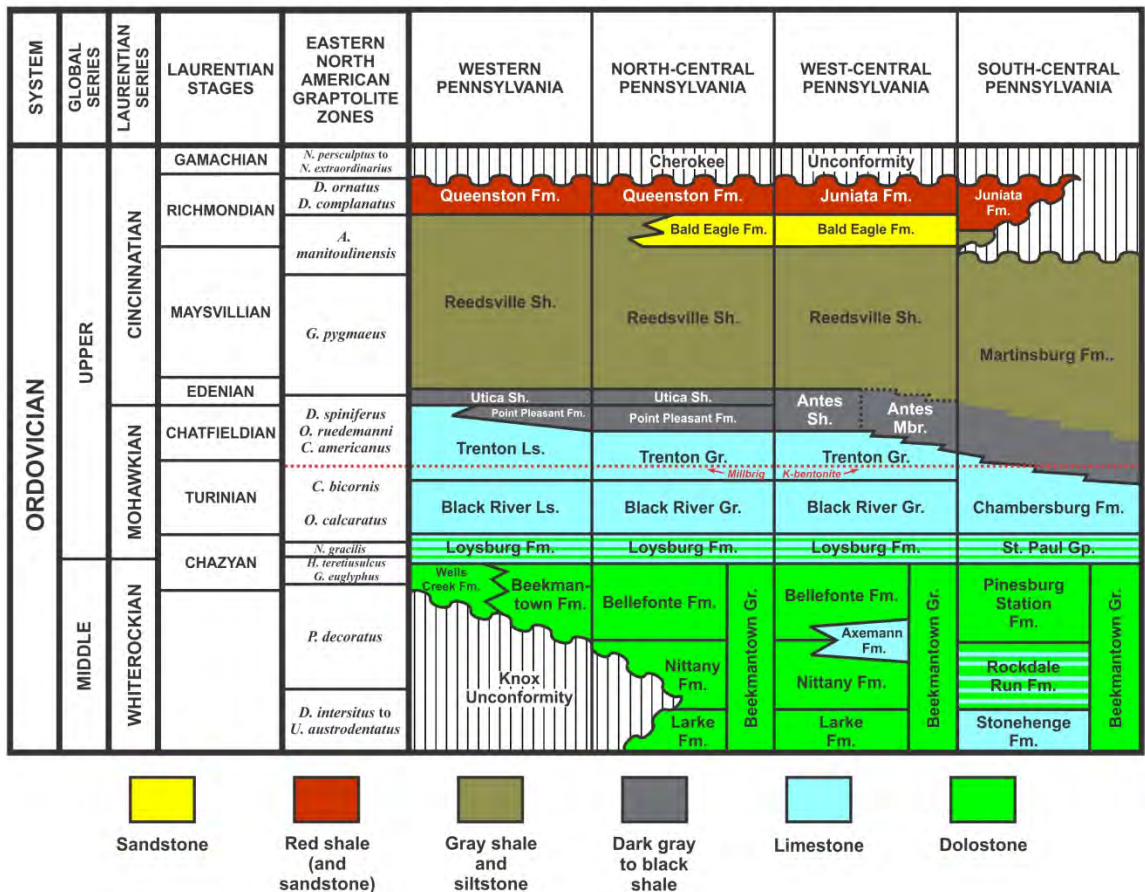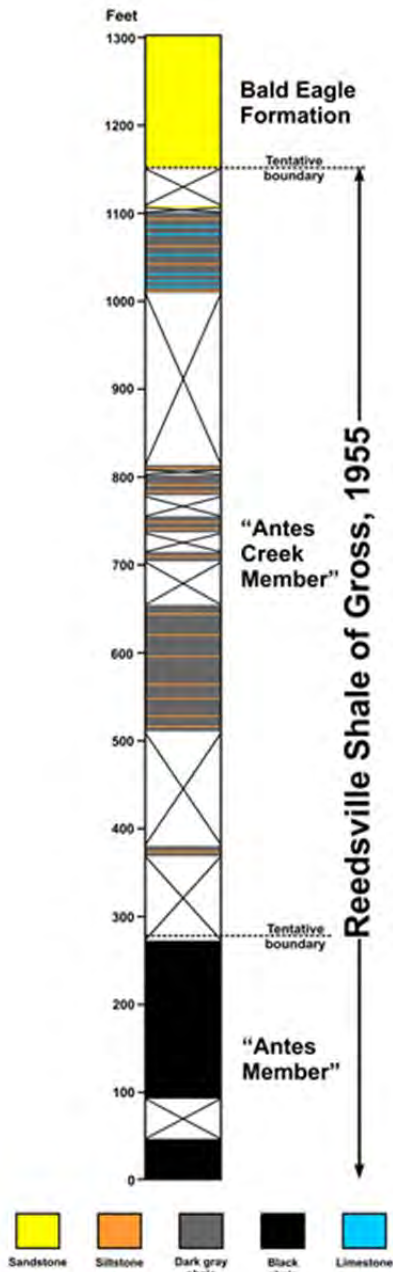Figure 3-3. Correlation of Upper Ordovician strata in Pennsylvania.

Figure 3-4. Generalized stratigraphic section of the Reedsville Formation and Antes Shale at Stop 3 (redrawn from Gross, 1955, fig. 4).

Reedsville Formation

Ulrich (1911) named the Reedsville Formation for exposures of dark gray shale and minor sandstones that are sandwiched between the sandstones of the Bald Eagle Formation and the carbonates of the Trenton Group (Coburn Limestone) at Reedsville in Mifflin County, PA. The First and Second Pennsylvania Geological Surveys (Rogers, 1858; Lesley, 1892) had followed New York's terminology to a point, calling the upper dark gray shales "Hudson River shales" (or slates) and the lower black shales "Utica shales" (or slates).[1] Since Ulrich's (1911) designation, 20th-century geologists working in central Pennsylvania have used the name Reedsville, and this name has found its way into the subsurface in numerous publications (e.g. Fettke, 1961).

Butts and Moore (1936; also Butts and others, 1939; and Butts, 1945), described the Reedsville as about 1,000 ft (305 m) of dark calcareous shales with thin layers of fossiliferous limestones. Gross (1955) and Washington (2009) included thin sandstones or siltstones that increase in thickness and coarseness upward, becoming more than 1.5 ft (0.5 m) thick near the top. The upper 50 ft (15 m) is gradational with the overlying Bald Eagle Formation (Butts and Moore, 1936; Washington, 2009). This part of the Reedsville, characterized by the brachiopod *Orthorhynchula linneyi* (James), is the *Orthorhynchula* zone of many authors (e.g., Woodward, 1951).

Based on outcrop studies published since 1945, it appears the Reedsville thickens to the south and east. It ranges from about 900 ft (275 m) thick in the northern part of the Ridge and Valley (Lycoming County – Faill and Wells, 1977) to about 2,500 ft (762 m) in the southern Ridge and Valley[2] (northwestern Fulton County – Pierce, 1966, although there is some question as to whether these rocks should be called Reedsville or Martinsburg – see Figure 3). The Reedsville thickens

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[1] Rogers (1858), following his bizarre method of using Latin hours of the day for stratigraphic nomenclature, actually designated the Reedsville Formation and Antes Shale as "Matinal shales" and "Matinal black shales", respectively. But he did recognize that his names correlated to the Hudson River and Utica of New York.

[2] Faill and Wells (1977) separated the Reedsville and Antes, whereas most other authors did not. Therefore, to be perfectly honest, an additional 330 ft (100 m) should be added to the Reedsville in Lycoming County in order to properly compare thicknesses across the state.

from about 1,000 ft (305 m) near the Allegheny front (Butts and Moore, 1936; Butts, 1945; Doden, 2005; Doden and Gold, 2008) to about 1,800 ft (549 m) in the central depocenter of Huntingdon, Mifflin, and Juniata counties (McElroy and Hoskins, 2007; 2008). In the subsurface, the Reedsville thickens from about 670 ft (204 m) in the northwestern corner of Erie County to over 1,600 ft (488 m) in western Bedford County (Wagner, 1958; Fettke, 1960, 1961).[3]

The Reedsville is equivalent to the Lorraine Shale of New York, the Kope Formation and lower part of Cincinnati Group of Ohio, and most of the upper Martinsburg Formation of southeastern/south-central Pennsylvania, Maryland, Virginia, West Virginia, and eastern Tennessee (see Harper, this guidebook, fig. 2).

The Reedsville Formation, although not overly fossiliferous, has a relatively abundant fauna that can be found, typically, in the scattered limestone and siltstone beds (Figure 4). Table 1 lists, and Figure 5 illustrates, the fossils that were found in the Reedsville at Antes Gap by Gross (1955) and Hoskins and others (1983). If you search the outcrop, you just might actually find some of these.

Table 3-1. Fossils from the Reedsville Shale at Antes Gap and nearby localities (also see Figure 5). Taxonomy has been substantially updated from original nomenclature.

Bryozoa
   Hallopora sigillaroides (Nicholson)
Brachiopoda
   Schizocrania filosa Hall
   Craniops cincinnatiensis (Hall)
   Rafinesquina alternata (Conrad)
   Sowerbyella rugosa (Meek)
   Cincinnetina multisecta (Meek)
   Cincinnetina meeki (Miller)
   Paucicrura rogata (Sardeson)
   Diceromyonia ignota (Sardeson)
   Orthorhynchula linneyi (James)
   Zygospira modesta (Say)
   Zygospira cincinnatiensis James
Bivalvia
   Ambonychia radiata (Hall)
   Ambonychia sp.
   Modiolopsis sp.
   Lyrodesma poststriatum Emmons
   Nucularca pectunculoides (Hall)

Rostrochonchia
   Ischyrinia cancellatus Ruedemann
Gastropoda
   Cyrtolites ornatus Conrad
   Sinuites cancellatus (Hall)
   Sinuites granistriatus (Ulrich)
   Bellerophon sp.
   Liospira sp.
   Hormotoma gracilis (Hall)
Cephalopoda
   Paractinoceras sp.
   Michelinoceras sp.
Trilobita
   Cryptolithus lorrainensis Ruedemann
   Flexicalymene meeki (Foerste)
Crinoidea
   Stems and columnals
Graptolithina
   Geniculograptus typicalis (Hall)

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[3] Because the Reedsville and Utica/Point Pleasant (Antes equivalent) can be recognized individually in the subsurface, the thicknesses of the two (three) formations should be added together for proper comparison with surface exposures in central Pennsylvania.
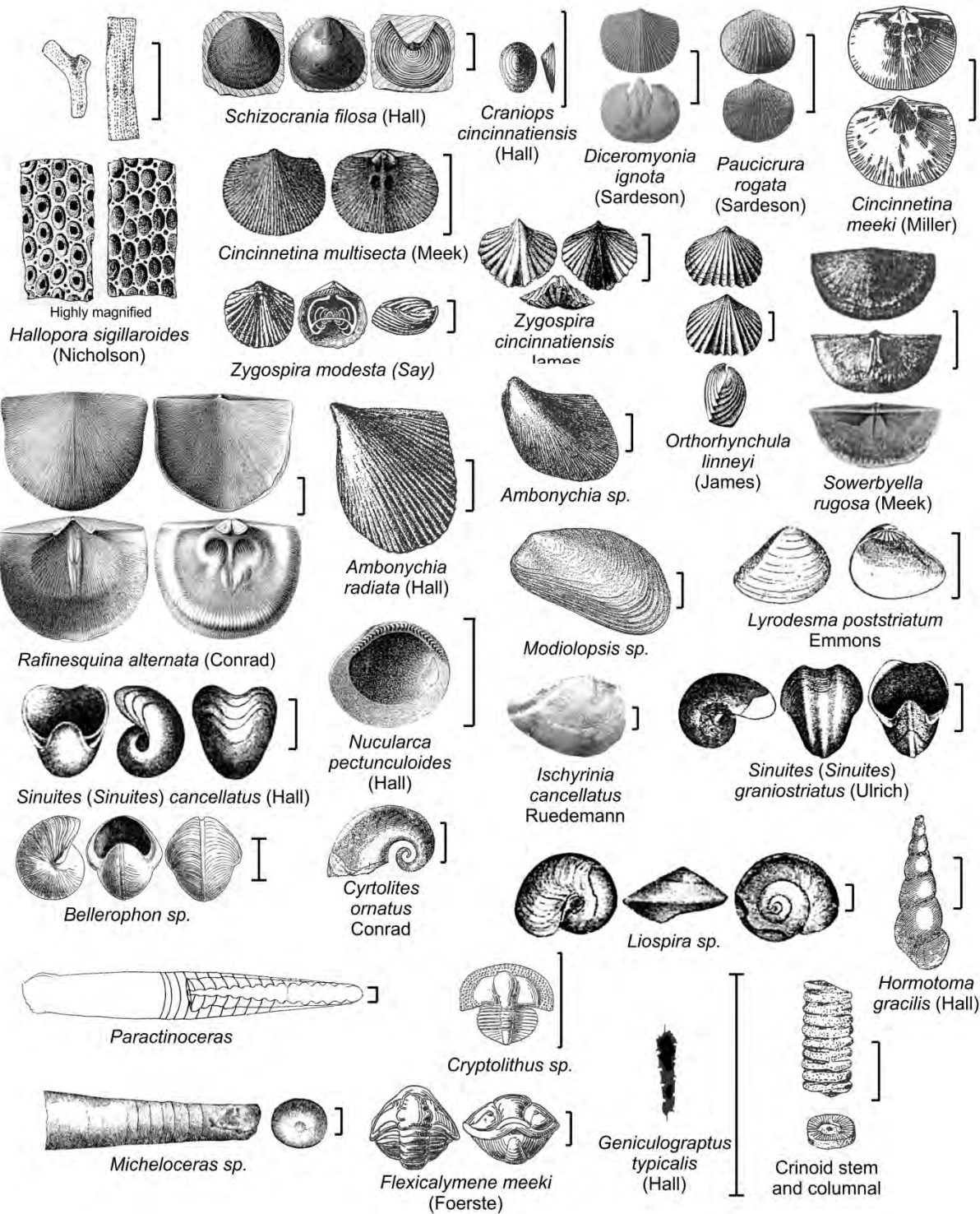
Figure 3-5. Fossils documented from the Reedsville Formation at Antes Gap and nearby localities. Illustrations from various publications, primarily from Meek (1873), Hall and Whitfield (1875), Nicholson (1875), Miller (1889), Lesley (1889-1890), Bassler (1919), Schuchert and Cooper (1932), Moore and others (1952), and Hoskins and others (1983). Bar scales = 10 mm.

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Antes Shale – Central Pennsylvania's Contribution to the Utica Facies

Kay (1944, p. 114) named the Antes Shale for about 400 ft (122 m) of dark-gray to black shale (Utica facies) lying between the Coburn Limestone and the Reedsville Formation along Antes Creek at Antes Gap, Clinton County, Pennsylvania (this stop). See Harper (this guidebook) for a discussion of the Utica facies rocks in the Appalachian basin. Faill and Wells (1977) determined that, lithologically, the Antes comprises about 330 ft (100 m) (somewhat less than thickness measured by Kay, 1944) of dark gray to black, fissile, massive- to thin-bedded, calcareous shale with interbeds of calcilutites, calcisiltites, and slightly fossiliferous calcarenites in the area of the type locality in north-central Pennsylvania. Limestone beds are concentrated in the middle of the formation where they comprise about 25 percent of the lithology. Only about 10 percent of the lower Antes consists of limestone, and the upper beds have hardly any at all. Faill and Wells (1977) found that the contact with the underlying Trenton was gradational over about 50 ft (15 m), unlike in New York where the contact in distinctly disconformable; the upper contact with the overlying Reedsville Shale occurs more rapidly as a change through about 10 ft (3 m) from black calcareous shale to olive-gray, non-calcareous, silty shale (Faill and Wells, 1977). A small borrow pit on the hillside across the creek from the intersection of PA routes 44 and 880 exposes typical Antes black shale (Figure 6).

Figure 3-6. Photograph of the Antes Shale in the borrow pit at Stop 3. Intern for scale.

Like most geologists working the Appalachian Ordovician in Pennsylvania, Butts and Moore (1936) and Butts and others (1939) did not separate the lower, very thin, brown to black shale containing the trilobite *Triarthus eatoni* (Hall) and graptolites from the upper sandy shales when describing the Reedsville Formation in the Bellefonte and Tyrone areas of Centre, Blair, and Huntingdon counties. Butts and Moore (1936) did, however, describe portions of the formation as being so highly carbonaceous and graphitic that it was once prospected for coal. The Antes is not as fossiliferous as the Reedsville. Table 2 and Figure 7 indicate the fossils that have been found in the Antes Shale at Stop 3. Have fun looking for them in the black shales.

Table 3-2. Fossils from the Antes Shale at Antes Gap (also see Figure 7). Taxonomy has been substantially updated from original nomenclature.

Brachiopoda

    Lingulella? riciniformis Hall
    Leptobolus insignis Hall
    Leptobolus latus Ruedemann
    Leptobolus walcotti Ruedemann
    Dalmanella sp.
    Sowerbyella sp.

Trilobita

    Triarthus eatoni (Hall)

Graptolithina

    Dicranograptus nicholsoni Hopkins
    Rectograptus amplexicaulis (Hall)
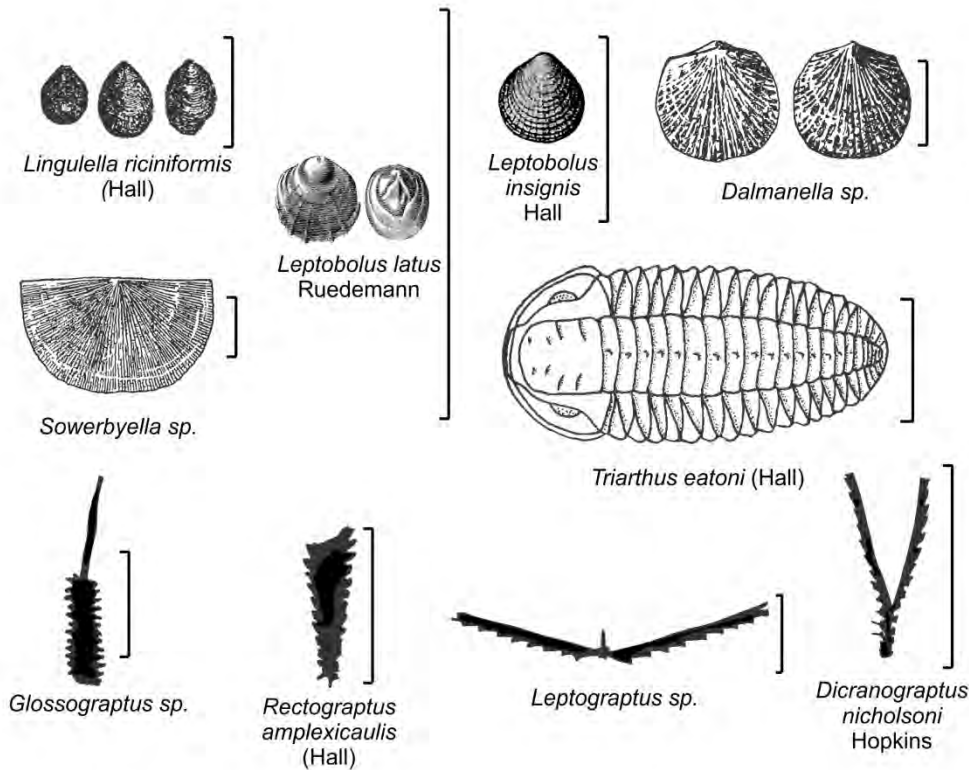   *Glossograptus* sp.
    Leptograptus sp.

Figure 3-7. Fossils documented from the Antes Shale at Antes Gap and nearby localities. Illustrations from various publications, primarily from Meek (1873), Lesley (1889-1890), Bassler (1919), Ruedemann (1925), Moore and others (1952), and Hoskins and others (1983). Bar scales = 10 mm.

Formation, Member, or What?

    Assuming you agree with Kay (1944), the Antes is the only formally named representative of the Utica facies in outcrop in Pennsylvania. It has definitely been recognized in the northern part of central Pennsylvania from Centre County northeast to Lycoming County (Thompson, 1963; Faill and Wells, 1977) and southeast to Mifflin County (Thompson, 1963). In fact, one of the best exposures of the Antes occurs at Reedsville, Mifflin County (Thompson, 1963; McElroy and others, 2007). Most workers, however, do not distinguish the Antes as a separate formation. Faill and Wells (1977) were unusual in calling it "Antes Formation". In fact, the name "Antes" is rarely used at all outside of the type area. Reports typically point out that the Reedsville contains black shale, rather than dark gray, and commonly has interbedded limestone beds toward the base. For example, Knowles (1966) described the basal Reedsville beds in Bedford County as 1- to 2-in (2.5- to 5-cm)-thick layers of dark-gray, fine-grained limestone alternating with 2- to 4-in (5- to 10-cm)-thick layers of calcareous shale. This same type of nomenclatural lack occurs to the south and east where the name "Reedsville Formation" is replaced by "Martinsburg Formation".

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So, should the Antes be considered a separate formation, as Kay (1944), Thompson (1963), and Faill and Wells (1977) indicated? If not, should it be considered a formal member of the Reedsville Formation, as suggested by Gross (1955)? Or should it be considered nothing more than a darker colored, less coarsely clastic facies at the base of the Reedsville as most authors have implied over the last 100 years? Before you make a decision, take a look at the outcrop at Stop 3 and read up on the Utica facies formations of New York State (for example, Lehmann and others, 1995; Baird and Brett, 2002; Brett and Baird, 2002). The geologists of the First and Second Pennsylvania Geological Surveys recognized the Antes was equivalent to the Utica shales of New York. Lesley (1892) actually used the name Utica slate (rather than shale, apparently in deference to the economic value of equivalent rocks in eastern Pennsylvania as roofing slates), indicating that, based on lithology and fossil content, it was a separate formation from the Hudson River slate (= Reedsville). As mentioned above, Faill and Wells (1977), in describing the Antes in the local area, found that calcilutites, calcisiltites, and slightly fossiliferous calcarenites were concentrated in the middle of the formation. Could it be that this part of the Antes correlates with the Dolgeville Formation of New York? If so, then the Antes type locality might represent the entire Utica Group of New York, with a lower Flat Creek Shale equivalent and an upper Indian Castle Shale, separated by the more Trenton-like Dolgeville (Figure 8). Perhaps if more decent outcrops could be located and studied, questions such as this would be moot.

Since I spent 35 years working with subsurface stratigraphy, I have found that, in western and north-central Pennsylvania, the Utica facies is easily distinguishable from the Reedsville, both in drill cuttings and geophysical logs. As a result, my preference is for the Antes to be a separate formation, Pennsylvania's outcrop contribution to the Appalachian basin's Utica facies.

And Then There's the Martinsburg Formation

Geiger and Keith (1891) first used the name "Martinsburg shale" in discussing the geology of the Harpers Ferry area of Virginia, but Darton (1892) actually was the first to describe the lithology. The Martinsburg consists of as much as 2,400 ft (732 m) of shale and sandstone in south-central Pennsylvania and adjacent Maryland and gets progressively thicker to the southeast (Woodward, 1951). The lower 650 ft (198 m) of the Martinsburg consists of black carbonaceous and calcareous shales interbedded with limestones in the basal 100 ft (30 m) that contains a distinctly "Trenton" fauna (Bassler, 1919).   The basal 10 ft (3 m) of black, muddy, rubbly to slabby limestone and shale of the Martinsburg has been called the "*Sinuites* bed" or "*Sinuites* zone" for the characteristic bellerophont gastropod (Figure 5) that occurs in it (Bassler, 1919; Craig, 1949; Woodward, 1951). Above this occurs about 100 ft (30 m) of black, platy to fissile, calcareous shale containing a fauna of trilobites (*Cryptolithus*) and graptolites (Woodward, 1951). This is succeeded by about 500 ft (152 m) of black, carbonaceous, fissile, unfossiliferous shale. This is obviously the Utica facies of the Martinsburg. Beares and others (2002) placed these lower beds of the Martinsburg at McConnellsburg, Fulton County, Pennsylvania, in the *Climacograptus bicornis* graptolite zone (Figure 3). Note that Beares and others, 2002 apparently agreed with Pierce, 1966 who referred the Ordovician clastic sequence above the carbonates at McConnellsburg to the "Reedsville" rather than "Martinsburg"; he called the Utica facies here "lowermost Reedsville" rather than Antes. Pierce (1966) could just as easily have used "Martinsburg" instead, but deferred to the existing Geologic Map of Pennsylvania for nomenclatural use in his mapping area.

Biostratigraphic Significance

Graptolites, which comprise the majority of the Antes fauna, are one of the major faunal groups used for Ordovician biostratigraphy. Kay (1944) found that the graptolite, *Dicranograptus nicholsoni* (Hall) (Figure 7), was abundant in the borrow pit at Stop 3. Butts and others (1939) listed *Geniculograptus typicalis* (Hall), *Dicranograptus nicholsoni* Hopkins*, and *Diplograptus v*espertinus (Ruedemann) from the basal

black shale of the Reedsville in the Tyrone area. Goldman and others (1994) considered *Dicranograptus nicholsoni* part of the constituent fauna of the *Diplacanthograptus spiniferus* graptolite zone. This places the Antes (Utica facies) at Stop 3 within the upper Chatfieldian Stage (upper Mohawkian Series) and/or lower Edenian Stage (lower Cincinnatian Series) of the Upper Ordovician (Figure 3). As such, the Antes in the northern part of central Pennsylvania can be correlated with the Lower Indian Castle Shale of New York (Figure 8).

The entire lower Martinsburg of eastern West Virginia and northern Virginia contains numerous K-bentonite beds (Craig, 1949), including one at the base of the Martinsburg that Kolata and others (1996) used chemical fingerprinting to identify as the Millbrig K-bentonite. The Millbrig is a significant ashfall that has been used for isochronous correlation throughout eastern North America and northern Europe. The Millbrig correlates with the top of the *Climacograptus bicornis* graptolite zone throughout eastern North America (Kolata and others, 1996), which places it within the upper Turinian Stage (middle Mohawkian Series) (Figure 3).   This places the basal black shales of the Martinsburg (Utica facies) in eastern West Virginia and northern Virginia in that stage as well. As stated above, Beares and others (2002) assigned the basal "Antes Shale" at McConnellsburg to the *Climacograptus bicornis* graptolite zone. Parris and Cruikshank (1992) also acknowledge that the basal beds of the Martinsburg in New Jersey and eastern Pennsylvania fall within the *Climacograptus bicornis* graptolite zone. Therefore, it appears that the basal black shales/slates of the Martinsburg correlate to the lower Trenton and/or upper Black River.
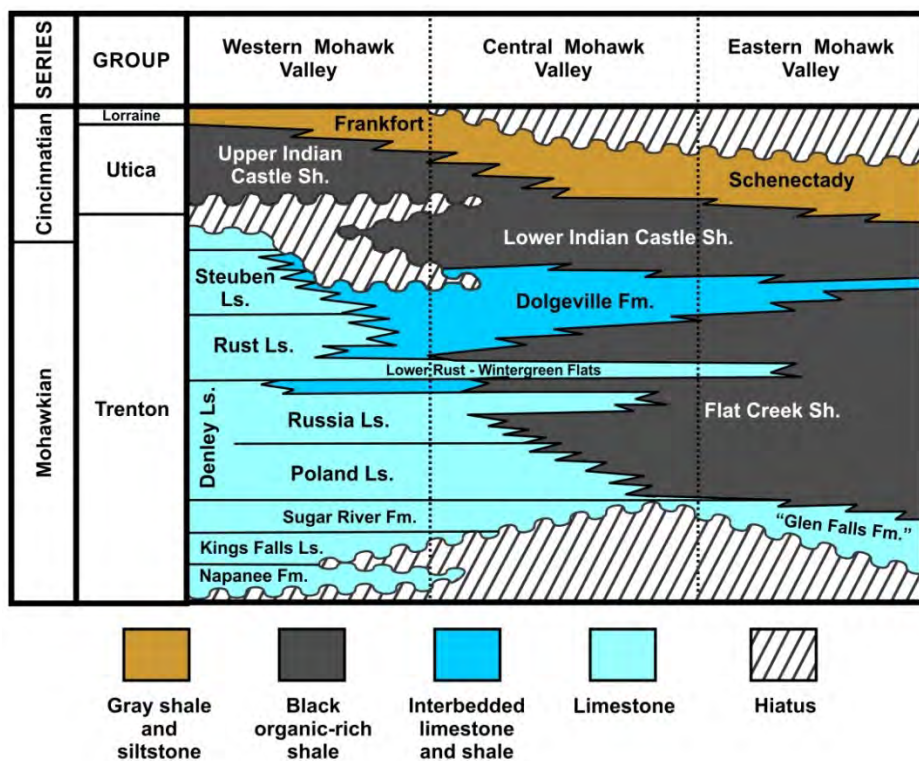

Figure 3-8. Stratigraphic diagram of Utica facies and associated formations in the Mohawk Valley of New York (modified from Brett and Baird, 2002). Is the zone of limestone beds in the middle of the Antes Shale described by Faill and Wells (1977) equivalent to the Dolgeville Formation of New York? If so, the Antes Shale might equate to the entire Utica Group (Flat Creek and Indian Castle shales) of our northern neighbors.

In the classic Union Furnace section adjacent to the Juniata River on the Centre-Huntingdon County border, the Millbrig K-bentonite occurs about 20 ft (6 m) above the contact between the Nealmont and Salona formations (lower Trenton) (Laughrey and others, 2004). At this locality, the Millbrig lies about 300 ft (100 m) below the base of the Antes Shale. This means that the Utica facies of the Martinsburg is significantly older than the Antes Shale in Centre and Lycoming counties. As such, the Martinsburg's Utica facies is actually older than the Flat Creek Shale of New York State (Figure 8), which is in the lower

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Corynoides americanus and *Orthograptus ruedemanni* zones (Goldman and others, 1994). The basal black shales of the Martinsburg, in fact, are most likely the oldest Utica facies rocks in the Appalachian basin!

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Stop 4: Lockport

Leader: Dr. Thomas C. Wynn, Lock Haven University

The outcrops at this stop record the first phase of the Catskill Delta Complex and range in age from Middle Devonian to Upper Devonian *(Figure 4-1)*. The formations present are as follows in stratigraphic order: Mahantango Formation (Shale Member and Tully Limestone Member), and the Harrell Formation, which is composed of the Burket Black Shale Member and the Upper Shale Member. The Brallier Formation overlies the Upper Shale Member of the Harrell Formation, but will not be looked at during this stop.
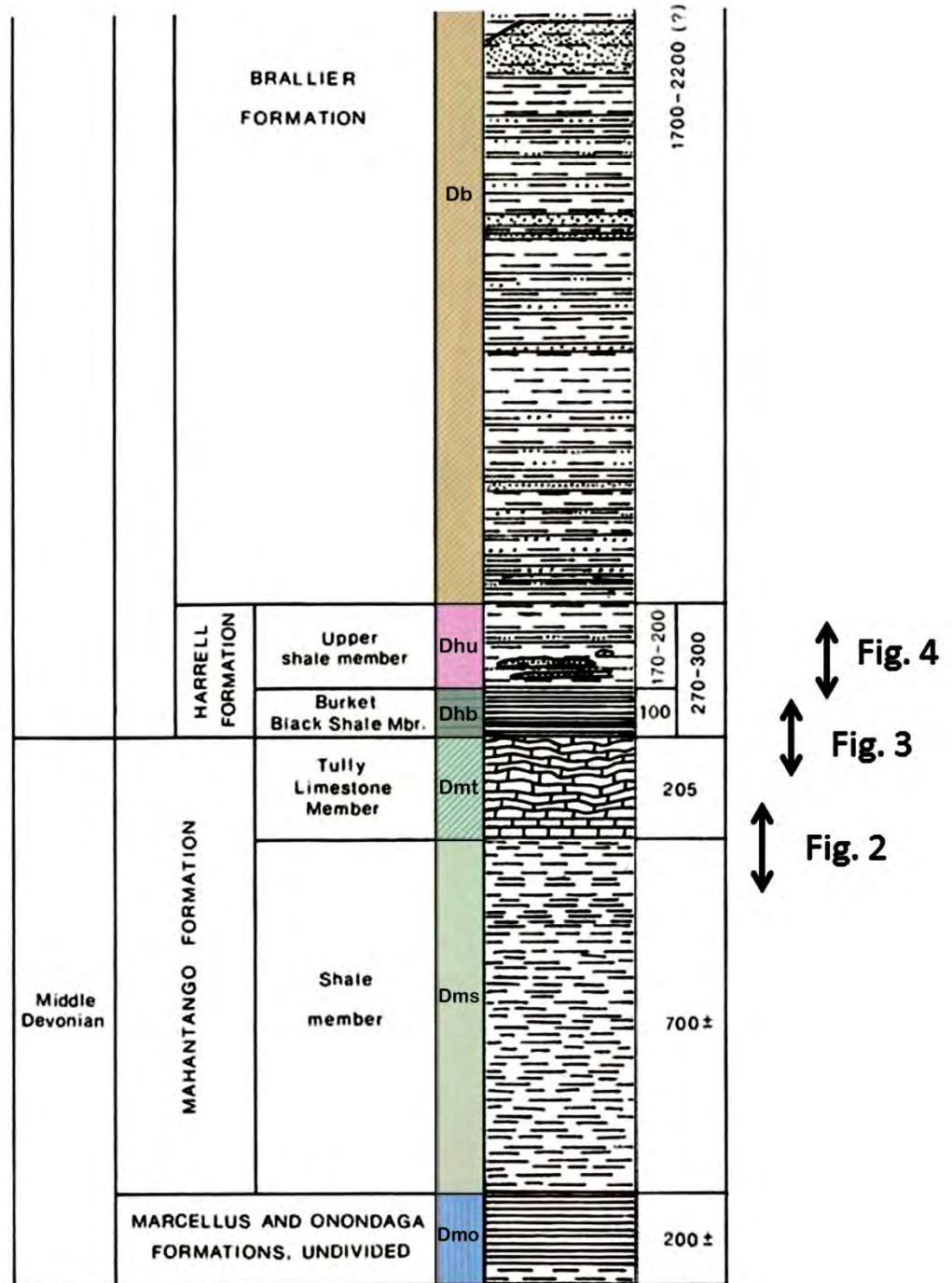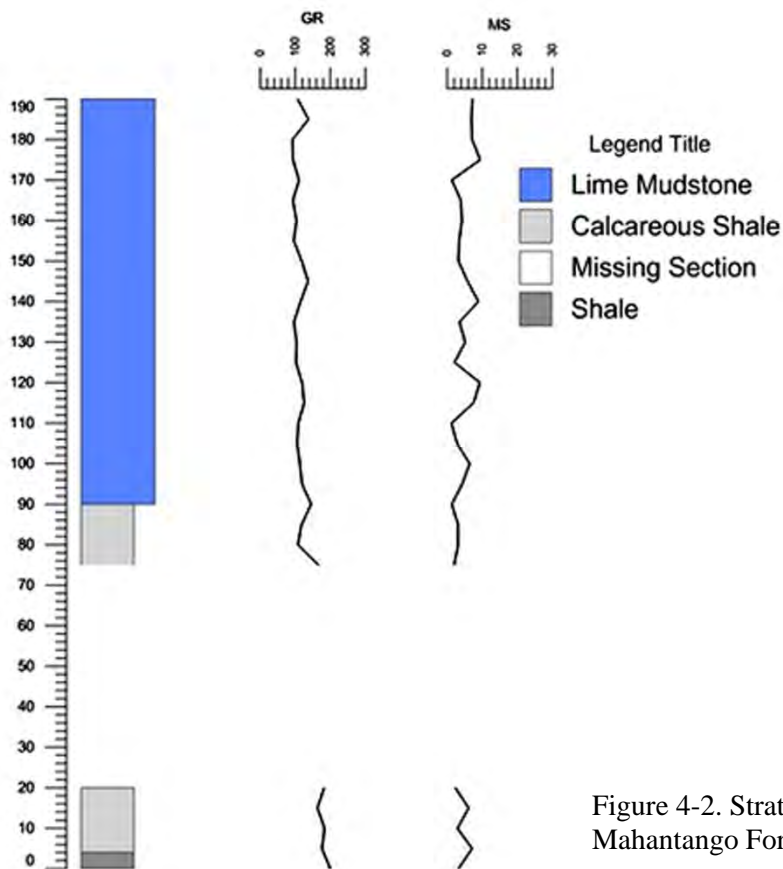

Figure 4-1. A portion the stratigraphic column from the Lock Haven Geologic map showing the location of the detailed stratigraphic columns for the stop (Modified from Taylor, 1977).

The Shale Member of the Mahantango Formation is composed of non-calcareous, platy, unfossiliferous gray shale, which grades into calcareous platy, unfossiliferous gray shale behind the Woodward Township sign. This change to calcareous shale is thought to be the gradational boundary from the Upper Shale Member to the Tully Limestone Member. The change to calcareous shale is thought to indicate a rise in sea-level. Siliciclastic materials where trapped farther to the East near shore as sea-level rose. Calcareous material is thought to have been transported from a carbonate platform located to the north in Central New York (Brett and Baird, 1985). The calcareous shale is interbedded with thin lime mudstones near the base of the Tully Limestone Member (Figure 4-2).

Figure 4-2. Stratigraphic column of transition from Shale Member of the Mahantango Formation to the Tully Limestone Member.

The lower portion of the Tully Limestone Member is barren of fossils, with the exception of rare trilobites and echinoderm fragments. The section of lime mudstone is believed to be the deepest water phase at this stop. Near the top of the Tully Limestone Member the lime mudstone grades into interbedded wackestone and argillaceous wackestone (Figure 4-3). This portion of the Tully is very fossiliferous. Fossils weather out from the wackestones and can be found on the "shaley" slope. Brachiopods, trilobite fragments, echinoderm fragments and corals are commonly found in this location. At the top of the Tully there is a thin layer of calcareous shale followed by a covered section. The covered section is most likely due to weathering of the calcareous shale. Above the covered section lies the Burket Black Shale. The Burket Shale Member is composed of very fissile black shale, which contains evidence of scouring. The abundant pyrite in the shale is why the Burket weathers a yellowish-brown color. Limestone nodules are found in a bed near the top of the Burket.
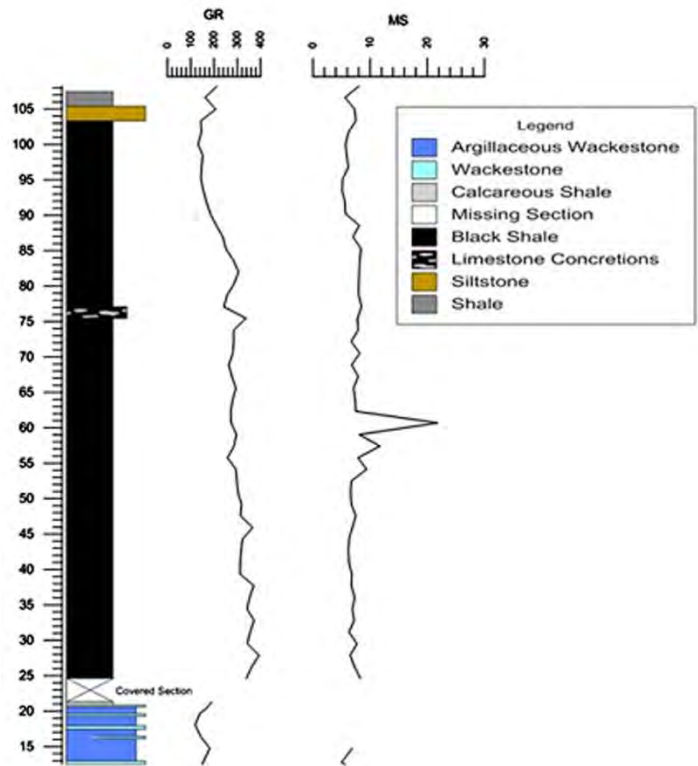

Figure 4-3. Stratigraphic column of transition from Tully Limestone Member of the Mahantango Formation to the Burket Black Shale Member of the Harrell Formation.

130

The boundary between the Burket Black Shale and the Upper Shale member is gradational and difficult to pin-point, but is thought to be near the location of the first siltstone *(Figure 4-4)*. The Upper Shale Member of the Harrell Formation is composed mostly of non-calcareous gray shale with two beds of limestone concretions in the lower portion. Near the top of the Upper Shale Member siltstone beds become more common. The Upper Shale Member represents a return of deltaic sediments and a lower sea-level.

Magnetic susceptibility (MS) and gamma-ray (GR) measurements have been used for correlation and as a proxy for sea-level variations for many years. Gamma-ray logs have been a staple in subsurface correlation for decades, but only recently have detailed outcrop gamma-ray studies been used to produce analogs for the subsurface. Magnetic susceptibility and gamma-ray (GR) measurements data was collected for sections of this stop at two feet intervals. With the increased use of MS for correlation and as a proxy for sea-level, MS may be a valuable tool in outcrop-to-subsurface correlations, along with GR.
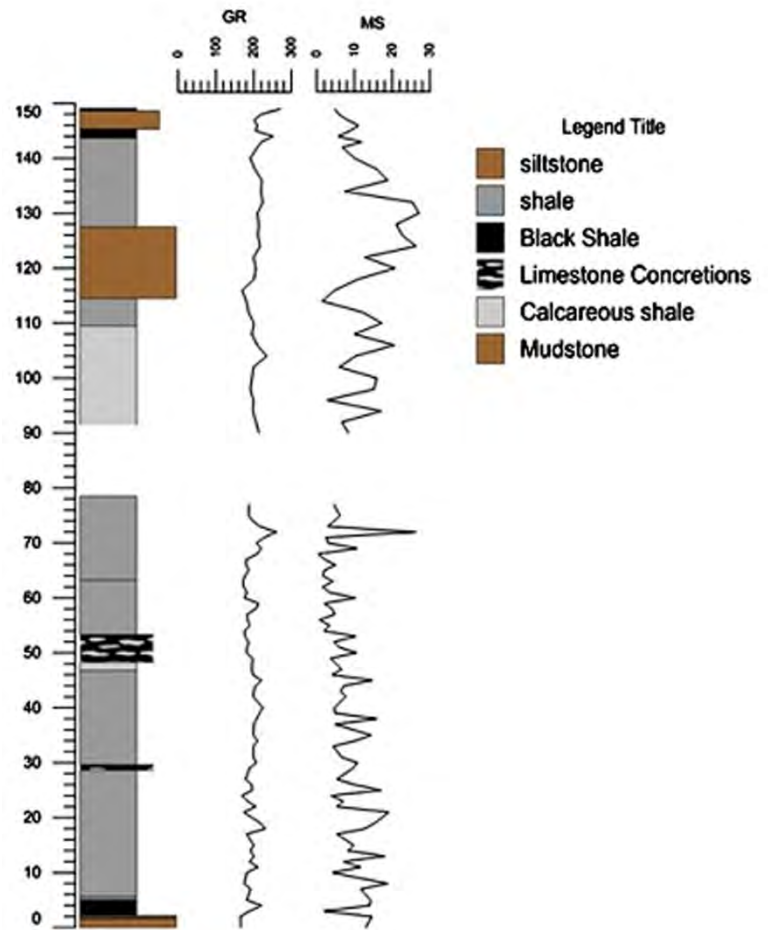

Figure 4-4. Stratigraphic column of the Upper Shale Member of the Harrell Formation.

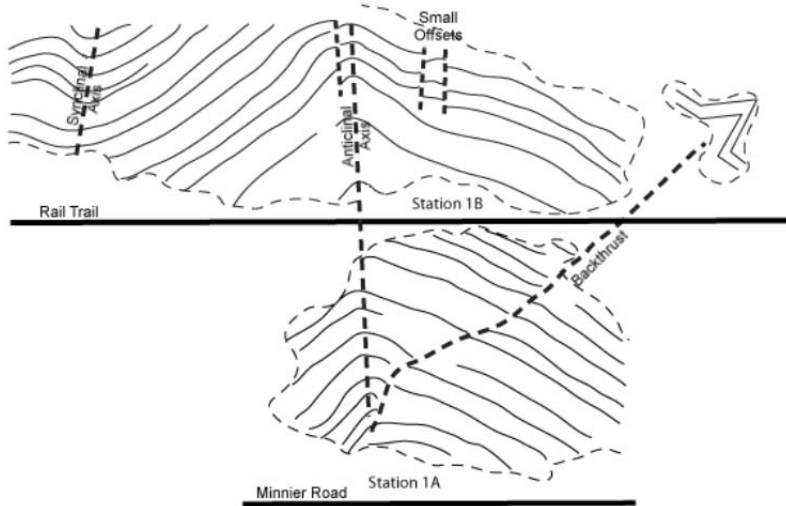# Stop 5: Pine Creek Rail Trail by Jersey Shore
Leader: Paul Washington

Figure 5-1 – Structure at Station 1 with the various parts of the outcrop in appropriate relative position.

The exposures along the Pine Creek Rail Trail north of Rte. 220 provide some of the best nearly continuous exposures of the structure in the transition zone. The exposures start along the north edge of a power line right of way and extend northward for approximately 3600 feet (1.1 km) before becoming widely spaced. At the south end, a road cut along the old road, now cut off by Rte. 220, provides an exposure below the southernmost of the exposures on the rail trail. This description will begin with that outcrop and then move up onto the rail trail itself.

Hoskins (1976) mapped the strata as Lock Haven formation over most of this distance but placed a fault (the Jersey Shore fault of Faill et al., 1977) near the southern end with Harrell and Brallier formations to the south of that fault. Based on a more detailed analysis of the structure, it appears that the fault is not accurately mapped, it probably lies just to the south of the exposures on the rail trail instead, and the only possible Brallier or Harrell strata exposed in this transect is in a thrust sheet klippe (probably a leading portion of the Jersey Shore thrust sheet) exposed at Station 7 in the axis of the large syncline approximately 1500 feet (450 m) north of the mapped fault.

Station 1a – Anticline in Roadcut

At first glance, the exposure along the road shows a classic fault-bend fold with more steeply dipping strata (approx. 50˚N) on the north limb and less steeply dipping strata (approx. 30˚ S) on the south limb, with a short nearly horizontal axial crest. This geometry is totally consistent with northward displacement up an underlying ramp, with the northern limb representing the portion of the hangingwall ramp that moved onto the footwall flat and the southern limb representing the portion of the hangingwall flat that lies atop the footwall ramp. Displacement on this ramp would be less than the length of the ramp.

Closer examination, however, reveals that the northern limb is displaced upward relative to the southern limb along a very high-angle backthrust. These types of small backthrusts occur where the frontal limb is pushed backward – usually in fault propagation folding. In this case, it appears to be related to subsequent folding (i.e. rotation of the leading portion) of the footwall strata.

Getting from Station 1A to Station 1B is difficult without trespassing on private property south of the powerline. However, these landowners are EXTREMELY jealous of their property rights, so DO NOT take that route without explicit permission. Alternatively, there is a VERY STEEP path by the parking area at the north end of the road.

If the route across private property is allowed, terrace deposits can be seen in the cut along the ramp up to the rail trail. These terrace deposits are consistent with Ramage et al.'s (1998) interpretation of delta terraces indicating an ice dammed lake system (Lake Lesley) in the West Branch valley. If these gravels are indeed related to such a lake, they would most likely be of Illinoian age.

Washington, P.(2013) Stop 5 – Pine Creek Rail Trail by Jersey Shore, *in*, Fleeger, G.M., Schmid, K., and Anthony, R., eds., A Tale of Two Provinces: the Nippenose Valley and Route 15 Corridor, 78th Annual Field Conference of Pennsylvania Geologists, Williamsport, PA, pp. 133 – 137.

Station 1B – Anticline on Rail Trail

just north of power line

The anticline from the Station 1A is seen on the rail trail, with the crest about 50 feet (15 m) north of the power line cut. The northern limb of this view of the anticline has a much lower dip (approx 37° N) than that along the road, indicating that this area has indeed been rotated nearly 15°along with the strata to the west. Within the crest of the anticline, small nearly vertical shear offsets can be seen accommodating some of this clockwise rotation. The main backthrust movement from the road-level exposure is seen as an overturned chevron fold along the edge of the power line cut.

The north side of the anticline merges into a syncline only approximately 15 feet (5 m) to the north of the anticlinal axis. This shows that this is a very small displacement thrust feature. The strata beyond the synclinal axis become very uniformly south dipping. Just north of the axis, there is a small wedge thrust accommodating some of the folding strain.

Station 2 – Spaced Cleavage

approx. 330 ft (100 m) north of Station 1b

Spaced cleavage is commonly ignored by those not familiar with the structure of fold-and-thrust belt foreland margins and plateaus, but it forms a penetrative shortening fabric that can be correlated across large areas of the foreland (Engelder and Engelder, 1979?). The spaced cleavage produces penciling in the shales. In the siltstones and sandstones, it mostly appears to be a regularly spaced fracture pattern. This fabric can be shown to pass into the strike-parallel joints mapped by Engelder (Engelder, 1979; Engelder and Geiser, 1979; Geiser and Engelder, 1983).

In this outcrop, two generations of spaced cleavage can be observed: a primary set striking about N70E and a secondary set striking about N20E. The first of these parallels the regional strike of the second phase of Alleghanian deformation, which is dominant in this area. The second parallels the regional trend of the main central Appalachian trend (phase 4) which dominates the structure south of Altoona.

It should be noted that the N70E cleavage planes are roughly perpendicular to bedding, indicating that they were formed prior to folding (Wood et al., 1969). As you proceed down the outcrop, you will see that the N90E (phase 1) and N70E spaced cleavage, which are the best developed in this area, are all pre-folding. This indicates that the folds here developed during phase 2 (since the folds are parallel to the phase 2 cleavage), and not during phase 1.

Just a few meters to the north, scoring on sandstones also provides evidence of spaced cleavage.

Station 3 – Major Anticline

approx. 240 ft (70 m) north of Station 2

Near the end of the outcrop (where it is set back from the trail), there are a series of chevron folds with the south limbs parallel to the dip of the strata that you have been passing and the north limbs vertical. These chevrons folds mark the south edge of the anticlinal crest; the north limb of the anticline can be seen about 250 feet (75 m) farther north (on the other side of the small valley) where the strata are all vertical. This fold is probably the western extension of the Old Lycoming anticline (Faill et al., 1977) reappearing from beneath the Jersey Shore fault.

The shape of this fold is consistent with fault propagation folding in which the fold develops as a ductile bead above a ramp that formed in brittle strata but does not continue through the more ductile strata on its up-dip end. The anticlinal crest marks the beginning of the ductile failure process with the fold growing by migration of the forelandward synclinal trough as the ductile failure process proceeds below, and the angle of folding indicates the relative speed of ramp displacement rate and ramp surface propagation. The small chevron folds in this outcrop indicate that the initial ductile failure occurred in a layered sequence with

brittle layers (down dip beneath the south-dipping limbs) interbedded with ductile layers (down dip behind the vertical north limbs). The ramp then propagated updip into ductile strata, as indicated by the continuous verticality of the beds on the north limb of the major fold. The vertical beds of the north limb continue north along the traverse for approximately 1000 feet (300 m). The height of the fold above the trail level is reconstructed to have been more than 1000 ft (300 m) high beneath the Jersey Shore fault.

Station 4 – Rotated Cleavage

by Mile-Marker Monument - approx. 165 ft (50 m) north of synclinal fold hinge

Spaced cleavage can be seen in the shales and siltstones just north of the monument. It should be noted that these cleavage planes are essentially perpendicular to bedding, meaning they formed prior to the folding and were rotated during the folding. By unrotating these beds, it can be shown that these are primarily phase 2 cleavage planes. As you continue past these vertical layers, you will commonly see the nearly horizontal cleavage of stage 2, but you will also see some west-dipping spaced cleavage planes that are remnants of stage 1 Alleghanian deformation. Where two cleavages are developed in the same layer, pencils are diamond-shaped.

Approximately 85 feet (25 m) farther north, there is a layer with spaced cleavage oriented perpendicular to the fold axis. This spaced cleavage is evidence of a pre-Alleghanian deformational event first recognized in the east end of the Nittany valley (Washington, 2009). Though only locally developed in the Devonian shales within the transition zone, the cleavage is the dominant cleavage within the Mid-Late Ordovician carbonates in the Nittany valley and is associated with thrusts and folds that are oriented transverse to the valley axis. Timing of this event is not yet established, but it must be no earlier than Late Devonian and no later than earliest Alleghanian.

Stations 5a and 5b – Shear vs. Creep

approx. 215 ft (65 m) and 400 ft (120 m) north of Station 4

At the north ends of the next two outcrops, the strata are rotated northward to an overturned dip. Some of this rotation in the first of these appears to carry into the strata to the south. The question is raised:

"Is this rotation the result of hillside creep or is it a result of tectonic shear?"

Rather than answer this in the guidebook, we present you with the question to debate in front of the outcrop.

Station 6 – Out-of-Sequence Thrusts

approx. 330 ft (100m) to 500 ft (150 m) north of Station 5b

As we approach the northern edge of the vertical limb, small north-verging thrusts can be seen cutting across the strata. Offsets are usually on the order of 1 m, and beds are slightly bent next to some of these faults. These faults obviously formed after the formation of the fold as the fold was tightening up. Because the displacement occurs within an already deformed part of the system, we call these out-of-sequence. Faults of this sort are common in steep fore-limbs of fault-bend folds.

Station 7 – Piatt Syncline

approx. 65 ft (20 m) north of Station 6

The transition from the vertical bedding to gently south-dipping strata occurs very abruptly. A closer look, however, reveals that the exposed syncline is not a simple syncline folding the vertical strata that you have been passing. Rather, the synclinal structure is seen only in strata that are lying obliquely atop the vertical strata (i.e. dipping approx. 60˚N) and cutting across those vertical strata. This contact is interpreted to be a footwall ramp rotated with the folding; movement on this ramp would have preceded the rotation, so it is really a north-verging ramp. The thrust sheet strata are possibly shales of the Braller formation, though no definitive determination has been made of their stratigraphic position.

The syncline appears to coincide with the transition from footwall ramp to footwall flat. So it is difficult to recognize the fault surface on the northern limb of the syncline. This syncline shows up farther east on Hoskins' (1976) map of the Jersey Shore quadrangle as the west end of a tongue of Catskill formation and is called the Piatt Syncline in the Williamsport quadrangle (Faill et al., 1977).

The thrust sheet is a small klippe of a much larger thrust sheet. It almost certainly represents the updip extension of the Jersey Shore thrust sheet, which crosses the rail trail near the power line just south of first outcrops; the Jersey Shore Fault is the largest of the thrust faults mapped in the transition zone in this area. The folding of this thrust sheet into the Piatt Syncline demonstrates that the fold-and-thrust deformation was progressing northward, with the syncline deforming the previously formed overlying thrust sheet. Footwall deformation beneath the Jersey Shore Fault is the probable explanation for the 25% thickening of the Catskill on the north limb of the next fold (north of this transect).

Station 8 – Flexures

approx. 1200 ft (360 m) north of Station 7

Outcrops over the next 150 m show numerous small flexures, with as much as 30° variations in bedding dip. The origins of these flexures cannot be absolutely explained by evidence at the outcrop, but they are consistent with the deformation on the trailing limb of fault-propagation folds (Mitra, 1990). This appears to be the trailing limb of a fault propagation fold centered approximately 4000 feet (1200 m) farther north (beyond the end of this traverse); this fold is apparently larger than the fold that we observed at Station 7.

Station 9 – Multiple Directions of Spaced Cleavage

approx. 95 ft (28 m) north of Station 8

Excellent exposure of N-S striking spaced cleavage in shale. This is the pre-Alleghanian cleavage discussed earlier (Station 4). The cleavage forms a lineation and penciling nearly parallel to the dip direction.

Lower in the outcrop is a strong N30 E cleavage in another shale layer. This appears to be the Alleghanian stage 4 cleavage. We have been seeing stage 2 cleavage throughout this series of outcrops, but stage 4 deformation appears to have affected this area as well.

Stage 2 and stage 4 are the two most important stages of the Alleghanian orogenesis in the central and northern Appalachians. Stage 2 structures can be seen throughout eastern Pennsylvania, southern New York and southern New England, and have been recorded as far north as the Champlain Valley of Vermont (Washington, 2011). Stage 4 structures dominate the main central Appalachian trend from Altoona, PA to Roanoke, VA.

The primary deformation in this area appears to be stage 2, with both the primary cleavage and all of the fold trends having stage 2 orientations. The presence of stage 4 structures indicates that the north edge of the main central Appalachian Alleghanian deformation did not end at Altoona but extended farther north and probably modified some of the preexisting structures. In this area, it is expected that it probably caused a right-lateral translation along stage 1 and 2 fault surfaces. The fact that we are close to the west edge of the stage 2 structural system would indicate that there was probably some westward movement up the lateral ramps at the west end of the stage 2 structural field. The western end of this fold system is approximately 5.7 mi (9 km) west (shown somewhat cryptically on Taylor, 1977).

Outcrop ends 160 ft (50 m) farther north.

Station 10 – Fault Gouge

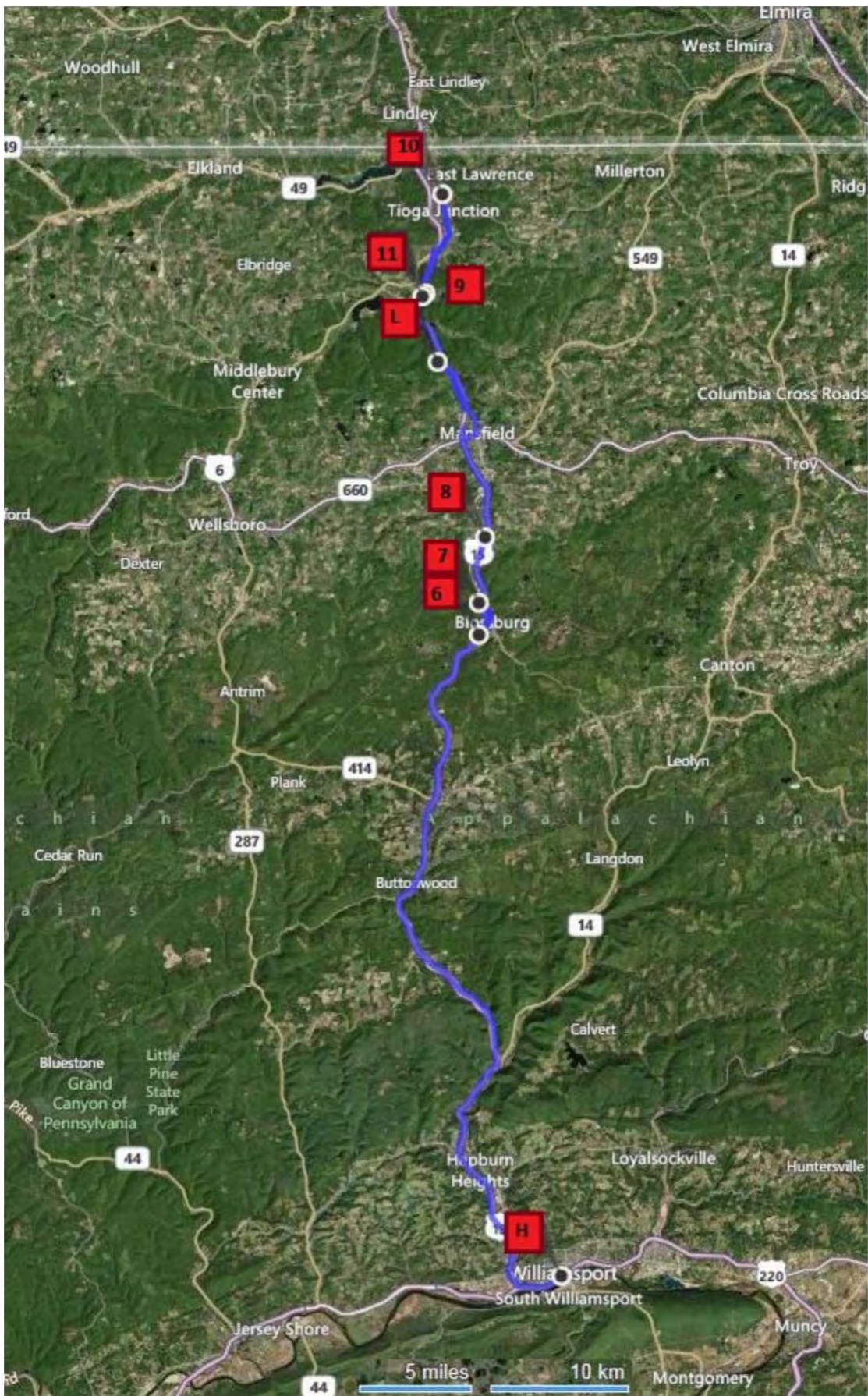### approx. 1240 ft (370 m) north of Station 9 (as end gate comes into view)

A strange nodular sandstone (approx. 2 m thick) in a shale matrix is a prominent feature in this outcrop. The layer appears to be fault gouge in a flat-on-flat setting with the sandstone nodules being tectonically rounded pieces of sandstone sitting in a highly deformed shale matrix. The fault must have experienced

significant displacement to produce such a thick and well developed gouge. Based on geometric reconstruction, it is proposed that this is the up-dip extension of the ramp that created the fault propagation fold we saw at Station 7.   If so, that fault motion would have caused 1000-1200 ft (300-360 m) of displacement. Another 1000 ft (300 m) of displacement due to movement in the footwall of the Jersey Shore fault could be postulated. Therefore, this fault gouge must have developed during movement of between 1000 and 2200 ft (300 to 660 m).

END OF TRAVERSE – approx. 450 ft (150 m) west of Station 10, follow crossroad downhill (west) to the parking lot. The next anticlinal axis crosses the trail 2150 ft (650 m) farther north. This anticline is probably the westward continuation of the Torbert Anticline (Faill et al., 1977; Faill and Wells, 1977), the outermost major anticline of the transition zone. The structural transition to the plateau occurs abruptly beneath the topographic front approximately two km north of the end of the traverse.

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Road Log- Day 2

Int	Cum	Description
	0.0	Right onto Court Street, bearing right onto Via Bella Street.
0.2	0.2	Procede through roundabout, continue on Bella Via.
0.2	0.4	Traffic light, straight onto US 15/220 ramp.
0.3	0.7	Merge onto US 15/220.
1.2	1.9	Bear right onto Exit 29, Mansfield/US 15. Merge onto Route 15 north.
1.2	3.1	Lycoming Creek on the right. Its source is a spring about half a mile east of Penbryn station (Carpenter's) on the Northern Central railroad. The stream is small at the beginning, but as it flows southward, it gathers strength from numerous tributaries until it passes through the western part of the city of Williamsport and reaches the Susquehanna River. The name is corrupted from the Delaware Legani-hanne, signifying sandy stream. On Scull's map it is written Lycaumick. It is plainly seen, therefore, how easy the transition was to Lycoming (Menginess, 1892).
0.7	3.8	Sign for Leganni Hanne (Lycoming Creek) Trail.
1.7	5.5	Roadcut exposes interbedded sandstones, siltstones, and shales of the Devonian Lock Haven Formation (Faill and Wells, 1977a).
0.2	5.7	Note the steady increase in the dip of bedrock to the south.
0.5	6.2	Essentially flat-lying Lock Haven strata takes a dive, bedding dips approximately 70 degrees north.
0.2	6.4	Kink fold. Near vertical beds rapidly transition northward to gently folded/near horizontal strata. This section is similar to what was observed at Stop 5 on Day 1.
0.6	7.0	Hepburnville Exit.
0.1	7.1	Say goodbye to the Lock Haven Formation for now.
1.2	8.3	Catskill redbeds.
0.4	8.7	Continued Catskill reds.
0.6	9.3	Back to the Lock Haven
0.7	10.0	Starting the topographic climb to the Plateau Province.
0.4	10.4	Catskill redbeds.
0.3	10.7	Catskill redbeds on the left. Note dissected hills on the right, cut by Lycoming Creek.
1.9	12.6	Excellent view of the dissected High Plateau on right and ahead.
2.0	14.6	Trout Run/Canton exit, SR 14 north.
1.1	15.7	Stacked medium-bedded sandstone, distributary channels of the Catskill overlain by overbank fining upward sequences of sandstone, shale, and mudstone. An abundance of plant root (rhizome) trace fossils can be found in the area of the rockfall (mileage 16.0).
1.2	16.9	Paleosol capped by levee or distributary channel facies.
0.5	17.4	Similar sequence to mileage 16.9.
2.0	19.4	Thick-bedded to massive, channel sandstone.
0.2	19.6	Continuation of thick to massive bedded channel sandstones.
0.1	19.7	Medium to thick-bedded reddish gray sandstones interbedded with silty shale and mudstone. Sedimentary structures such as ripple marks and plant trace fossils are common.

0.3	20.0	At the ramp, overbank deposits along with intraformational conglomerates, plant trace fossils; abundant plant fragments as lag deposits from greenish gray sandstone high on the outcrop; fish bones and scales scattered about; most occurring in the berm just south of the Cogan House exit sign
1.7	21.7	Steam Valley exit/SR 184.
0.2	21.9	Turkey Ranch Restaurant on left at the top of the Plateau.
1.6	23.5	Still in the Catskill!
1.2	24.7	And again….
0.3	25.0	English Center/Buttonwood exit, SR 284.
1.7	26.7	Catskill channel/distributary channel/overbank facies on left.
0.4	27.1	Same
0.4	27.5	Ditto
1.1	28.6	Ditto.
0.4	29.0	Tioga County line.
0.6	29.6	Morris/Liberty exit, SR 414.
2.1	31.7	Sebring exit. Note how the road has flattened out. We be in the Plateau for sure.
0.8	32.5	Catskill
1.8	34.3	Catskill sandstone, agglomerate.
1.6	35.9	Conglomerates of the Pennsylvanian Pottsville Formation on the left.
0.1	36.0	Entering Bloss Township.
0.5	36.5	Section of Mississippian Burgoon (Pocono) Formation or the Huntley Mountain (Spechty Kopf) Formation
1.0	37.5	Spoil pile on the hillside (left) leftover from mining the Blossburg coal. The lead mineral galena can be found within siderite concretions on the pile.
0.8	38.3	An excellent exposure on the left illustrates the complexity of stratigraphic correlations in the Blossburg area. The presence of paleosols complicates the interpretation as to whether the middle sandstone unit is the Mississippian Mauch Chunk or Burgoon Formation. The basal sandstone is interpreted to be the Huntley Mountain Formation - or is it? More at Stops 6 and 7.
0.6	38.9	Take the Blossburg exit.
0.5	39.4	At the stop sign, turn left, park on the right after passing beneath Route 15.
0.1	39.5	Disembark busses and head up the "Wrong Way" for **Stop 6. Blossburg Exit Section, Huntley Mountain (Mississippian and Devonian) and Pottsville (Pennsylvanian) Formations** Leader- Clifford H. Dodge See page 143.
0.0	39.5	Leave Stop 6. Retrace route east beneath route 15 bridges to stop sign. Kwik Fill Food Express and Gas is across the road.
0.0	39.5	Turn left (north) onto Bloss Mountain Road towards Blossburg.
0.9	40.4	Enter Blossburg.
1.0	41.4	Pass beneath US 15. Park on right berm. **Stop 7. Sandstone and shale of the lower Huntley Mountain Formation (Late Devonian- Early Mississippian).** Leader: Brett McLaurin See page 155. AMD stream on the right is a colorful orange.
0.0	41.4	Leave Stop 7. Return (south) through Blossburg.
1.8	43.2	Turn right and continue onto Route 15 north/Mansfield
1.3	44.5	Huntley Mountain Formation on the left (Stop 7).

0.4	44.9	Crossing Tioga Creek.
0.7	45.6	Back into the Catskill. See McLaurin write up.
0.9	46.5	More Dck.
0.5	47.0	Redbeds; mostly fine-grained sandstones, shales, and mudstones; the usual plant trace fossils; rare intraformational conglomerates
1.2	48.2	Pull off onto berm. **Stop 8. Lock Haven Formation fossils.** Leader- Bill Kochanov See page 159.
0.4	48.6	Lock Haven Formation. Note reddish coloration on left (west) side. Is this the initial taint of red Catskill?
0.6	49.2	Marine Lock Haven in the lower half of the section. Upper half shows Catskill redbeds.
0.6	49.8	Lock Haven Formation.
0.5	50.3	Covington/Canoe Creek exit. SR 660/15 Business. Arby's and Wendy's.
2.5	52.8	Entering Mansfield.
0.3	53.1	Mansfield/Wellsboro exit, US 6. All off for the Grand Canyon of Pennsylvania!
0.1	53.2	Traffic light. Turn left onto US 6/Wellsboro Street.
0.3	53.5	Crossing Tioga River.
0.3	53.8	Traffic light. Left onto North Main Street. Yorkholo Brewing Company on the left, next to Chango's Cantina.
1.1	54.9	Intersection with US 15. Stay straight on Appalachian Thruway.
2.0	56.9	Right to picnic area, Lambs Creek Recreational Area.
0.4	57.3	Parking lot. **Lunch**.
0.4	57.7	Leave lunch. Left onto Appalachian Thruway.
2.0	59.7	Intersection with US 15. Turn left onto US 15 north.
4.4	64.1	Crossing Tioga River.
0.6	64.7	Tioga Reservoir.
1.4	66.1	Tioga/Hammond Dam on left.
0.5	66.6	Pull off onto berm. **Stop 9. Contact between the upper Lock Haven Formation (Late Devonian) and the lower Catskill Formation (Late Devonian).** Leader: Brett McLaurin See page 167.
0.4	67.0	Last gasp of the red Catskill; in comes the greenish gray Lock Haven Formation.
1.2	68.2	Tioga/Tioga Junction exit, SR 287. Take it.
0.6	68.8	Stop sign. Turn right onto SR 287 north.
2.2	71.0	Continue on SR 287 through the traffic light at Tioga Junction (intersection with SR 328).
0.8	71.8	Turn right at entrance of Cross Excavation gravel pit.
0.8	72.6	**Stop 10. Esker cross-section.** Leaders: Duane Braun and Aaron Bierly See page 171. WARNING when exiting buses, do NOT approach any of the headwalls as they are not stable!
2.2	74.8	Return south on SR 287.
3.2	78.0	Turn left onto ramp for US 15 South.
0.5	78.5	Merge onto US 15 South.
1.4	79.9	Exit right into the rest area.
0.7	80.6	Arrive at rest area.

Stop 11. Overlook of the Tioga River - Crooked Creek valleys and the Tioga-Hammond dams
Leaders: Duane Braun and Aaron Bierly See page 175.

0.2	80.8	Leave rest stop and continue south.
0.5	81.3	Merge onto highway. Immediately after, there is a hollow to the left which was a locality of slope failure due to road construction undercutting a glacial varve.
2.8	84.1	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. 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.
3.0	87.1	Crossing the Tioga River.
32.2	119.3	If you look off to your left (east side) on top of mountain you may be able to see the top portion of a drilling rig.
15.6	134.9	Merge with SR 220 N/I-180/SR 15.
1.5	136.4	Take exit 27A (US 15 South).
0.3	136.7	Turn left at light onto Market Street.
0.1	136.8	Turn left onto Church Street.
0.1	136.9	Turn Left onto Court Street.
0.1	137.0	Turn right onto Pine Street and Arrive back at Holiday Inn.

Stop 6: Blossburg Exit Section, Huntley Mountain (Mississippian and Devonian) and Pottsville (Pennsylvanian) Formations.

Leader: Clifford H. Dodge

LOCATION

The Blossburg Exit Section was measured on May 14 and 16, 2013, along a road cut of U.S. Route 15 South (west side of road) at the exit ramp to Blossburg, Pennsylvania (Figure 6-1). It is located in the Blossburg 7.5-minute quadrangle, Borough of Blossburg, Tioga County, at GPS coordinates (X = longitude, Y = latitude, in decimal degrees, NAD 83; Z = altitude, in meters (m), NGVD 88), for the base of section, X = -77.072956, Y = 41.675981, Z = 444.5; and for the top of section, X = -77.074144, Y = 41.675768, Z = 487.5.
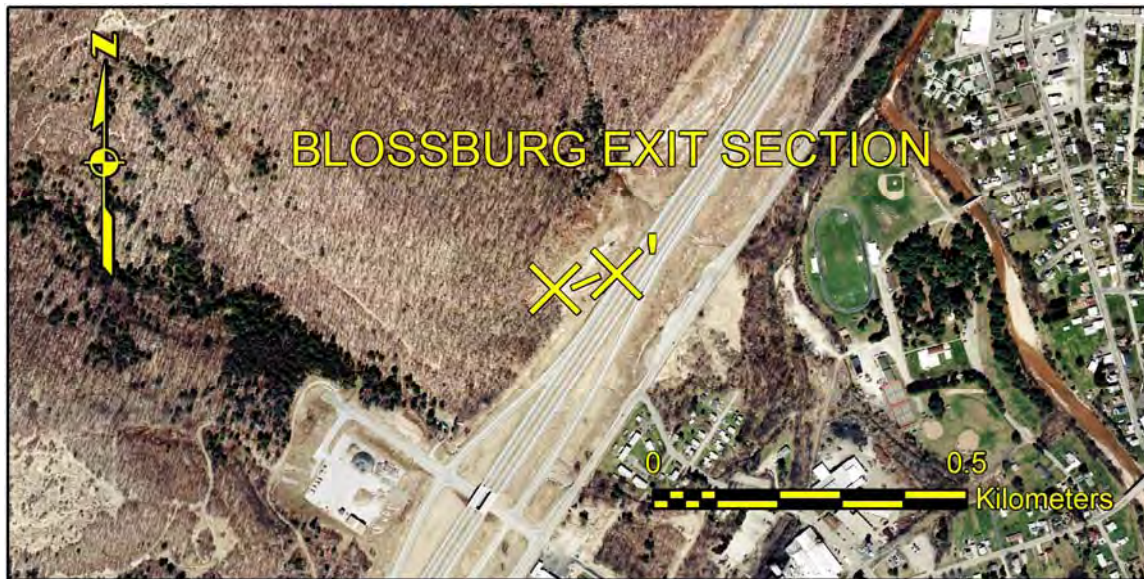
Figure 6-1. Orthophotograph showing location of Blossburg Exit Section. Top and base of Section denoted by yellow "X" and "X'," respectively.

Introduction

Reconstruction of the U.S. Route 15 corridor between Williamsport and the New York border during the past two decades has led to the creation of numerous magnificent road cuts along the state highway. These exposures provide an unprecedented view of the bedrock geology of central Lycoming and Tioga Counties. From 2002 to 2004, the portion of U.S. Route 15 between Bloss Mountain (south) and Maple Hill (north) was rebuilt, which transects the east-west-trending Blossburg syncline (Blossburg-Antrim coalfield) and includes Stops 6 through 8, Day 2, of the 2013 Field Conference.

The Blossburg Exit Section includes the youngest bedrock geologic units to be examined on Day 2 of the Conference and consists of nonmarine (fluvial) upper Huntley Mountain Formation (Lower Mississippian) and overlying nonmarine (fluvial) lower Pottsville Formation (Middle Pennsylvanian) (Figure 6-2). A major regional unconformity is present at the base of the Pottsville. The 43-m Section is noteworthy for the contrasting fluvial styles of the two formations, occurrence of the so-called "Patton" redbeds near the top of the Huntley Mountain Formation, and presence of conglomerate and coal in the lower Pottsville Formation. The Section is situated about 100 m north of the axis of the Blossburg syncline (McLaurin and Dodge, 2012).
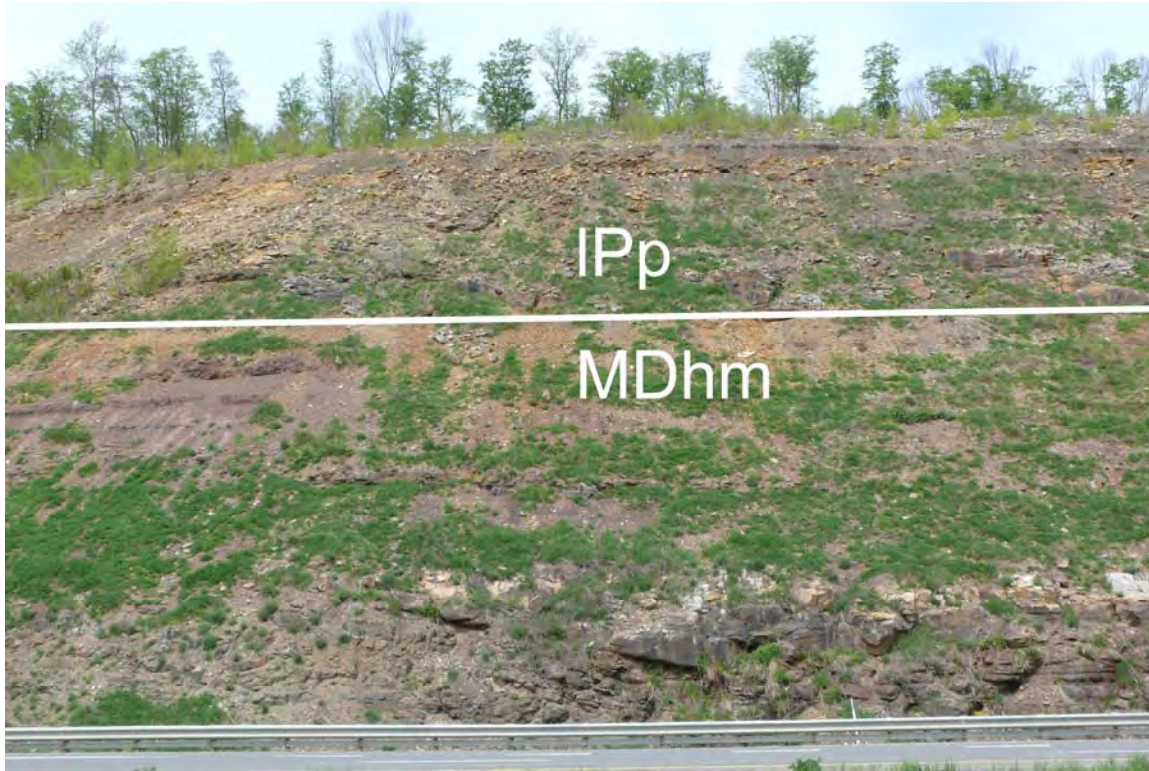
Figure 6-2. Photograph of Blossburg Exit Section showing contact between Pottsville Formation (IPp) and Huntley Mountain Formation (MDhm).

REGIONAL GEOLOGY

Huntley Mountain Formation

Berg and Edmunds (1979) named the Huntley Mountain Formation (Lower Mississippian and Upper Devonian) for exposures along Huntley Mountain (type section) near the village of Waterville, Lycoming County. The formation represents the dominantly nonmarine clastic sequence in north-central Pennsylvania that is transitional between the Catskill Formation (Upper Devonian) and Burgoon Sandstone (Lower Mississippian), both in terms of lithologic characteristics and interpreted depositional environments.

The Huntley Mountain Formation is about 175 to 200 m thick. It is characterized by thick sequences of sandstone (amalgamated) that are greenish gray to light olive gray, very fine to fine-grained, impure, thin to medium bedded, planar bedded to dominantly low-angle trough cross-bedded, and flaggy to slabby, and contains some thin beds of grayish-red siltstone, silt shale, and clay shale. Minor lithologies that may occur include intraformational conglomerate (commonly calcareous), extraformational conglomerate (including the conglomerate facies of the thin Cedar Run marine zone), nonred fine-grained clastics, and pisolith beds. The Cedar Run marine zone, situated slightly above the middle of the formation, is not known to extend any farther east than western Lycoming County, a distance of about 30 km from the Blossburg Exit Section. The formation has been informally subdivided in places into a "lower sandstone sequence" and "upper sandstone sequence," based on the presence of the Cedar Run marine conglomerate (key bed), which was used to make the break (Colton, 1963).

Berg and Edmunds (1979) noted that the Huntley Mountain contains upward of nine major fining-upward (grain size) cycles that range in thickness from about 7 to 31 m. The cycles consist of a lower coarse member that includes sandstone and much subordinate conglomerate, if present, and an upper fine

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member of dominantly siltstone, shale, or claystone that may be red, nonred, or both. Overall, fining-upward cycles tend to be thicker and less well organized than those of the underlying upper Catskill Formation. For the most part, the lower coarse members of the Huntley Mountain cycles are thicker than those of the Catskill, and the upper fine members are thinner than the equivalent members of the Catskill cycles (Berg and Edmunds, 1979). The Burgoon Sandstone does not contain fining-upward cycles.

As a general rule, lower Huntley Mountain sandstones are more grossly similar to sandstones of the upper Catskill (i.e., less-pure composition, finer grain size, thinner bedded, and lower angle trough cross-bedding), whereas upper Huntley Mountain sandstones may take on more characteristics of the overlying Burgoon Sandstone (i.e., cleaner composition, coarser grain size, thicker bedded, and higher angle trough cross-bedding). The lower Huntley Mountain also tends to contain more redbeds than the upper part, though this still accounts for less than 20 percent or so of the interval. As there are many exceptions to these generalizations, interpretation and correlation of small, isolated outcrops may be difficult or impossible without additional information when field mapping the bedrock geology of the Tioga County area.

Previous workers believed that two of the redbed units were regionally persistent and identified them by formal names. The lower unit, about 15-30 m above the base of the Huntley Mountain Formation, was loosely referred to as the "Mount Pleasant" red shale (Willard, 1946; and Ebright, 1952). The upper unit, occurring at or just below the top of the Huntley Mountain, was named the "Patton" red shale (Ebright, 1952; Colton, 1968). However, designations of these red units are only speculation, as correlations with the Mount Pleasant Shale of White (1881, p. 59, 63) in Wayne County and the type Patton Shale of Campbell (1904) in Jefferson County have never been made. Indeed, recent core drilling of the Huntley Mountain Formation in north-central Pennsylvania by the Pennsylvania Geological Survey suggests that none of the red units are persistent over any great extent, and the so-called "Patton" is often not present. Recognizing that the "Patton" might not be as laterally continuous as previously believed, Berg and Edmunds (1979, p. 34) correctly suggested that "the apparent ubiquity of the 'Patton' just below the Burgoon may be due to the fact that natural exposures of the base of the Burgoon are best developed and most visible where an underlying red shale is present and has been eroded to undercut the overlying sandstone and produce a striking outcrop."

With the exception of the Cedar Run marine zone, which may represent estuarine deposits, Huntley Mountain sedimentation resulted from deposition in a fluvial environment on an alluvial plain (Berg and Edmunds, 1979). The Catskill Formation and Burgoon Sandstone in north-central Pennsylvania are also interpreted as fluvial deposits. The environment of deposition of the upper Catskill is inferred to be high-sinuosity meandering-stream systems (Berg and Edmunds, 1979; Berg, 1999), whereas the depositional environment of the Burgoon is interpreted as high-gradient braided-stream systems (Cotter, 1978; Berg and Edmunds, 1979). Huntley Mountain cycles are interpreted to be mostly low-sinuosity meandering-stream deposits. Huntley Mountain rivers appeared to have carried a greater average sand load than did those of the Catskill (Berg and Edmunds, 1979). The source areas for Huntley Mountain sediments lay to the north (cratonic) and northeast (orogenic) and were uplifted during the late phase of the Acadian orogeny as the supercontinent Gondwana (Africa portion) continued to converge with Laurussia (includes North America) (Berg, 1999; Faill, 1999; Blakey, 2008).

The interpreted paleoclimate of the Huntley Mountain Formation began (Late Famennian Age) as a period of increased wetness (amount and/or frequency of precipitation) and humid conditions coupled with dramatic cooling that allowed for the development of widespread glaciation at the middle south paleolatitudes of the Appalachian basin (Brezinski and others, 2009). Later on (Early Tournaisian Age), precipitation decreased slightly and temperatures rebounded to more temperate levels. A relatively short-term climate shift to somewhat drier conditions is thought by some to have occurred near the end of Huntley Mountain deposition, based on the presumed persistence of the so-called "Patton" redbeds (Brezinski and

others, 2009). Berg and Edmunds (1979, p. 57) also suggested increased precipitation during Huntley Mountain deposition relative to the Catskill.

Pottsville Formation

Lesley (1876) introduced the name Pottsville as a synonym for the older term "Seral Conglomerate" of the First Geological Survey of Pennsylvania (1836–1858), recognizing the excellent development and exposure of the clean, coarse sandstones and conglomerates near the town of Pottsville, Schuylkill County, Pennsylvania, in the Southern Anthracite coalfield. In the Blossburg syncline (Tioga County), the Pottsville Formation consists of a complex, heterolithic, mostly nonmarine sequence of predominantly light- to medium-gray sandstone and extraformational conglomerate (quartz arenite), with much subordinate medium-gray siltstone, medium-dark-gray to dark-gray shale, olive-gray claystone (paleosol), and coal. The sandstones are generally medium to very coarse grained and conglomeratic in places, clean, siliceous, well cemented, thin to thick bedded, and trough cross-bedded. They contain scattered fossil plant impressions and compressions. The formation is more commonly conglomeratic in the lower part. The conglomerates occur as very thin to medium interbeds and lenses and contain clasts (extraformational) of milky-white quartz up to large and locally very large pebbles. Dodge (1995) reported rare, restricted-marine invertebrate fossils (*Lingula* sp.) about 6 m above the Bloss coal (upper Pottsville) in core from a Pennsylvania Geological Survey drill hole near the former mining village of Fall Brook, about 6.5 km east of Blossburg. The lower contact of the Pottsville Formation is sharp and erosional. A major regional unconformity is present at the base. The thickness of the Pottsville is approximately 46 m.

Three coals of historical economic importance are present within the Pottsville Formation and include in ascending order (oldest to youngest, using long-established local nomenclature) the Kidney, Bear Creek, and Bloss. The Kidney and Bear Creek coals are highly lenticular and generally range in thickness from 0 to about 0.8 m. They were rarely mined in the past on a commercial scale. The Kidney and Bear Creek occur about 16 to 24 m and 5 to 9 m, respectively, below the Bloss. The stratigraphic position of the Bloss coal is about 11 to 17 m below the top of the Pottsville. The Bloss coal was formerly the most important economic seam of the area, where it was extensively deep and strip mined. It was prized for its coal quality, generally medium in ash and low to medium in sulfur. The Bloss ranges in thickness from about 0.8 to 1.5 m, but locally forms a coal complex where it has split into two principal benches that may be separated from one another by as much as 1.5 to 3 m. Each bench varies in thickness from about 0.6 to 1.1 m, but where one bench is thicker (i.e., toward the upper limit of its thickness range) the other is thinner and may contain more partings. The thinner bench was generally uneconomical to deep mine along with the thicker bench, owing to bed thickness, seam impurities, or distance between the two benches. The Bloss coal is the most important key bed within the coal measures of the Blossburg-Antrim coalfield and has been used for correlation and structure contouring of the coal basin (McLaurin and Dodge, 2012). The Bloss coal complex is now considered by the Pennsylvania Geological Survey as equivalent to the Upper Mercer coal complex of the Main Bituminous coalfield.

The lower part of the Pottsville was deposited as sediments by gravelly or sandy braided-stream systems crossing an alluvial plain (Meckel, 1964; Dodge, 1992). Regional changes in sediment load/gradient and stream competency were primarily in response to changes in base level of the fluvial system brought on by glacial-eustatic fluctuations of sea level (Heckel, 1994; 1995). Upper Pottsville rocks (i.e., Bloss coal and above) represent depositional environments chiefly characterized by high-sinuosity meandering-stream systems in a proximal-alluvial-plain or coastal-plain/paralic setting. Peat accumulated in the coal swamps mostly during times of near-maximum to maximum transgressions (highstands) that were controlled by glacial eustasy (Heckel, 1994; 1995). Pottsville sediments for this area were derived mostly from a cratonal source to the north during the initial suturing between Gondwana and Laurussia to form Pangea (Meckel, 1964; Edmunds and others, 1999; Blakey, 2008).

Paleoclimate reconstructions of the Carboniferous by Cecil (1992) and Cecil and Eble (1992) suggest that the Pottsville Formation was humid, mostly ever-wet tropical, with some transition to more wet-dry (seasonal) tropical at the top. During the deposition of the Pottsville, the paleolatitude was about 5 to 10 degrees south of the equator (Edmunds and others, 1979; Blakey, 2008).

The pre-Pottsville regional unconformity is believed to have begun no later than the Early Bashkirian (Early Morrowan of North American usage) or early Early Pennsylvanian (Englund, 1979; Edmunds and others, 1999). The erosional surface has been interpreted as the result of eustatic sea-level decline associated with the world-wide mid-Carboniferous erosional hiatus (Saunders and Ramsbottom, 1986; Blake and Beuthin, 2008) or tectonic uplift due to a peripheral bulge (Ettensohn and Chestnut, 1989). Edmunds (unpublished manuscript) provided evidence that both mechanisms may be involved in Pennsylvania, resulting in two closely timed sequential unconformities occurring in the Late Serpukhovian (Late Chesterian of North America), for eustacy, and Early Bashkirian (Early Morrowan), for structural arching.

BLOSSBURG EXIT SECTION

<u>Discussion</u>

For purposes of discussion, the geology of the Blossburg Exit Section is subdivided into three informal fluvial sequences (Figure 6-3). Sequence 1 (lowest interval) and Sequence 2 pertain to the Huntley Mountain Formation, whereas Sequence 3 encompasses the entire exposure of the Pottsville Formation. Sequence 2 includes the so-called "Patton" redbeds. Sequences 1 and 2 are genetically linked, comprising the greater part of the uppermost fining-upward cycle of the Huntley Mountain preserved in this area and representing meandering-stream deposits. In contrast, Sequence 3 consists of sandy and gravelly braided-stream deposits.

The Burgoon Sandstone (Lower Mississippian) and Mauch Chunk Formation (Upper Mississippian) that overlie the Huntley Mountain farther south are missing here, owing to erosion on the pre-Pottsville regional unconformity. Based on his recent (2011) geologic mapping of the English Center quadrangle, Lycoming County, Dodge (in preparation) estimated that the regional inclination (corrected for subsequent deformation) of the pre-Pottsville bedrock units strikes about N70°E and dips 4.2 to 4.7 m/km. Using these values and known nearby Huntley Mountain thicknesses determined from recent core drilling by the Pennsylvania Geological Survey, I extrapolated the loss of section northward from English Center to the Blossburg area. It is estimated that the top 5 to15 m of the Huntley Mountain Formation is missing from the Blossburg Exit Section. (This does not include possible effects of local relief on the unconformity.) If the upper estimate is correct, the stratigraphic interval traditionally assigned to the "Patton" has been completely removed by erosion here!

<u>Sequence 1</u>

Sequence 1 (Units 1–14, 7.9 m thick) includes the channel-phase sandstones of the fining-upward cycle (Figure 6-3). The base of the cycle is not exposed. The sequence consists of interbedded sandstone and silt shale. The sandstones are mostly light gray to light olive gray; very fine to fine grained; impure; planar bedded, laminated, ripple bedded, or locally trough cross-bedded; and micaceous. Sandstone interbeds are mostly medium to thick. Lower contacts are sharp, gradational, or erosional. The silt shales were deposited during times of decreased stage. They are generally gray to olive gray, fissile to subfissile, and ripple laminated to ripple bedded in places. A few scattered fossil root traces occur upward. Silt-shale interbeds are mostly thin to medium. Lower contacts are sharp.

Several units stand out for individual commentary. The lowest redbed of the Section occurs as Unit 11, a silt shale that is brownish gray, slightly mottled with light olive gray, chippy, hackly, and with a few

scattered horizontal fossil root traces. Unit 12 is a channel-form sandstone, which is lenticular and locally thickens southward from 0.4 m to 1.5 m, cutting out underlying Units 9–11. Several meters farther to the south, these underlying units can be observed dipping northward toward the channel. Unit 13 is more sheet-like, continuing across the outcrop, and is easily recognizable by its distinctively lighter color. This unit consists of sandstone that is light yellowish gray, fine to medium grained, well cemented, much cleaner than underlying sandstones, planar bedded to trough cross-bedded, and thin to thick bedded. It is slightly calcareous to calcareous in the bottom 0.6 m (secondary calcite can be observed on the sandstone surfaces of this interval as well), the only calcareous portion of the Section. The base of Unit 13 is erosional. Given these observations (but only this one outcrop), what interpretations are most plausible for the origin for this unit and the calcite cement in the lower part? Unit 14 is really a continuation of the underlying unit compositionally, but it is strongly rootworked (plant bioturbation) and interpreted to be a very weakly developed paleosol. (Soil-forming processes intensified in overlying Unit 15.) The sandstone has almost no internal bedding or parting, appears "lumpy," and fragments to hackly to rubbly. The upper half is mottled light yellowish gray and grayish red. Occurring toward the base of the unit are a few scattered granules to small pebbles (to 5 mm in diameter) of white and rose quartz. The clasts are rounded to well rounded. How might the presence of extraformational pebbles affect any interpretation for the origin of Unit 13 (and Unit 14)?

Sequence 2

Sequence 2 (Units 15–23, 11.3 m thick) consists of the overbank (interfluve) deposits of the fining-upward cycle (Figure 6-3). Because the top of the sequence is truncated by a regional unconformity, it is not clear if Unit 23 is part of the cycle under discussion or the base of another cycle. The sequence is made up of interbedded red clayey siltstone to siltstone and light-olive-gray sandstone. The siltstones represent stacked paleosols, comprising half the thickness of the sequence, and the sandstones are mostly coarse-phase flood-basin (crevasse-splay) deposits. The siltstones are grayish red, contain a few scattered horizontal to subvertical fossil root traces, exhibit minor pedogenic slickensides in places, are noncalcareous, lack internal bedding or parting, and are hackly. Lower contacts are sharp. No distinct vertical profiles were observed. Individual paleosols range in thickness from 0.8-3.5 m. Although not studied in detail, the paleosols are best classified as ancient vertisols. The redbeds appear to reflect periodic wetting and drying (seasonal precipitation?) under semitropical climatic conditions (Brezinski and others, 2009). The sandstones are mostly yellowish gray, very fine to locally fine grained, very thin to thin bedded, and slabby. Lower contacts are generally sharp.

Unit 16 is particularly distinctive and noteworthy. It consists of sandstone that is yellowish gray, very fine to fine grained overall but with a few red chert (jasper) pebbles near its base, well cemented, relatively clean, burrowed, ripple bedded, very thin to thin bedded, and slabby. The lower contact is erosional with up to 0.5 m of local relief. The presence of red chert indicates a shift in provenance for at least some of the sediments. A few scattered dusky-red to grayish-red chert granules to small pebbles (to 7 mm in diameter) occur in the lower 0.2 m of Unit 16 where it is of normal thickness and within the scoured depressions as well where it thickens. Many of the extraformational clasts appear as "floating" (i.e., are isolated). The chert pebbles are rounded to well rounded and range in shape from subspheroidal to mostly elongated to discoidal. Unit 16 was also located in the next road cut about 350 m north of here, where red chert sand (most grain sizes) and pebbles appear more common and occur in the bottom 0.3 m of the unit. The red chert also occurs as "floating" sand grains (medium to very coarse) and clasts (granules to small pebbles) in the upper meter or so of underlying Unit 15. What processes can account for the deposition of the chert clasts in the much finer grained red siltstone?

Unit 16 is also bioturbated and strongly burrowed in places. Most of the fossil burrow traces are horizontal, occurring predominantly on the base of beds and protruding into the underlying strata (preserved in convex hyporelief). The burrows are generally <0.5 cm wide and <5 cm long, cylindrical, unbranching,

and straight to sinuous. They are identified as the trace fossil *Palaeophycus* sp. (Figure 6-4), which is interpreted to represent open dwelling or feeding burrows constructed by a predaceous or suspension-feeding worm-like animal (Pemberton and Frey, 1982). Larger burrows of *Palaeophycus* (up to 1 cm wide
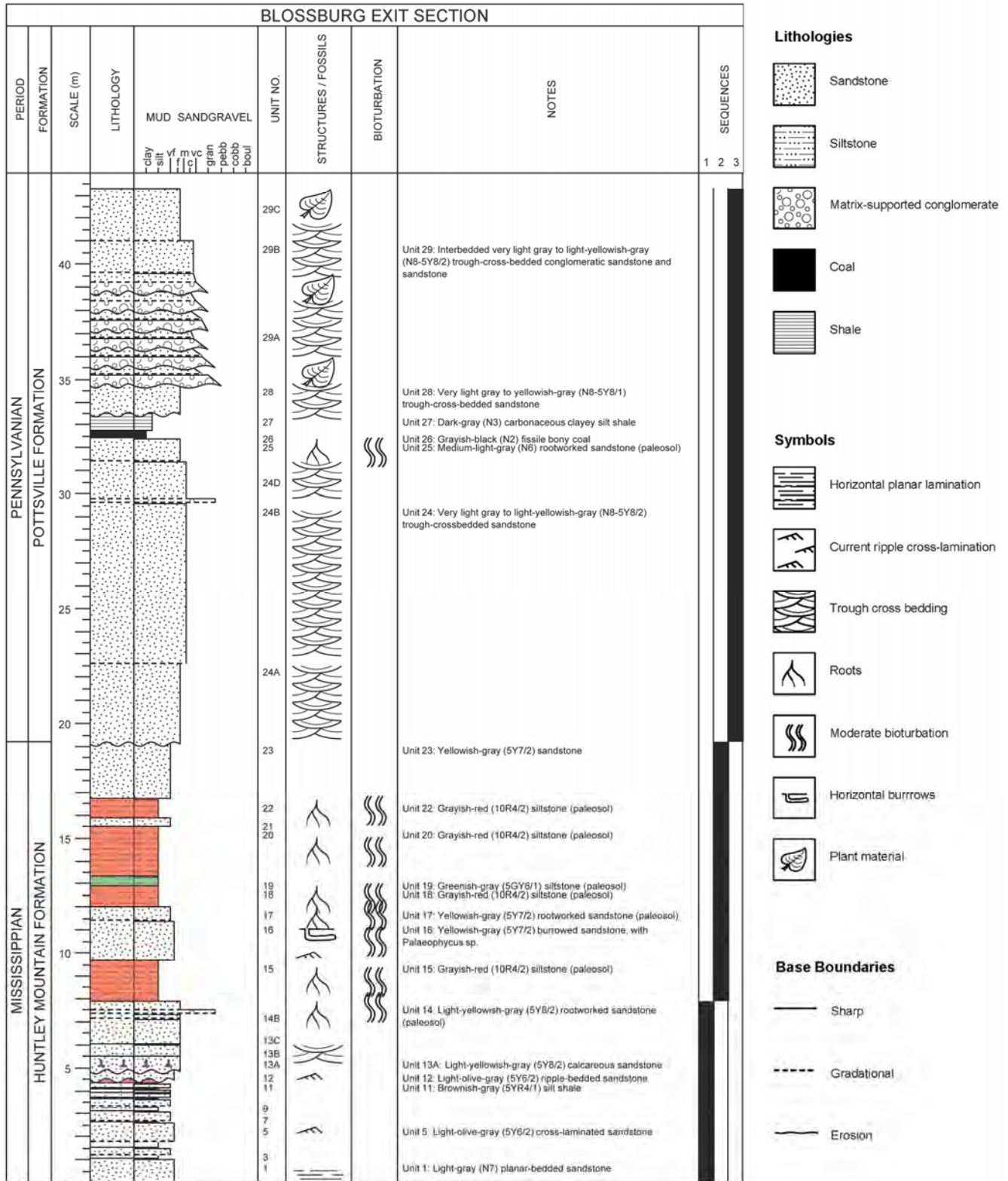

Figure 6-3. Graphic log of Blossburg Exit Section, and explanation of symbols (based on SedLog v.3.0 freeware). Red and green beds colored for clarity.

by 14 cm long) were observed at the next road cut to the north of here. Smaller horizontal to subvertical burrows also occur in places on the top of beds (convex epirelief). No body fossils are present.

Also occurring in Unit 16 are few to common asymmetrical ripple beds that are straight to sinuous crested, with rounded crests. Ripple height (H) is approximately 0.4-0.6 cm and length (L) about 4.5-6 cm. Systematic measurements of H and L were not attempted, and therefore an average ripple index (L/H) was not computed. Nevertheless, the estimated range in values for H and L suggests a probable average ripple index that falls within the range for combined-flow ripples.

Figure 6-4. Trace fossil *Palaeophycus* sp. in convex hyporelief on slab of float from Unit 16.

Unit 16 may represent a crevasse-splay deposit that resulted from levee breach associated with a major flooding (storm) event. An increased nutrient load in these sediments supported a variety of infaunal organisms. *Palaeophycus* is known to occur in a number of depositional environments, from marine to continental, including alluvial. Trace fossils are not uncommon in the Huntley Mountain (Berg and Edmunds, 1979), although *Palaeophycus* is not known to have been reported previously. Are there other plausible interpretations for the depositional environment of this unit? Extraformational discoidal pebbles, for example, are commonly associated with (known to originate in) some marginal-marine environments. The only other known occurrence of jasper pebbles in the Huntley Mountain is associated with the Cedar Run marine conglomerate (Berg and Edmunds, 1979). Could these two distinct occurrences of red chert pebbles share the same source area? The estimated stratigraphic interval (vertical distance) between the Cedar Run and Unit 16 is about 65 m. Also, no other sandstone unit in this sequence (or outcrop) contains trace fossils. Can a case be made that Unit 16 was actually deposited in an estuarine or other marginal-marine environment during a short-duration transgressive event? Does the observed type of ripple bedding support such an interpretation? Clearly, more work needs to be done on this fascinating unit.

Sequence 3

Sequence 3 (units 24–29, 24.1 m thick) includes braid-plain sandstones and conglomerates and minor impure coal (Figure 6-3). The base of the sequence is unconformable. Local relief on the unconformity is around 2 to 4 m, based on field measurements and drill-hole records.

Unit 24 makes up the lower half of the sequence and consists of sandstone (quartz arenite) that is very light gray to light yellowish gray, fine to medium (dominant) grained with scattered coarse grained upward (i.e., unit coarsens upward overall), clean, well cemented with silica, trough cross-bedded with very thin to thin foreset beds and medium to thick cross-bed sets, and slabby to blocky. One thin bed toward the top contains a few scattered milky-white quartz granules to small pebbles (to 5 mm in diameter). Cut-and-fill structures are present. Only a few scattered fossil plant compressions were observed locally.

The unit grades upward to a strongly rootworked (with carbonized fossil rootlets), hackly, very fine to medium-grained, medium-gray sandstone that has no internal bedding or parting (Unit 25). This represents a weakly developed paleosol that supported the vegetation from which the overlying coal developed.

Unit 26 consists of thin, fissile, nonbanded, grayish-black bony coal. It is lenticular and grades locally to a highly carbonaceous clay shale (i.e., rock having <50 weight percent carbon). The coal formed from vegetation deposited in small, relatively short-lived mires that occurred in places on the braid-plain landscape. Braided-stream systems are characterized by rapid changes in discharge, high sediment load, ephemeral channel patterns, and paucity of fine-grained overbank deposits. Consequently, the interfluves associated with Unit 26 tended to be rather limited in area and duration and generally lacked accommodation space and isolation for thicker accumulations of peat to form that were free from influxes of clastic sediments (i.e., free of impurities). Nevertheless, based on subsurface data, small mires (lenticular coals) are recognized to occur repeatedly at this horizon in this area for the first time in the Pottsville, suggesting a temporary decrease in stream load/gradient of the braid system that was in response to an increase in base level controlled by glacial-eustasy (Heckel, 1994; 1995). Unit 26 correlates with the Kidney coal (long-established local nomenclature), which was better developed and formerly mined commercially from 1926-1940 at the Flower Run deep mine, Blossburg Coal Company, about 3.2 km southwest of here.

Unit 27 is composed of chippy to papery, fissile, dark-gray carbonaceous clayey silt shale. No invertebrate fossils were observed. The unit marked the end of peat accumulation, as shifting channels supplied fine-grained sediments that overwhelmed and slowly buried the small coal swamps. Vegetation continued to grow and accumulate for a time, as suggested by the high organic content of the shale.

Thereafter, as indicated by Unit 28 (composed of the same sandstone as Unit 24), the fluvial system reverted back to its original state.

This was followed by the deposition of Unit 29, which consists of interbedded very light gray to light-yellowish-gray conglomeratic sandstone and sandstone, with a few very thin to thin interbeds and lenses of conglomerate. The upper 3 m or so is mostly sandstone. The unit is grossly fining upward. The matrix consists of medium- to very coarse grained sand, and the clasts are composed of milky-white quartz granules to medium pebbles (to 12 mm in diameter). Clasts are mostly matrix supported. The unit is clean, well cemented with silica, and slabby to locally blocky. It is well trough cross-bedded, with very thin to medium foreset beds and medium to thick cross-bed sets. Cut-and-fill structures are common. Small to large fossil plant compressions and impressions (mostly as bark and stems) are few to common. Unit 29 represents an increase in sediment load/gradient and greater stream competency, with some relaxation near the top of the interval. The primary factor controlling the change in character of these sediments is believed to be an initial drop in base level of the fluvial system as a result of sea-level change, followed by some increase (glacial-eustasy). Intensity of precipitation (storms) could have had some influence as well.

The stratigraphic position of the famous Bloss coal is about 5 m above the top of the Section. The Bloss is interpreted as representing a coal-swamp environment associated with a high-sinuosity meandering-stream system on a coastal plain at a time of near-maximum transgression. The Bloss is relatively thin in this immediate area to the west of the Tioga River and was not mined here commercially.

ACKNOWLEDGEMENTS

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STOP 7: Sandstone and shale of the lower Huntley Mountain Formation (Late Devonian–Early Mississippian)
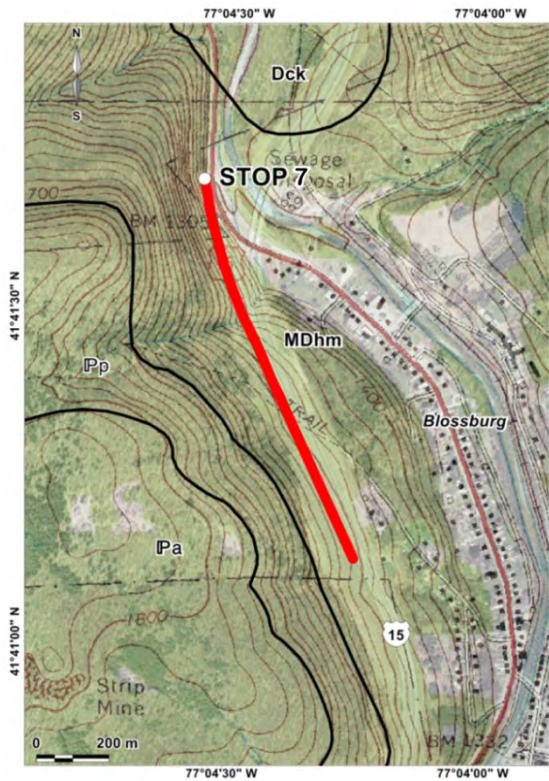
Leader: Brett McLaurin

Figure 7-1. Location map for Stop 7. From Blossburg 7½' quadrangle.

Location and Geologic Setting

This abandoned on-ramp along US 15 S/I-99 is on the north limb of the Blossburg syncline and exposes the south dipping Devonian-Mississippian Huntley Mountain Formation (Figure 7-1). The Huntley Mountain Formation was defined by Berg and Edmunds (1979) to encompass the transition from the red beds of the Catskill Formation (Devonian) to the Burgoon Sandstone (Mississippian) (Figure 7-2). The Huntley Mountain Formation is correlative to the Pocono and Spechty Kopf formations of the Valley and Ridge Province (Berg et al., 1993), the Rockwell Formation in southern Pennsylvania and Maryland (Berg and Edmunds, 1979; Brezinski et al., 2009), and the Price Formation of West Virginia (Brezinski et al., 2009). In the Blossburg area the Huntley Mountain Formation is dis-conformably overlain by the Pottsville Formation (McLaurin and Dodge, 2012).

Figure 7-2. Composite stratigraphic column for the northern Tioga county area of the northern tier of Pennsylvania. Modified from Berg et al. (1993).

At this field conference stop the lower contact with the Catskill Formation is not exposed, but the gray sandstone and red shales of the uppermost Catskill Formation (Duncannon?) can be seen from this point to the north-northeast along US 15 N (weather depending). The northern end of the cut at this stop probably represents the lowest ~ 20-30 m of the Huntley Mountain Formation. Overall, a total of 115 m of the Huntley Mountain Formation is present at this locality. The total thickness of the Huntley Mountain Formation in the Blossburg quadrangle is approximately 160 m (Figure 7-3).
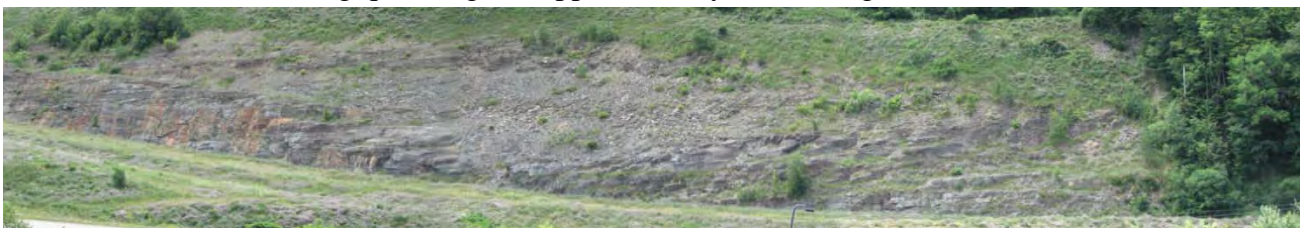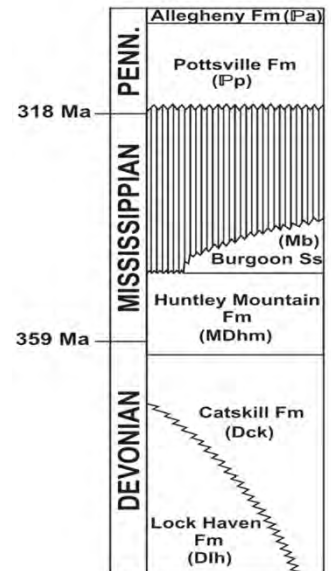

Figure 7-1. Overview of the lower part of the Huntley Mountain Formation at Stop 7. View is to the west.
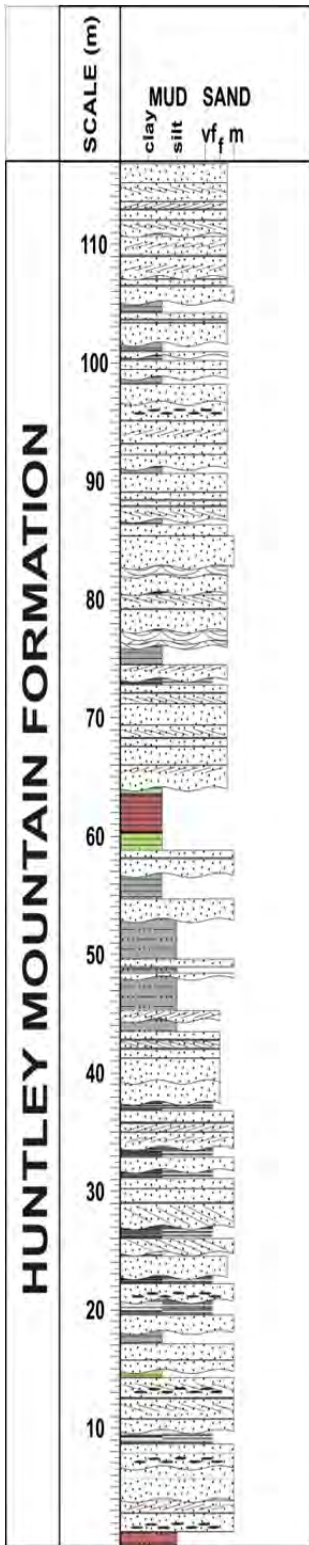
Figure 7-2. Measured section through the lower part of the Huntley Mountain Formation.

Lithology

In this part of the Appalachian Plateau the Huntley Mountain Formation overall, is a sandstone-dominated unit with thin, poorly preserved shale intervals (Figure 7-4). Sandstones are often gray-green, trough and planar-tabular crossbedded with a size range from very fine to medium-grained sand. The cross beds often give the Huntley Mountain Formation a slabby appearance. The sandstone geometry is dominantly sheet-like with planar erosion surfaces and the occasional channel-shaped erosion surface (Figure 7-5).

Figure 7-5. Channel and sheet sandstones with zones of heterolithic barforms.

Intraformational conglomerate is present in thicker sandstone units, typically at the base of fining-upwards cycles, and is composed of carbonate fragments (also previously described as calcareous breccias). These are primarily observed in the lowest 20 – 30 m of the Huntley Mountain Formation. Extraformational conglomerate is rare and was observed interbedded with red shales just below the contact with the overlying Pottsville Formation. Clast size is up to 1 cm in diameter and clasts are composed of quartz and red chert. This succession was examined at field conference Stop 6. Finer-grained intervals occur as 1) heterolithic accretionary zones of interbedded 20 – 30 cm thick sandstone with 3-5 cm green-gray shales and 2) gray, green and red siltstone and shale. The lowest 45 m of the roadcut contain a few ~ 1 m thick shale and siltstone intervals, but is otherwise sandstone dominated. These finer-grained zones typically are laterally restricted, having been eroded by overlying sandstone units. From approximately 45 – 65 m above the base of the section, a thickening of the siltstone and shale is observed. There are three intervals of gray shale/siltstone each of which is > 2 m thick and separated by thinner 25 cm to 1.5 m sandstones. Above this is ~ 5 m thick of interbedded red and green shale (Figure 7-6).

Figure 7-6. Red shales truncated by channel sandstone. This interval of shale may be correlative to the Mount Pleasant Shale.

This zone is truncated and cut out by overlying channel sandstone. This 20 m of finer-grained lithology may be partially correlative to the Mount Pleasant Red Shale as observed in the Huntley Mountain Formation to the southwest in the Waterville quadrangle (Colton, 1968).

Above this fine-grained succession the Huntley Mountain Formation returns to a sandstone-dominated lithology with thinner shales 50 cm – 1 m thick. In addition, there are less of the heterolithic acccretionary units that are observed in the lower part of the Huntley Mountain. The red shale zone observed at Stop 6 may be correlative to the Patton Red Shale as described by Colton (1968).

Age

As noted by Berg and Edmunds (1979) the Huntley Mountain Formation spans the Devonian – Mississippian boundary, but its specific location within the succession is unknown. Within a nonmarine unit such as this, fossils are sparse and what does occur is limited to plant remains such as Archaeopteris? (Figure 7-7). The question of the position of the Devonian-Carboniferous boundary in this road cut was addressed through palynological analysis (Don Woodrow and John Richardson, unpublished PAGS report). In their analysis, fourteen samples from fine-grained intervals were examined to identify the spore assemblage. The presence of species Retispora lepidophyta

Figure 7-7. Archaetopteris? from the Huntley Mountain Formation at Stop 7.

in the uppermost samples, near the top of the roadcut, indicates that much of the Huntley Mountain Formation is Late Devonian in age (late Famennian). Thus, the Devonian-Mississippian boundary is not present in this part of the succession and is probably higher up within the red shales (Stop 6).

Depositional Environment

The Huntley Mountain Formation has traditionally been considered to be the product of a meandering stream that was probably transitional to a braided system (Berg and Edmund, 1979). While evidence for at least moderate sinuosity is present, a modification of architectural-element analysis approach (Miall, 1985) is preferred where individual building blocks of the fluvial system are interpreted. In this system (McLaurin and Steel, 2007), thicker sandstones would represent the channel, heterolithic accretionary intervals would be the barforms and the finer-grained shales and siltstones are floodplain deposits. In the lowest 35 - 40 m of the road cut there are a number of fining-upward packages preserved. Most of these, however, preserve the channel and overlying barform; floodplain deposits are seldom preserved. From approximately 43 – 64 m a significant shift is observed where the succession becomes dominated by floodplain deposits. This transition is significant because the fluvial system is experiencing prolonged stability that is allowing thicker intervals of floodplain fines to accumulate. Whether this is a product of allocyclic or autocyclic controls is uncertain. It is interesting, though, that there is a shift in the architecture above this finer-grained zone. The sandbodies appear more amalgamated and there is less preservation of bar deposits. The red shales above this interval, again, would indicate stability of the fluvial system and establishment of significant floodplain deposits.

Brezinski et al. (2009) interpret greater wetness towards the end of the Devonian associated with Gondwanan glaciation (Figure 7-8).
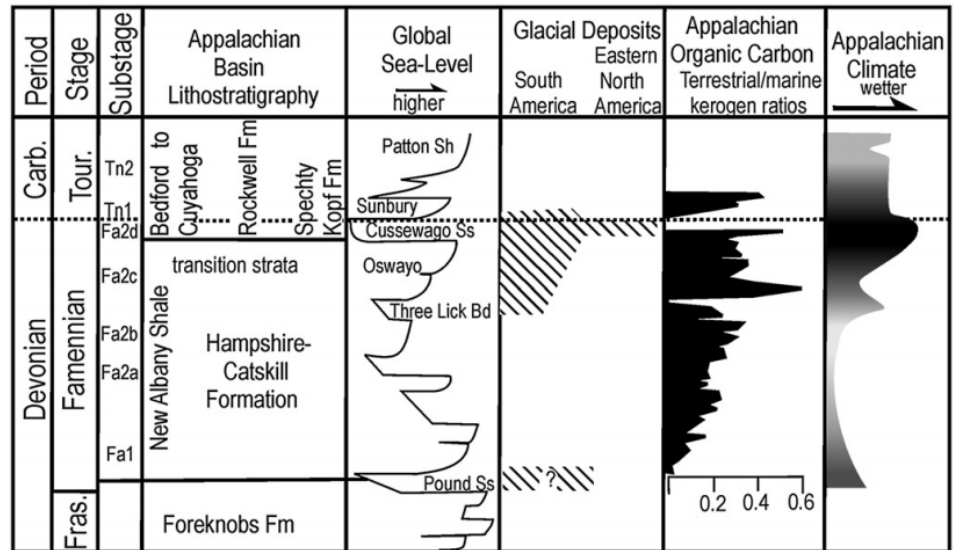

Figure 7-8. Climatic interpretation for the latest Devonian - Carboniferous transition. From Brezinski et al. (2009).

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Stop 8. Lock Haven Fossils.

William E. Kochanov, Pennsylvania Geological Survey

This stop is a brief examination of the paleontology of the Devonian Lock Haven Formation. The unit outcrops here and at other locations along US 15 north of Blossburg.

The majority of the outcrop is covered by loose rock of various sizes with very few outcrops peeking through the rubble (Figure 8-1). Be careful as you walk up and along the side of the hill.

An alternate locality has also been identified at the top of this hill and south along a secondary road approximately 0.3 miles south.

Figure 8-1. Hillside exposure of the Lock Haven Formation.

The Rocks

The Lock Haven Formation was introduced by Faill and others (1977) replacing the previously used name, Chemung. Its type section is described from an exposure along the Susquehanna River north of Lock Haven, Clinton County.

The upper and lower contacts of the Lock Haven are reported to be gradational (Faill and Wells, 1977) for the Saladasburg and Cogan Station quadrangles south of Blossburg where the total thickness of the Lock Haven is given as 3,876 feet (1,181 m). The Lock Haven is approximately 100 m thick but at this exposure we are looking at the uppermost portion (approximately 10 to 15 m). From this point, as one approaches the New York border, we are just about riding the Lock Haven/Catskill contact.

At this locality, bedrock lithologies alternate between, gray to greenish-gray, very-fine to fine-grained, micaceous, sandstone, siltstone, and shale. The rock weathers to a yellowish-gray or reddish-gray depending on the degree of oxidation from residual iron within the rock. Bedding is relatively thin, averaging 4 to 15 cm in thickness. Based on outcrop exposures and float, the bedding contacts can be smooth but can have some irregularities due to the abundant fossils occurring along bedding surfaces. The bedding contacts also tend to be slightly undulating in part.

An outlying outcrop of the Lock Haven, approximately 0.3 miles north of this exposure on US 15, shows the bedding to be more planar (will refer to this section as 8a). The beds still alternate between cycles of very-fine, micaceous, sandstone, siltstone, and shale. Fossils are more sporadic in the lower 2/3's of the exposure. The upper bench is very fossiliferous much like Stop 8.

It is interesting to note the presence of redbeds within the greenish-gray Lock Haven at the section at 8a. This is best observed on the western side of the southbound lanes of US 15 (Figure 8-2). Perhaps the presence of the redbeds signifies the introduction of terrestrial sediments along with freshwater into the marine Lock Haven environment. This may have put a pause in the proliferation of sessile (attached to the substrate), invertebrates due to the changes in substrate and water chemistry. Burrowing organisms may have been better adapted to ride out these changes and "mind the farm" until the marine waters returned. Storm surge may have introduced a shell pavement for the establishment of successive brachiopod/pelecypod communities.

Do these red beds mark the beginning of the Catskill Formation? It has been observed in recent environments where reddish sediment enters a marine setting. Figure 8-3 is a satellite view of the Betsiboka River in Madagascar (NASA, 2004). The red sediment is derived from the erosion of iron-rich lateritic soils. The source of the iron is likely residuum from the dissolution of limestone common to the Island. Even though the reason for such severe erosion on Madagascar is related to deforestation, one could use it as a comparative model where lateritic alluvial plain deposits during Late Devonian time were periodically washed into marine (Lock Haven) settings.

At Stop 8 sedimentation cycles could start with a Lock Haven marine transgression punctuated with periodic storm events and "regressive" events sparked by the occasional terrestrial slug of sediment and freshwater. A tug-of-war so to speak, marking the transition between marine and terrestrial before the Catskill finally establishes its regional domination of the late Devonian.

Figure 8-2. Catskill-like redbeds (brackets) within the marine Lock Haven beds on the west side of US 15 south.

Figure 8-3. Satellite image of the Betsiboka River on the NW coast of Madagascar choked with red sediment and being deposited into the Indian Ocean (NASA, 2002).

The Fossils

One is initially overwhelmed with the abundance of fossils at this site. Fossils can range in occurrence from thin beds of jumbled masses (coquinites?) with no preferred orientation, to fossils oriented right side up; apparently preserved in life positions. The alternation of orientation infers periods of quiescence with higher energy surges of wave and/or currents probably due to storm conditions or perhaps strong tidal influences.

The calcitic shell material, along with identifiable shell ornamentation, has been removed on many specimens through weathering or spalled off as a result of rock breakage. However, careful searching and splitting can result in finding specimens with sufficient detail for a generic identification. Brachiopods (very abundant) and bivalves (common) are the predominant fossils found at this site and cover entire bedding surfaces.

The faunal associations have a consistent core of brachiopod types with two or three dominant genera, *Orthospirifer*. *Productella* and *Tylothyris* (Figure 8-4). They can probably account for 80 percent of the brachiopods present.

Bivalves tend to occur as singles as opposed to *en masse* like the brachiopods and therefore require a bit more careful searching as they may only be partially exposed. In addition, with much of the calcitic shell material lacking and degree of weathering due to open-air exposure, it is oftentimes difficult to recognize the internal casts as being pelecypods.

Crinoid columnals are a bit rare but can be found. Also rare are bryozoans (the "Stictapora" of Faill and Wells, 1977), occurring as small, approximately 0.25 cm in diameter, "sticks" and simple branches.

Trace fossils are typically capped by layers of fossil-shell hash and are represented by burrowers with the form genera *Planolites and Palaeophycus* being the most abundant. The trace fossils appear to be more common in finer-grained sediments and perhaps play the role as pioneers as there are associations of *planolites* with a sparse, low diversity brachiopod/bivalve fauna.

The presence of Planolites places the environmental setting in the *cruziana* ichnofacies sublittoral zone (neritic zone of Crimes, 1975). Sublittoral being defined as the zone between low tide and approximately 200 m in depth.

Cyclic changes in water depth, substrate, and salinity would be the determining factors in faunal associations. On a more regional scale, tectonic activity would in turn influence paleogeography, climatic, and erosional cycles. The relative thinning of the overlying Catskill section observed north of Blossburg as well as the unconformable surfaces noted at the Blossburg cuts along US 15 (Stop 7 and mile marker 38.3) suggests that paleoenvironmental changes are influenced by a progradation of Catskill-type sediments (the high ground) onto the Lock Haven marine shelf (the low ground).

Table 1. Generic listing of fossils at Stop 8.

Brachiopods	Bivalves
Cyrtospirifer sp.	*Grammysia sp.*
Orthospirifer sp.	*Leptodesma sp.*
Athyris sp.	*Mytilarca sp.*
Camarotoechia sp.	*Cypricardella sp.*
Leiorhynchus sp.	*Paleoneilo sp.*
Tylothyris sp.	*Nuculoidea sp.*
Productella sp.	*Cornellites sp.*
Ambocoelia sp.	
Spinatrypa sp.	Crinoid columnals
Douvillina sp.	
Gypidula sp.	Trilobites
Dalmanella sp.	*Trimerus sp.*
Bryozoa indet.	Trace Fossils
"Stictapora" sp.	*Beaconites sp.*
	Palaeophycus sp.
	Planolites sp.

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Bryozoa

"Stictapora"

Brachiopods

Tylothyris sp.

Brachiopods

Productella sp.

Devonchonetes sp.

Machaeraria (Camarotoechia sp.)

Costistrophonella? and Cyrtospirifer sp.

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Brachiopods

Cyrotospirifer sp.

Dalmanella sp.

Bivalves

Paleoneilo sp.

Leptodesma sp.

Echinoderms

Crinoid columnals

Vertebrates

Bone, indet.

Trace Fossils

Palaeophycus sp.

Palaeophycus sp.

Palaeophycus sp.

Planolites sp.

Planolites sp.

Planolites sp.

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STOP 9: Contact between the upper Lock Haven Formation (Late Devonian) and the lower Catskill Formation (Late Devonian)

Leader: Brett McLaurin

Location and Geologic Setting

This field conference stop is in northern Tioga County, 10 km south of the state border of New York. The road cut is on the east side of US 15N, across from the Pennsylvania Welcome Center and can be traced along a distance of greater than 1 km (Figure 9-1). This locality is north of the Caledonia Syncline axis and south of the Sabinsville Anticline (Faill, 2011). This exposure provides a spectacular view of the southeast dipping uppermost part of the Lock Haven Formation with the transition to the lower Catskill Formation. For the sake of simplicity the base of the Catskill Formation is placed at the base of the first significant zone of reddish sandstone and shale which contrasts significantly with the grayish-green sandstone and shale of the main body of the Lock Haven Formation (Figure 9-2). Approximately 84 m of the Lock Haven Formation is present in the road cut (Figure 9-3). The contact with the Catskill Formation begins an 80 m thick zone of alternating red sandstone and shale with green-gray sandstone and shale (Figure 9-4). The uppermost 10 m of the measured section represents true "red" Catskill up into the transition with the Huntley Mountain Formation. In this succession there is at least 90 m of Catskill Formation present (Figure 9-5). McLaurin (2010) noted that the Catskill Formation was approximately 260 m thick further south in the Mansfield quadrangle.
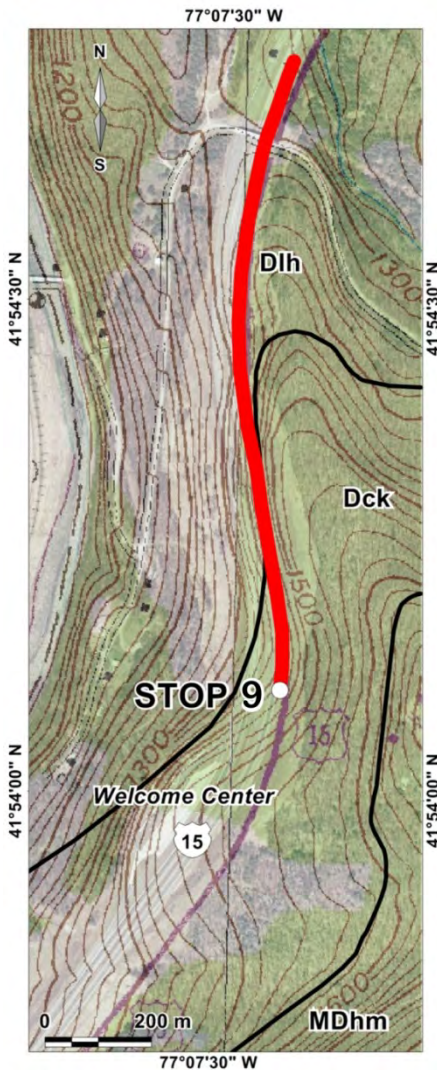

Figure 9- 2. View of the Tioga road cut with the Lock Haven-Catskill contact at the base of the lowest significant red shale and sandstone unit.
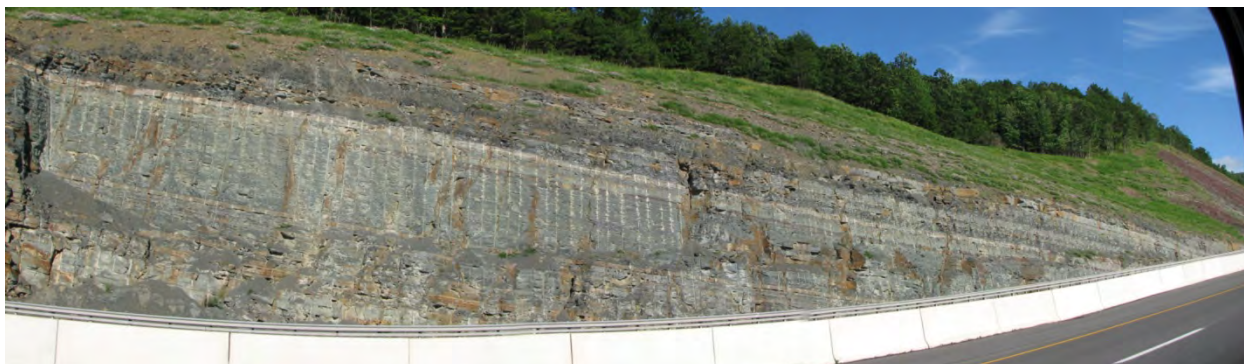
Figure 9- 3. Uppermost sandstone and shale of the Lock Haven Formation. Contact with the Catskill Formation is on the right side of the photo.
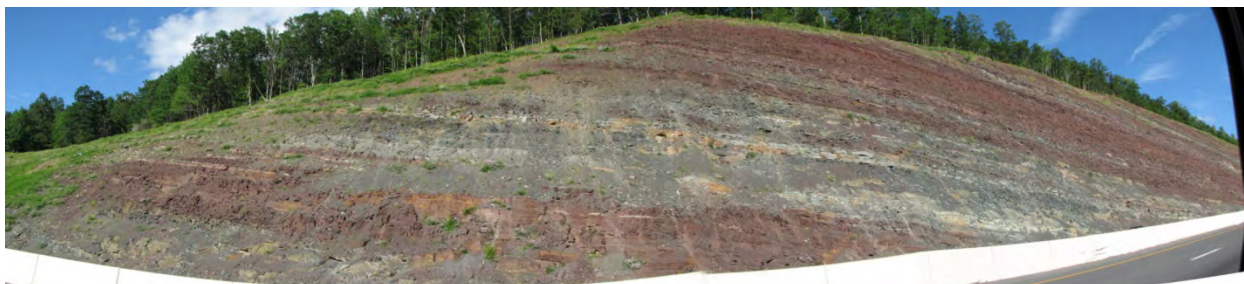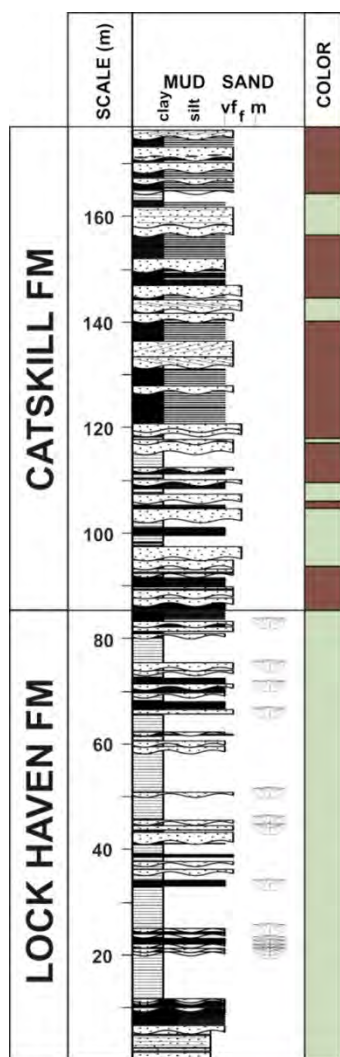

Figure 9- 4. View of the contact between the Lock Haven and Catskill formations.

Figure 9- 5. Measured section through the upper Lock Haven and Catskill Formations. Fossil symbols denote the location of brachiopod occurrences. The colored bar illustrates the alternation of colors observed in the lower part of the Catskill.

Lithology

Lock Haven Formation. The Lock Haven Formation along this road cut begins with a succession of gray-green, silty, very fine-grained sandstone that is laminated and slightly rippled with zones of concentrated brachiopod fossils (Figure 9-5). There are frequently zones of interbedded very fine sandstone (5 – 10 cm thick) and clayey silt (2 – 5 cm thick). A significant portion (~50%) of the Lock Haven Formation here is thicker zones of gray clay-shale that is up to 8 m thick. Punctuated within these shale zones are thinner (30 – 50 cm) reddish-brown "shell hash/lag" containing broken and abraded brachiopod fossils. There is also the occasional 1 – 2 cm thick very fine-grained laminated to massive sandstone layer present. Sandstone and interbedded sandstone/shale increase in frequency towards the top of the formation. Sandstones slightly coarsen to very fine to fine-grained sand and are dominantly laminated. The contact with the overlying Catskill Formation occurs within a 3 m thick interval of interbedded very fine-grained sandstone and shale (Figure 9-6). Above this the occurrence of brachiopod fossils diminishes significantly.

Catskill Formation. The lower Catskill Formation consists of 80m of alternating red sandstone/shale with gray-green sandstone/shale. The red intervals are characterized by trough crossbedded to cross-
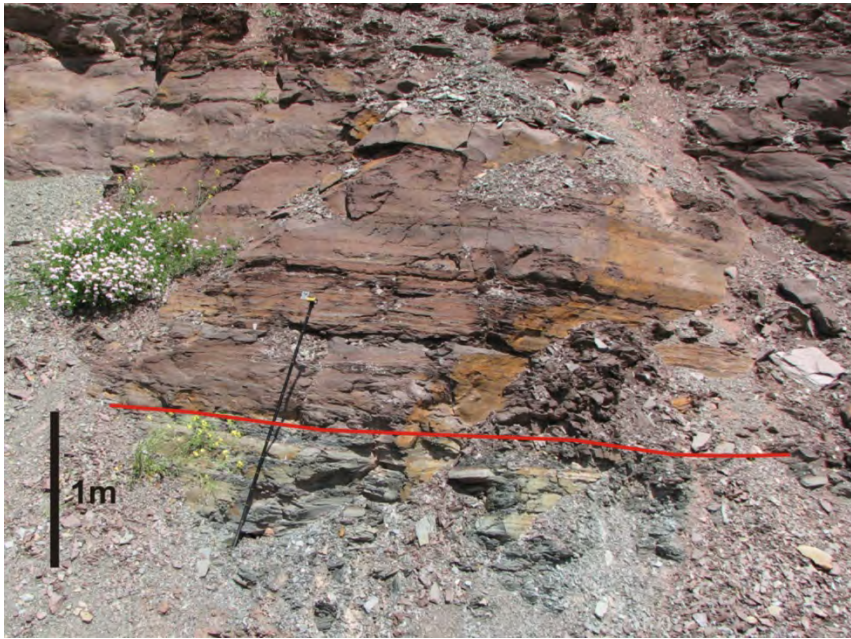
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Figure 9- 6. Close-up of contact between the Lock Haven and the Catskill Formation.

laminated very fine to fine-grained sandstone. The upper parts of the sandstones may be rippled. Plant fragments are present in the sandstone along with structures that appear to be burrows. There is also interbedded very-fine sandstone (1 – 2 cm) and shale (3 – 5 cm). The transition to the gray-green intervals is marked by a slight coarsening of the sandstones to fine-grained sand. The sandstone is largely massive, though some minor cross-lamination is present. At the base of some of these gray sandstones, there are zones of alteration that is a yellow color. In these zones, there is often evidence of silicification and carbonized plant remains. Overall, there are five of these red-gray/green couplets that range in thickness from 5 to 24 m thick. The gray-green packages are primarily sandstone and shale while the red packages show an increase in zones of interbedded thin sandstone (10 – 15 cm) and shale (5 – 10 cm). Vertically, the occurrence of rip-up clasts, pebbles, and cross-bedding increases in both the red and gray-green zones. From a lithofacies perspective there is little difference in the grain size, lithology, and sedimentary structures within these different colored intervals.

Age

The Lock Haven and Catskill formations are both Late Devonian in age. Fossil assemblages from the type section of the Lock Haven Formation contain brachiopod fauna (Faill et al., 1977) that are characteristic of the Givetian – Fammenian substages of the Late Devonian (Paleobiology Database, http://www.paleodb.org).   Specifically, Orthospirifer mesastrialis, Douvillina cayuta, and Nervostrophia nervosa do not extend into the Fammenian and were observed some 400 m below the contact with the Catskill Formation near Lock Haven. Other species, Cyrtospirifer disjunctus and Tylothyris mesacostalils are extant during the Fammenian and were observed by Faill et al. (1977) 60 – 70 m below the Catskill Formation. It is likely that the Frasnian-Fammenian boundary is somewhere in the upper part of the Lock Haven Formation. The age of the Catskill Formation is based on a combination of palynomorphs and vertebrate fossils that have been identified at localities near Covington, PA (Blossburg quadrangle) and at Red Hill, near Hyner, PA (Daeschler and Cressler, 2006; Long and Daeschler, 2013). Palynomorph samples from this stop contain an assemblage characteristic of Fammenian substage 2b (Fa2b). Other localities show assemblages from Fammenian substage 2c (Fa2c) (Daeschler and Cressler, 2006). Thus, the Catskill Formation is of late Famennian age.

Depositional Environment

Lock Haven Formation. The depositional environment interpretation for the uppermost Lock Haven Formation follows that of Castle (2000) who recognizes that marine deposition occurred along proximal and distal portions of a foreland ramp setting. The proximal ramp lithofacies consists of laminated sandstone

that was probably deposited along the middle to upper shoreface. The proximal ramp deposits overlie finer-grained distal ramp deposits that are predominantly gray shale. The shale intervals were deposited below storm wave base. The brachiopod shell lags are also distal ramp deposits and are the result of wave reworking possibly by storm activity. A transitional zone between the distal and proximal ramp is marked by interstratified sandstone and mudstone. The arrangement of the distal and proximal ramp deposits reflects an overall regression up into the Catskill Formation.

Catskill Formation. The Catskill Formation, noted by the appearance of red shales and sandstone is a much sandier unit than the underlying Lock Haven. The lowermost zone of alternating red sandstone/shale with gray-green sandstone/shale is similar to the "Irish Valley motif" (Walker and Harms, 1971) described to the south in the Salladasburg and Cogan Station quadrangles within the lower Catskill (Faill et al., 1977). This motif is described as reflecting alternations of marine (greenish-gray zones) and nonmarine (red zones). While there is an obvious alternation of colors in the lowest Catskill from red to green, the greenish zones lack a strong marine signature as described in the more classic Irish Valley motif. Slane and Rygel (2009) interpret the Catskill Formation here as being deposited in the lower delta plain with transitional zones probably being deposited within a muddy tidal flat. The uppermost 10 m of measured section mark the end of the "Irish Valley" style alternation and a transition to red Catskill that is similar to the Sherman Creek Member (Faill et al., 1977) that is fluvially-dominated.

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Stop 10. Esker cross-section (41° 58' 13.52", 77° 06' 30.0")

Leader- Duane Braun

This stop is in an esker on the floor of the Tioga River valley that fed deltas to the south of it that were deposited in Glacial Lake Tioga (Figure 10-1). 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. There are a number of "bright colored" crystalline clasts from Canada or carbonate clasts from New York. Pit faces are changing continuously so one can't predict what we'll be able to see when we arrive. Immediately to the south of the esker two large, coalesced para-glacial to post glacial fans from east side tributaries occupy the valley.

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 10) both had a significant amount of carbonate and crystalline erratics while the kame deposit southeast of Tioga Junction 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 Fuller and Alden (1903) and Coates (1966). 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.

Braun, D.D. (2013) Stop 10- Esker cross section, *in*, Fleeger, G.M., Schmid, K., and Anthony, R., eds., A Tale of Two Provinces: the Nippenose Valley and Route 15 Corridor, 78th Annual Field Conference of Pennsylvania Geologists, Williamsport, PA, pp. 171 – 173.
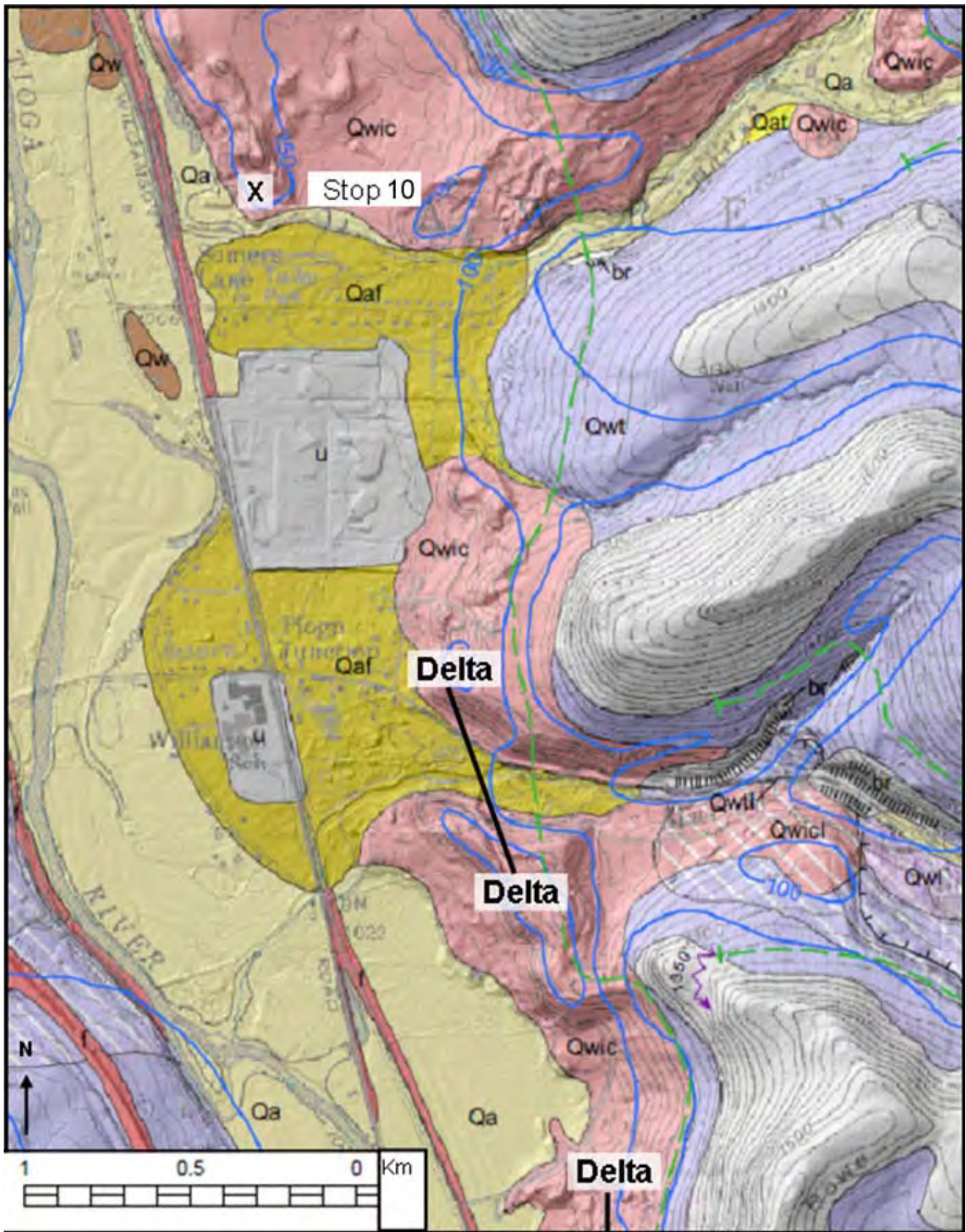
Figure 10-1. Stop 10. Esker ridge composed of ice-contact stratified drift (Qwic). Qwic patch to south is the remnants of a delta built into Glacial Lake Tioga that was fed by the esker. 2009). A second delta lies just to the south of the map. (Jackson Summit Surficial Geology quadrangle, Braun 2009)

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Stop 11 Overlook of the Tioga River - Crooked Creek valleys and the Tioga-Hammond dams (41° 54' 01.28", 77° 07' 34.90")

Leader- Duane Braun. Aaron Bierly

At this site at about 20,000 BP, one would have been standing at ice margin 3 (Fig. 6.1) looking at the ice front curving across the Tioga River-Crooked Creek confluence and watching meltwater pour westward across the ridge between the valleys in the 1,220 feet elevation sluiceway channels.
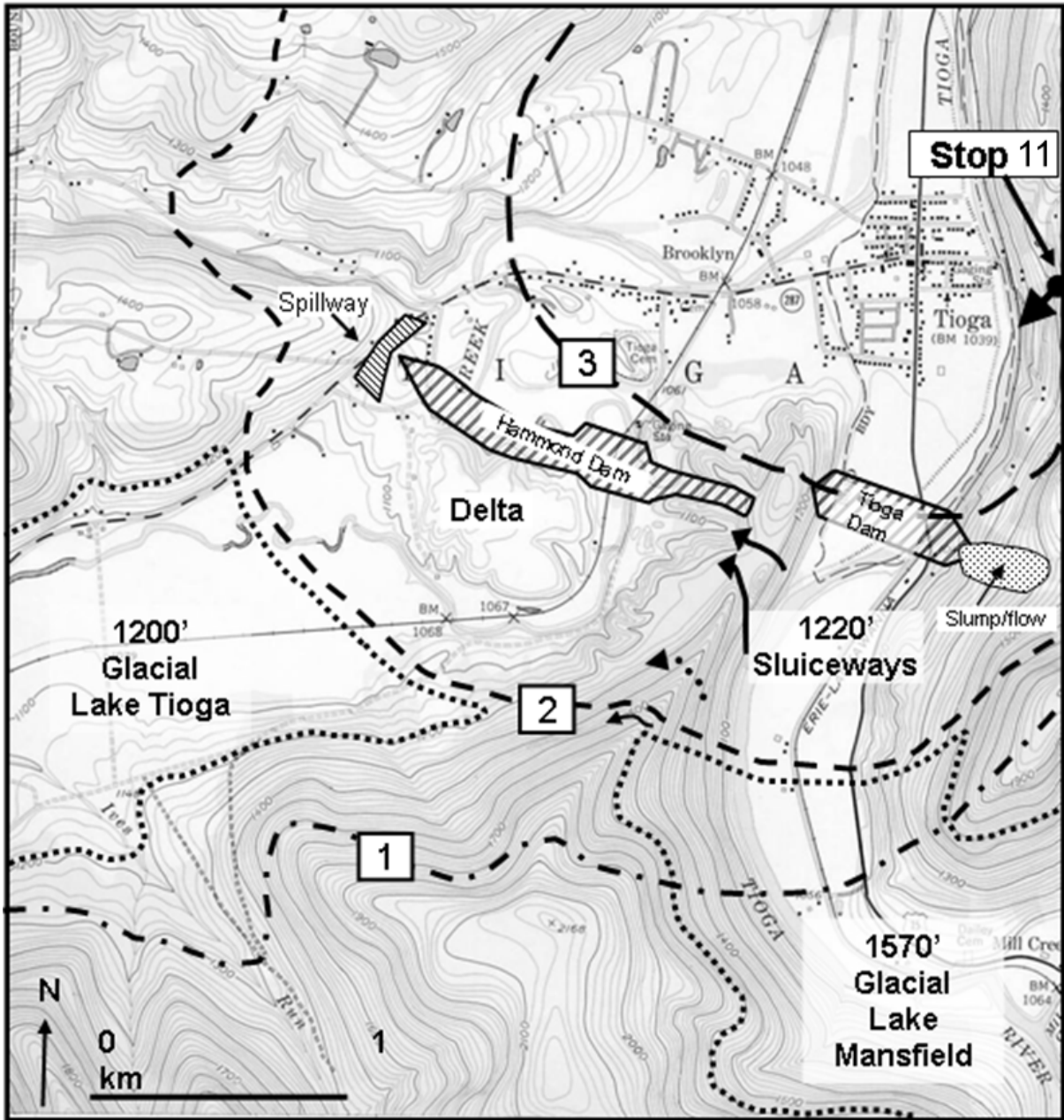

Figure 6-1. Stop 11. Ice margin positions when Glacial Lake Mansfied started draining westward into Glacial Lake Tioga. Ice margin 1 is probably the last or most northerly ice position that held in 1560 feet Glacial Lake Mansfield without any ice marginal or subglacial meltwater drainage westward to 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).

Braun, D.D..(2013) Stop 11- Overlook of the Tioga River - Crooked Creek valleys and the Tioga-Hammond dams, *in*, Fleeger, G.M., Schmid, K., and Anthony, R., eds., A Tale of Two Provinces: the Nippenose Valley and Route 15 Corridor, 78th Annual Field Conference of Pennsylvania Geologists, Williamsport, PA, pp. 175 – 179.

The cutting of the 1220 feet elevation sluiceways

At this stop we are at looking down at the deep connecting channel between the Tioga-Hammond dams (Figure 6-1, 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 (Figure 6-1, 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 (Braun, 1989). 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 (Figure 6-1, south of ice margin 3). When ice receded to the north side of the synclinal mountain (Figure6-1, 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 (Figure 6-1, 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 glacially breached Tioga-West Branch Susquehanna divide, Pine Creek Gorge, the so-called Grand Canyon of Pennsylvania). 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 in the headwater valley of Pine Creek and Glacial Lake Cowanesque in the Cowanesque valley. 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. 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 (Figure 6-2) (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 Figure 6-2). The lake sediments are overlain and cut out by the sandy, bouldery gravel that is more than 100 feet thick 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
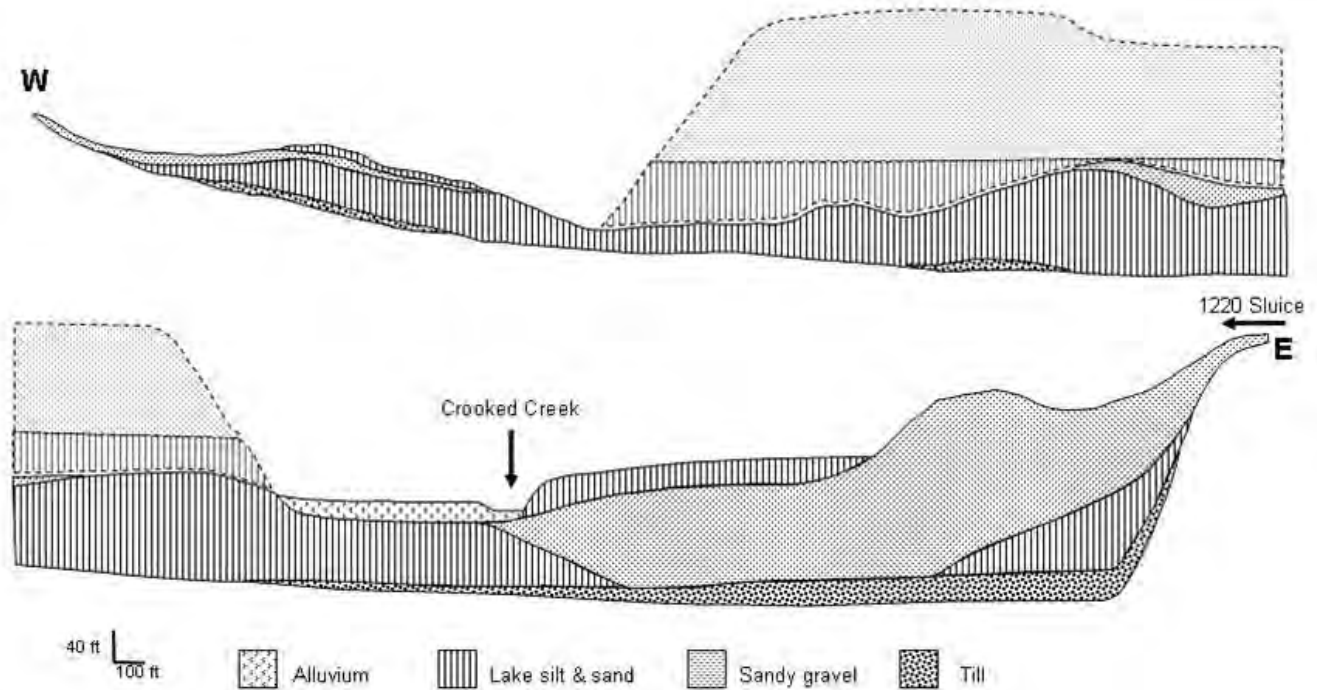

Figure 6-2 Simplified composite cross-section of glacial deposits under the centerline of the Hammond Dam (solid line outlines)(Engineer Corps, Soil Profile H-2) and of the delta south of the Hammond Dam (dashed line outlines)(Engineer Corps, Soil Profile H7-H9). Borehole sites and thin lenses have been 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 (dashed line portion of the cross-section). See figure 6.1 for topographic setting.

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 6-2). 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 6-1 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 6-1) 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. 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.

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. 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 (Figure 6-1, 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 (Figure 6-1, 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.

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.

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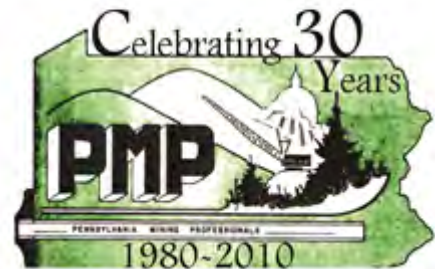
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Wondering where a guy can get a decent meal.
Taken 8/21/2013 at the
Lamb's Creek Recreational picnic area; Lunch Stop, Day 2.