

89th Annual Field Conference of Pennsylvania Geologists

Glacial Retreat, Quaternary Paleoclimates, Paleozoic Bedrock, and Modern Hazards; Exploring Northeastern Pennsylvania's Geologic Past



GUIDEBOOK

Wilkes University



Pennsylvania
Department of Conservation
and Natural Resources
Bureau of Geological Survey



www.fcopg.org

Guidebook for the 89th Annual Field Conference of Pennsylvania Geologists

October 1-3, 2026

Headquarters

Wilkes-Barre Inn and Suites
880 Kidder Street, Wilkes-Barre, Pennsylvania

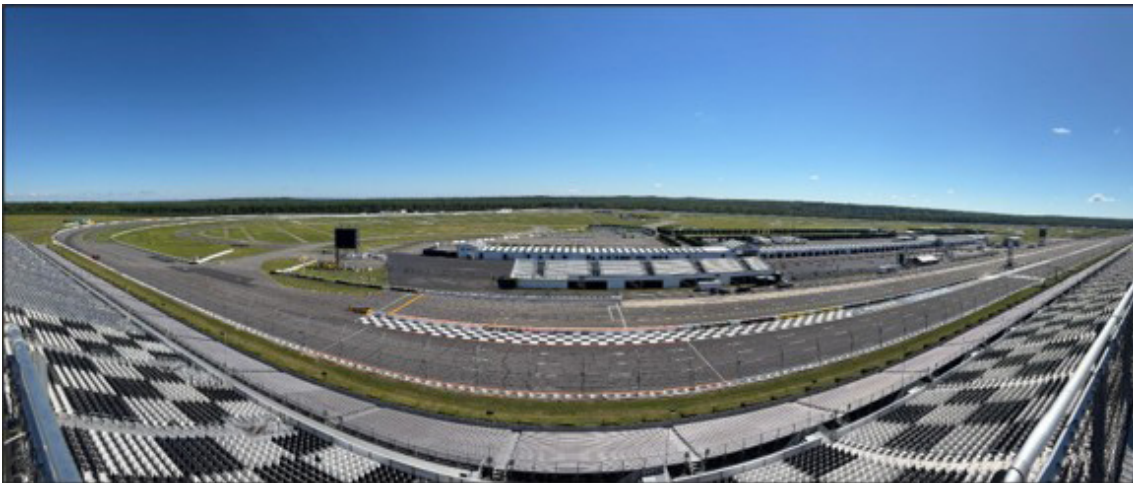


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Panorama of the Pocono Raceway and surrounding terrain including the late-Wisconsinan terminal moraine (Stop 6 and 7) from the Main Grain Stand.

Welcome

Welcome to the 89th Annual Field Conference of Pennsylvania Geologists headquartered in the incised valleys, water gaps, and glaciated terrane of the Wyoming-Lackawanna Valley in Wilkes-Barre, Pennsylvania. This northeastern Pennsylvania location offers us glacial sediments, landslides, anthracite mining, and late-Paleozoic rocks to investigate. This year's conference is hosted by Wilkes University and offers an international flair, with guest leaders from the United Kingdom and Denmark joining us in our exploration of Quaternary landforms and sediments, and Devonian through Carboniferous age rocks, which we share on both sides of the Atlantic. The Field Conference was last in this area for our 62nd Conference in 1997 (headquartered in Scranton, PA).

On the first day, our stops will have us exploring lacustrine sediments at Nuangola Lake, an outcrop discussion of diamictite within the Devonian/Mississippian Specty Kopf Formation, and lunch with geomorphology and age-dating talks at the Seven Tubs Recreation Area. Two additional afternoon stops will focus on ancient landslides situated along the hillslope of Shickshinny Mountain adjacent to the North Branch of the Susquehanna River.

Our evening program back at headquarters will include our annual business session and a keynote address from Dr. Dorothy Peteet, Senior Research Scientist at NASA/Goddard Institute for Space Studies and Adjunct Professor at Columbia University. Dr. Peteet will give a presentation entitled "Timing and patterns of deglaciation across the eastern Laurentide margin – What was the Landscape Like?" that focuses on late-Quaternary paleoenvironments in the northeast United States.

Our second day of field stops begin at the Mauch Chunk (Mississippian) and Pottsville (Pennsylvanian) Formation contact to reveal structurally important bedrock features contributing to the 2024 Toby Creek Mine Subsidence. Our second stop will have us at Pocono Raceway for lunch, our group photo, and a discussion of recent surficial mapping in the Blakeslee Quadrangle, south of the late-Wisconsinan age terminal moraine complex in an area of Illinoian-aged till. Our final stop will have us at the Wildlands Conservancy Maple Tract Preserve to observe the late-Wisconsinan age terminal moraine complex, including findings of recent work by our conference leaders, Dr. Matt Finkenbinder (Wilkes University), Dr. Tim Lane (Aarhus University), and Dr. Kathryn Adamson (Manchester Metropolitan University), in dating glacial erratic boulders in this area. Our next conference is planned for the Broad Top Region in Huntingdon County, Pennsylvania. Please consider leading a Pre-Conference Trip in 2027 if you have one in this area you would like to share.

Enjoy!

Ted Tesler, PG
FCOPG Chairman



Acknowledgements

The Department of Conservation and Natural Resources, Bureau of Geological Survey, Field Conference Committee, would especially like to thank our 2026 Conference Leaders:

Matt Finkenbinder, Associate Professor of Earth System Science, Wilkes University
Bobak (Bobby) Karimi, Associate Professor of Earth System Science, Wilkes University
Nick Fedorchuk, Chairman and Associate Professor, Connecticut State University
Kathryn Adamson, Senior Lecturer in Physical Geography, Manchester Metropolitan University
Tim Lane, Assistant Professor in the Department of Geoscience, AARHUS University
Alistair Monteath, Peatland Paleobiologist, British Antarctic Survey
David McGovern, PG, DEP Bureau of Abandoned Mine Reclamation
Bill Kochanov, DCNR Bureau of Geological Survey, Retired
Ben May and George Ewald, Pocono Raceway
Dorothy Peteet, Senior Research Scientist, NASA/Goddard Institute for Space Studies

And the following:

Wilkes Barre Inn & Suites
Martz Bus Company
Flagger Force, LLC.
The Grove at Nuangola
Pocono Raceway
Wildlands Conservancy
Luzerne Lumber
Garden Drive-In
PA Fish & Boat Commission
Lackawanna County Coal Mine Tour
LaFratte's Catering
Wegmans
Gunn Mowery Insurance
Pennsylvania Department of Conservation and Natural Resources

And, the generous gifts and sponsorships provided by our Conference Members, Volunteers, and Patrons who continue to support and advance the rich history of the Field Conference of Pennsylvania Geologists.



Keynote Address



Dr. Dorothy M. Peteet is a Senior Research Scientist at NASA/ Goddard Institute for Space Studies and Adjunct Professor, Columbia University. She directs the Paleoecology Division of the New Core Lab at Lamont Doherty Earth Observatory of Columbia and in collaboration with GISS climate modelers is studying the Late Pleistocene and Holocene archives of lakes and wetlands (peatlands, salt marshes, tidal freshwater marshes, bogs, fens). Documenting past vegetational change using pollen and spores, plant and animal macrofossils, loss-on-ignition, carbon, and charcoal in conjunction with accelerator mass spectrometry (AMS) radiocarbon dating and x-ray fluorescence (XRF), her research provides local and regional records of vegetational and climate history.

Abstract:

Timing and patterns of deglaciation across the eastern Laurentide margin – What was the landscape like?

We trace the deglaciation retreat of the eastern Laurentide ice sheet northward from southern Connecticut across the Hudson River and into New Jersey and Pennsylvania utilizing lake and bog stratigraphic lithologic sequences with AMS radiocarbon dating of identified plant macrofossils in glacial clays and silts of less than 5% organic matter. Glacial retreat began as early as 17-16 cal yr BP but the majority of the dates are 14-15 cal yr BP. Details of the lithologic and botanical stratigraphy will be presented, with ongoing thoughts about the tundra and boreal forest composition and rates of migration northward from the margin.

Comparison of these sites with new sites in Pennsylvania and sites in Europe along with ^{10}Be cosmogenic ages for deglaciation will be discussed, as well as the context of the timing of northern hemispheric insolation, sea level rise, and meltwater pulses down the Mississippi and the Dnieper and Volga Rivers in Europe.



Contents

Welcome	i
Acknowledgements	ii
Keynote Address	iii

Route and Stop Descriptions

Day 1	5
Stop 1a: Nuangola Lake – Sediment Coring Demonstration	5
Stop 1b: Nuangola Lake – Historic Volcanic Ash Record	5
Stop 1c: Nuangola Lake - Paleoenvironmental Record	5
Stop 1 – Nuangola Lake	6
Stop 2 –Solomon Gap roadcut	22
Spechty Kopf Outcrop at Solomon Gap	22
Stop 3A: Seven Tubs Nature Area - Exposure Dating of Pocono Formation Bedrock Outcrops.....	41
Stop 3B: Seven Tubs Nature Area - Origin of the Seven Tubs in Wheelbarrow Run and Stream Gauging Study	41
Seven Tubs Recreation Area	41
Stop 4A: Landslide Mapping - Garden Drive In Small Landslide	55
Stop 4b: Landslide Mapping - Union Township Boat Launch Large Landslide	55
Landslides of Shickshinny Mountain	55
Return to Wilkes Barre Inn and Suites. End of Day 1	68
Day 2	69
Stop 5a: Luzerne Outcrop	69
Stop 5b: Toby Creek Collapse and Remediation	69
Stop 5c: Toby Creek Geophysical and Hydrology Investigation.....	69
The Toby Creek Subsidence	70



Stop 6: Pocono Raceway - Surficial Geological Mapping of the Blakeslee Quadrangle	83
Pocono Raceway	84
Stop 7: Pocono Lake Moraine - Surficial geological mapping and exposure dating of moraine boulders	98
Pocono Lake Moraine Mapping and Exposure Dating.....	98
Return to Wilkes Barre Inn and Suites. End of Day 2	106



Glacial Retreat, Quaternary Paleoclimates, Paleozoic Bedrock, and Modern Hazards: Exploring Northeastern Pennsylvania's Geologic Past

Map of the Conference Headquarters and Field Stops

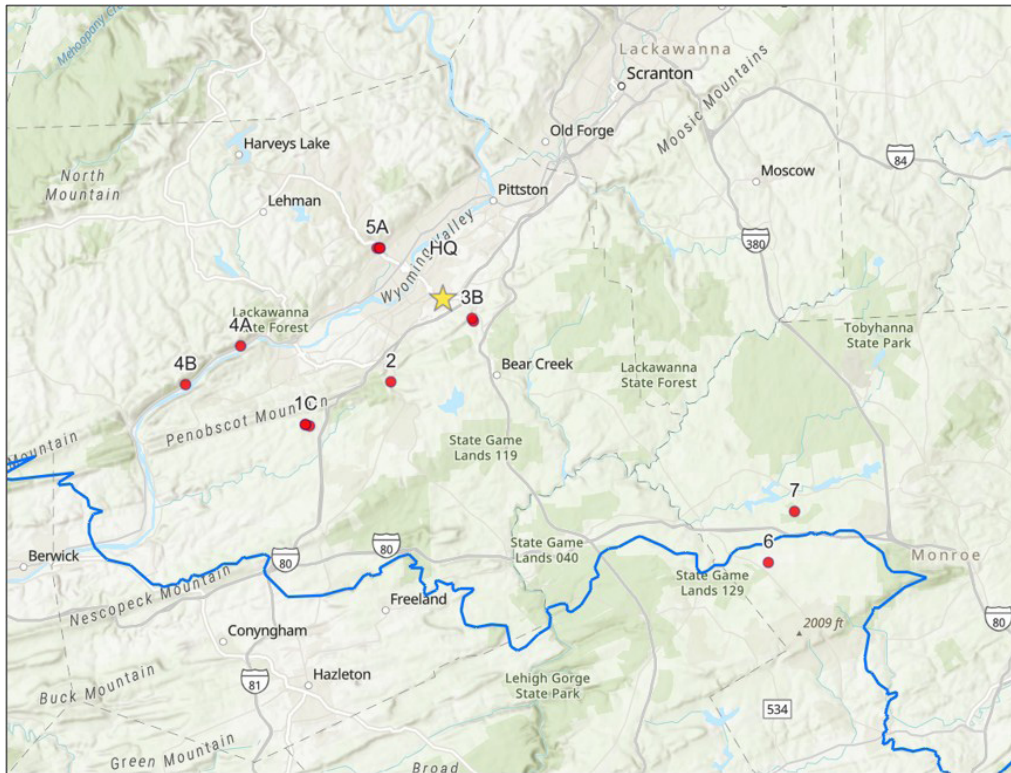


Figure 1. Location map for the 2026 FCOPG.

Headquarters: Wilkes-Barre Inn and Suites, Wilkes-Barre

Stop 1: Nuangola Lake, Nuangola Borough

Stop 2: Solomon Gap, Hanover Township

Stop 3: Seven Tubs Recreation Area, Wilkes-Barre

Stop 4: Shickshinny Mountain, Hunlock Creek

Stop 5: Toby Creek, Luzerne Borough

Stop 6: Pocono Raceway, Long Pond

Stop 7: State Game Lands 318, Long Pond



2026 Field Conference of Pennsylvania Geologists

Matthew Finkenbinder

Introduction

This year's field conference in northeast Pennsylvania has a broad focus on regional topics. The idea for the conference largely originated from a series of undergraduate student focused research projects advised by professors at Wilkes University, Southern Connecticut State University, Aarhus University (Denmark), and Manchester Metropolitan University (United Kingdom). In addition, the field stops are based on research projects advised or co-advised by the field stop leaders and completed by graduate students at Liverpool John Moores University (United Kingdom) and Southampton University (United Kingdom). The field stops cover a wide range of topics related to late-Quaternary paleoenvironmental change and glaciation, surficial geological mapping and radiometric dating of glacial deposits, the sedimentology and stratigraphy of late-Devonian bedrock, landslide mapping and susceptibility analysis, and remediation and post-collapse investigation of the April, 2024 Toby Creek mine collapse.

On Day 1 of the conference, we will start by visiting Nuangola Lake to discuss a recently published paleoclimate record based on an analysis of bog sediment cores (Finkenbinder et al., 2026) (Stop 1) (**Figure 1**). The lake is located ~10 km north of the late-Wisconsinan age terminal moraine complex, and therefore radiocarbon dating and proxy analysis of the sediments provides an estimate for the timing of deglaciation and post-glacial climate and environmental conditions for the region. The field stop will include a sediment coring demonstration, presentation of proxy data along with paleoenvironmental interpretations, and presentation of an invisible volcanic ash (cryptotephra) record from the sediments (Monteath et al., 2025).

The next field stop includes a newish outcrop of the late-Devonian age Spechty Kopf Formation at Solomon Gap (Stop 2) (**Figure 1**), where an ongoing sedimentological, stratigraphic, and geochemical investigation seeks to evaluate the paleoenvironmental interpretation of this rock unit. The origin of the Spechty Kopf Formation diamictites have been interpreted as sub-aqueous debris flows (Sevon et al., 1997) induced by tectonic events or tsunamis, while more recent studies invoke glacial processes associated with a late-Devonian cooling event (Brezinski et al., 2008). New geochemical datasets and proxies of the degree of chemical weathering of mudstones within the Spechty Kopf Formation outcrop here (and from other localities across central and northeast Pennsylvania) add a new perspective to this ongoing debate!

After lunch at the Seven Tubs Recreation Area in Bear Creek Township, the origin of the 7 bedrock potholes (whirlpools) along Wheelbarrow Run will be reviewed (Braun, 1997), and an ongoing stream gauging study will be discussed that provides baseline hydrologic data for Laurel Run (Stop 3) (**Figure 1**). Glacially polished bedrock outcrops of the Mississippian age Pocono Formation will also be visited to discuss a cosmogenic exposure dating project that seeks to generate age constraints on glacial deposits in the region. The last field stops for Day 1 include visits to two landslide deposits located at the base of the Shickshinny Mountain along the Susquehanna River (Stop 4) (**Figure 1**), where a recent landslide mapping and susceptibility analysis was completed (Karimi et al., 2019), and ongoing geophysical and dendrochronology investigations are evaluating more recent slope failures.

On Day 2 of the conference, we will visit a mine collapse that occurred along Toby Creek in Luzerne (Stop 5) (**Figure 1**) that resulted in significant stream loss and immediate repair by the Pennsylvania Department of Environmental Protection. An adjacent outcrop of the Mississippian age Mauch Chunk Formation and Pennsylvanian age Pottsville Formation (last visited during the 1997 Field Conference) (Edmunds and Gillmeister, 1997) will be visited to evaluate the structural features (fractures and

faults) that contribute to losing stream flow. A geophysical and surface water hydrology investigation of the creek and select regional streams will be discussed, that further evaluates whether streams are naturally gaining or losing water flow as they enter the Wyoming-Lackawanna valley.

The next field stop is at the Pocono Raceway in Long Pond and includes lunch at the 300-level (Terrace Club) of the Main Grandstand, which provides a sweeping view of the racetrack and surrounding terrain. The late-Wisconsinan age terminal moraine complex of the Laurentide Ice Sheet is visible across the track and Turn 2 (the Tunnel Turn) and is locally known as the Fern Ridge. After lunch, the results of a recently published surficial geological map of the Blakeslee 7.5-Minute Quadrangle (Winter et al., 2025) will be presented from the Main Grandstand (Stop 6) (**Figure 1**), with a focus on ice sheet marginal landforms and deposits. Supplemental investigations including radiometric dating of bog sediment cores and cosmogenic exposure dating of erratic moraine boulders will be presented and discussed in the context of regional datasets defining the age of the moraine and subsequent timing of deglaciation (Balco et al., 2002; Balter-Kennedy et al., 2024; Corbett et al., 2017; Finkenbinder et al., 2026; Peteet et al., 2012; Prince et al., 2024). In addition, Pocono Raceway personnel will give a short overview of development and use of the track including a history of paving.

The last field stop for Day 2 involves a visit to the late-Wisconsinan age terminal moraine complex at State Game Lands 318 (Stop 7) (**Figure 1**) (or Wildlands Conservancy's Maple Tract Preserve, depending on access), to discuss an ongoing surficial geology mapping project of the Pocono Pines 7.5-Minute Quadrangle. At this stop we will observe the classic hummocky topography and poorly drained landscape encompassing the moraine, and we will discuss results from cosmogenic exposure dating of erratic boulders, which provide additional age constraints for moraine stabilization and deglaciation.

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Route and Stop Descriptions

* All QR codes by stops take you to Google Maps on your phone or tablet to that location.



Bathrooms



Coffee and Donuts



Lunch

Day 1

Start: 08:00 AM : Wilkes-Barre Inn and Suites, 800 Kidder Street, Wilkes-Barre PA 18702
(41.25107, -75.84039)



Stop 1a: Nuangola Lake – Sediment Coring Demonstration

(41.15532, -75.97868) Wilkes University Bog and Natural Area

Sediment cores will be collected from the boardwalk or boat platform using a modified square-rod Livingstone piston corer.

Stop 1b: Nuangola Lake – Historic Volcanic Ash Record

(41.15532, -75.97868) Nuangola Pavilion

PowerPoint presentation of the volcanic ash record from the Nuangola Lake sediment cores and regional synthesis of records from eastern North America.



Stop 1c: Nuangola Lake - Paleoenvironmental Record

(41.15532, -75.97868) Nuangola Grove Theatre

PowerPoint presentation of the Nuangola Lake sediment core geochronology and proxy data. Presentation of sediment cores collected from the lake.

PLEASE DO NOT TOUCH THE CORES.



Stop 1 – Nuangola Lake

Matthew S. Finkenbinder¹, Timothy P. Lane², Kathryn R. Adamson³, and Alistair Monteath⁴

¹ Biology and Earth Systems Sciences, Wilkes University, Wilkes-Barre, PA, USA

² Department of Geoscience and iClimate, Aarhus University, Denmark

³ Natural Sciences, Manchester Metropolitan University, Manchester, UK

⁴ British Antarctic Survey, Natural Environment Research Council, Cambridge, UK

This location includes separate stops where the group will be divided into three groups and rotate between each stop.

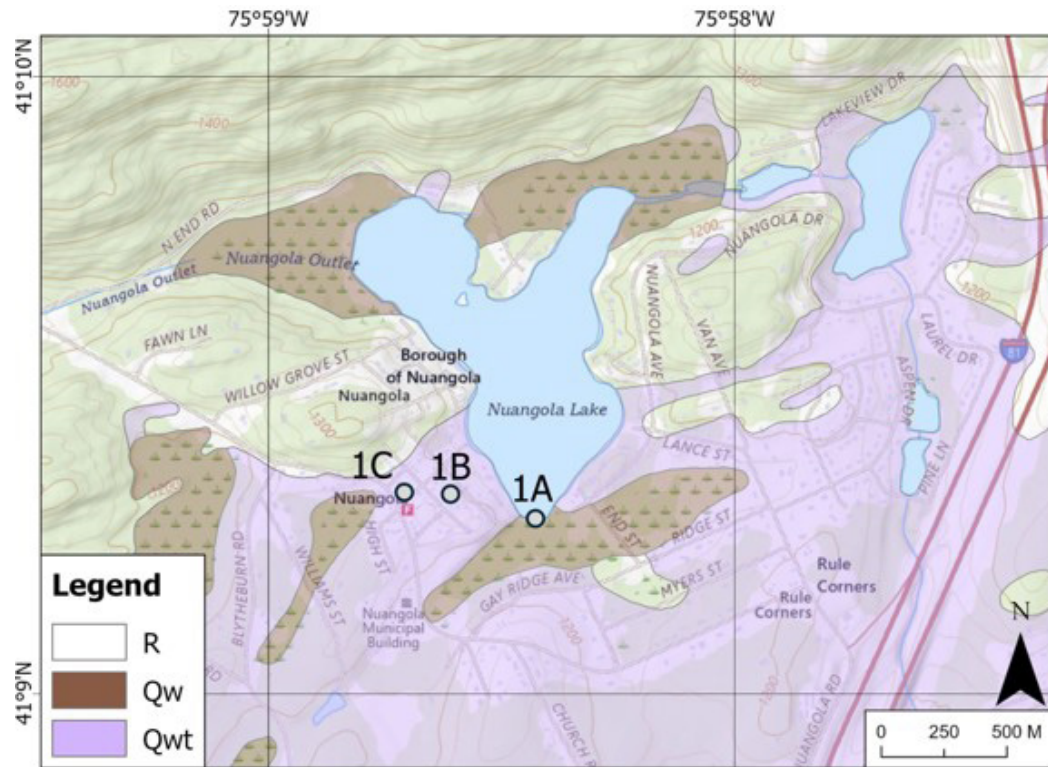


Figure 1. Surficial geology map of Nuangola Lake (Stop 1) showing the location of sites visited at the lake including: 1A) coring demonstration along the boardwalk of Nuangola Bog, 1B) presentation of the Nuangola Bog sediment core cryptotephra record at the Nuangola Pavillion, and 1C) presentation of the Nuangola Bog sediment core paleorecord at the Grove Theatre. Surficial geology units include bedrock (R), wetlands (Qw), and Wisconsin age till (Qwt) (Braun, 2008).

Introduction

Nuangola Lake (41.159° N; 75.974° W; 353 m ASL), is a relatively small (0.40 km² surface area), open-basin lake (**Figure 1**, **Figure 2**) located along the southern edge of Penobscot Mountain in Luzerne County, Pennsylvania. The lake is located ~10 km north of the late Wisconsin age terminal moraine complex of the Laurentide Ice Sheet (LIS) (Crowl & Sevon, 1980) in a 2.98 km² watershed (**Figure 3**). The lake and surrounding catchment are situated in the Susquehanna Lowland and Anthracite Valley sections of the Ridge and Valley Physiographic Province, a region characterized by low to moderate relief, steeply dipping and folded strata with many faults, with the origin of both sections due to fluvial and glacial erosion, and glacial deposition (Sevon, 2000). Local bedrock consists of the late Devonian age Duncannon and Sherman Creek members of the Catskill Formation; primarily sandstone, siltstone,



Figure 2. Photograph of Nuangola Lake from the boardwalk along the bog and southern shoreline of the lake.

and mudstone (Berg et al., 1980), while the surficial geological materials in the catchment consist of Wisconsin age till (Braun, 2008).

Bedrock exposures along and near the lake shoreline indicate that the lake basin was formed by glacier erosion and ice scour during LIS advance and retreat across the site. Present-day catchment vegetation consists of a mixed deciduous and conifer forest that includes white oak (*Quercus alba*), red oak (*Quercus rubra*), red maple (*Acer rubrum*), white pine (*Pinus strobus*), eastern hemlock (*Tsuga canadensis*), and sweet birch (*Betula lenta*). A scrub-shrub wetland (bog) along the southern edge of the lake (encompassing 4 parcels of land owned by Wilkes University) (**Figure 4**) represents a former part of the lake basin that has been infilled with sediments since deglaciation. The bog vegetation is dominated by leatherleaf (*Chamaedaphne calyculata*), cranberry (*Vaccinium corymbosum*), and whorled loosestrife (*Decodon verticillatus*) - all underlain by peat moss (*Sphagnum* species).

Past Research of Nuangola Lake Sediments

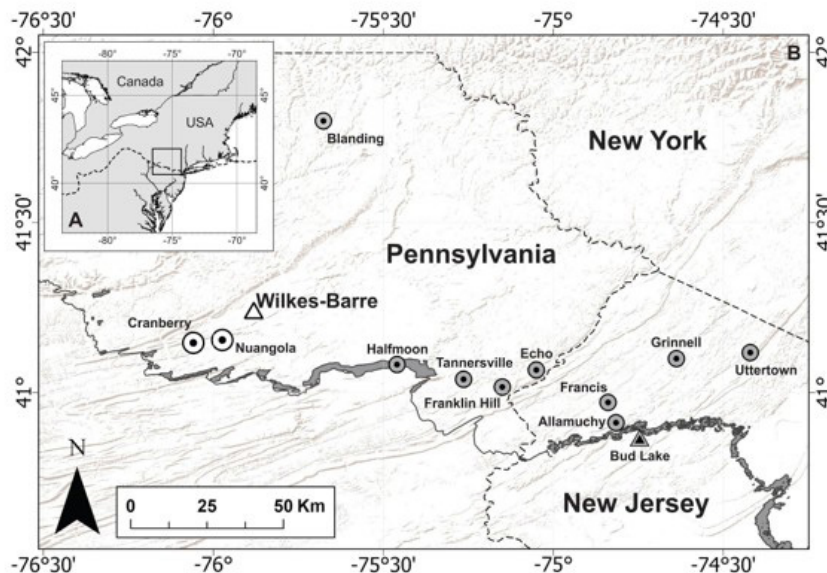
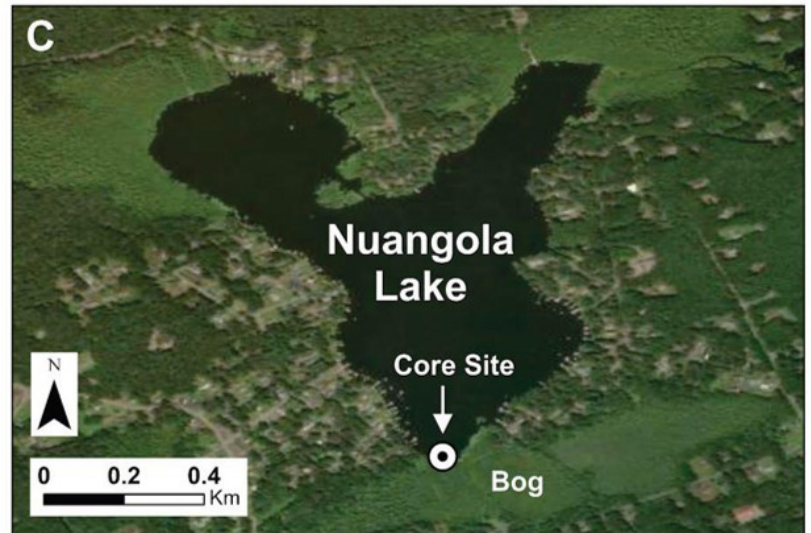


Figure 3. (A) Map of the northeast United States and Canada showing the late-Wisconsinan extent of the Laurentide Ice Sheet (dashed black line) and rectangle defining the extent of the (B) regional shaded relief map of northeast Pennsylvania and the surrounding states (black dashed lines) of New York and New Jersey showing select cities (white triangle), the maximum extent of the Laurentide Ice Sheet during the Last Glacial Maximum (LGM) defined by mapped terminal moraines and ice limits in Pennsylvania (Crowl and Sevon, 1980) and New Jersey (Stone et al., 2002) (gray lines and polygons), and the location of the study site (white circle with black dot) Nuangola Lake and other lake/bog (gray circle with black dot) and moraine sites (gray triangle with black inset triangle). Source: Finkenbinder et al. (2026).

Figure 4. Aerial photo of Nuangola Lake and the surrounding area showing the location of the composite A-17/18/19 core site at the south edge of the lake and bordering Nuangola Bog. Source: Finkenbinder et al. (2026).



There is an extended research history of Nuangola Lake (and bog) sediment cores by Wilkes University faculty and students. Reif & Rodda (1974) collected discontinuous sediment samples every meter to a depth of 11-meters below the surface of Nuangola Bog using an Everhart coring device. The sampling device yielded a 10-inch length by 0.75-inch-wide sample beginning at 1-meter below the bog down to 11-meters. The sediments were analyzed for basic geochemical parameters (organic matter, water content, silica, and oxides as weight percent) and dissolved ions (Ca, Cr, Co, Cu, Fe, Mn, Ni, K, and Zn in ppm). Results from this initial study identified basal minerogenic and clay-rich sediments overlain by 9 meters of organic-rich sediments. Bulk sediment conventional radiocarbon (beta counting via liquid scintillation counting) from 7 meters and 8 meters depth yielded ages of $8,365 \pm 145$ ^{14}C yr BP and $8,760 \pm 145$ ^{14}C yr BP, respectively. Calibration of the ages using the Calib 8.2 radiocarbon calibration program (Stuiver & Reimer, 1993) and the IntCAL20 calibration curve (Reimer et al., 2020) yield median calendar ages of 9,330 years before 1950 CE (cal yr BP) and 9,820 cal yr BP, respectively. Downcore changes in metals were attributed to variations in weathering and oxidation of local bedrock (Catskill Formation sandstones, siltstones, and shales) over time.

Case et al. (1989) collected surface sediment samples from several lakes in Luzerne County and deeper sediment cores from Nuangola Lake to analyze long-term patterns in Lead deposition. Surface sediment samples, representing the upper 15 cm of modern sediments, were collected with a brass Ekman dredge to analyze modern Lead levels in regional lakes and ponds. Discontinuous sediment samples were collected every 0.5 to 1.0 meter to a depth of 9-meters below the sediment-water interface using a piston-coring device at Nuangola Lake to analyze longer-term variations in Lead deposition. The concentration of Lead in sediment samples was analyzed via a reflux digestion using nitric acid and hydrogen peroxide and measurement via atomic absorption spectrophotometry. Results from this study showed elevated levels of Lead in surface sediments from regional lakes and much lower and stable Lead levels in the deeper and older sediments at Nuangola Lake. Case et al. (1989) concluded that the relatively high levels of Lead in recently deposited sediments were not derived from weathering of parent materials, but rather from transport of anthropogenic Lead from the atmosphere.

Recent Research of Nuangola Lake Sediments

The recent studies of Nuangola Lake sediments have focused on the development of a record of paleoenvironmental change via multi-proxy analysis and radiocarbon dating of sediments. In addition, and given the location of the lake and proximity to the late-Wisconsinan ice margin, the basal radiocarbon ages provide constraints on the timing of LIS retreat and deglaciation of the region. Our

sediment coring efforts have focused on the outermost edge of the bog along the edge of the lake, along a boardwalk that traverses the Nuangola Bog property. A small bridge that crosses an old channel across the bog, dug sometime in the early 1900's, has been the site of our field work and sediment core collection (**Figure 5**). We (Matt Finkenbinder and Wilkes University undergraduate students) collected sediment cores from the bog using a modified square-rod Livingstone corer (Wright et al., 1984) that when successfully deployed acquires 1-meter length and 5-cm diameter cores. A total of 11 cores were first collected in March of 2017 (A-17 cores) in PVC casing with no overlap. Subsequently, an additional 12 cores were collected in May of 2018 (A-18 cores) from a parallel core site with 50 cm overlap from the original core site. Additional surface cores were later collected in June of 2019 (A-19) from another parallel core site to ensure complete recovery of the core sequence. Sediment cores were transported to the Department of Biology and Earth Systems Sciences at Wilkes University for processing and analysis.



Figure 5. Photograph of the bridge and boardwalk that crosses Nuangola Bog, and the site of sediment core collection for paleo studies.

The cores were split lengthwise and described – notable sedimentary structures, grain-size, and Munsell colors were characterized for each core. Dry bulk density was measured at 1.0-2.5 cm intervals using a 1 cm³ piston sampler, subsequently oven dried for 48 hours at 80° C, and weighed thereafter. Loss-on-ignition (LOI) analysis was conducted at 1.0-2.5 cm intervals at 550° C for 4 hours to estimate the weight percent organic matter (Dean, 1974). Magnetic susceptibility was measured on all split cores at continuous 1 cm intervals using a Bartington MS2E1 high-resolution surface sensor. The archive sediment cores were imaged using a Geotek® core scanner fit with a Geoscan V high-resolution camera in the Department of Geology and Environmental Science at the University of Pittsburgh. Sediment color values were collected at continuous 0.1 cm intervals and sediment grayscale values were calculated from raw RGB values. A composite sediment core was developed for the Nuangola A-17/A-18/A-19 core sequence field measurements and by aligning visible stratigraphic markers common to overlapping cores. Additional proxy analysis of the cores has included measurement of bulk sediment organic matter total organic carbon (TOC), total nitrogen (TN), and stable carbon isotopes ($\delta^{13}\text{C}$), biogenic silica (BSi), grain size, and elemental abundances. Age control of the sediment cores was developed using Accelerator Mass Spectrometry (AMS) radiocarbon analysis of charcoal and plant macrofossils. The upper portion of the Nuangola Lake core sequence is further constrained using the known ages of independently dated cryptotephra deposits (Monteath et al., 2025).

Three groups of undergraduate level Environmental Science and Geology students at Wilkes University have completed capstone Senior Projects and written theses based on analysis of the cores. Cassie Papp (Environmental Science graduate from 2018) conducted the initial analysis of the A-17 core sequence and characterized the cores at 5 cm intervals for dry bulk density, loss-on-ignition, and magnetic susceptibility. Michael Delpais and Angus Fortune (Environmental Science graduates from 2020) expanded on the earlier work by completing analysis of the A-18 and A-19 core sequences and characterized the cores at continuous 1 cm intervals for dry bulk density, loss-on-ignition, and magnetic susceptibility. They also developed a 11.7-meter length composite core sequence by splicing individual cores together, measured biogenic silica at 5-20 cm intervals, and collected additional plant macrofossil samples for radiocarbon dating. More recently, Megan Novak and Angelica Delgado Rodriguez (Environmental Science graduates from 2025) analyzed the bulk sediment organic matter proxy data from the core sequence. All data were subsequently compiled into a manuscript that sheds light into the timing of Laurentide Ice Sheet (LIS) retreat along with post-glacial climate and environmental conditions from the region (Finkenbinder et al., 2026).

Field Stops

Stop #1A – Coring Demonstration at Nuangola Lake (41.1545° N; 75.9733° W)

Leader: Matt Finkenbinder (Wilkes University) with assistance from Erika Wintersteen (Pennsylvania DEP) and Ryan Wysocki (Pennsylvania DEP)

This field stop involves a coring demonstration along the boardwalk that traverses the Nuangola Bog. 1-meter length cores will be recovered from the bog using a modified square-rod Livingstone corer (Wright et al., 1984) and 1.5-meter core rods. Cores will be extruded into pre-cut 2-inch inside diameter PVC tubes using a winch system. All field equipment was purchased from the Continental Scientific Drilling (CSD) Facility (formerly LacCore and Continental Scientific Drilling Coordination Office), School of Earth and Environmental Sciences, University of Minnesota.

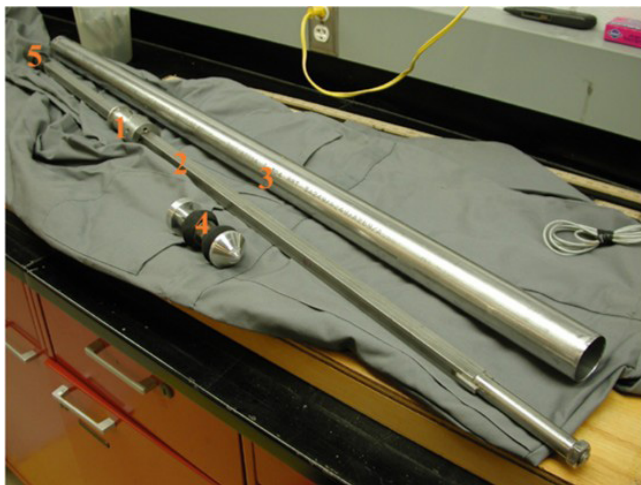
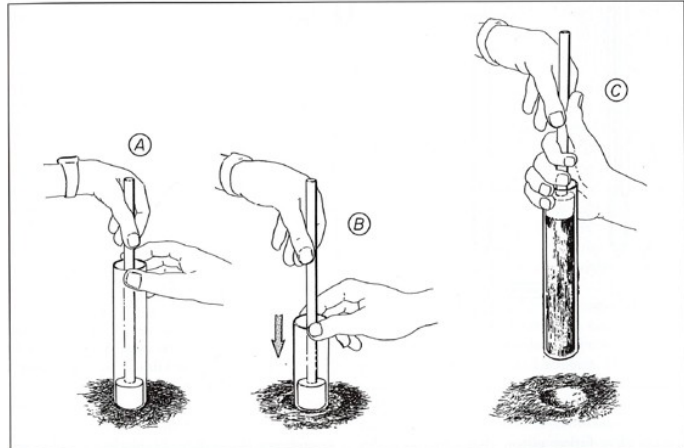


Figure 6. The components of a disassembled square-rod Livingstone corer include a drive head (1), square-rod (2), core barrel (3), piston (4), and a threaded connector (5) that attaches and connects the sampler with drive rods. A steel cable (not shown) connects the piston and threads through the drive head and is used to manage and stabilize the position of the piston when actively coring. Source: Limnological Research Center Core Facility SOP series for Bolivia-Livingstone corer (Myrbo and Wright, 2008).

The Livingstone-type drive rod piston corer (**Figure 6**) can be used in water up to about 30 meters deep to collect successive 1-meter drives down hole of soft to consolidated lake sediment. The cores are extruded between drives into split tubes and stabilized for transport and storage. Operation is most efficient with a team of 3-4 people, though it is possible with fewer people in certain circumstances. The corer can be pushed into soft sediments to a specific target depth, while keeping the sampler closed using the square rod and piston. Once the corer is deployed to the target depth (i.e. where you want to start coring), the square rod is pulled up and rotated 90 degrees and set to a flange on the drive head, which then allows the user to push the sampling barrel past the piston to collect soft sediments. Subsequently, the stationary piston creates a vacuum and prevents the sediments from compacting or expanding as the barrel is retrieved from the sediment (**Figure 7**). Once at the surface, the sediment cores are extruded into core barrels or PVC pipes either manually or with a winch and secured for transport.

Figure 7. Schematic diagram showing the basic principles used in piston coring, (A) the corer is positioned in the sediment with the piston at the bottom, (B) the piston is held stationary and the core tube is pushed past it into the sediment, and (C) the core section is pulled back up the surface with the core tube and piston locked together. Source: Smol (2002).



The main limitations in using the Livingstone corer are related to water depth and person power. The sampler works best in shallow water settings (< 30 meters but ideally 10 meters) where the sediment is fine-grained (silt and clay) and not too cohesive or dense. Coarse grained sediments (sand and gravel) can be impossible to sample, given the lack of cohesion and/or particle size, which limits the ability of unconsolidated sediments to stick together and fit within the barrel. When encountering highly cohesive or dense sediments, retrieving the sampler becomes very difficult and breaking suction requires teamwork, patience, and of course strength. Installing a wider diameter PVC casing, through the water column and into the upper sediments, and coring inside the casing helps achieve greater penetration and acquiring deeper sediments, by taking advantage of an active borehole (which ultimately collapses quickly in soft sediments). Overall, the primary objective of this stop is to illustrate the inherent fun (**Figure 8**) and difficulty in collecting undisturbed sediment cores from lakes and bogs, and to highlight the required skillset and knowledge base involved in completing this type of geological fieldwork.



Figure 8. Photograph of Wilkes University faculty (Matt Finkenbinder) and students extruding a sediment core sample at the Nuangola Bog in April of 2019.

Stop #1B – Cryptotephra Deposits at Nuangola Lake (41.1553° N; 75.9768° W)

Leader: Alistair Monteath (British Antarctic Survey)

This stop includes a presentation on cryptotephra (non-visible volcanic ash) deposits preserved in Nuangola Lake and across northeastern North America (Monteath et al., 2025) at the Nuangola Lake Association Pavillion. These isochrons provide chronology for the Holocene portion of the Nuangola Lake sediment record (**Figure 9**; Finkenbinder et al., 2026) as well as hinting at past changes in atmospheric circulation. Cryptotephra deposits were first identified in northeastern North America at Nordan's Pond Bog, Newfoundland (Pyne-O'Donnell et al., 2012). Prior to this, it was assumed that such far-travelled cryptotephra could only be identified in ultra-clean archives such as ice cores; however, in this peat record cryptotephra deposits were correlated with volcanic sources in the Pacific northwest - up to 7000 km away. Since this pioneering work, numerous studies have explored the peatland and lake archives of northeastern North America, looking for volcanic ash (e.g., Mackay et al., 2016; Jensen et al., 2021). The result is a tephrostratigraphic framework including 45 distinct isochrons, founded on more than 20 sites. Some of these isochrons are derived from relatively minor eruptions while others are from caldera forming events that can now be traced into the North Atlantic, the Greenland ice cores and western Europe (e.g., Jensen et al., 2014; Monteath et al., 2023).

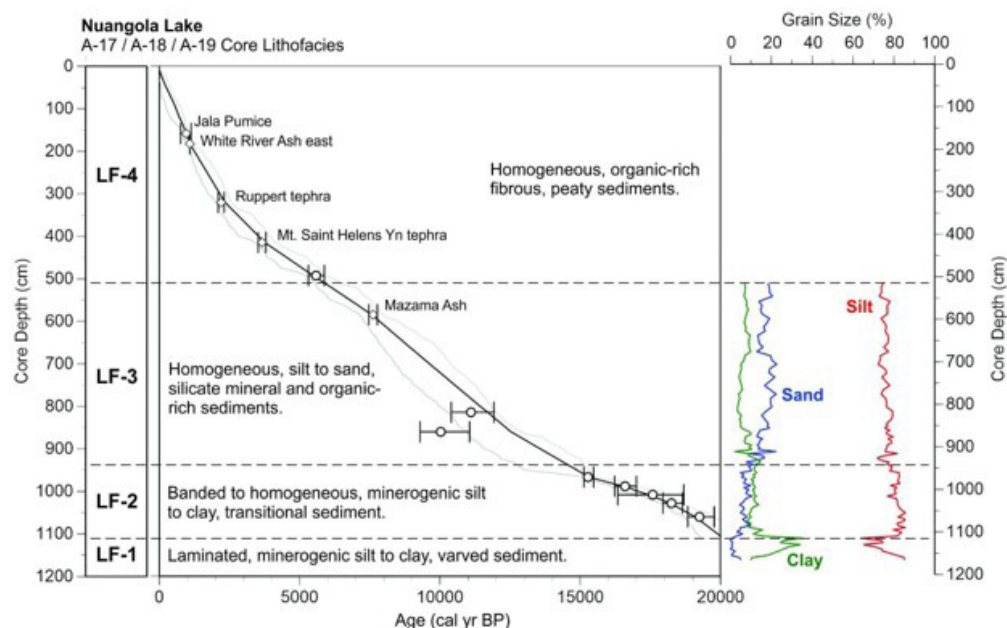


Figure 9. Nuangola Lake stratigraphic column with clastic grain size data and age model for the composite core sequence developed from Accelerator Mass Spectrometry radiocarbon and cryptotephra inferred ages (Monteath et al., 2025) using a cubic spline interpolator. The radiocarbon dates (open circles) and cryptotephra-inferred ages (open diamonds) are presented as the median calibrated age (yr BP) with the associated 2-sigma uncertainty for each age control point (error bars). Grain size data are shown as the percent of clay (green line), silt (red line), and sand (blue line). Source: Finkenbinder et al. (2026).

Cryptotephra shards preserved in the peat and lake deposits of northwestern North America are typically around 40 μm in length - that's about half the diameter of a human hair. So, how do we find this near-invisible ash? The methods follow a simple principle of removing as much material as possible, leaving behind the volcanic ash which can be quantified under a high-powered microscope. First samples are dried, weighed, ashed and weighed again to remove organic material and calculate loss on ignition. They are then placed in 10% HCL overnight, to remove carbonates, and sieved at 80 and 15 μm to exclude anything too big or too small to be identifiable tephra. Finally, heavy density liquids are used to separate tephra from minerals, biogenic silica and other remains that have higher or lower densities than the volcanic glass (Blockley et al., 2005). Once found, individual glass shards can be mounted and analyzed with electron probe microanalysis (EPMA) to quantify major-minor element compositions. This geochemical fingerprint can then be compared with reference material to find the source volcano (Lowe et al., 2017).

In the Nuangola Lake sediment record, we correlated cryptotephra deposits with five explosive eruptions that took place during the Holocene. The oldest is indistinguishable from the caldera forming eruption of Mt Mazama (Crater Lake) which took place around 7600 years ago (**Figure 10**). Above this, younger cryptotephra deposits are correlated with eruptions of Mt St Helens (Yn tephra) in Washington State, Mt Churchill (White River Ash eastern lobe) in Alaska and Volcán Ceboruco in Mexico (Jala Pumice) (**Figure 10**). One final cryptotephra deposit was correlated with Ruppert tephra (**Figure 10**) which has no known origin and is simply named after the lake in Alaska in which it was found (Monteath et al., 2017).

One noticeable feature of the Nuangola Lake cryptotephra record is that none of the volcanic deposits predate the Early-Mid Holocene transition occurring at ~8200 years ago (**Figure 10**). Although there are a handful of exceptions, this pattern is broadly repeated throughout the tephrostratigraphic framework of northeastern North America; of the 45 cryptotephra isochrons identified in the region, only five are predate 8000 cal yr BP. Monteath et al. (2025) explored this pattern and rejected hypotheses related to eruption frequency and preservation bias. This left changing atmospheric transport patterns and tephra dispersal as the only remaining explanation. Comparison with wider proxy records suggests that this trend can be explained by continental changes in atmospheric circulation that occurred as the Laurentide ice sheet collapsed, indicating that distal cryptotephra records are sensitive to long-term atmospheric reorganization. Further work will undoubtedly find new cryptotephra isochrons in the region and test whether the findings of Monteath et al. (2025) remain accurate. Let the hunt begin!

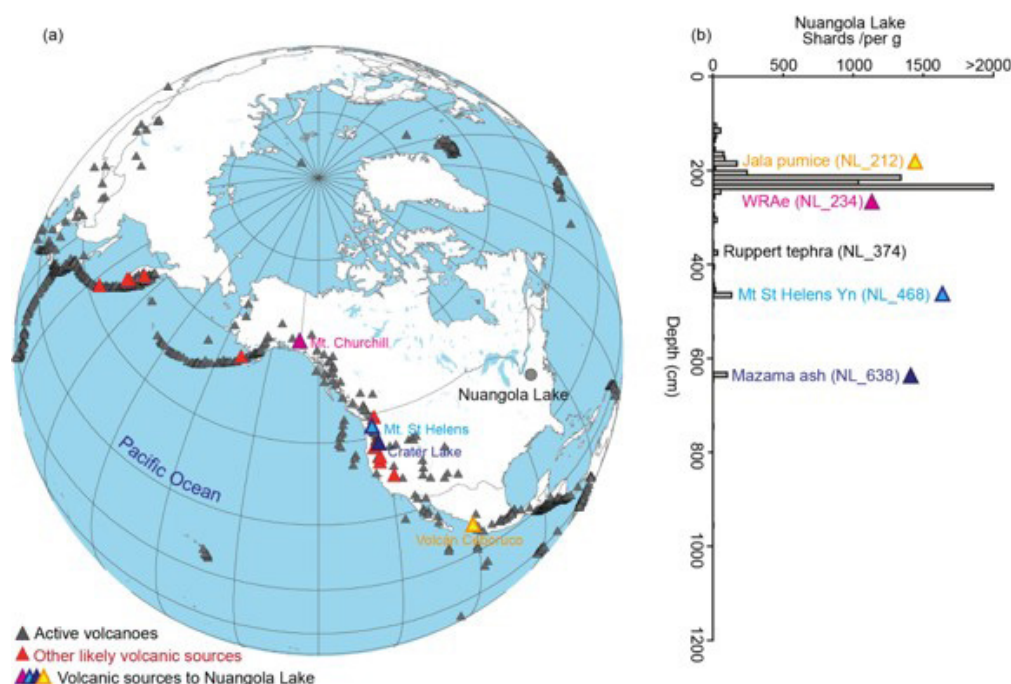


Figure 10. a) Volcanic regions affecting northeastern North America (left panel). Black triangles represent active (Quaternary) volcanoes, red triangles represent volcanic sources to northeastern North America and multi-colored triangles represent volcanic sources to Nuangola Lake. b) Volcanic ash counts (shards per gram) plotted against depth (cm) for the Nuangola Lake sediment sequence. Identified cryptotephra deposits include the Mazama ash at 638 cm, Mt. St. Helens Yn ash at 468 cm, White River ash eastern lobe at 234 cm, Ruppert tephra at 374 cm and Jala pumice at 212 cm.

Stop #1C – The Nuangola Lake Paleo Record (41.1556° N; 75.9786° W)

Leaders: Timothy Lane (Aarhus University) and Kathryn Adamson (Manchester Metropolitan University)

PLEASE DO NOT TOUCH THE CORES.

This field stop involves a presentation of the Nuangola Lake composite sediment core sequence and the resultant paleoenvironmental record (Finkenbinder et al., 2026) at the Nuangola Grove Theatre. The 11.7-meter length sediment core sequence is derived from a series of overlapping 1-meter length drives from the A-17, A-18, and A-19 parallel core sites from the bog. The cores are presented here for viewing and inspection by the conference participants. The cores consist of soft, unconsolidated sediments that are covered with saran wrap to minimize water loss. When not in use, the cores are secured in saran wrap and stored in Ocean Drilling Program d-tubes in a 4° C cold room in the Department of Biology and Earth Systems Sciences at Wilkes University.

The Nuangola Lake composite sediment core extends from the bog surface to a depth of 11.7 meters (**Figure 9**). The composite core age-depth model is based on eight (8) AMS radiocarbon ages from charcoal (characterized by very low bulk density and a platy or elongate, twig-like geometry) and plant macrofossils and five diagnostic cryptotephra deposits (Monteath et al., 2025) (**Figure 9**). The age-depth model reveals a general increase in age with depth and variations in sediment accumulation rates between 0.017 cm/year and 0.200 cm/year. A basal median calibrated radiocarbon age of ~19,260 cal yr BP (~18,840 to ~19,780 cal yr BP range) provides a minimum age for the onset of sedimentation in the Nuangola Lake basin. The composite core sequence is divided into four distinct lithologic units (LF-1 to LF-4) based on variations in core sedimentology and stratigraphic characteristics (**Figure 11**) along with physical and geochemical proxy data (**Figure 12**).

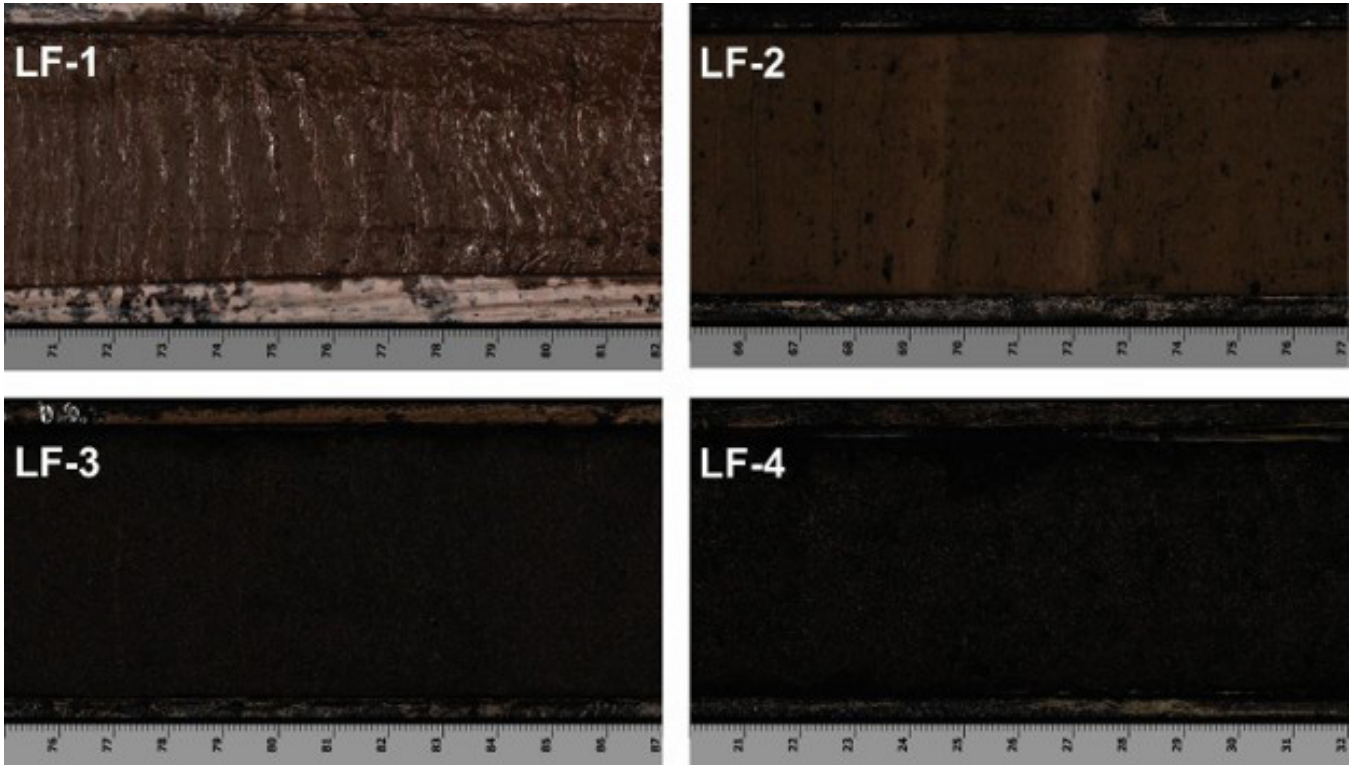


Figure 11. High resolution line scan images of the Nuangola Lake composite core sequence showing representative examples of the four distinct lithofacies, including LF-1 (A-17 Drive 11 70-72 cm, total depth = 1122-1134 cm) showing laminated and minerogenic clay rich varved sediments, LF-2 (A-17 Drive 10 65-77 cm, total depth = 1013-1025 cm) showing banded to homogeneous minerogenic clay and sandy sediments, LF-3 (A-17 Drive 5 75-87 cm, total depth = 523-535 cm) showing homogeneous and fine-grained organic rich sediments, and LF-4 (A-17 Drive 1 20-32 cm, total depth = 68-80 cm) showing homogeneous and organic rich, fibrous peaty sediments. Source: Finkenbinder et al. (2026).

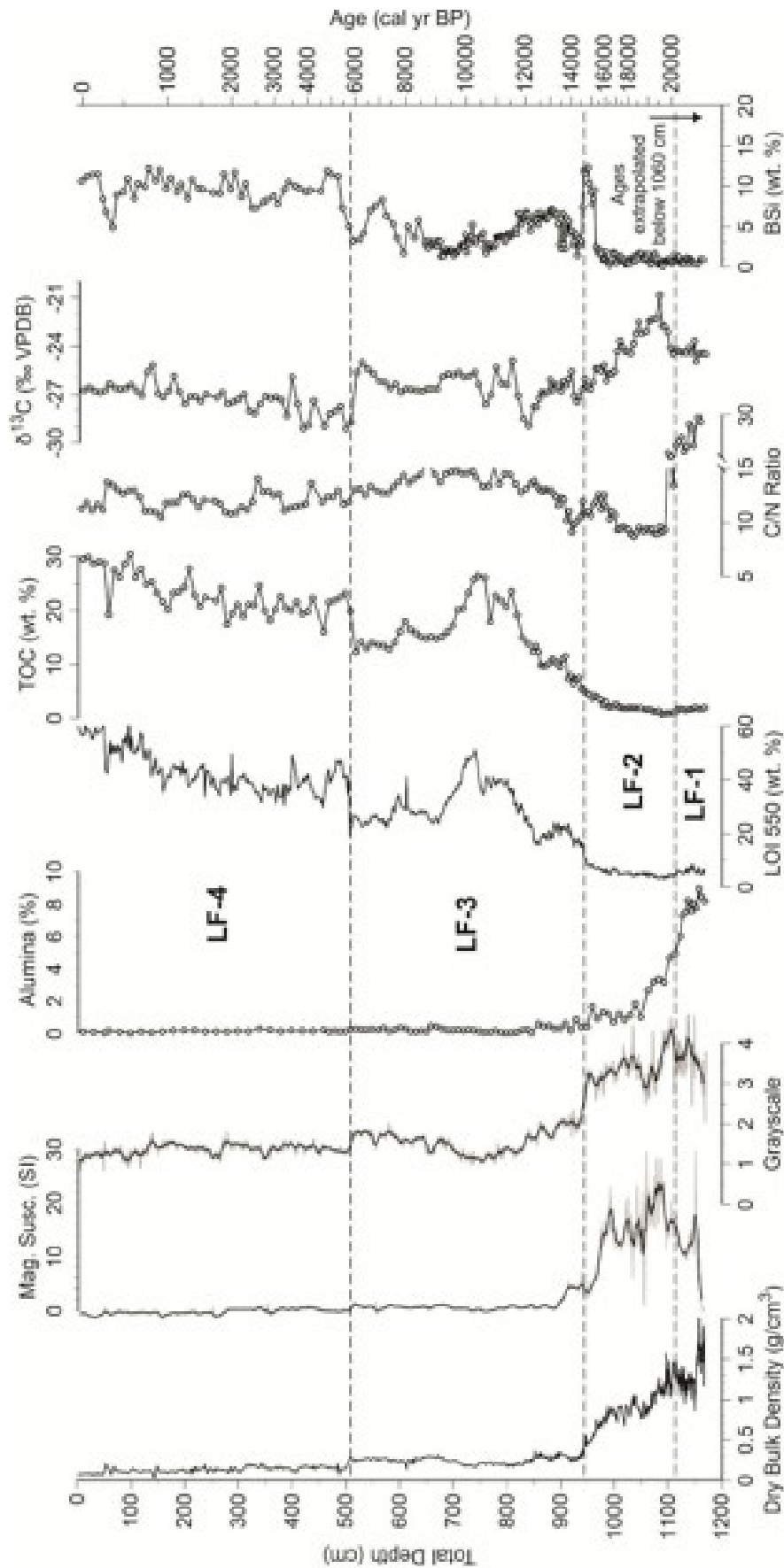


Figure 12. Nuangola Lake composite core physical, chemical, and biological proxy data plotted against core depth (cm) and age (cal yr BP) with lithofacies zones identified by variations in proxy data. Source: Finkenhinder et al. (2026).

Lithologic Unit 1 (LF-1)

LF-1 begins at the base of the core sequence (1170 cm) and extends up-section to a depth of 1115 cm. The stratigraphically lowest radiocarbon age obtained (from 1060-1061 cm in LF-2 sediments) indicates the sediments were deposited sometime prior to ~19,260 cal yr BP (~18,840 to ~19,780 cal yr BP range). The sediments consist of reddish gray (5YR 5/2) finely laminated, minerogenic silt to clay, with a small amount (5 to 10 %) of granules (**Figure 11**). The laminations include couplets consisting of a silt-rich layer (< 1 cm thickness) and thinner clay-rich layer (< 0.2 cm thickness) with sharp basal and top contacts. Layer counting indicates ~70 lamination couplets that vary in thickness from <1 mm up to 10 mm. LF-1 sediments are characterized by relatively high dry bulk density, magnetic susceptibility, grayscale, and alumina values, and relatively low sediment organic and biogenic silica content (**Figure 12**).

The basal sediments reflect a proglacial lake environment where the presence of the retreating LIS within the catchment resulted in a high flux of meltwater with suspended glacial flour into the lake basin. The presence of a small amount of coarse-grained (granules) mineral sediments likely reflects ice-rafted debris input, whereby icebergs calved from the retreating ice front, melted, and deposited dropstones. These interpretations are reinforced by stratigraphic observations of laminated couplets (rhythmites), which are likely varves representing seasonal differences in sedimentation processes. Proxy data supports the interpretation of a proglacial lake environment characterized by a high flux of minerogenic sediments and overall low in-lake and terrestrial productivity.

Lithologic Unit 2 (LF-2)

LF-2 spans a depth range of 1115 to 949 cm. The age-depth model indicates that LF-2 was deposited from sometime prior to ~19,260 cal yr BP until ~14,410 cal yr BP. The sediments consist of banded to homogeneous, dark yellowish brown (10 YR 4/2) to olive gray (5 Y 3/2) minerogenic silt to clay (**Figure 11**). Dry bulk density, magnetic susceptibility, grayscale, and alumina values are relatively high and decrease up-section while sediment organic content values are relatively low and stable (**Figure 12**). Biogenic silica values are relatively low until a major increase and peak in values between ~14,890 and ~14,680 cal yr BP, before decreasing at the LF-3 boundary.

The abrupt change in sedimentological, physical, and geochemical properties at the base of LF-2 represents a substantial change in sediment delivery pathways to the lake. LF-2 proxy data reflects a transitional post-glacial environment defined by retreat of the LIS out of the lake catchment. The decrease in clay content is inferred to coincide with deglaciation of the catchment and cessation of glacial meltwater influence on the lake basin. The age model indicates that catchment deglaciation occurred sometime prior to ~19,260 cal yr BP (~18,840 to ~19,780 cal yr BP range). Relatively high bulk density, magnetic susceptibility, and alumina values indicate sustained high levels of clastic mineral inputs to the lake and overall low organic productivity in a recently deglaciated environment. Following initially unstable catchment conditions, slow soil formation and vegetation colonization in the catchment and surrounding area, likely stabilized the landscape and led to decreased clastic sediment inputs into the lake basin.

Lithologic Unit 3 (LF-3)

LF-3 spans a total depth range of 949 to 509 cm. The age-depth model indicates that this unit was deposited between ~14,410 and ~5920 cal yr BP. The sediments consist of homogeneous, dark yellowish brown (10 YR 4/2) to dusky yellowish brown (10 YR 2/2), fine-grained (silt to clay) organic rich sediments (**Figure 11**) characterized by relatively low and generally stable dry bulk density, magnetic susceptibility, grayscale, and alumina values (**Figure 12**). Organic content is low at the base, gradually increases up-section, peak between ~10,000 to ~10,900 cal yr BP, and then decreases to the LF-4

boundary. Biogenic silica values are low to intermediate and vary on millennial timescales with a broad minimum between ~10,000 to ~11,000 cal yr BP (**Figure 12**).

The transition to fine-grained (silt to clay) (LF-3) sediments characterized by gradually increasing organic and biogenic silica content and low dry bulk density, magnetic susceptibility, and alumina values indicates generally high and stable water levels. Higher and variable sediment organic content and biogenic silica values indicate greater aquatic and terrestrial ecosystem productivity. Collectively, proxy data to represent post-glacial conditions characterized by an open lake environment with higher and more variable levels of ecosystem productivity compared with older sediments. Increasing and variable organic content and BSi likely reflect some combination of warmer and/or wetter conditions where catchment soils were mature, supporting dense, terrestrial vegetation cover.

Lithologic Unit 4 (LF-4)

LF-4 sediments begin at 509 cm total depth and extend up-section to the bog-water interface (0 cm). The age-depth model indicates these sediments were deposited between ~5920 cal yr BP and 2017 CE. The sediments consist of homogeneous, dark yellowish brown (10 YR 4/2) to dusky yellowish brown (10 YR 2/2), organic rich and fibrous, peaty sediments (**Figure 11**) largely comprising partially decomposed plant matter, derived from present day wetland vegetation types including mosses and shrubs. LF-4 sediments are characterized by very low and stable dry bulk density, magnetic susceptibility, grayscale, and alumina values (**Figure 12**). Sediment organic content values are relatively high and gradually increase up-section to the core top and biogenic silica values are the highest on average throughout the entire core sequence.

The rapid transition from fine-grained organic rich lake sediments to organic rich and fibrous peaty sediments in Nuangola Lake occurred at ~5900 cal yr BP (**Figure 12**). The sedimentology and proxy data suggest the sediments reflect a wetland (bog) environment characterized by sustained high water levels and water saturated (shallow water) conditions. This interpretation of the depositional setting is consistent with the site of core collection; a scrub-shrub wetland (bog) along the southern edge of Nuangola Lake. The abrupt nature of the transition to peaty sediments could reflect a lowering of water levels and expansion of the bog along the southern edge of the lake, or terrestrialization and gradual infilling of the lake basin and a reduction in accommodation space.

Deglaciation of Northeast Pennsylvania

The proximity of the Nuangola Lake basin to the late-Wisconsinan ice margin, combined with the new basal radiocarbon ages, provides new constraints on the deglacial history of the region. The basal calibrated median radiocarbon age from Nuangola Lake indicates the onset of post-glacial sedimentation sometime prior to ~19,260 cal yr BP (~18,840 to ~19,780 cal yr BP range) (**Figure 13**). However, given the sample was isolated from post-glacial sediments ~54 cm above the transition from glaciolacustrine sediments (LF-1), the age represents a minimum limiting time of deglaciation. Extrapolation of the age model beyond the basal radiocarbon age, to the contact with glaciolacustrine sediments (LF-1), yields an age of ~20,270 cal yr BP for deglaciation of the Nuangola Lake basin. This extrapolated age is highly speculative and does not account for the typically high sedimentation rates in deglaciated catchments, due to highly unstable catchment materials and a lack of vegetative cover.

The basal radiocarbon age from Nuangola Lake represents the oldest known age for proglacial lakes in Pennsylvania. The calibrated median radiocarbon age of ~19,260 cal yr BP (~18,840 to ~19,780 cal yr BP range) lies between the current regional estimates of deglaciation from the late-Wisconsinan terminal moraine complex bound by basal radiocarbon ages from lakes and bogs (J. Cotter et al., 1986; J. P. Cotter, 1983; Crowl, 1980; Peteet et al., 2012) and cosmogenic exposure ages on moraine

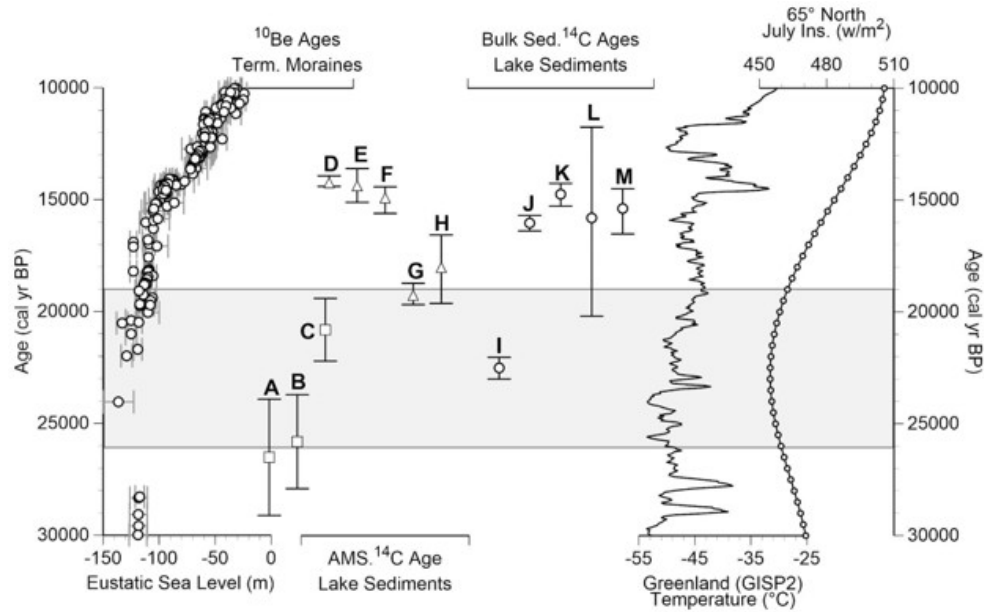


Figure 13. Comparison of Last Glacial Maximum proxy records and the culmination age of glacial moraines and age of deglaciation from regional lake and bog records, including eustatic sea level; ^{10}Be cosmogenic exposure dates (black square symbols) from Martha's Vineyard Island, Massachusetts, Bud Lake, New Jersey (B), and the Ronkonkoma moraine, New York (C); accelerator mass spectrometry ^{14}C basal ages (black triangle symbols) from lakes including Tannersville Bog, Pennsylvania (D), Allamuchy Pond, New Jersey (E), Uttertown Bog, New Jersey (F), Nuangola Bog, Pennsylvania (G) Finkenbinder et al. (2026), Cranberry Pond, Pennsylvania (H) (Finkenbinder et al. (2026); bulk sediment traditional ^{14}C basal ages (black circle symbols) from lakes including Francis Lake, New Jersey (I), Tannersville Bog, Pennsylvania (J), Halfmoon Lake, Pennsylvania (K), Echo Lake, Pennsylvania (L), and Franklin Hill, Pennsylvania (M); Greenland Ice Sheet Project 2 (GISP2) temperature; and 65° North July insolation. The gray bar depicts the time range for the Last Glacial Maximum. Source: Finkenbinder et al. (2026).

boulders and proximal bedrock outcrops in southern New England and adjacent New York and New Jersey (Balco et al., 2002; Balter-Kennedy et al., 2024; Corbett et al., 2017) (**Figure 13**). While the exact reason for this is unclear, it could be related to several factors including 1) spatial-temporal variations in LIS retreat, 2) the investigation of proglacial lake sediments collected from a bedrock basin located in close proximity to the ice margin and 3) the application of radiocarbon dating micro-charcoal fragments in low organic content post-glacial sediments.

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Stop 2: Spechty Kopf Outcrop at Solomon Gap

(41.18782, -75.89258) Route 309 (Wilkes Barre Township Boulevard)

This stop will examine an outcrop of the late Devonian (Famennian) Spechty Kopf Fm. and early Mississippian (Tournaisian) Pocono Formation.

Paleoenvironmental Interpretation of the Spechty Kopf Formation, Route 309 Solomon Gap Roadcut

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Abstract

The Route 309 roadcut at Solomon Gap, northeastern Pennsylvania, exposes Upper Devonian to Lower Mississippian strata of the Spechty Kopf and Pocono Formations. The Spechty Kopf Formation contains a diamictite interval whose origin has been the subject of sustained debate, with interpretations ranging from subaqueous debris flows to direct glacial emplacement by alpine glaciers advancing from the Acadian Highlands. A glacial origin would make these deposits from the earliest pulse of late Paleozoic ice age (LPIA) glaciation, and one of the only examples of LPIA glaciation at low paleolatitudes. Analysis of the Spechty Kopf Formation at this locality identifies three co-genetic facies: a soft-sediment deformation facies comprised of slides/slumps, ball and pillow structures, and boudinaged sandstone bodies; a sandy diamictite facies; and a laminite and lonestone facies with isolated pebble-sized clasts and current-generated sedimentary structures. Together these facies are interpreted here as part of a fan delta system in which gravity-driven mass transport processes delivered sediment downslope into a standing body of freshwater. The absence of striated or faceted clasts, lack of a subglacial deformation fabric, lonestone geometry consistent with debris flow outrunner clasts rather than ice-rafted dropstones, stable Chemical Index of Alteration (CIA) values, and freshwater strontium/barium ratios collectively suggest that the diamictites at this location were emplaced by subaqueous debris flows into a lacustrine environment rather than directly by overriding ice. The overlying Pocono Formation records an abrupt transition to amalgamated braided fluvial deposition, marking a fall in base level and renewed progradation of coarse clastics from the Acadian Highlands.

Stop 2 –Solomon Gap roadcut

Route 309 -- eastern side of highway, Mountain Top, Pennsylvania, 41.18713° N, 75.89169° W

Introduction

This site is a well-exposed roadcut at Solomon Gap along Route 309 on the eastern side of the highway just north of Mountain Top, Pennsylvania (**Figures 1 and 2**). The outcrop exposes approximately 150 meters of Late Devonian (Famennian) to Early Mississippian (Tournaisian) strata spanning the enigmatic Spechty Kopf Formation and the overlying Pocono Formation. Critically, the Spechty Kopf Formation at this location contains a poorly-sorted, matrix-supported diamictite interval that has been at the center of an ongoing debate in Appalachian Basin geology: are these diamictites

the product of entirely non-glacial debris flows, direct subglacial emplacement (tillites), or were they deposited by sediment-gravity flows with possible indirect glacial influence?

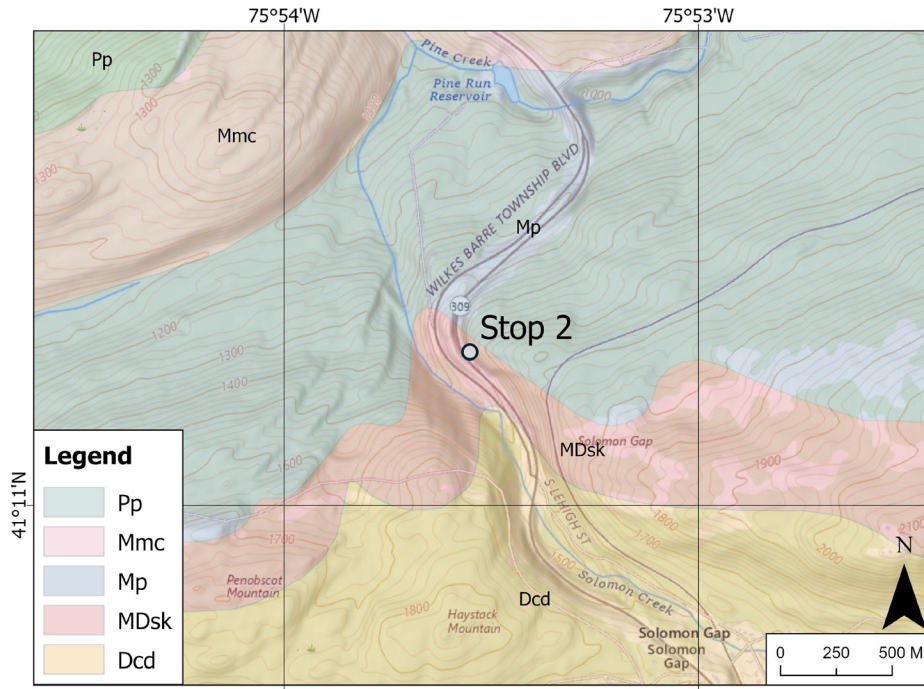


Figure 1. Map of Stop 2 – Solomon Gap outcrop location and relevant stratigraphic units. Circle is Stop 2 location. Dcd = Catskill Formation, Duncannon Member; MDsk = Spechtly Kopf Formation; Mp = Pocono Formation; Mmc = Mauch Chunk Formation; Pp = Pottsville Formation.

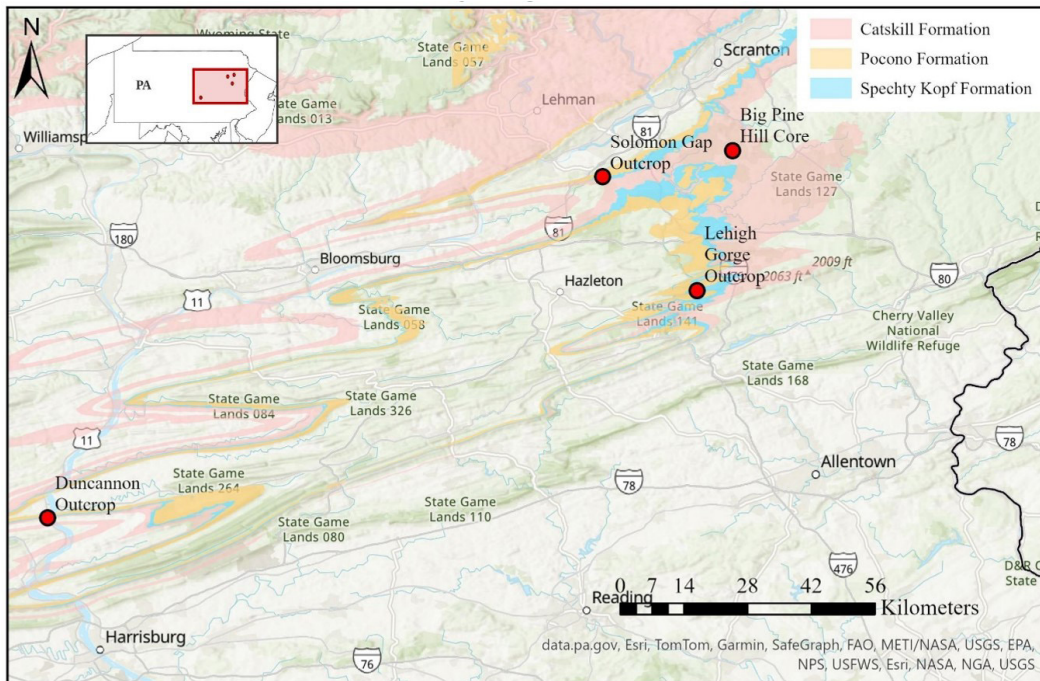


Figure 2. Map of broader study region and relevant stratigraphic units. Red circles are study localities. Fieldtrip stop is at the Solomon Gap outcrop. Other localities (Duncannon outcrop, Lehigh Gorge outcrop, Big Pine Hill core) are mentioned in this guidebook chapter.

This stop provides an opportunity to examine the sedimentary evidence bearing on this question firsthand. This includes the full range of lithofacies within the Spechty Kopf Formation, including the diamictite, associated soft-sediment deformation structures, and the fine-grained interval with isolated clasts (lonestones=non-genetic term). The overlying Pocono Formation is equally well-exposed and provides a clear contrast in depositional style to the underlying Spechty Kopf.

The Late Paleozoic Ice Age (LPIA; late Devonian–late Permian) began approximately 360 million years ago and is relevant context for evaluating the origin of these diamictite deposits. The LPIA was the longest icehouse of the Phanerozoic and is the only icehouse interval in Earth's history prior to the Cenozoic to have occurred on a biologically complex Earth with terrestrial flora and fauna (e.g. Montañez and Poulsen, 2013; Rosa and Isbell, 2021). The earliest, high-latitude evidence for LPIA glaciation is found in the late Devonian (Fammenian) in South America (Caputo, 1985; Rosa and Isbell, 2021). This Late Devonian cooling on the southern supercontinent of Gondwana is well-documented, but whether equivalent alpine glaciation extended to the subtropical highlands of Euramerica, including the Acadian Mountains of the Appalachian region, has been vigorously debated (**Figure 3**). The diamictites of the Spechty Kopf Formation, as seen at this outcrop, are among the most-discussed evidence for or against low paleolatitude LPIA glaciation, making them globally significant.

Geologic Setting

During the Late Devonian, the Appalachian Basin occupied a paleolatitude of approximately 30–35° S within the subtropical southern hemisphere (**Figure 3**; e.g. Scotese, 2002, Torsvik and Cocks, 2004). The basin formed as a peripheral foreland during the Acadian Orogeny, driven by collision of the Avalon terrane against the eastern margin of Euramerica (e.g. Ettensohn, 1985; Faill, 1985), producing the Acadian Mountains whose erosional products were shed westward as a series of clastic wedges throughout the Middle and Late Devonian. The largest of these is the Catskill Delta complex, a thick eastward-thickening prism of alluvial, fluvial, and deltaic sediments underlying much of central and northeastern Pennsylvania, represented in the uppermost Devonian by the Duncannon Member of the Catskill Formation (**Figure 4**; Sevon, 1985).

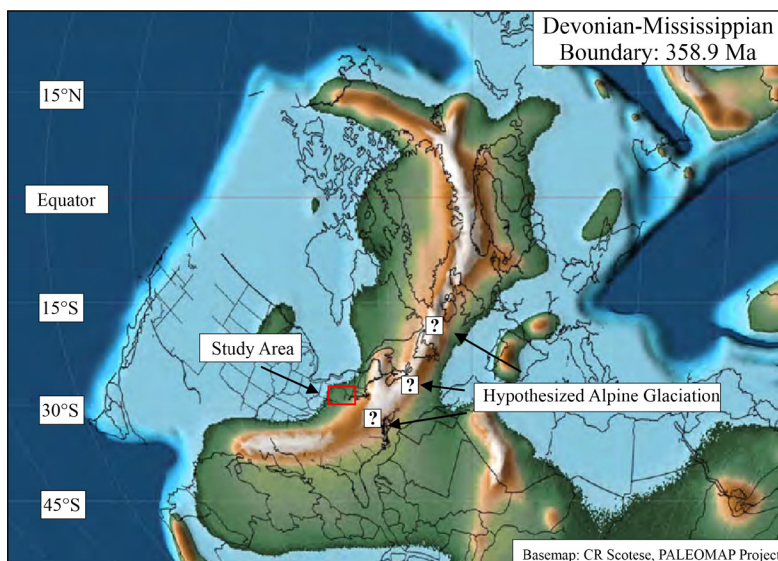


Figure 3. Paleogeographic setting of the study area (red box) at the Devonian-Mississippian boundary. Note: depicted alpine glaciation is disputed. Modified from Scotese (2014).

		Mississippian	Tournaisian	Spore Zonation	Eastern OH	Western PA	East-central WV	Western MD	South-central PA	Northeastern PA
					VI-CM?	VI-CM?	VI-CM?	VI-CM?	VI-CM?	VI-CM?
Devonian	Famennian	LN	LE	VH	Cuyahoga Fm.	Cuyahoga Gp.	Price Fm.	Purslane Fm.	Burgoon Fm.	Pocono Fm.
					Sunbury Sh.	Murrysville sand	Riddlesburg Sh.	Rockwell Fm.	Rockwell Fm.	
					Berea Ss	Cussewago Ss	Riddlesburg Sh.	Riddlesburg Sh.	Riddlesburg Sh.	
					Bedford Sh.		Rockwell Mbr.	<i>Diamictite interval</i>		Spechty Kopf Fm.
					Cleveland Sh. Mbr.	Oswayo Fm.		Oswayo Mbr.		
Devonian	VH	LN	LE	VH	Huron Sh. Mbr.	Venango Gp.	Hampshire Fm.	Hampshire Fm.	Catskill Fm.	Catskill Fm.
					Ohio Shale					

Figure 4. Regional lithostratigraphy, miospore zonations, and position of the Spechty Kopf Formation diamictite interval across the central Appalachian Basin. Modified from Brezinski et al. (2008) and Ettensohn et al. (2020a). VH — *Vallatisporites hystricosus* Zone; LE — *Retispora lepidophyta* – *Indotriradites explanatus* Zone; LN — *Retispora lepidophyta* – *Verrucosisporites nitidus* Zone; VI — *Vallatisporites verrucosus* – *Retusotriletes incohatus* Zone; VI-CM? — uncertain range from the *Vallatisporites verrucosus* – *Retusotriletes incohatus* Zone through the *Schopfites claviger* – *Auroraspora macra* Zone. Green shading indicates the stratigraphic position of the diamictite interval.

The Catskill Formation is unconformably overlain by the Spechty Kopf Formation (**Figure 4**), with the basal contact reflecting substantial incision; relief of up to 24 m has been measured over short lateral distances in some regions (Costolnick, 1987). The Spechty Kopf spans the latest Devonian (Famennian) into the earliest Mississippian (Tournaisian), with palynology placing it within the *R. lepidophyta*-*Indotriradites explanatus* (LE) and/or *R. lepidophyta*-*V. nitidus* (LN) miospore biozones (**Figure 4**; Richardson and McGregor, 1986; Sevon et al., 1997; Ettensohn et al., 2020a, 2020b, 2023). This interval coincides regionally with a significant regression recorded by the Berea Sandstone, followed by a major transgression depositing the Sunbury Shale and Riddlesburg Shale Member in basinal and nearshore settings respectively (**Figure 4**). It is globally correlative with the Hangenberg mass extinction event (Kaiser et al., 2016). The Spechty Kopf is a lithologically distinctive unit transitional between the underlying red-bed Catskill Formation and the overlying gray Pocono Formation, occupying the same lithostratigraphic position as the Huntley Mountain Formation of north-central Pennsylvania, the Rockwell Formation of south-central Pennsylvania and Maryland, and the Rockwell Member of the Price Formation in West Virginia (**Figure 4**; Berg et al., 1980). It is dominated in northeastern Pennsylvania by a vertically ordered succession of diamictite, pebbly mudstone, rhythmites/laminites, and sandstone, and is unconformably overlain by the multi-storied channel sandstones and quartz pebble conglomerates of the Pocono Formation.

Debate Over Origin of the Appalachian Basin Diamictites

The diamictite interval present within the Spechty Kopf has been traced across an outcrop belt (spanning lithostratigraphic equivalent units) approximately 400 km long and 40 km wide from northeastern Pennsylvania into Maryland and West Virginia (Brezinski et al., 2008). The diamictite-bearing succession has been interpreted at various times as the product of subaqueous, non-glacial debris flows (e.g. Sevon, 1969, Bjerstedt and Kammer, 1988; Suter, 1991; Sevon and Woodrow, 2009), resedimented alpine glacial deposits (Sevon, 1973), tsunami deposits following a Late Devonian bolide impact (Woodrow et al., 1990; Sevon et al., 1997; Woodrow and Richardson, 2006), or direct glacial emplacement by alpine glaciers advancing from the Acadian Highlands (e.g. Cecil et al., 2002, 2004; Dennis, 2005a, 2005b, 2005c, 2007; Brezinski et al., 2008, 2009, 2010; Brezinski and Cecil, 2015; Ettensohn et al., 2020a, 2020b, 2023). Proponents of a non-glacial debris flow origin cite dewatering structures, diapiric features, and flow rolls as evidence of subaqueous emplacement (e.g. Bjerstedt and Kammer, 1988; Suter, 1991; Sevon et al., 1997). In northeastern Pennsylvania, the absence of marine fauna or trace fossils led Sevon et al. (1997) to propose deposition in a freshwater lake eroded

into the Catskill alluvial plain, with the diamictite representing an initial debris flow, the overlying pebbly mudstone its waning phase, and the rhythmically bedded laminite a prodelta turbidite deposit. He suggested that flexural uplift of a peripheral forebulge, generated north of Scranton in response to Acadian orogenic loading, obstructed fluvial drainage across the Catskill alluvial plain and thereby impounded the water that formed the lakes. Sevon et al. (1997) further argued that palynological data place the sequence along its entire outcrop belt within the same narrow miospore subzone, implying near-simultaneous, regionally extensive deposition requiring a single triggering event, and proposed seismicity related to the Charlevoix impact crater (Quebec; ~46 km diameter, 360 ± 25 Ma) as a plausible mechanism.

Alternatively, a collection of authors interpret the diamictites as lodgement tills directly emplaced by alpine glaciers (e.g. Brezinski et al., 2008, 2009, 2010; Brezinski and Cecil, 2015; Ettensohn et al., 2020a, 2020b, 2023), noting their temporal coincidence with the earliest pulse of LPIA glaciation on Gondwana, widespread distribution, and narrow time window of deposition (Caputo, 1985). Brezinski et al. (2008, 2010) cite striated/faceted clasts, overlying rhythmites with lonestones interpreted as varves with ice-rafted dropstones, and soft-sediment deformation attributed to overriding ice. Most recently, Ettensohn et al. (2020a, 2020b, 2023) report multiple several-ton granitic boulders of Appalachian origin embedded in situ within the age-equivalent Cleveland Member of the Ohio Shale in eastern Kentucky, nearly 500 km west of the inferred glacial source. They interpret these as iceberg rafted dropstones. In the correlative Rockwell Member of the Price Formation in West Virginia, they describe a similar three-part succession with lonestones and diamictite, which they interpret as mudstones with dropstones deposited before glacial advance, diamictites with shear fabrics they viewed as a lodgement till, and glaciolacustrine or glaciomarine sediments with dropstones deposited after the ice retreated.

Description of Units at the Solomon Gap Outcrop

Spechty Kopf Formation (~0–45 m)

F1 — Soft-Sediment Deformation Facies

Fine- to medium-grained sublithic sandstone blocks (**Figure 5**) ranging from less than a meter to tens of meters in maximum dimension are enclosed within a finer-grained matrix of mudrock to fine sandstone. Two block types are recognized: semi-coherent blocks retaining some original bedding, and strongly deformed blocks exhibiting disrupted internal fabric. Ball and pillow structures can also be observed within this facies, indicating rapid loading and a water-saturated substrate (**Figures 5a and 6a**). Some sandstone blocks display semi-brittle extensional deformation expressed as boudinage, in which competent sandstone beds are thinned and disaggregated into disjointed segments within the surrounding matrix, reflecting competence contrast between the sandstone and enclosing sediment at the time of deformation (**Figures 5b and 6b**). Matrix bedding is deflected around sandstone blocks, with beds rotated progressively toward vertical along block margins (**Figures 5a and 6b**). Where blocks are closely spaced, elevated pore-fluid pressures within the water-saturated matrix drove upward injection of fine-grained material between them, producing flame structures and diapirs that record rapid fluid escape during or immediately following emplacement. Several isolated recumbent fold noses with broadly consistent axial orientations are present, and rafted beds of finely laminated sandstone can be observed. In some horizons, convolute bedding passes laterally into sandy diamictite (**Figure 6a**). This facies is interpreted as subaqueous mass transport deposits (MTDs) representing a continuum of gravitational failure processes from coherent sliding through slumping (e.g. Shanmugam et al., 1994; Talling et al., 2012). The coexistence of coherent blocks with preserved bedding alongside strongly boudinaged and folded blocks reflects variable strain intensity across the deposit, consistent with heterogeneous deformation typical of subaqueous MTD complexes. The fold axis orientations suggest a NNW transport direction in agreement with regional paleocurrent indicators, and the lateral transition from convolute bedding into sandy diamictite records the threshold between plastic deformation and complete disaggregation within a single depositional event.

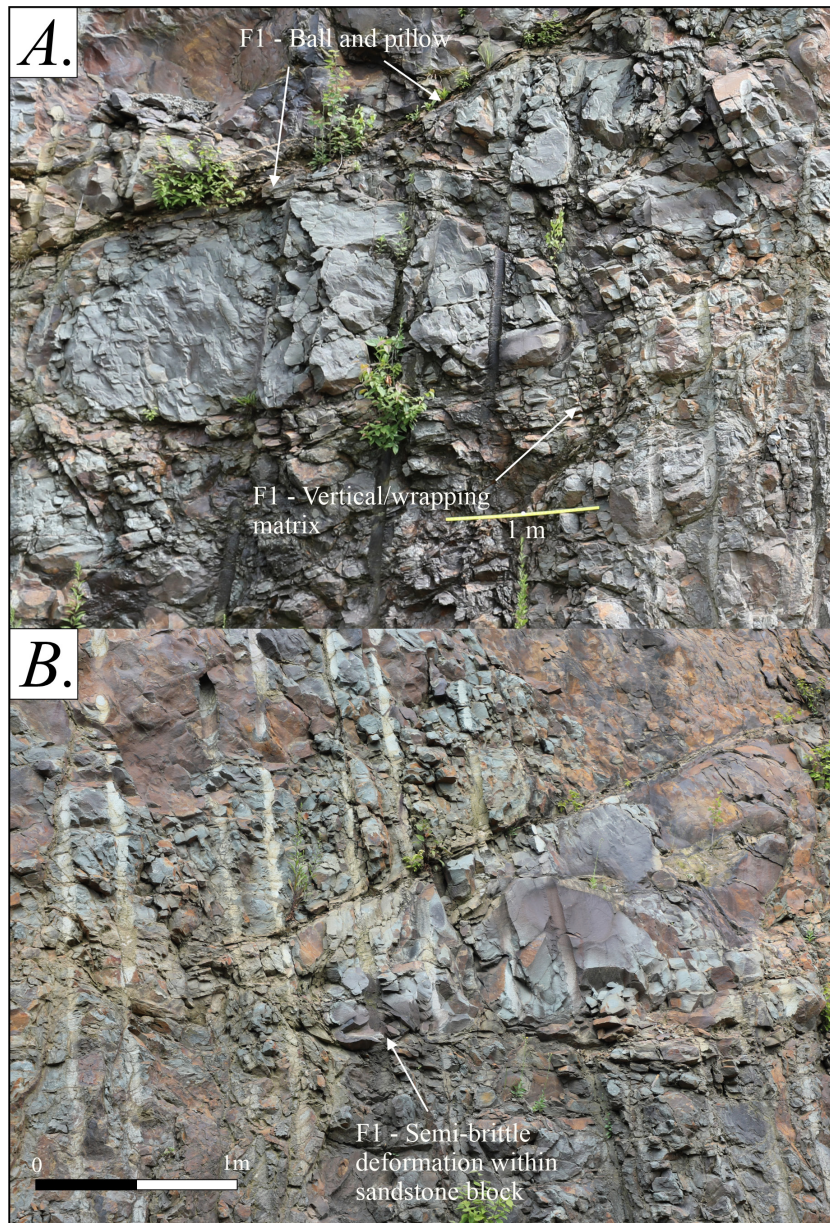


Figure 5. F1-Soft sediment deformation facies. A.) Ball/Pillow structures from rapid deposition on water-saturated substrate. B.) Sandstone slide block with semi-brittle deformation on edge.

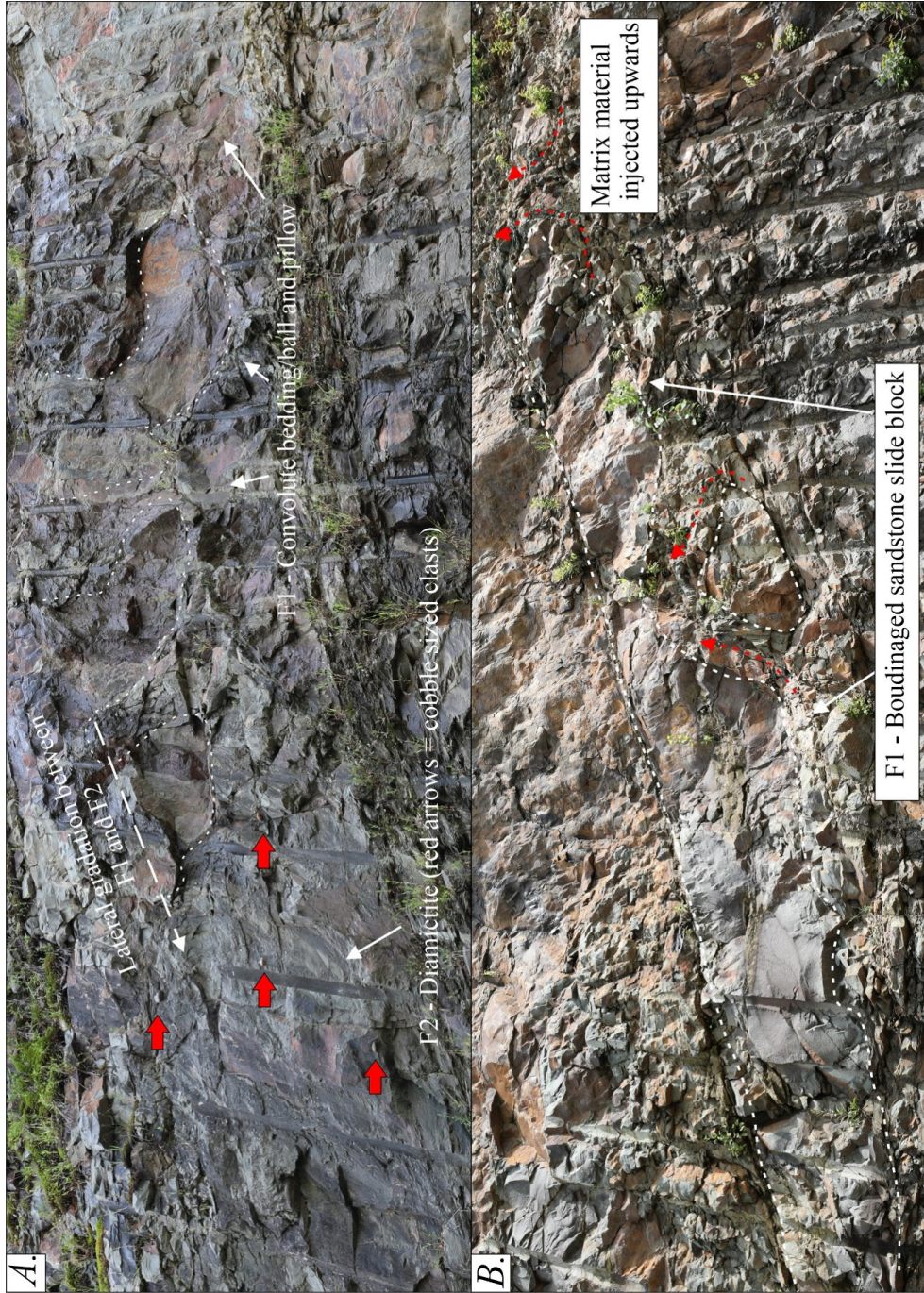


Figure 6. F1-Soft sediment deformation facies. A.) Lateral gradation of F1 convolute bedding/ball and pillow structures into the F2 - Sandy diamictite facies. Red arrows are location of cobbles sized clasts. Dotted white lines around ball and pillow structures. B.) Boudinaged slide blocks of sandstone (circled by white dotted lines) with vertically injected matrix material between/around blocks (shown with red dashed arrows).

F2 — Sandy Diamictite Facies

The F2 facies is comprised of three discrete diamictite packages 1–5 m thick consisting of massive to poorly stratified, clast-rich sandy diamictite (5–40% clasts; Hambrey and Glasser, 2003). Individual packages thicken and pinch out laterally, grading into F1 or F3 through a progressive decrease in clast content without a sharp bounding surface. Clasts range from granule to cobble size and are dominantly comprised of rounded to well-rounded (some angular) quartz and sandstone (**Figure 7a**), but an exotic population of metamorphic and igneous lithologies not present in the underlying Catskill Formation is also present. Rafted and folded mudrock or sandstone beds from F3 occur locally within the diamictite (**Figures 7b-d**). No striated or faceted clast surfaces were identified despite systematic examination. The matrix is a massive to faintly-structured gray sandy mudrock texturally continuous with the enclosing F1 matrix. Fresh core exposure at the nearby Big Pine Hill locality reveals a red matrix color similar to the underlying Catskill Formation, suggesting significant incorporation of reworked Catskill material. This facies is interpreted as cohesive debris flows representing the fully disaggregated end-member of the MTD continuum established in F1, in which clasts were carried in suspension within a mud-rich matrix with sufficient yield strength to inhibit clast settling and fabric development. The deformed, rafted beds and lateral gradation into F1 without sharp bounding surfaces indicates that debris flows and sliding-slumping were contemporaneous processes. The rounded exotic clast population requires derivation from lithologies stratigraphically or geographically removed from the local succession, consistent with fluvial transport from upslope basement sources, though glacial erosion followed by secondary fluvial reworking cannot be excluded.

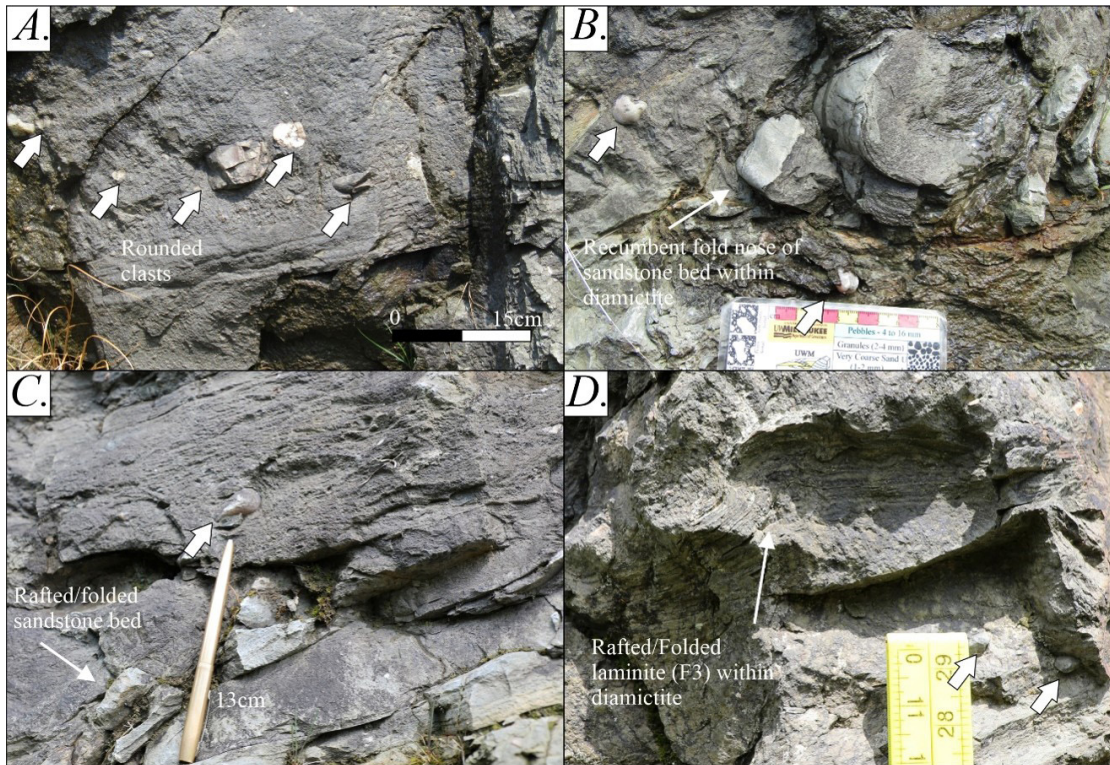


Figure 7. F2 Sandy Diamictite Facies. White arrows indicate clasts. A.) Rounded clasts within the diamictite. B.) Recumbent fold nose of sandstone bed within diamictite. C-D.) Rafted/folded beds within the diamictite.

F3 — Laminite and Lonestone Facies

Packages 2–4 m thick of F3 always occur in association with F1 and/or F2, sometimes with lateral gradational contacts (**Figure 8a**). Laminated mudrocks and thinly bedded, fine sandstones (1–2 cm thick) contain isolated pebble-sized clasts that depress and deform underlying laminae without piercing/disrupting multiple sets and are draped conformably by overlying laminae (**Figure 8c,d**). Current ripple and climbing ripple cross-laminae indicate episodic waning flows with NNW paleocurrent directions (**Figure 8b,d**). Rare plant fragments occur within the fine-grained matrix. The uppermost Spechty Kopf defines an upward-coarsening package in which laminites pass into thicker sandstone beds (~25–50 cm) with load casts at their bases.

This facies records suspension settling and weak tractive bottom flows in a low-energy subaqueous environment, interpreted as the dilute distal expression of the same mass transport events as F1 and F2 (e.g. Talling et al., 2012). The thin traction beds are interpreted as small turbidites generated by the turbulent, low-density flow front of the parent debris flows outrunning the main MTD body. Isolated clasts are interpreted as outrunner clasts decoupled from decelerating debris flow fronts (e.g. Postma et al., 1988) rather than ice-rafted dropstones, as they do not pierce multiple laminae sets like dropstones. The occurrence of plant fragments is inconsistent with an ice-contact proglacial setting. The upward-coarsening sandstone package at the top of the Spechty Kopf Formation may record a drop in base level and the final progradation of the subaqueous fan delta front prior to the erosional truncation and onset of subaerial braided fluvial deposition recorded by F4.

Summary: The Spechty Kopf at this outcrop is interpreted as the subaqueous front of a fan delta system, dominated by gravity-driven mass transport of coarse sediment into a standing body of freshwater (**Figure 9a,b**; e.g. Ethridge and Wescott, 1984; Postma, 1984; Shanmugam et al., 1994; Wang et al., 2025).

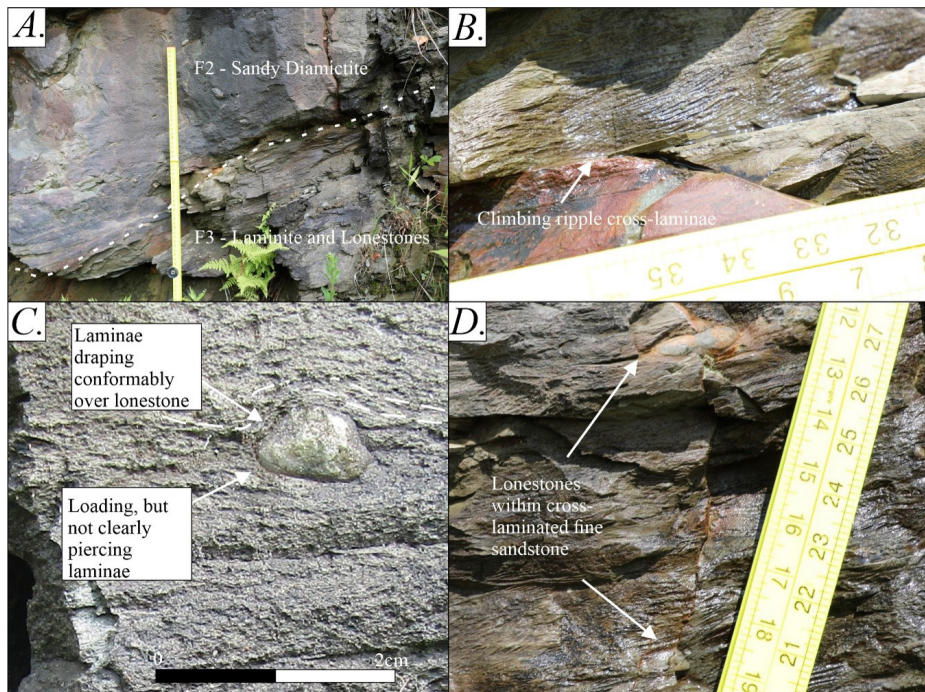
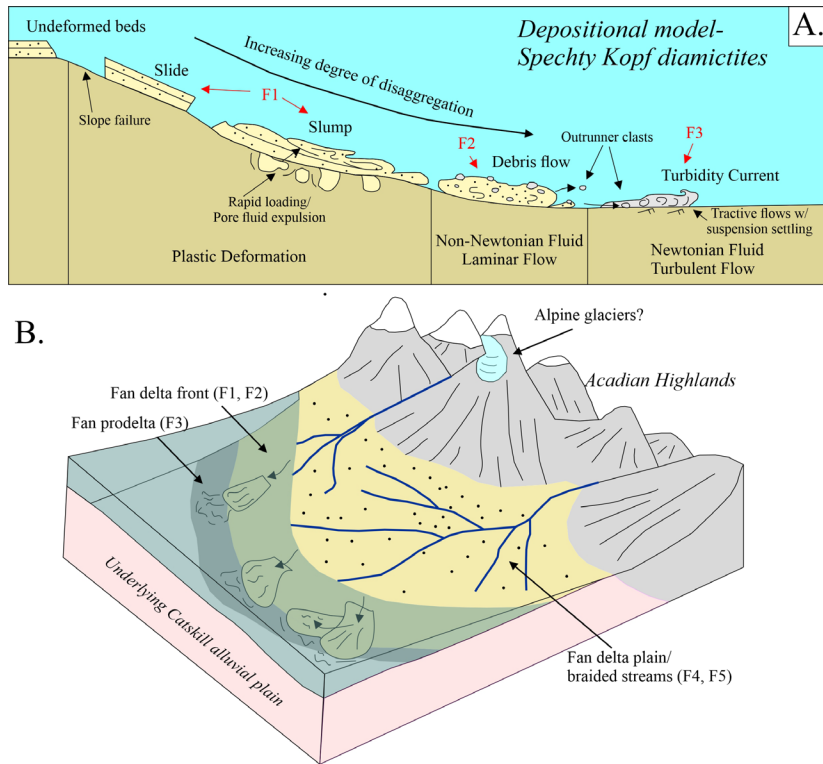


Figure 8. F3 – Laminite and Lonestone facies. A.) White dashed line shows gradational contact between F3 and F2 Sandy Diamictite Facies. B.) Thin climbing ripple cross-laminated sandstone beds within F3. C.) Lonestone that does not pierce the underlying laminae and shows conformably draping laminae, indicating it was not dropped into the substrate from above. D.) Lonestones resting within cross-laminated fine sandstone.

Summary: The Spechty Kopf at this outcrop is interpreted as the subaqueous front of a fan delta system, dominated by gravity-driven mass transport of coarse sediment into a standing body of freshwater (**Figure 8**; e.g. Ethridge and Wescott, 1984; Postma, 1984; Shanmugam et al., 1994; Wang et al., 2025).

Figure 9. Depositional model for the Spechty Kopf Formation diamictites and associated facies. A.) Gravity-driven mass transport of coarse sediment into a standing body of water can transform coherent slides into slumps, debris flows, and turbidity currents with progressive disaggregation and increased fluid mixing during downslope translation. Coherent slide blocks (F1) grade into sandy diamictite (F2) as internal shear increases, while the dilute turbulent flow front separates from the main debris flow body to generate the small turbidites recorded by the laminite and lonestone facies (F3). Lonestones within F3 represent outrunner clasts decoupled from decelerating debris flow fronts rather than ice-rafted dropstones. Modified from Shanmugam et al. (1994). B.) Simplified model of typical fan delta facies association with location of described facies (F1-F5). Modified from Zhao et al. (2018).



Pocono Formation (~45–150 m)

F4 — Conglomerate Facies

Interbedded trough and planar cross-bedded quartz pebble conglomerate and medium lithic sandstone in normally graded beds (10–50 cm) rest sharply on the Spechty Kopf along an erosional channelized contact (**Figure 10a,b**). This facies is interpreted as a pulsing, high-energy gravel-bearing braided fluvial system representing a base level fall and progradation of the Pocono clastic wedge (**Figure 9b**). Rounded quartz pebbles indicate long-distance fluvial transport from metamorphic and igneous sources in the Acadian highlands.

F5 — Multistoried Channelform Sandstone Facies

Multistoried channelform sandstone bodies 1–5 m thick and tens of meters in lateral extent are defined by erosional basal contacts with up to 3–5 m of relief, above which basal lags of rip-up clasts or conglomerate are commonly developed (**Figure 10c,d**). Trough cross-stratification, upper plane beds, and lateral accretionary surfaces are the dominant internal structures (**Figure 10d,e**). Individual bodies are occasionally separated by discontinuous mudrock drapes less than 10 cm thick (**Figure 10f**). Concretions can be seen in the uppermost beds of the section. This facies is interpreted as a sandy braided channel belt in which repeated cycles of channel incision, dune migration, bar accretion, and avulsion reflect a sediment supply-dominated system with high avulsion frequency (**Figure 9b**; e.g. Miall, 1977). Thin discontinuous mudrock drapes record the low preservation potential of overbank fines in a laterally mobile system where repeated flood reworking and channel migration prevented significant floodplain accumulation. The overall multistoried amalgamated architecture records a system in which accommodation space was continuously consumed by lateral channel migration rather than vertical aggradation.

Summary: The transition to the Pocono Formation records a shift to high-energy, subaerial, braided fluvial conditions, marked by an erosional contact cutting into the uppermost Spechty Kopf beds (Figures 9b, 11).

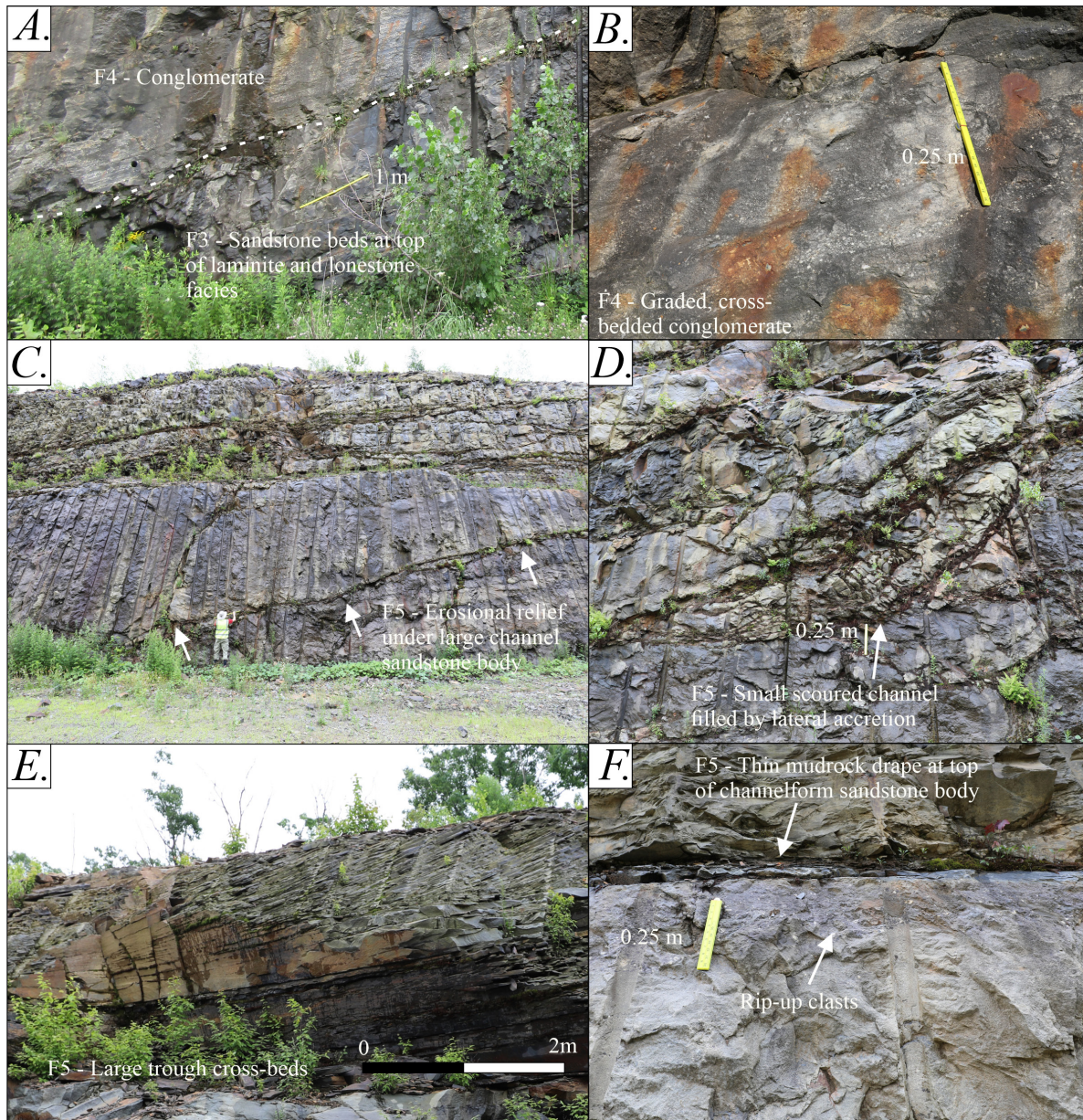


Figure 10. F4-Conglomerate Facies and F5-Multistoried Channelform Sandstone Facies. A.) Erosional contact (white dashed line) between F3 sandstone beds of the uppermost Spechty Kopf Fm. and quartz pebble conglomerate (F4) of the Pocono Fm. B.) Normally graded and cross-bedded quartz pebble conglomerate of F4. C.) Erosional relief under large channel sandstone of F5. D.) Small scour and fill channelform sandstone body in F5 showing fill by lateral accretion. E.) Large trough cross-beds and bar forms in F5 at the top of the measured section. F.) Thin mudrock drape at top of F5 channelform sandstone body. Rip-up clasts can be seen in the underlying sandstone.

Solomon Gap Outcrop

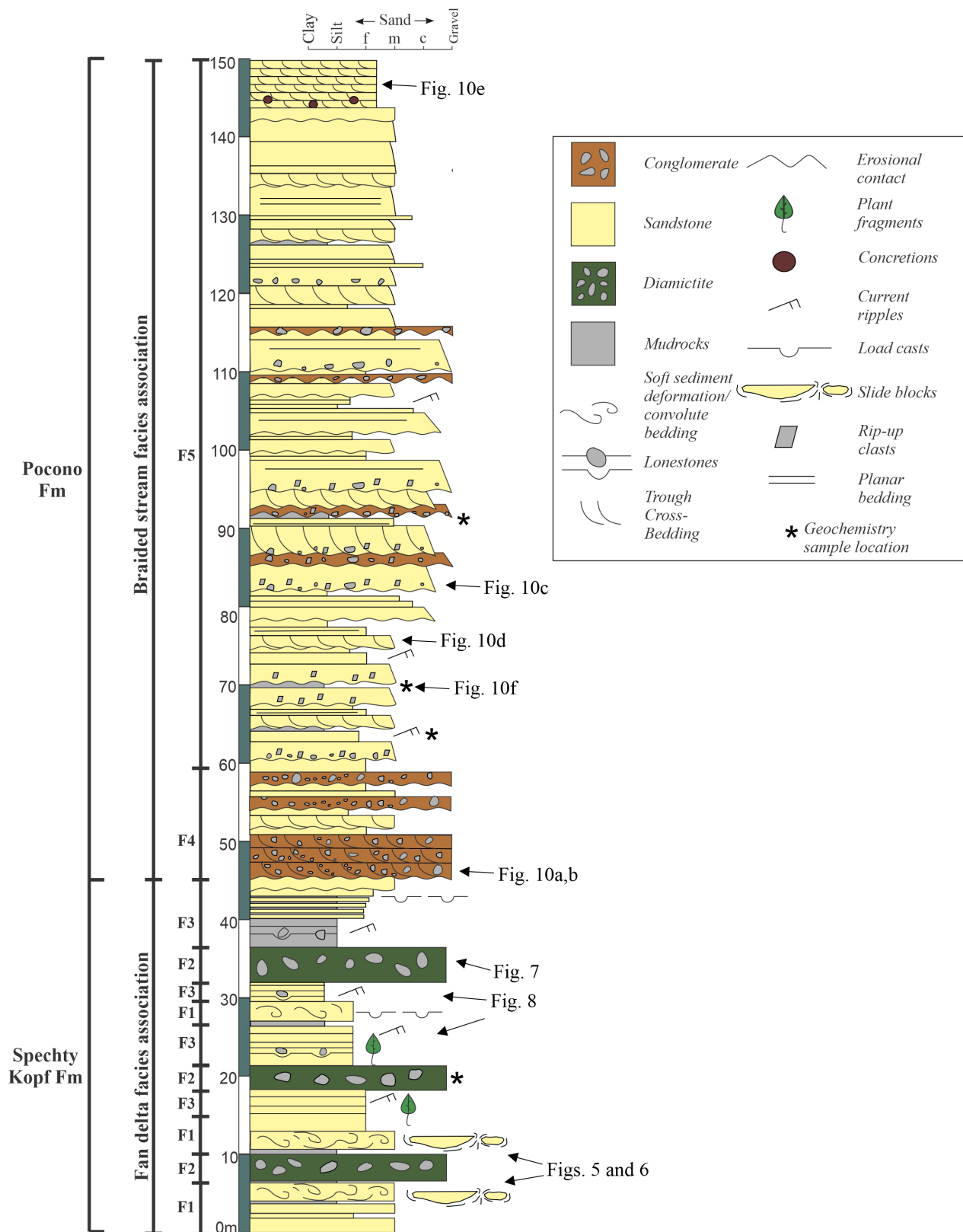


Figure 11. Stratigraphic column of measured section at the Solomon Gap outcrop.

Geochemical Data

Mudrock and diamictite matrix samples from the Catskill, Spechty Kopf, and Pocono formations were collected at Solomon Gap and three additional localities (Duncannon, Lehigh Gorge, Big Pine Hill core – see **Figure 2** map) and analyzed by wavelength dispersive X-ray fluorescence spectrometry for major element oxides and trace element concentrations.

Paleoweathering Analysis: These data were used to calculate the Chemical Index of Alteration (CIA; Nesbitt and Young, 1982; Wang et al., 2020), a proxy for paleochemical weathering intensity (**Figure 12**). The CIA is defined as:

$$\text{CIA} = [\text{Al}_2\text{O}_3 / (\text{Al}_2\text{O}_3 + \text{CaO}^* + \text{Na}_2\text{O} + \text{K}_2\text{O})] \times 100$$

where CaO* represents calcium in silicate phases only. In a dry, glacially dominated landscape with minimal chemical weathering, CIA values typically range from approximately 50–70. CIA values above 80 typically indicate humid conditions with significant chemical weathering of feldspars and consequent enrichment in aluminous clay minerals (Nesbitt and Young, 1982). For context, fresh granites typically yield values of 45–55, while strongly weathered kaolinite approaches values of 100. The CIA values from the Duncannon outcrop consistently plot highest into the typical humid range suggesting more intense paleoweathering at that location or further transport distance of sediments from the source (**Figure 12**). Big Pine Hill Core records the lowest mean CIA values across all formations (**Figure 12**; Spechty Kopf: 76.9; Catskill/Pocono: 76.7). Most samples fall in the typical seasonal/temperate range and no samples from any locality or formation approach the typical arid/glacial range (CIA < 70), and the overall consistency of CIA values across localities and formations suggests broadly similar paleoweathering conditions during deposition of the Spechty Kopf, Catskill, and Pocono Formations (**Figure 12**).

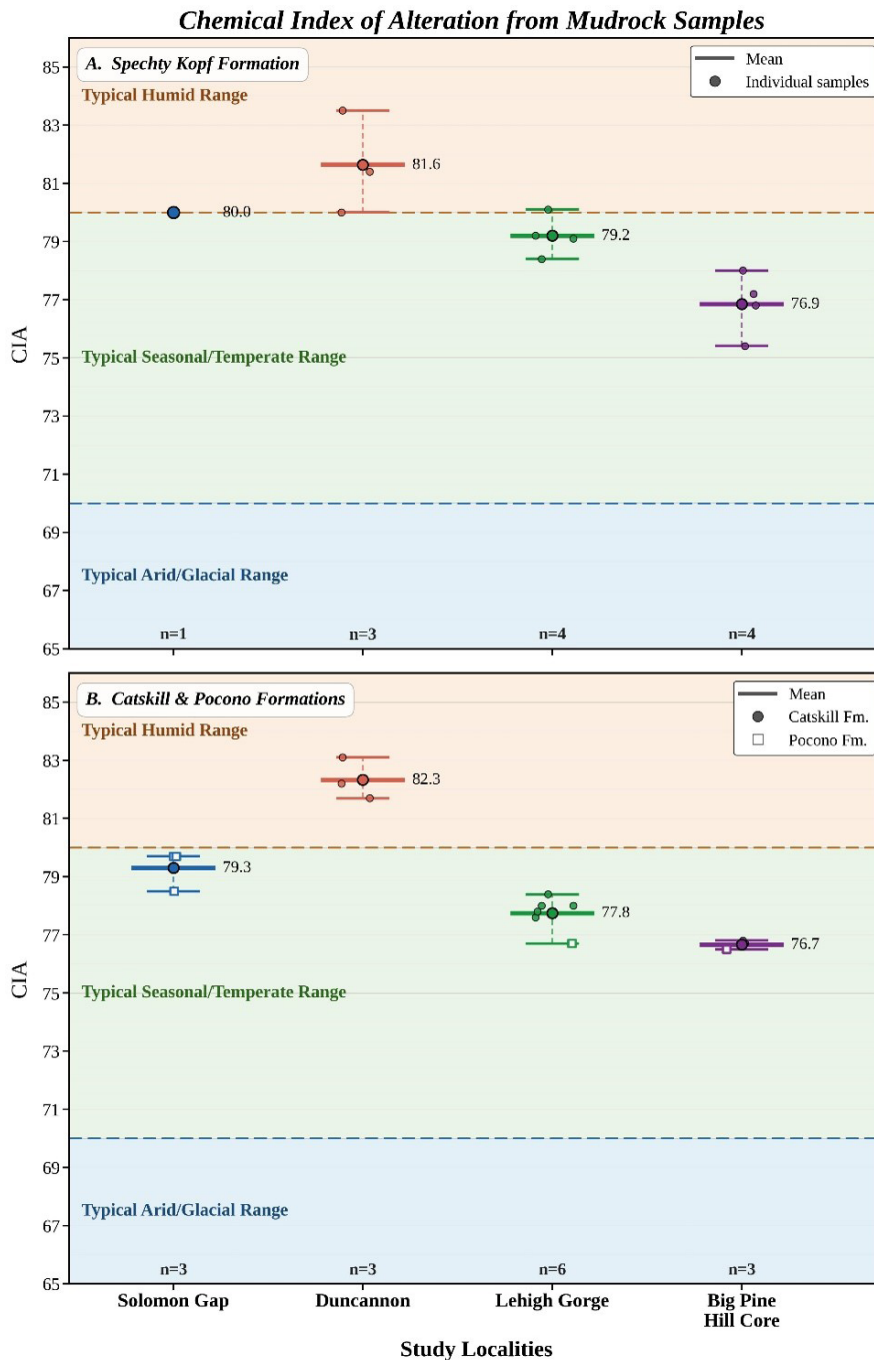


Figure 12. Paleoweathering (CIA - chemical index of alteration) data from mudrocks/diamicite matrix from the Solomon Gap Outcrop plus additional study locations at Duncannon, Lehigh Gorge, and a core from Big Pine Hill (see Figure 2 map). Red shaded area represents values typical of humid paleoclimates, green shaded area represents typical seasonal or temperate values, and the blue shaded area represents typical arid/glacial paleoclimates. A.) Spechty Kopf Fm. values. B.) Catskill and Pocono Fm. values. Threshold values from Nesbitt and Young (1982).

Paleosalinity Analysis: Sr/Ba ratios from all four localities plot well below the freshwater threshold of Wei and Algeo (2020), with Sr values of approximately 35–100 ppm and Ba values of approximately 300–950 ppm, supporting deposition of the Spechty Kopf under continental freshwater conditions with no marine influence at any locality (**Figure 13**).

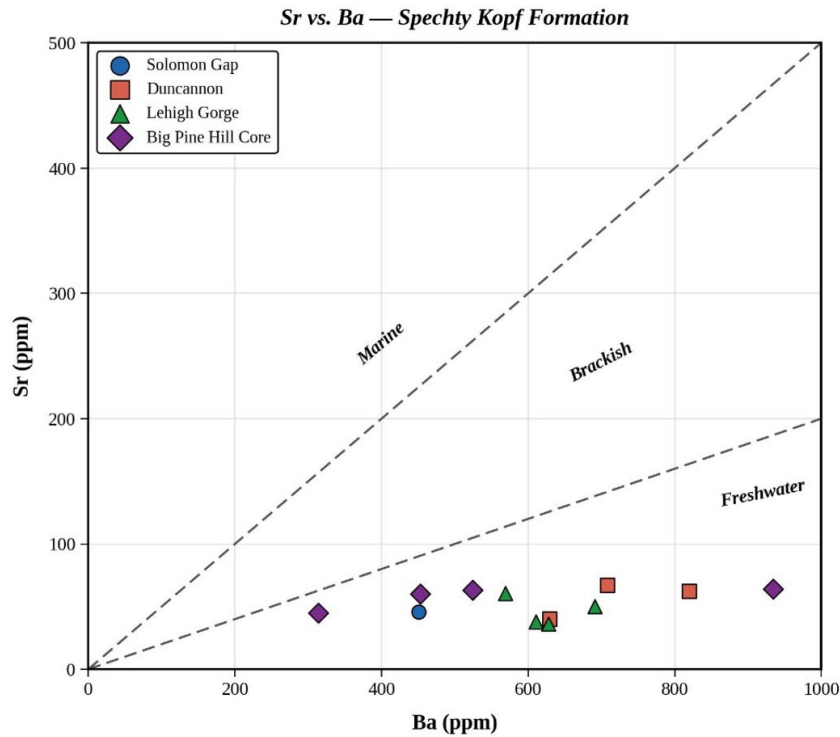


Figure 13. Sr/Ba values of mudrocks and diamictite matrix from the Spechty Kopf Fm. at Solomon Gap, Duncannon, Lehigh Gorge, and the Big Pine Hill Core. Threshold values for freshwater, brackish, or marine paleoenvironments established by Wei and Algeo (2020).

Discussion: Glacial vs. Non-Glacial Origin

The facies assemblage of the Spechty Kopf Formation at Solomon Gap is consistent with a fan delta facies model, in which coarse sediments were delivered down a depositional slope through mass transport processes/sediment gravity flows into a lacustrine environment. This interpretation is broadly in agreement with that of Sevon et al. (1997) and is supported by the following lines of evidence:

1. **Evidence for subaqueous deposition in a standing body of water:** The laminite facies (F3) records suspension settling and weak tractive bottom currents diagnostic of a low-energy subaqueous environment, mass transport deposits of F1 pass laterally and gradationally into F3 without sharp bounding surfaces, and load casts, ball and pillow structures, flame structures, and diapirs throughout both facies indicate rapid deposition into a water-saturated substrate, together indicating that F1–F3 are co-genetic products of a single subaqueous depositional system.
2. **Lonestone geometry inconsistent with ice-rafted dropstones:** Isolated clasts in F3 are found within cross-laminated strata, depress underlying laminae without piercing them and are draped conformably, consistent with outrunner clasts from a decelerating debris flow (Postma et al., 1988) rather than ice-rafted dropstones impacting the substrate.
3. **Rounded clast population and absence of striated or faceted clasts:** Clasts within F2 are dominantly rounded to sub-rounded, and no striated or faceted surfaces were identified despite systematic examination, in contrast to localities where such clasts have been reported (Brezinski et al., 2008).
4. **Absence of subglacial deforming bed:** Lodgement tills often show a vertical increase in deformation, a pervasive clast fabric parallel to ice flow, ductile shear zones, recumbent folds with consistent vergence, and a sharp erosional basal contact, often with striae or grooves (e.g. Busfield and Le Heron, 2013; Fedorchuk et al., 2019). None of these are observed at Solomon Gap, where

diamictite bases are irregular and gradational into F1 or F3.

5. **No CIA decline across the Spechty Kopf:** CIA values cluster between 70 and 83 across all three formations at all four localities, with no drop within the Spechty Kopf. Even if the diamictite largely comprises reworked Catskill sediment, dilution by fresh glacial flour would be expected to drive values down. The absence of this signal argues against direct glaciogenic sediment input.
6. **Freshwater geochemistry supports lacustrine deposition:** Sr/Ba ratios from all four localities plot well below the freshwater threshold of Wei and Algeo (2020), directly supporting the lacustrine model of Sevon et al. (1997) in which debris flows were deposited into freshwater lakes that formed in the deeply eroded Catskill alluvial plain.

The triggering mechanism for slope failure may have been some combination of rapid sediment loading on oversteepened fan delta slopes, seismicity related to Acadian/Neoacadian tectonic activity, or the destabilization of glacially derived sediment during advance or retreat of alpine glaciers in the Acadian Highlands. The evidence above argues against direct glacial emplacement (lodgement till) of the Spechty Kopf diamictites but does not necessarily preclude an indirect glacial origin or possible glaciation in other parts of the basin. The Acadian Mountains were likely elevated to 2,500–3,500 m, likely sufficient for alpine glaciation at 30–35° S (e.g. Hastenrath, 2009; Ettensohn et al., 2023), and several observations support the glacial hypothesis:

1. **Abrupt coarse clastic pulse:** The Spechty Kopf records a sudden regional influx of coarse sediment whose temporal coincidence with the earliest LPIA glaciation on Gondwana (e.g. Caputo, 1985) raises the possibility of glacially enhanced erosion in the uplands, although this influx may also be the product of tectonic uplift.
2. **In situ boulders in the coeval Ohio Shale:** Several-ton granitic boulders of Appalachian origin occur in situ in the age-equivalent Cleveland Member of the Ohio Shale in eastern Kentucky (Ettensohn et al., 2020a, 2020b). The occurrence of these boulders in a marine shale is difficult to explain without ice-rafting. Glaciers would have had to reach sea level in some areas.
3. **Exotic clast compositions:** Many diamictite clasts represent lithologies absent from the Catskill Formation and overlying Paleozoic units (Sevon et al., 1997), consistent with either deep fluvial incision into Acadian basement or glacial erosion followed by meltwater reworking.
4. **Striated and faceted clasts in correlative units:** Brezinski et al. (2008) document striated and faceted clasts within the diamictites of the correlative Rockwell Formation in Maryland and West Virginia, providing possible evidence of subglacial abrasion at those localities. Their absence at Solomon Gap may reflect local variability in transport history rather than a fundamental difference in sediment source.

Conclusions

The evidence at this stop argues against direct glacial emplacement of the Spechty Kopf diamictites (lodgement till hypothesis). If glaciers were present in the Acadian Mountains, the most parsimonious interpretation is one in which glaciers were confined to high-elevation settings in most areas and contributed sediment to the foreland indirectly through meltwater and fluvial reworking rather than extending as sheet ice to low elevations for any extended period. At the same time, renewed tectonic activity in the Acadian source area provides a viable non-glacial explanation for the abrupt coarse clastic pulse that defines the base of the Spechty Kopf, and distinguishing between these two possibilities or determining the degree to which both contributed will require broader work across the full outcrop belt, as well as more precise geochronological constraints on the timing of sediment delivery.

Acknowledgements

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Stop 3A: Seven Tubs Nature Area - Origin of the Seven Tubs in Wheelbarrow Run and Stream Gauging Study

(41.235596, -75.810406) Seven Tubs Recreation Area



The Seven Tubs will be observed and theories for their origin will be discussed. A stream gauging station and data will be introduced and shown as a means to understand flood frequency relationships.



Stop 3B: Seven Tubs Nature Area - Exposure Dating of Pocono Formation Bedrock Outcrops

(41.235596, -75.810406) Seven Tubs Recreation Area

Pocono Formation outcrops will be visited. New 10-Be exposure ages that constrain the timing of deglaciation will be presented.

Stop 3 – Seven Tubs Recreation Area

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This location includes two separate stops that will be visited by the entire group at the same time. A short hike to the field stops involves steep and uneven terrain along the valleys of Wheelbarrow Run and Laurel Run, and the hillslope of the Bear Creek (or Penobscot) Mountain.

Introduction

The Seven Tubs Recreation Area is a 121-acre site located in Plains Township, Luzerne County (**Figure 1**). The Seven Tubs is part of and managed by the Pinchot State Forest. The centerpiece and most-visited attraction at the recreation area is a series of bedrock whirlpools developed on Wheelbarrow Run, a small tributary of Laurel Run. The site was last visited at the 1997 Field Conference where several bedrock exposures were examined and ideas relating to the origin of the Seven Tubs were discussed. The field stops this year are related to 2 ongoing field-based projects at the Seven Tubs including 1) stream-gauging of Laurel Run to establish a baseline hydrologic dataset for the stream and 2) cosmogenic exposure dating of glacially polished bedrock outcrops to determine the time of Laurentide Ice Sheet retreat from the landscape.

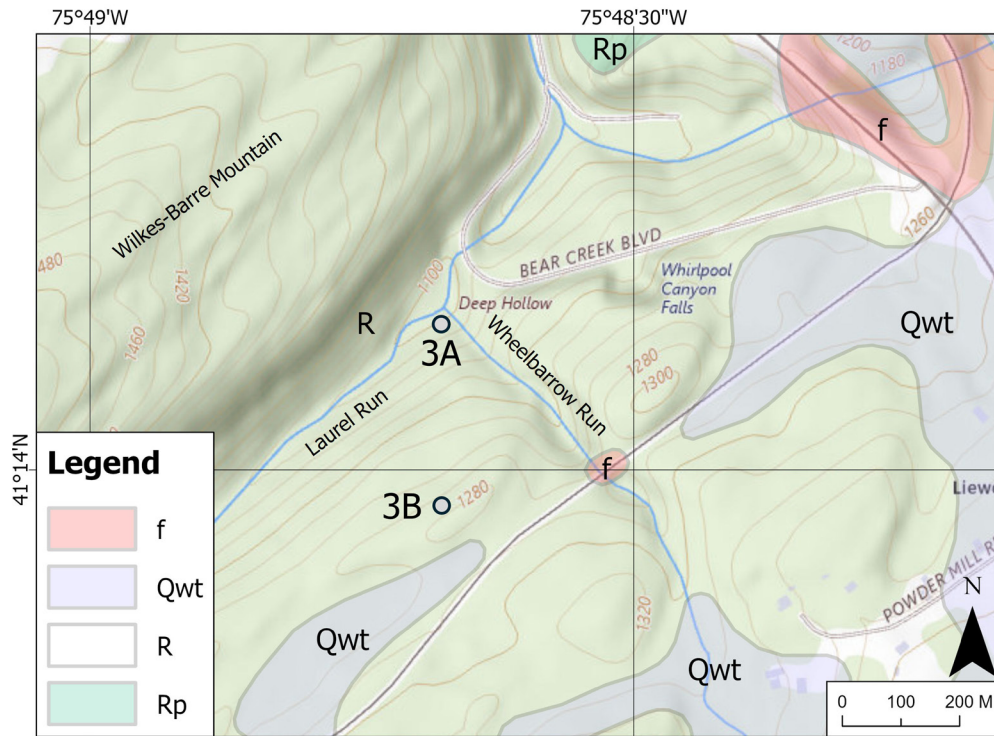


Figure 1. Surficial geology map of the Seven Tubs Recreation Area (Stop 3) showing the location of sites visited including: 3A) the lowermost pothole along Wheelbarrow Run at the confluence with Laurel Run and 3B) glacially abraded outcrops of the Pocono Formation sandstones located southwest of Wheelbarrow Run. Surficial geology units include bedrock (R), artificial fill (f), Wisconsinan age till (Qwt), and sandstone and/or shale pit (Rp) (Braun, 2008).

Physiography and Geology of the Seven Tubs Recreation Area

The Seven Tubs Recreation Area is located along the eastern hillslope of the Wyoming-Lackawanna Valley. Laurel Run is the primary stream contained in the recreation area that flows in a northeast-southwest trending valley between the Bear Creek Mountain (to the east) and the Wilkes-Barre Mountain (to the west). Several small tributaries draining off the hillslope of the Bear Creek Mountain flow into Laurel Run, including Wheelbarrow Run that contains the Seven Tubs potholes (**Figure 2**). Below the confluence of Wheelbarrow Run, Laurel Run flows to the northwest through a water gap cut through the Wilkes-Barre Mountain and then flows into the Wyoming-Lackawanna valley emptying into Mill Creek and eventually the Susquehanna River.

Bedrock geology in the recreation area consists of the Mississippian age Pocono Formation and Mauch Chunk Formation, and the Pennsylvanian age Pottsville Formation and the lowermost Llewellyn Formation (Berg et al., 1980; Sevon, 1976). The Pocono Formation is composed of light- to olive-gray, fine- to medium-grained, cross-bedded sandstone, siltstone, and conglomerate (Geyer & Wilshusen, 1982). It has some subordinate dark shale, siltstone, and very thin coaly horizons and plant fossils are common. The Mauch Chunk Formation consists of grayish-red shale and siltstone, brown, gray, and white sandstone, and some conglomerate (Geyer & Wilshusen, 1982). The Pottsville Formation consists of well- to very well cemented, medium-grained to conglomeratic sandstone beds with minor amounts of siltstone, claystone, shale, thin discontinuous coals (Geyer & Wilshusen, 1982). The Llewellyn Formation consists of interbedded sandstone, siltstone, shale, conglomerate, and coal. The Llewellyn contains abundant plant fossils and most of the minable coal beds in the Wyoming-Lackawanna Valley (Geyer & Wilshusen, 1982). The strata dip steeply across the area

to the northwest, and some units are folded and faulted, in particular the resistant sandstones and conglomerates in the Pocono and Pottsville Formations (Bergin, 1973; Sevon, 1976). Additional details on the bedrock geology of the site are contained in the 1997 Field Conference guidebook. Surficial geological materials are mapped as Wisconsinan age till and shallow bedrock (**Figure 1**), with glacial striae mapped on polished bedrock surfaces that define paleo-ice flow directions (Braun, 2008). Polished bedrock surfaces of Pocono Formation sandstones along the hillslope of Bear Creek Mountain are the target for cosmogenic exposure dating at the stop (**Figure 1**).



Figure 2. Photograph of the lowermost pothole (i.e. the Seventh Tub) developed along Wheelbarrow Run at the Seven Tubs Recreation Area

Review of the Origin of the Seven Tubs

Duane Braun (retired professor at Bloomsburg University) presented ideas relating to the origin of the Seven Tubs in the 1997 Field Conference guidebook (Braun, 1997). A summary of the narrative is presented here for context.

The origin of The Tubs has traditionally been attributed either to glacial meltwater or to erosion by the modern stream. Potholes form when high-velocity, turbulent water creates vertical vortices that rotate sediment and larger clasts against resistant bedrock, gradually grinding out circular depressions via the process of abrasion. Because potholes at The Tubs reach depths exceeding 10 feet, their development likely required hundreds to thousands of years of repeated high-energy flow events rather than rapid formation over a few decades. Although their location within a glaciated region initially suggested a glacial meltwater origin, geological evidence indicates that meltwater drainage at The Tubs would have been brief and limited. Unlike the nearby Archbald Potholes, which are aligned with former ice and meltwater flow, The Tubs are oriented transverse to glacial movement, and mapping of glacial deposits by Braun (2008) in the region suggests that any downslope meltwater capable of carving them would have existed for only a few years during ice retreat.

The evidence therefore supports the conclusion that the modern stream has excavated The Tubs since deglaciation (speculated to be ~17,000 years at the time). The remarkable size of the potholes likely resulted from a unique combination of geomorphic and geologic factors: highly resistant and well-jointed sandstone and conglomerate of the Pocono Formation, a steep joint-slot gorge that concentrates flow into localized vortices, and the largest drainage basin of any comparable stream on the south flank of the Lackawanna synclinorium. The gorge's zig-zag pattern along multiple joint directions forces water into abrupt channel turns, creating ideal conditions for pothole erosion and increasing the necessary stream power to incise vertically into bedrock. Combined with a drainage area of 1.6 square miles, channel widths of 5–10 feet, and a steep gradient of 0.2 ft/ft, these characteristics generate sufficient energy (stream power) to transport coarse-grained bedload sediments during high discharge events and produce the largest and deepest potholes in the region.

Field Stops

Stop #3A – Stream Gauging at Laurel Run (41.2349° N, 75.8111° W)

Leader: Matt Finkenbinder, Sam Winter, Aidan Hogan, and Faith Morcombe (Wilkes University)

A gauging station was installed on Laurel Run in July of 2025 to establish baseline hydrologic data for the stream (**Figure 3**). The motivation to gauge the stream was based on several factors, including 1) observations of substantially low water levels and flows during drought conditions throughout the spring and summer of 2022, 2) proximity of the stream to the Wilkes campus and use of the site for teaching hydrologic field techniques, 3) a desire to quantify seasonal to annual trends in streamflow and longer-term flood frequency relationships, and 4) impacts of streamflow conditions on aquatic ecosystems. Laurel Run and its tributaries are cold-water streams that are classified as natural trout reproduction or Class A streams and therefore contain a thriving population of sustainably reproducing brook trout (Pennsylvania Fish and Boat Commission, 2025). Low water levels combined with high water temperatures and low dissolved oxygen levels, experienced through the summer months, likely adversely impact aquatic ecosystems in the Seven Tubs. As a result, generating baseline hydrologic data for the stream over time will provide insights into event scale and longer conditions that will allow for comparison against future changes driven by land use and/or climate changes. Furthermore, the vast majority of gauged streams in Pennsylvania have watershed areas > 10 square kilometers (Chaplin, 2005), and therefore small mountain streams are substantially underrepresented in gauging networks.



Figure 3. Photograph of the stilling well and gauging station installed on Laurel Run for continuous measurements of stage (water levels) and temperatures at 5-minute intervals.

The gauge is constructed of a screened 4-inch diameter PVC pipe (**Figure 4**) and in-situ water and atmospheric pressure loggers to monitor stage (water level) and temperature variations at 5-minute intervals. The gauge is affixed to the downstream end of a concrete culvert located approximately 100 meters downstream of the confluence of Wheelbarrow Run. Access to the gauged site for the group is difficult due to very steep, uneven terrain and space limitations. We will therefore discuss this project at the confluence of Wheelbarrow Run with Laurel Run while viewing the lowermost pothole (e.g. the Seventh Tub).

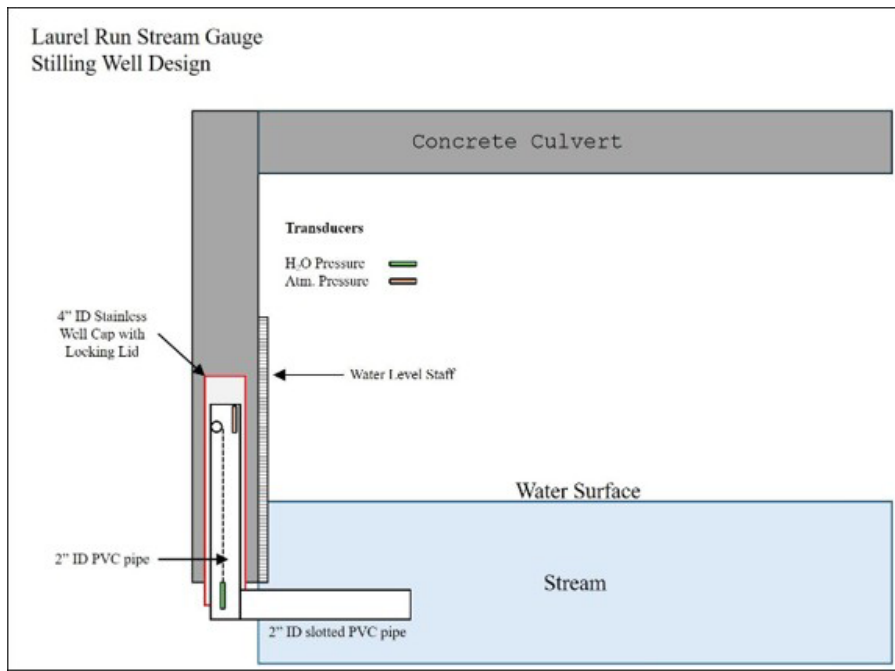


Figure 4. Schematic showing the stream gauge and still well design.

Preliminary discharge and in-situ water quality (temperature and dissolved oxygen) measurements were collected throughout the summer of 2025 by Wilkes University students Sam Winter, Aidan Hogan, and Faith Morcombe as part of a summer research project (**Figure 5**). The water quality measurements were acquired at the gauging station directly from the stream using a handheld Oakton dissolved oxygen meter and values of temperature (° C) and dissolved oxygen (mg/L) were recorded. The discharge measurements were acquired using a U.S. Geological Survey pygmy current meter and an AquaCalc Pro Plus device fixed to a 4-foot length staff gauge. Pressure logger data was periodically retrieved from the gauging station and analyzed to calculate stage (water level) changes over time. To calculate stage, air pressure data were first subtracted from the water pressure data, to derive a pressure anomaly value in units of pounds per square inch (PSI). The pressure anomaly data were then converted to a water depth in units of feet according to the following equation:

$$\text{Depth (feet)} = \text{pressure (PSI)} * 2.31$$

A discharge rating curve was then created using simple linear regression in Excel by plotting stage (feet) vs. discharge (cubic feet per second or cfs) and adding a linear best-fit line (**Figure 6**). The equation of the best-fit line was then used to calculate the corresponding discharge for the stage measurements at continuous 5-minute intervals. Preliminary data from the gauging station, for the period July 2025 through April 2026, shows several notable features including 1) baseflow recession over the summer of 2025, 2) groundwater recharge over the fall of 2025 through spring of 2026, and 3) several storm hydrographs defined by a rapid increase in discharge followed by a slower recession (**Figure 7**). A series of storm hydrographs and sustained high flows during March and early April of 2026 represent flushing of stored water in the catchment due to rising temperatures and the spring melt (**Figure 7**).



Figure 5. Wilkes University students Aidan Hogan, Sam Winter, and Faith Morcombe collecting streamflow data at Laurel Run in July of 2025.

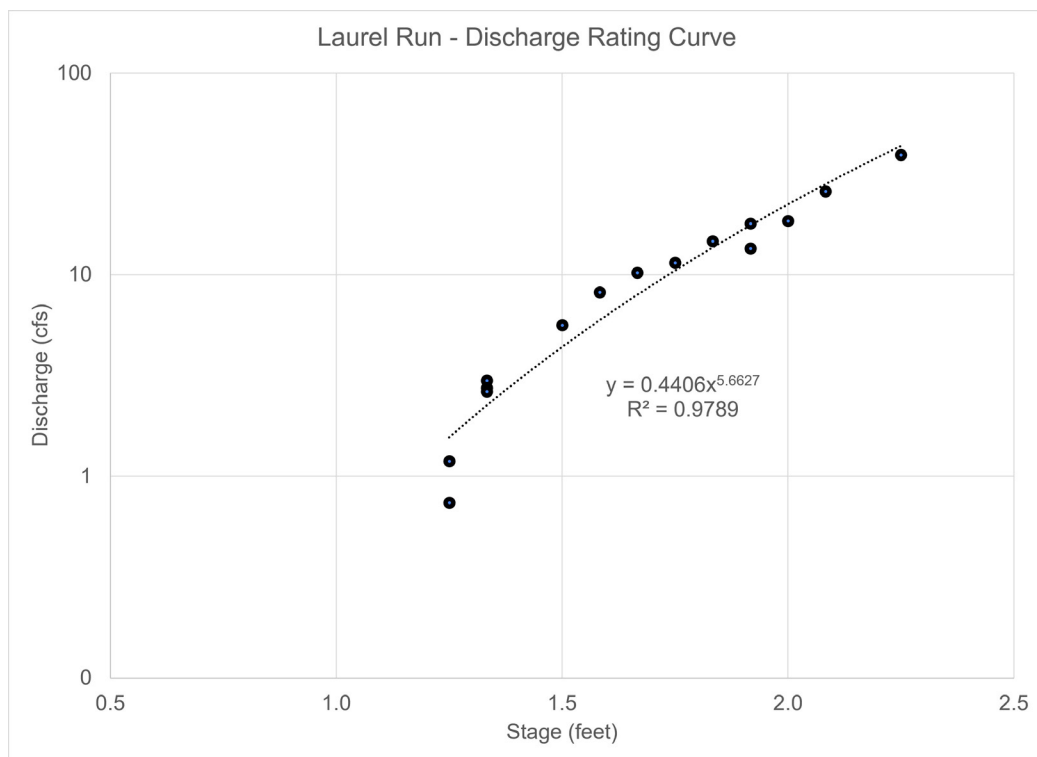


Figure 6. Discharge rating curve for Laurel Run and best-fit line used to convert stage (water level) data to discharge (cubic feet per second).

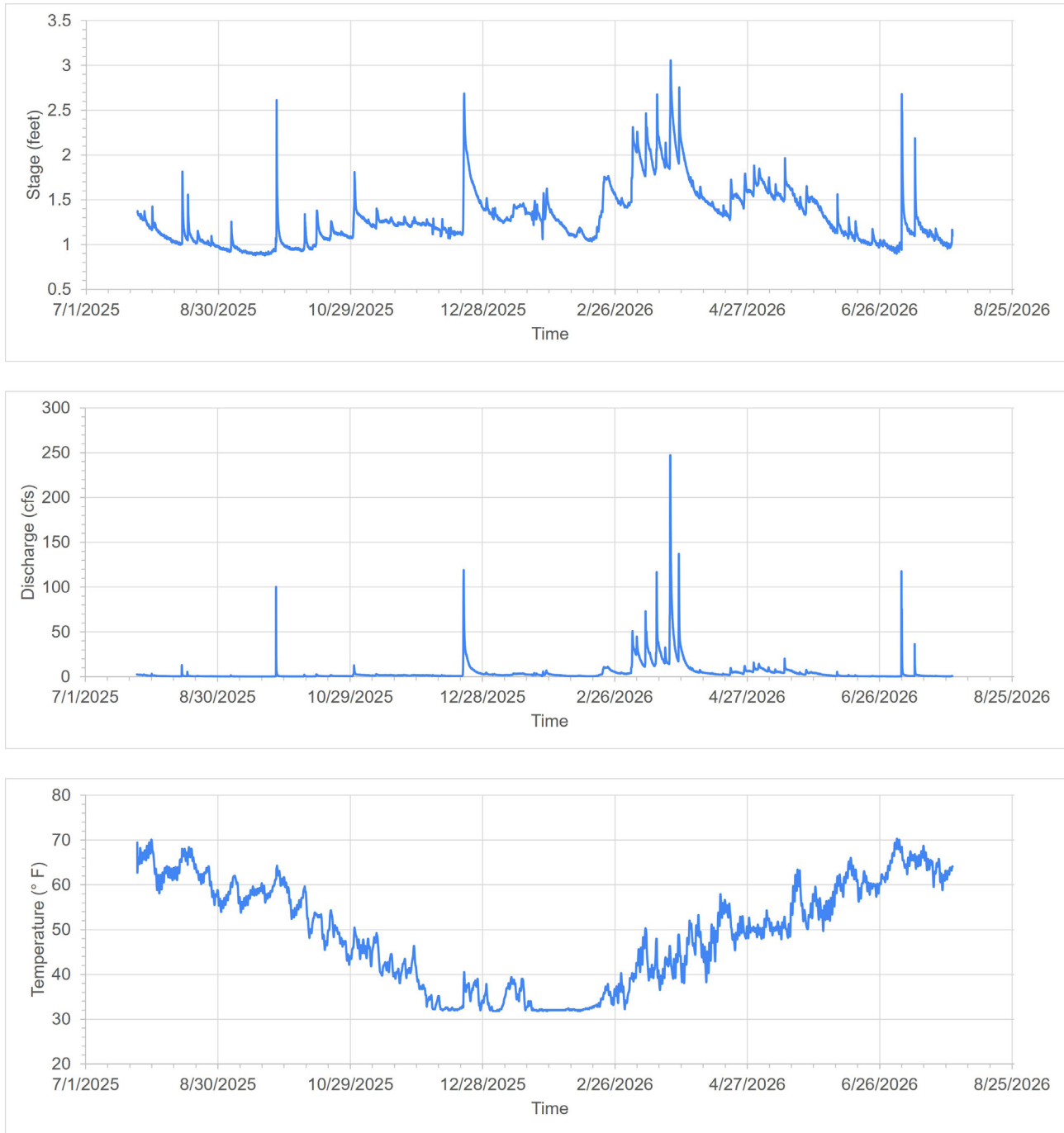


Figure 7. Laurel Run gauge data from July of 2025 through May of 2026 including stage (feet), discharge (cfs), and water temperature (° F).

Future work by Wilkes University students will focus on collecting additional discharge measurements during extreme hydrologic conditions (high flows) to capture high magnitude, low frequency events that are often under-represented in rating curves. In addition, the amount of suspended sediment transport through the stream during high-flow events is unknown. Therefore, we intend to collect suspended sediment samples from the stream to develop a sediment rating curve, and combined with daily discharge data, will quantify annual lowering (erosion) rates (Walling, 1983) to better understand the conditions necessary to produce the Seven Tubs potholes along Wheelbarrow Run.

Stop #3B – Cosmogenic Exposure Dating of Bedrock (41.2330° N, 75.8106° W)

Leader: Timothy Lane (Aarhus University), Kathryn Adamson (Manchester Metropolitan University), and Matt Finkenbinder (Wilkes University)

This field stop involves an introduction to cosmogenic nuclide dating (or surface exposure dating), a method which can directly determine the timing of events and rates of change in the Earth's surface, using measurement of cosmogenic nuclide production through cosmic ray-induced reactions in rocks and sediment.

Cosmogenic Nuclide Dating

Cosmic radiation is constantly bombarding the Earth, and those which penetrate the Earth's magnetic field trigger a cascade of secondary cosmic rays (mostly high energy nucleons and mesons). These rays attenuate through the Earth's atmosphere before hitting the Earth's surface, where they are further attenuated depending on the density of the surface. This triggers reactions in target minerals (quartz for this geological setting) in the rock that produces cosmogenic nuclides. The neutron flux causing these reactions attenuates rapidly in rock, reaching <1% *three metres* below the surface. Given that these primary rays are produced at a known (though time variable) rate, sampling rock from the top of a bedrock or boulder surface allows us to calculate the length of time that location has been exposed. In quartz bearing rocks the most commonly used nuclides for studies over the last several glacial cycles are ^{10}Be and ^{26}Al , which have half-lives of 1389 ka and 708 ka respectively. Studies routinely use ^{10}Be , but due to their differential production and decay rates, the two nuclides (often ^{10}Be and ^{26}Al) can be paired together to provide more information on exposure history. This is especially relevant in settings where the exposure history is complex (periods of thin ice cover and exposure), or allows identification of locations when glacial erosion has not been sufficient to remove the top three metres of material (**Figure 8**).

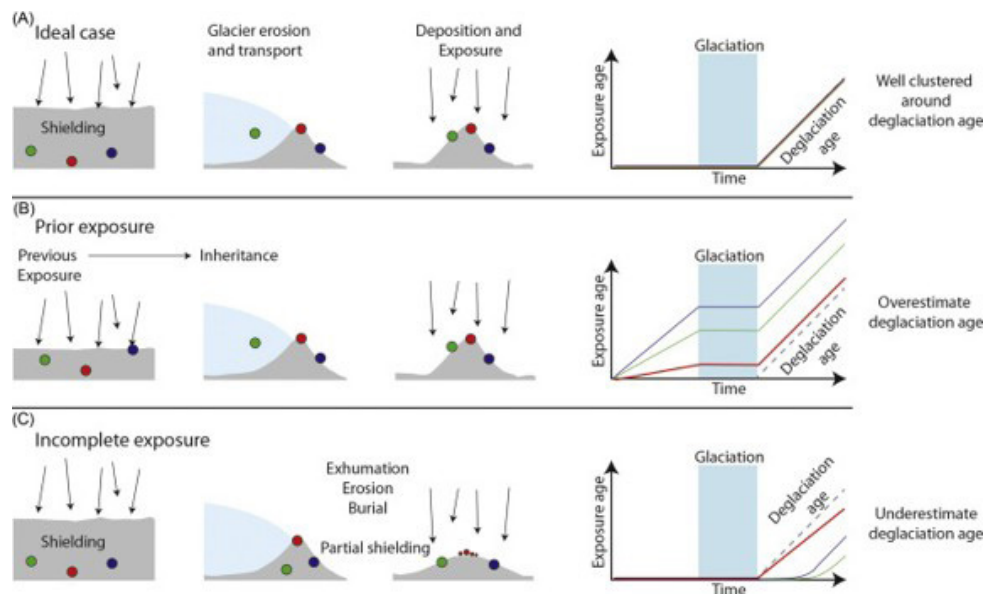


Figure 8. Principles of prior exposure and incomplete exposure and resulting apparent exposure ages. (A) The sample has been completely shielded from cosmic rays prior to glacialiation, and then continuously exposed since deglaciation. (B) The sample is exposed to cosmic rays prior to glacialiation and experiences no post-glacial shielding (prior exposure). The apparent exposure age exceeds the deglaciation age. (C) If a sample is incompletely shielded from cosmic rays prior to glacialiation and partially shielded following deglaciation (incomplete exposure), the apparent exposure age will be younger than the deglaciation age. Reproduced from Davies et al. 2022.

Calculating Exposure Age

In the example presented here (and discussed at Stop 6 and elaborated further at Stop 7), we will discuss the use of cosmogenic nuclide dating to constrain the exposure of bedrock surfaces or boulders following the retreat of the Laurentide Ice Sheet during the last deglaciation. These methods are especially valuable in constraining the timing of deglaciation as they do not rely on the presence of organic material for radiocarbon dating, but instead just require suitable outcrops of bedrock, or well situated erratic boulders.

The concentration of nuclides in the host mineral can be calculated with:

$$N = \frac{P}{\lambda} (1 - e^{-\lambda t})$$

When solving for exposure age (t), this gives us:

$$t = -\frac{1}{\lambda} \ln \left(1 - \frac{N\lambda}{P} \right)$$

Where:

t: Exposure age of the rock surface (years).

N: Measured concentration of the cosmogenic nuclide (atoms per gram of rock).

P: Local production rate of the nuclide at the sampling site (atoms per gram per year). This varies significantly based on elevation, geomagnetic latitude, and sample shielding.

λ : Radioactive decay constant of the specific cosmogenic nuclide (year⁻¹).

Therefore, once we measure concentration (N), and choose a preferred production rate for our site (P), a simple exposure age can be calculated. Whilst nuclide concentration can be precisely measured by accelerator mass spectrometry, producing a precise exposure age, there are many geological assumptions and sampling uncertainties which need to be considered and where possible, minimized.

Inheritance: Unless specifically investigating periods of prior exposure or areas covered by cold-based ice, we only want samples which have a nuclide inventory representing the last period of exposure (e.g. since the LGM retreat). However, if they have not been eroded to >3 m depth the most recent period of erosion (e.g. the last glacial cycle) they will have “inherited” nuclides from before they were covered. This can elevate the age, as it is a sum of the time period since exposure (which we want to measure) and any prior exposure (**Figure 8B**).

Erosion: Depending on the landscape, surface erosion can vary widely depending on climate, lithology, and local context. For most nuclides, this will reduce the age as, although the sample is exposed for the “correct” amount of time, part of the nuclide inventory will be removed. If the average surface erosion over the exposure period is known or can be estimated this can be corrected for. Otherwise, this is difficult to do and will result in the exposure ages being a minimum.

Sediment, vegetation, snow cover, and exhumation: Following exposure from the retreating ice sheet, many processes can act to suppress nuclide production through physical shielding (**Figure 8C**). This includes cover by sediment, thick vegetation, or persistent snow (over several metres thick). All of these will reduce nuclide production, and are difficult to quantify, and may appear absent in the present day, so are not considered. Thus, it is very important to consider these factors when sampling.

Production rate: The production rate of a specific nuclide varies through time, driven by changes in incoming cosmic rays and Earth's atmospheric conditions. Thus, production rates need to account for both changes in time, and location of the samples on the Earth's surface. Nuclide production increases with altitude as a thinner atmosphere provides less interaction with cosmic rays. Cosmogenic nuclide production is lower towards the equator because the magnetic field more strongly deflects incoming cosmic rays at low latitudes (Nishiizumi et al., 1989, Gosse and Phillips, 2001, Stroeve et al., 2015). This can be addressed with good locational data, and well produced localised production rates. At present, the Sea-Level, High-Latitude production rate for in-situ cosmogenic ^{10}Be in quartz ranges from approximately 3.6 to 4.3 atoms/g/year. There is a global production rate (Borchers et al., 2016), or a number of regional and local production rates (**Figure. 9**). Each has its benefits and drawbacks, but it is important that the production rate used is presented in any studies. Here we will be using the global production rate of 4.09 ± 0.20 at/g/yr (Borchers et al., 2016).

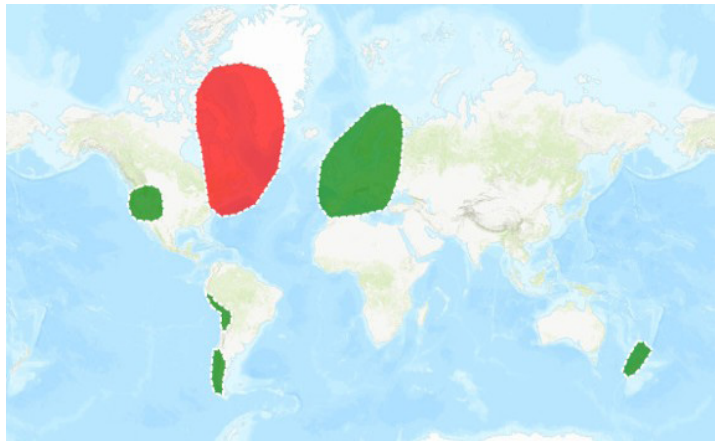


Figure 9. Screenshot from the Cosmic Ray Exposure Program (CREP), used to calculate exposure ages, with the different regional production rates shown. The regional production rate for NEPA (Arctic-NE North America, Young et al., 2013) is shown in red.

Sampling for Cosmogenic Nuclide Dating

Taking samples for dating is a time-consuming task which requires a large amount of planning, strategizing, and sometimes luck. Care is needed to attempt to understand the likely geological history of the surface or boulder before you take it. As outlined above, cover by sediment or vegetation can dramatically lower the sample's age, and unexpected inheritance can dramatically raise it. Below and at the fieldstop we present the key steps when sampling with the aim of reconstructing ice sheet retreat histories.

- When choosing a sample site, an understanding of sample context is needed. Has the sample been exposed following ice retreat? Could it have been exposed by other, non-glacial processes? If it is a bedrock sample, is it striated, or is there other evidence it was eroded sub-glacially? If it is a boulder, is it certain it was emplaced by the ice sheet, or has it rolled into position from a nearby slope? Could the boulder have rolled from one part of the moraine to another, post-depositionally?
- Is the sample surface sufficiently above the surrounding terrain? Ideally there will be 50-100 cm of "topography" between the location to be sampled and the lower lying ground. This will help ensure sediment and snow cover is minimised due to windsweeping.

- Can replicate samples be taken? There is often geological scatter across samples, so taking > three samples allows for outliers to be estimated.
- Once satisfied, the sample itself can be taken. The ideal surface will be flat and relatively horizontal to minimize effects of self-shielding. The sample site should be away from edges (50 cm ideally), and not close to large joints or cracks.
- A sample (roughly 1 kg of rock for ^{10}Be analysis) can then be taken, either using a circular saw, explosives, or a hammer and chisel. Care should be taken to avoid damage to the surface, though some is inevitable.

Constraining Ice Retreat

As discussed at Stop 1, constraining the timing of LIS retreat is important for understanding ice sheet behavior and the forcing factors which caused the onset and pace of deglaciation. Whilst much of the northern and southern LIS margins are now well-constrained (Dalton, 2022), the response of the large interlobate region of the southeastern LIS margin between the Erie and Hudson lobes (including the Salamanca re-entrant and northeast Pennsylvania) is still poorly understood. This region is important because it is characterized by well-preserved ice marginal deposits (Braun, 2011). Using a combination of cosmogenic nuclide dating presented here and at Stop 6 and Stop 7, alongside basal radiocarbon ages from Nuangola Lake (Stop 1), Cranberry Pond (Finkenbinder et al., 2026), and Lost Lakes Bog (Winter et al., 2025), work at the Seven Tubs aimed to (1) establish the timing of deglaciation at the southeastern LIS margin and (2) test current hypotheses of LIS deglacial behavior, and implications for regional paleoenvironment and landscape evolution following deglaciation. At present (and as discussed at Stop 1), there is a ~7,000 to 10,000 year offset between ^{10}Be exposure ages from the New England and New Jersey LGM terminal moraines and the AMS ^{14}C ages from lakes and bogs in northeast Pennsylvania and adjacent New Jersey, which cannot be explained by environmental lag-times alone. As a result, the onset of deglaciation in this region remains debated.

The Seven Tubs is situated about 20 km north of the Last Glacial Maximum LIS terminal moraine. Five bedrock samples (ST-01 through ST-05) were taken from exposures of the Mississippian age Pocono Formation at the Seven Tubs to help constrain LIS retreat across this location (**Figure 10**). Samples were taken from prominent areas of glacially abraded bedrock, displaying evidence of glacial polish and striations (**Figure 11**). Sample locations were raised above the surrounding terrain, to ensure minimal postglacial sediment cover or vegetation growth (**Figure 11D**). Samples were measured at the National Environmental Isotope Facility (East Kilbride, UK) and exposure ages were calculated using the global production rate (Borchers et al., 2016) and the LSDn time-dependent scaling framework (Lifton et al., 2014) in years (yr) prior to sampling (2018 CE).

The five bedrock samples from Seven Tubs return ^{10}Be exposure ages of 24,000 to 31,000 yr with a mean exposure age of 28,000 yr (**Figure 12**). These ages, acquired from bedrock surfaces located ~20 km north of the maximum extent of the Laurentide Ice Sheet in Pennsylvania are incompatible with knowledge of regional deglaciation, which is thought to have occurred 24,000-20,000 years ago (Halsted et al. 2024), and disagree with basal age from Nuangola Lake (Stop 1C) that suggests deglaciation before ~19,260 cal yr BP (Finkenbinder et al., 2026) (**Figure 12**). In addition, the Seven Tubs mean exposure age of 28,000 yr is substantially older than our preliminary ^{10}Be exposure ages from the Pocono Lake terminal moraine (discussed later at Stop 6 and Stop 7) (**Figure 12**). The samples are uniform in that they are all older than plausible, and, because they are bedrock samples, we suggest that erosion during the LGM was not sufficient to remove >3 meters of bedrock, fully “resetting” their nuclide inventory. The apparent ‘too old’ exposure ages from the Seven Tubs are therefore due to inheritance of (^{10}Be) radionuclides from previous exposure due to a combination of relatively thin ice cover and/or a limited duration of ice cover that resulted in minimal subglacial erosion.

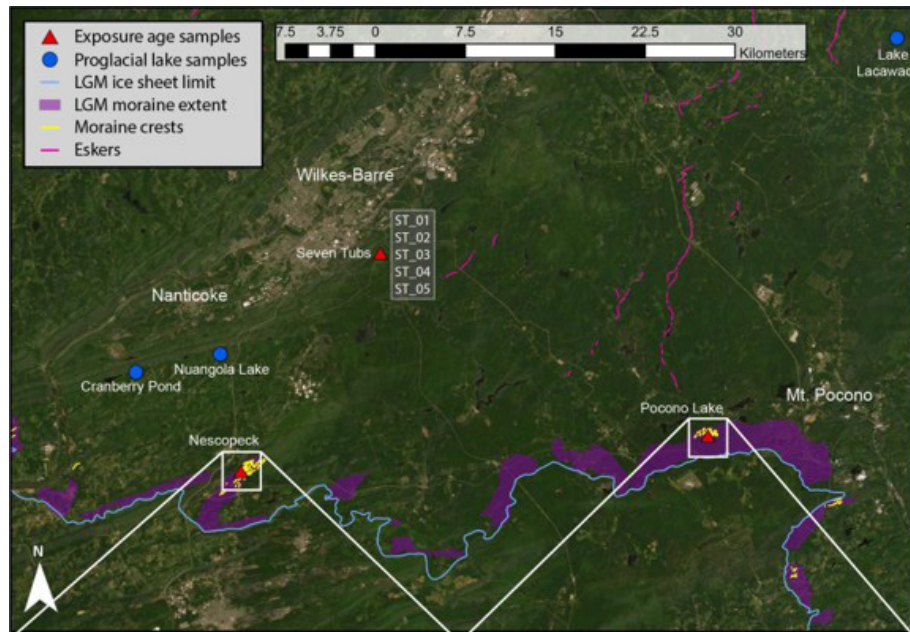


Figure 10. Map of the study sites in northeast Pennsylvania showing the Last Glacial Maximum extent of the Laurentide Ice Sheet (extensive purple moraine, with maximal limit indicated by blue line) (Crowl and Sevon, 1980), the location of boulder/bedrock samples collected for cosmogenic surface exposure dating at the Nescopeck Moraine (not presented here), Pocono Lake Moraine, and Seven Tubs bedrock outcrop (red triangles), and lake study sites Nuangola Lake, Cranberry Pond, and Lake Lacawac (blue circles).



Figure 11. Photographs of glacially abraded and polished Pocono Formation outcrops at the Seven Tubs Recreation Area sampled for cosmogenic exposure dating.

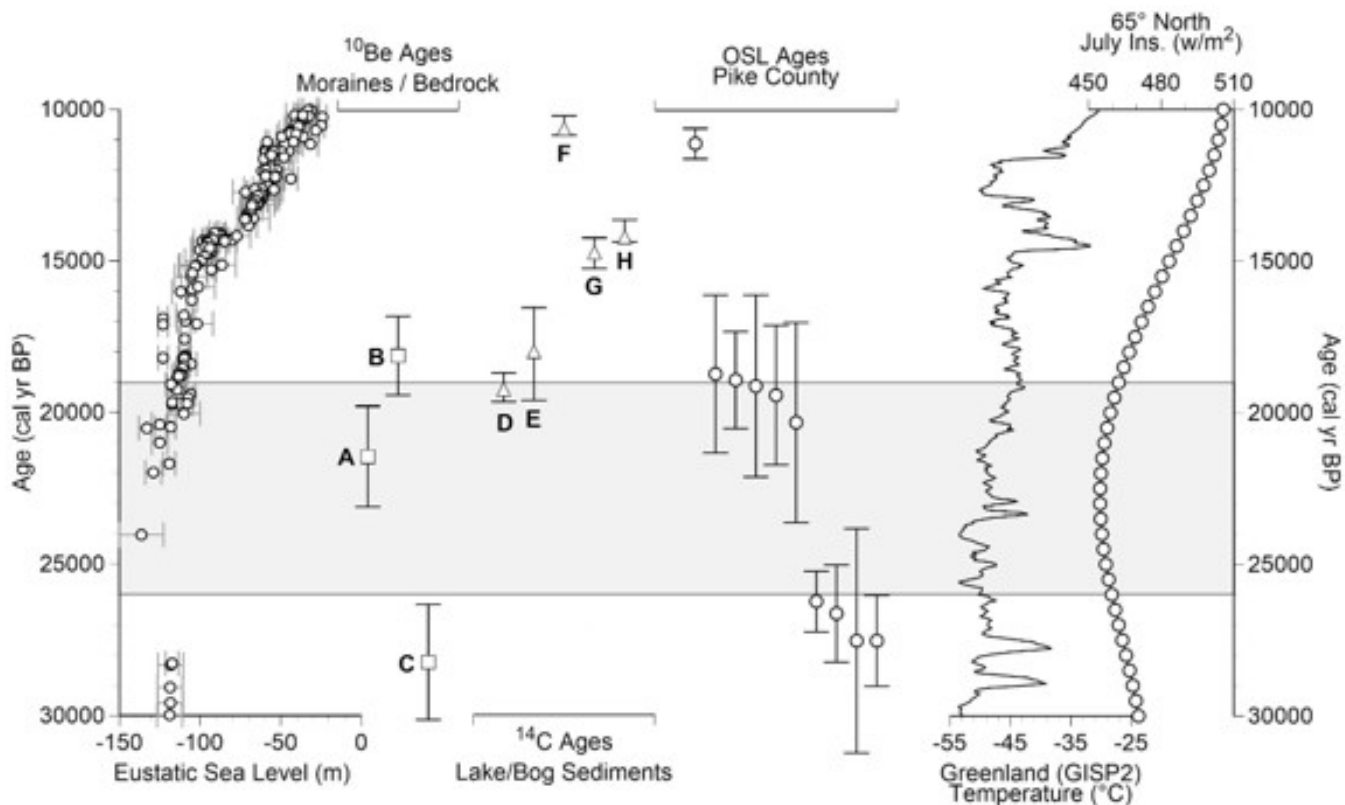


Figure 12. Comparison of Last Glacial Maximum proxy records and the culmination age of glacial moraines and age of deglaciation from regional lake and bog records, including eustatic sea level (Lambeck et al., 2014); ^{10}Be cosmogenic exposure dates (black square symbols) from (A) the Pocono Lake moraine at the Maple Tract Preserve (Winter et al., 2025), (B) the Pocono Lake moraine at State Game Lands 318, and (C) the Seven Tubs Recreation Area; Basal radiocarbon ages (black open triangle symbols) from lakes including (D) Nuangola Lake, Pennsylvania (Finkenbinder et al., 2026), (E) Cranberry Pond, Pennsylvania (Finkenbinder et al., 2026), (F) Lost Lakes bog, Pennsylvania (Winter et al., 2026), Halfmoon Lake, Pennsylvania (Cotter et al., 1986), and Tannersville Bog, Pennsylvania (Peteet et al., 2012); Optically Stimulated Luminescence (OSL) ages on glacial sand and gravel deposits from Pike County, Pennsylvania (Johnson and Finkenbinder, 2023); Greenland Ice Sheet Project 2 (GISP2) temperature (Alley, 2000); and 65° North July insolation (Laskar et al., 2004). The gray bar depicts the time range for the Last Glacial Maximum (Clark et al., 2009).

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Stop 4A: Landslide Mapping - Garden Drive In Small Landslide

(41.21444, -76.041509) Garden Drive In, along Route 11 Gardens PA

Presentation on historical landslides, estimates of initiation, and the development of a landslide inventory and susceptibility model for northeast Pennsylvania.
Observation of the toe deposit of a small landslide.



Stop 4b: Landslide Mapping - Union Township Boat Launch Large Landslide

(41.17953, -76.1052) Union Township Boat Launch off Route 11.

Presentation on risk assessment of the landslides on Shickshinny Mountain with a demonstration of tree ring data acquisition for activation/reactivation studies.
Observation of the toe deposit of a large landslide.

Stop 4 – Landslides of Shickshinny Mountain

Bobak Karimi¹

¹ Biology and Earth Systems Sciences, Wilkes University, Wilkes-Barre, PA, USA

This location includes two separate stops that will be visited by the entire group at the same time. There are short hikes along the paved (gravel), flat path of Warrior Trail to the toe of two landslides.

Introduction

Shickshinny Mountain, located along the western bank of the Susquehanna River in Luzerne County, Pennsylvania, is one of the most landslide-prone slopes in northeastern Pennsylvania. The steep mountainside parallels both the Susquehanna River and U.S. Route 11 and contains numerous mapped and suspected landslides representing multiple stages of development. At least one historic landslide partially obstructed the Susquehanna River, demonstrating the potential for large slope failures to alter river flow and impact surrounding communities (Braun and Gillmeister, 1989). Recent investigations have identified approximately 19 mapped and suspected landslides distributed along the slope (**Figure 4.1**), ranging from dormant and relict failures to areas exhibiting evidence of continued movement.

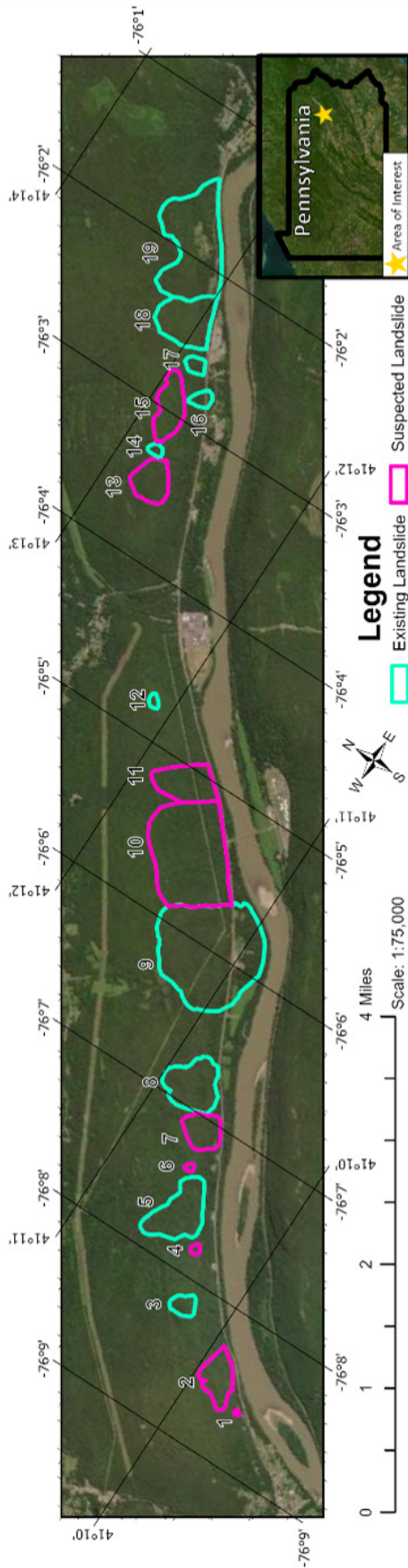


Figure 4.1. Shickshinny Mountain landslides categorized into landslides known to exist, and those suspected to exist based on surficial features.

The combination of steep relief, continued river incision, and unfavorable geologic conditions has produced repeated episodes of slope instability. Continued movement threatens nearby transportation corridors, utilities, and the Susquehanna River, while large-scale failures have the potential to introduce substantial volumes of sediment and debris into the river. In extreme cases, future failures may temporarily dam the Susquehanna River, producing upstream flooding and downstream hydraulic impacts. Because these hazards extend beyond the immediate slope, understanding the distribution, activity, and potential consequences of landslides is critical for effective hazard mitigation and emergency planning.

The investigation presented in this chapter was designed to address three primary questions:

1. Where are the active, dormant, and suspected landslides located, and what does their distribution reveal about slope evolution?
2. When were these landslides initiated or reactivated, and what evidence exists for continued movement?
3. What are the potential consequences of future slope failure to nearby infrastructure, the Susquehanna River, and surrounding communities?

These questions form the basis for the field excursion and guide the observations presented at Stops 4A and 4B. To address these questions, an integrated investigation was conducted combining field mapping, historical imagery, dendrochronologic analysis, hydrologic modeling, and transportation network analysis. Together, these methods were used to develop a comprehensive landslide inventory, reconstruct the history of slope movement, and evaluate the potential consequences of future failures.

The field stops presented in this chapter progress from documenting the geomorphic evidence and historical development of landslides along Shickshinny Mountain (Stop 4A) to evaluating their potential impacts using modern hazard and risk assessment techniques (Stop 4B).

Geologic Setting

Shickshinny Mountain forms the northern limb of the Lackawanna Synclinorium, a broad northeast-plunging fold within the Appalachian Plateau of northeastern Pennsylvania. Along the southern slope of the mountain, bedrock units strike approximately east-northeast to northeast and dip 19–28° toward the southeast. These strata are exposed along the mountainside and exert a strong influence on slope morphology and stability.

The southern slope of Shickshinny Mountain commonly parallels the dip of the underlying bedrock and, in many locations, exceeds the dip angle of the strata. This dip-slope geometry creates favorable conditions for downslope movement by promoting failure along bedding planes and other pre-existing structural discontinuities. Slope stability is further influenced by the physical characteristics of the underlying bedrock and surficial materials. Weathering and fracturing of bedrock, unfavorable bedding orientation, joint and fracture networks, contrasts in permeability, groundwater saturation, and erosion at the base of the slope all act to reduce resisting forces or increase the driving forces responsible for slope failure (Popescu, 1994). Failure commonly occurs along pre-existing discontinuities, including bedding planes, joints, fractures, and contacts between unconsolidated material and competent bedrock. Along Shickshinny Mountain, these conditions are further enhanced by continued undercutting of the slope by the Susquehanna River and localized excavation associated with U.S. Route 11, both of which reduce support at the toe of the slope and increase the likelihood of translational and rotational slope failures as well as reactivation of existing landslides (Braun and Gillmeister, 1989; Lipoma et al., 2014).

Lithology exerts a primary control on both the distribution and style of landslides along Shickshinny Mountain. West of Hunlock Creek, the majority of mapped landslides occur within the resistant sandstones and conglomerates of the **Pocono Formation**, whereas landslides east of Hunlock Creek are developed primarily within the **Duncannon Member of the Catskill Formation**, consisting of interbedded red mudstones and gray sandstones. The alternation of competent sandstone beds with weaker mudstone intervals provides favorable conditions for sliding along mechanically weak horizons. The largest landslide complex spans both the **Mauch Chunk Formation** and the underlying **Pocono Formation**, with a younger nested landslide developed entirely within the larger failure. This nested geometry provides evidence for multiple episodes of slope movement and illustrates the long-term evolution of the Shickshinny Mountain landslide complex.

Braun and Gillmeister (1989) proposed that incision of the Susquehanna River through late Wisconsinan outwash deposits progressively steepened the valley wall and initiated instability along the dip slope. Because the largest landslide extends onto and partially diverts the modern river channel, they further suggested that its development postdates establishment of the present valley floor, providing an estimated age of approximately 2–4 ka. This estimate represents a preliminary geomorphic age constraint rather than a precise chronology of landslide activity.

Collectively, regional folding, dip-slope geometry, lithologic contrasts, and continued river incision have produced conditions favorable for repeated slope instability. These geologic controls explain both the concentration of landslides along Shickshinny Mountain and the variety of failure mechanisms observed throughout the study area.

Previous Investigations of the Shickshinny Mountain Landslides

The first systematic descriptions of the Shickshinny Mountain landslides were presented by Braun and Gillmeister (1989). These investigations identified approximately 13 landslide deposits or potential rock-block slides along the mountain using topographic information, imagery, and field observations. The identified features were interpreted primarily as large translational failures associated with the steep slope above the Susquehanna River and U.S. Route 11 (Braun and Gillmeister, 1989; Lipoma et al., 2014).

The earlier work also recognized the potential consequences of renewed movement. Failures along the lower slope could affect U.S. Route 11, nearby property, and utilities, while a sufficiently large landslide could enter and partially obstruct the Susquehanna River. Such an obstruction could impound water upstream and, if the resulting landslide dam failed, generate downstream flooding (Lipoma et al., 2014). One of the largest mapped landslides has a prominent toe that appears to have displaced the river channel locally, illustrating the long-term interaction between slope movement and river morphology.

Braun and Gillmeister (1989) also documented the geometry of the largest landslide on Shickshinny Mountain, describing it as a large rock block slide involving nearly the entire valley wall. The landslide extends approximately 650 m downslope and 750 m along strike, with an estimated thickness of 20–27 m and an estimated volume of 15.5–21 million m³. The lobate toe extends approximately 380 m onto the Susquehanna River floodplain and locally constricted the river channel, demonstrating the capacity of large prehistoric landslides to significantly modify valley morphology.

Despite this early recognition, the available work was largely limited to preliminary mapping, surface characterization, broad descriptions of failure type, and generalized estimates of landslide age. Little information was available concerning subsurface geometry or the timing of individual activation and

reactivation events. Building upon the original inventory developed by Braun and Gillmeister (1989), this investigation expanded the mapped inventory to approximately 19 landslides and suspected landslide features through detailed field mapping, interpretation of historical aerial imagery and LiDAR-derived terrain models, and geomorphic analysis. Additional investigations including dendrochronology, hydrologic modeling, and transportation network analysis were conducted to better constrain landslide activity and evaluate the potential consequences of future slope failures.

Mass Movement Classification & Landslide Evolution

Landslides are a type of mass movement in which rock, soil, or debris moves downslope primarily under the influence of gravity. Although numerous classification systems have been developed based on material type, movement mechanism, and style of deformation (Varnes, 1978; Hungr et al., 2014), the term *landslide* is used throughout this chapter to broadly encompass the variety of slope failures observed along Shickshinny Mountain (**Figure 4.1**).

One of the most widely adopted classification systems was developed by Varnes (1978), who classified mass movements according to both the material involved (rock, debris, or earth) and the dominant style of movement (**Table 4.1**). This framework recognizes six principal categories: **falls**, **topples**, **slides** (rotational and translational), **lateral spreads**, **flows**, and **complex movements**, the latter representing combinations of two or more movement types (**Figure 4.2**). Because movement style is strongly influenced by material properties, different types of mass movements exhibit distinct rates of movement, failure mechanisms, and geomorphic expressions. Rock slopes commonly fail as falls, topples, or slides, whereas debris- and earth-dominated failures are more likely to behave as flows due to the lower shear strength and greater mobility of unconsolidated material.

Table 4.1. Abbreviated and modified version of Varnes (1978) slope movement classification. This system takes into account both type of movement and the material being moved.

Type of Movement		Type of Material	
		Bedrock	Engineering Soils
			Predominantly Coarse Predominantly Fine
Falls		Rock fall	Debris fall Earth fall
Topples		Rock topple	Debris topple Earth topple
Slide	Rotational	Rock slide	Debris slide Earth slide
	Translational		
Lateral Spreads		Rock spread	Debris spread Earth spread
Flows		Rock flow	Debris flow Earth flow
		(deep creep)	(soil creep)
Complex		Combination of two or more principal types of movement	

Each landslide type produces characteristic geomorphic features that can be recognized in the field and used to infer failure mechanisms and landslide evolution. Headscarps, displaced blocks, hummocky topography, tension cracks, pressure ridges, and depositional toes provide valuable evidence for interpreting failure mechanisms and reconstructing the evolution of a landslide complex. The landslides

Varnes' (1978) classification of subaerial slope movements

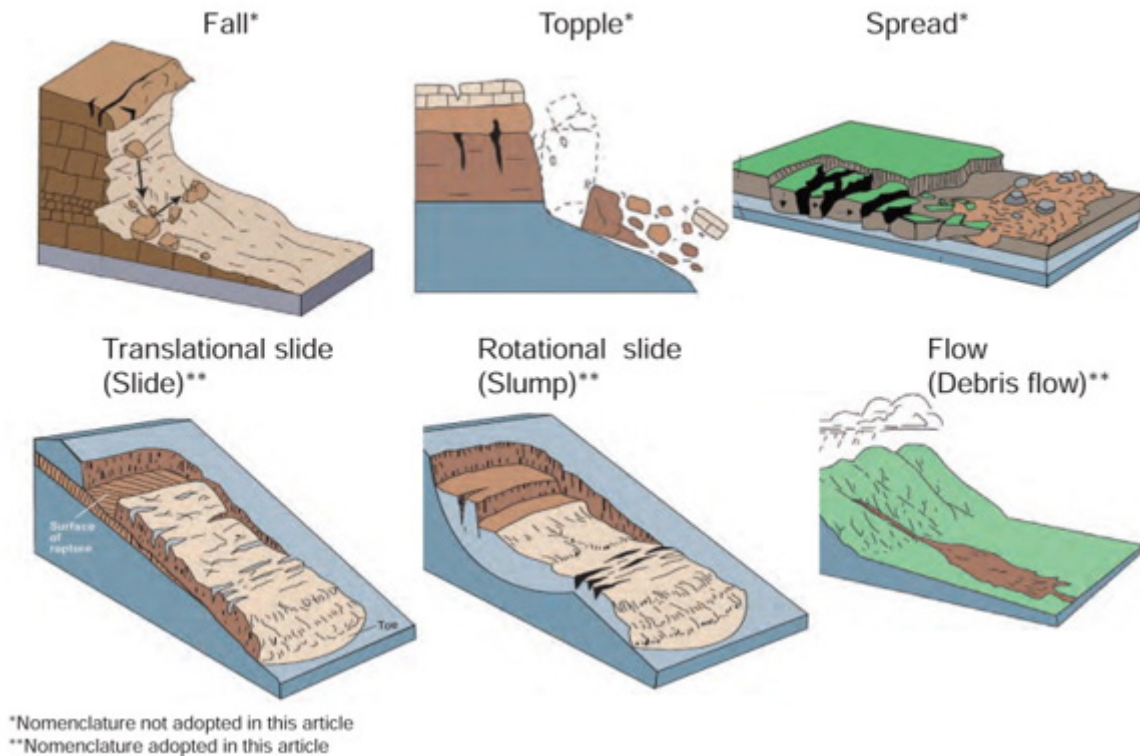


Figure 4.2. Different types of slope movements and how they are classified by Varnes (1978). From Shanmugam and Wang (2015).

along Shickshinny Mountain are dominated by rock block slides, many of which exhibit translational movement along bedding-controlled failure surfaces. Several landslides also display evidence of rotational deformation and complex movement associated with subsequent reactivation.

Slope instability generally develops through the progressive reduction of slope strength resulting from geologic structure, groundwater conditions, weathering, river erosion, and human activities. Failure is commonly initiated by triggering events such as prolonged or intense precipitation, rapid snowmelt, seismic activity, and anthropogenic disturbance (Gelisli et al., 2015). Because landslides often evolve over decades to centuries, recognizing their stage of development is essential for understanding both their history and future behavior.

Rotaru et al. (2007) described four generalized stages of landslide evolution:

1. **Pre-failure** – The slope remains largely intact but exhibits subtle deformation, including soil creep, tilted trees, and minor surface irregularities.
2. **Failure** – A continuous shear surface develops, producing distinct geomorphic features such as headscarps and tension cracks.
3. **Post-failure** – Displaced material moves downslope until motion ceases, forming hummocky topography and depositional toes.
4. **Reactivation** – Previously developed failure surfaces become active again, producing renewed movement that may occur episodically in response to changing environmental conditions.

Many landslides along Shickshinny Mountain display geomorphic evidence representing multiple stages of development simultaneously, reflecting repeated episodes of movement over time. Identifying these

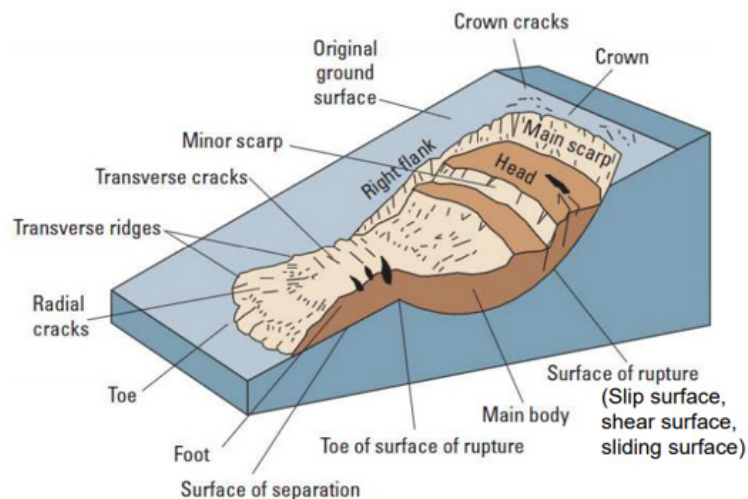
geomorphic indicators provides the foundation for developing a comprehensive landslide inventory, reconstructing landslide history, and evaluating present-day slope activity.

Geomorphic Indicators of Landslide Activity

Landslide inventories are commonly developed through a combination of high-resolution terrain analysis and field verification. Digital elevation and terrain models can reveal features such as headscarps, internal scarps, slide flanks, tension cracks, displaced blocks, and depositional toes, while field observations help confirm feature boundaries, characterize displaced material, and identify smaller features that may not be resolved in remotely sensed data (Mantovani et al., 1996; Metternicht et al., 2005; Nichol et al., 2006).

The principal geomorphic indicators of sliding include a crown and headscarp near the upper limit of failure, minor scarps and transverse cracks within the displaced mass, lateral flanks, hummocky terrain, transverse or pressure ridges, and a bulging or depositional toe at the base of the slope (Varnes,

Figure 4.3. Depiction of general landslide surfae and subsurface characteristics. From "Landslide Types and Processes," by U.S. Geological Survey, 2004, [usgs.gov](http://pubs.usgs.gov/fs/2004/3072/). In the Public Domain.



<http://pubs.usgs.gov/fs/2004/3072/>

1978). Rotational failures may produce backward-tilted blocks and curved rupture surfaces, whereas translational failures typically move along more planar surfaces and may exhibit less internal rotation (Varnes, 1978; Cruden and Varnes, 1996) (**Figure 4.3**).

The distribution and preservation of these geomorphic features provide insight into the relative age and activity of individual landslides. Fresh headscarps, open tension cracks, recently tilted trees, and disrupted drainage commonly indicate recent or ongoing movement, whereas subdued scarps, rounded topography, and mature vegetation generally reflect older, dormant, or relict failures. Consequently, interpretation of geomorphic indicators forms the basis for landslide inventory development and guides the selection of locations for detailed field investigation.

Vegetation and drainage can provide additional evidence of recent or continuing movement. Tilted or curved tree trunks, damaged trees, exposed roots, disrupted drainage, localized seeps, and tension cracks may indicate deformation or reactivation. Curved trunks are particularly useful because trees may resume vertical growth after being tilted, preserving a visible record of earlier movement. These surface indicators are central to delineating landslide boundaries, distinguishing different stages of activity, and selecting trees for dendrogeomorphic analysis (Rotaru et al., 2007; Stoffel et al., 2013). The landslide inventory shown in **Figure 4.1** was developed through interpretation of these geomorphic

indicators using LiDAR-derived terrain models, historical aerial imagery, and extensive field verification.

Although the geomorphic indicators described above are characteristic of many landslides, the landslides along Shickshinny Mountain commonly exhibit atypical morphology. Many of the mapped landslides display well-defined headscarps and evidence of substantial displacement in their upper portions, yet lack clearly developed depositional toes. Instead, the lower portions of many failures are characterized by broad, bulging slope morphology with limited downslope translation. This geometry suggests that movement has been constrained near the base of the slope, likely due to confinement by adjacent landslide deposits, valley geometry, or resistance offered by intact bedrock and the narrow Susquehanna River floodplain. Consequently, many of the landslides appear to have accommodated deformation through internal compression rather than extensive runout, resulting in stacked or overlapping landslide complexes that preserve evidence of repeated reactivation.

Risk Management Framework

Effective landslide risk management requires consideration of both the likelihood of slope failure and the consequences should failure occur. The framework adopted for this investigation follows the approach outlined by Macciotta et al. (2016), in which risk is evaluated through four interconnected components (**Figure 4.4**):

1. **Hazard Assessment** — Identification and characterization of landslides, including their location, activity, magnitude, and potential triggering mechanisms.
2. **Consequence Analysis** — Evaluation of the potential impacts of slope failure on transportation infrastructure, utilities, river systems, ecosystems, and nearby communities.
3. **Risk Estimation** — Integration of landslide hazard with the severity of potential consequences to estimate overall risk.
4. **Risk Evaluation** — Comparison of estimated risk with acceptable thresholds to determine where monitoring, mitigation, or additional investigation is warranted.
- 5.

The research presented in this chapter applies this framework through the integration of landslide inventory development, dendrochronologic analysis, hydrologic modeling of potential Susquehanna River damming, and transportation network analysis. Together, these complementary approaches provide a comprehensive assessment of both landslide hazard and the potential societal impacts of future slope failures. The results of these investigations are presented at **Stop 4B**, where they are



Figure 4.4. Overview of landslide risk assessment. Adapted from Macciotta et al. (2015). Included in this image are research contributions towards the overall risk assessment.

discussed in the context of regional hazard mitigation and infrastructure resilience.

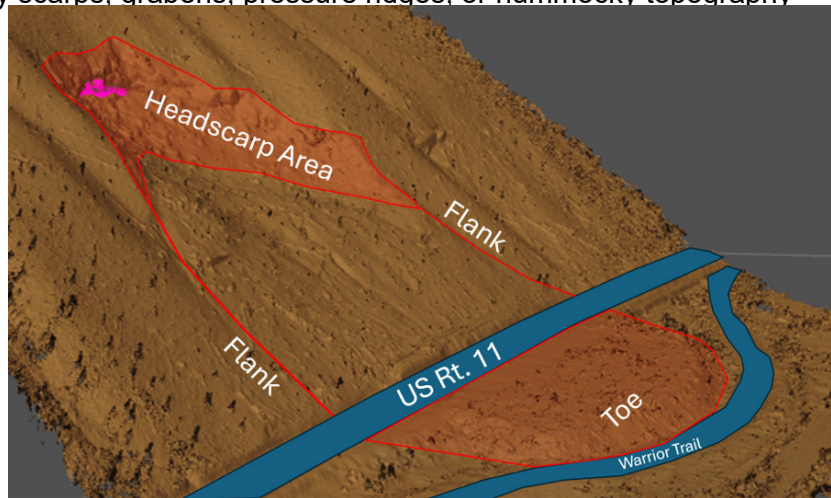
Field Stops

Stop #4A – Landslide Morphology and Slope Evolution Along Shickshinny Mountain

Leader: Bobak Karimi (Wilkes University)

The 1947 rock block slide represents the youngest documented landslide on Shickshinny Mountain and provides an excellent example of an early-stage rock block slide. Because of its relatively recent age, many of the geomorphic features associated with initial failure remain well preserved. The landslide is characterized by a distinct headscarp, clearly defined lateral flanks, and a small depositional toe (**Figure 4.5**). Unlike many larger landslides along the mountain, however, no readily identifiable internal geomorphic features such as secondary scarps, grabens, pressure ridges, or hummocky topography

Figure 4.5. Oblique perspective of a digital terrain model derived using photogrammetry of drone-captured imagery from a DJI phantom 4 pro v2.0, rendered using agisoft metashape software. Landslide features, US Rt. 11 and the Warrior Trail have been highlighted.



are present. This relatively simple morphology suggests that the landslide represents a comparatively young failure that has experienced little subsequent deformation or reactivation.

The 1947 landslide provides a useful reference for interpreting the broader landslide complex developed along Shickshinny Mountain. Although textbook landslides are commonly illustrated with well-developed headscarps, internal benches, and prominent depositional toes, the morphology of the Shickshinny Mountain landslides is considerably more variable. Many of the larger prehistoric landslides preserve impressive headscarps with tens of meters of displacement and laterally extensive flanks, yet lack clearly defined depositional toes. Instead, the lower portions of many landslides exhibit broad zones of compressional bulging where displaced material appears to have shortened rather than translated freely downslope.

In contrast to the relatively simple morphology of the 1947 landslide, many of the larger prehistoric landslides exhibit evidence of prolonged – and possibly repeated – deformation. These landslides commonly preserve well-defined headscarps and laterally extensive flanks, yet many lack the prominent depositional toes expected of fully developed rock slides. Instead, the lower portions of the landslides are characterized by broad zones of compressional bulging. This morphology suggests that deformation has been concentrated within the upper and middle portions of the landslide while movement near the toe has been comparatively limited. The reduced displacement at the toe likely reflects increased resisting forces resulting from a decrease in slope angle, greater confinement of the displaced material, and, in several locations, anthropogenic reinforcement of the slope through retaining walls constructed along U.S. Route 11 that further inhibit movement. Consequently, continued

movement is more readily expressed upslope through displacement at headscarps and localized internal deformation than by substantial advancement of the landslide toe.

Many of the larger landslides also contain nested failures, secondary scarps, and overlapping slide boundaries, indicating that slope evolution has occurred through multiple episodes of movement rather than a single catastrophic event. The largest landslide complex, for example, contains a younger landslide developed entirely within the older failure, demonstrating that reactivation and localized instability continue to modify the slope long after the initial failure. More on this phenomenon will be presented at Stop 4B.

Consequently, mapping landslides along Shickshinny Mountain requires more than identifying obvious depositional toes. Landslide boundaries were delineated using a combination of geomorphic indicators, including headscarps, lateral flanks, rotated and tilted blocks, compressional bulges, drainage disruptions, vegetation anomalies, and topographic expression. This integrated approach allowed recognition of landslides representing multiple stages of development and ultimately expanded the inventory beyond previously mapped failures (**Figure 4.1**).

Stop #4B – Evaluating Landslide Activity and Risk Along Shickshinny Mountain

Leader: Bobak Karimi (Wilkes University), Sean Papke

The largest landslide complex on Shickshinny Mountain provides an opportunity to examine how multiple complementary datasets can be integrated to evaluate landslide hazards. While previous investigations documented the size and geomorphic characteristics of this remarkable rock block slide (Braun and Gillmeister, 1989), several important questions remained unanswered regarding its internal structure, present-day activity, and potential impacts should renewed movement occur. Recent investigations combined geophysical surveys, dendrogeomorphic analysis, and hazard modeling to address the three primary questions introduced at the beginning of this chapter.

Question 1. Where are the active, dormant, and suspected landslides located, and what does their distribution reveal about slope evolution?

Field mapping established the distribution and geomorphic characteristics of nineteen mapped and suspected landslides along Shickshinny Mountain (**Figure 4.1**). To better understand their internal geometry, two electrical resistivity tomography (ERT) profiles and two Multichannel Analysis of Surface Waves (MASW) profiles were completed across the largest landslide, with one survey located near the headscarp and a second farther downslope. Together, these methods provided estimates of the depth and morphology of the slip surface, distinguished displaced material from intact bedrock, and constrained the thickness of the landslide mass. Estimated landslide volumes closely matched those previously calculated by Braun and Gillmeister (1989), providing independent geophysical support for the interpreted extent of the landslide. The geophysical data further indicate that the landslide remains a coherent rock block slide rather than a shallow surficial failure, with movement controlled by a deep-seated failure surface.

Question 2. When were these landslides initiated or reactivated, and what evidence exists for continued movement?

Although the largest landslide formed thousands of years ago, multiple lines of evidence indicate that portions of the Shickshinny Mountain slope remain active today. To reconstruct the timing of landslide movement, an eccentricity-based dendrochronological analysis was conducted across three representative landslides corresponding to pre-failure (LS 10), post-failure (LS 9), and syn-failure (LS 19) stages of development. A total of 126 increment cores were collected from 63 angiosperm trees growing on the landslides (**Table 4.2, Figure 4.6**). Analysis of annual growth eccentricity identified 60 disturbance events, including 14 dated landslide-related movement events between 1995 and 2022

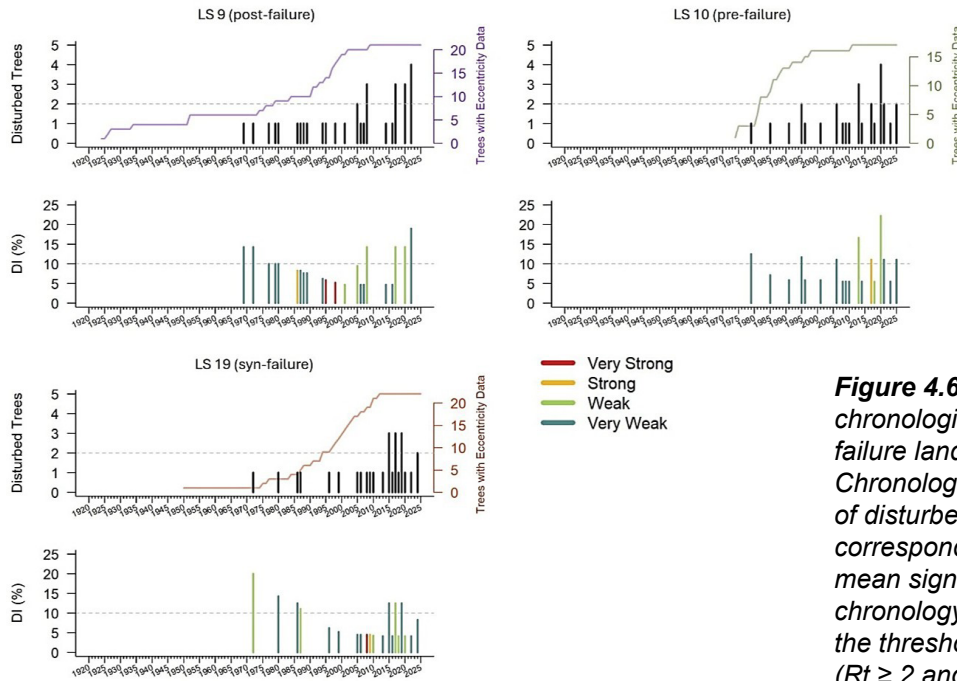


Figure 4.6. Disturbance signal chronologies for syn, post, and pre failure landslides from sampled trees. Chronologies include the number of disturbed trees, Rt (top) and corresponding disturbance index DI with mean signal strength (bottom) for each chronology. Grey dotted lines represent the threshold for landslide event signals ($Rt \geq 2$ and $DI \geq 10\%$). Sample depth (trees alive with eccentricity values) is shown for each landslide.

(Figure 4.7).

Table 4.2. Sampled Trees Within Three Landslides Selected for Dendrochronology.

Species	LS 9 (Post-Failure)	LS 10 (Pre-Failure)	LS 19 (Syn-Failure)	Total	Maximum Age (years)	Minimum Age (years)	Mean Age (years)*
<i>Acer plantanoides</i>	1	-	-	1	-	-	45
<i>Liriodendron tulipifera</i>	2	1	-	3	46	25	35 (SD: 10.5)
<i>Quercus alba</i>	1	-	-	1	-	-	42
<i>Quercus montana</i>	9	4	6	19	113	32	66 (SD: 23.5)
<i>Quercus rubra</i>	1	2	-	3	101	46	66 (SD: 30.1)
<i>Betula lenta</i>	6	11	18	35	92	15	38 (SD: 15.1)
<i>Fagus grandiflora</i>	1	-	-	1	-	-	31
<i>Acer saccharum</i>	-	-	1	1	-	-	26
Total	21	18	24	63	-	-	-

* If only one tree of a species was sampled, the age of that tree is reported in lieu of mean tree age.

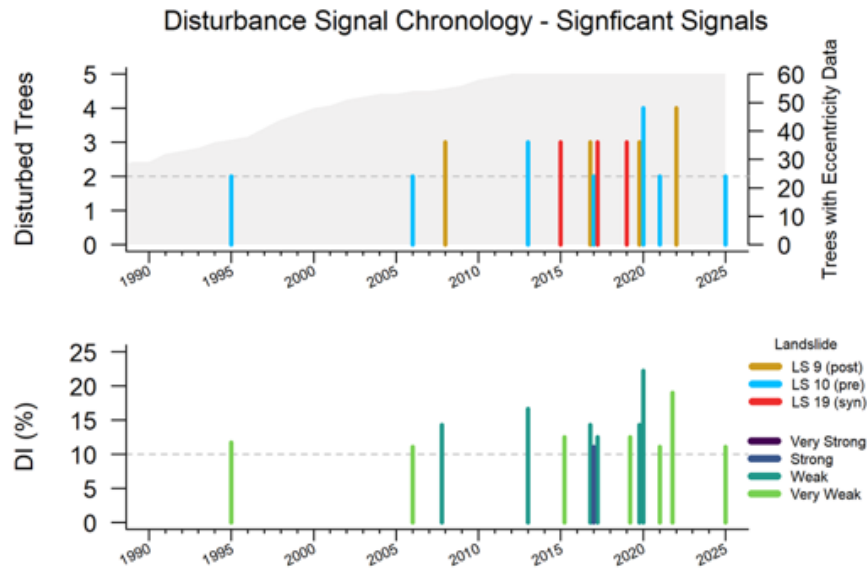


Figure 4.7. Disturbance Intensity across sampled landslides. Left to right: post-, pre-, and syn-failure landslides. Disturbance frequency (events per century, epc) interpolated with kernel density estimation. Increased disturbance frequency is concentrated around mass bulges and headscarp regions.

Distinct patterns of activity were observed among the three landslides. LS 9, representing a post-failure landslide, recorded four disturbance events (2008, 2017, 2020, and 2022), indicating continued reactivation of an older landslide. LS 10, classified as pre-failure, exhibited the greatest frequency of movement, with disturbance years identified in 1995, 2006, 2013, 2017, 2020, 2021, and 2022, suggesting progressive deformation prior to development of a major failure. In contrast, LS 19, representing the youngest syn-failure landslide, recorded only three disturbance years (2015, 2017, and 2019), which may reflect either comparatively lower activity or a more recent initiation of movement. Notably, disturbance events in 2017 were recorded on all three landslides, while 2020 was recorded on both LS 9 and LS 10, suggesting that landslide activity may respond synchronously to regional

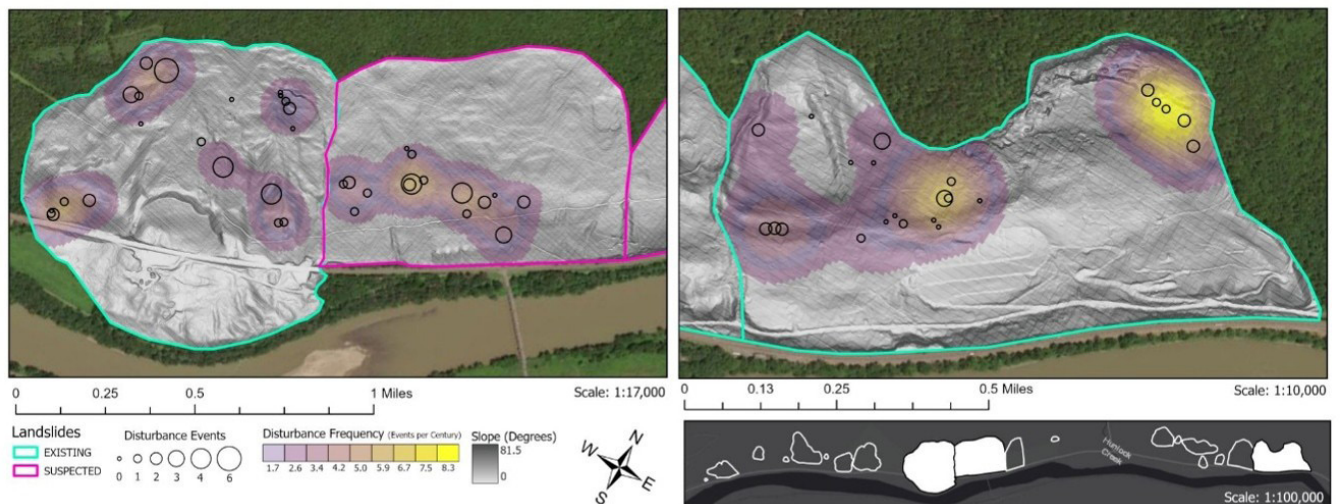


Figure 4.8. Landslide event signal chronology. A filtered chronology depicting significant signals meeting the threshold ($R_t \geq 2$ and $DIt \geq 10\%$) indicative of landslide event related disturbances. The number of disturbed trees (top) with labeling corresponding to the landslide the signal was detected in. Disturbance index DI (bottom) with labeling corresponding to mean signal strength.

hydrologic conditions rather than occurring solely as isolated slope failures.

Spatial analysis of disturbance frequency further supports this interpretation. Disturbance rates ranged from 1.7 to 8.3 events per century (epc) and were consistently concentrated near headscarps and zones of compressional bulging (**Figure 4.8**). These observations agree with field mapping presented at Stop 4A, where many landslides exhibit substantial displacement and repeated deformation near their upper margins but comparatively limited movement at their toes. The correspondence between tree-ring disturbances and these geomorphic features indicates that deformation remains concentrated in these portions of the landslide complexes and identifies them as areas most likely to experience continued movement.

While dendrochronology reconstructs the timing of past movement(s), it does not directly evaluate the

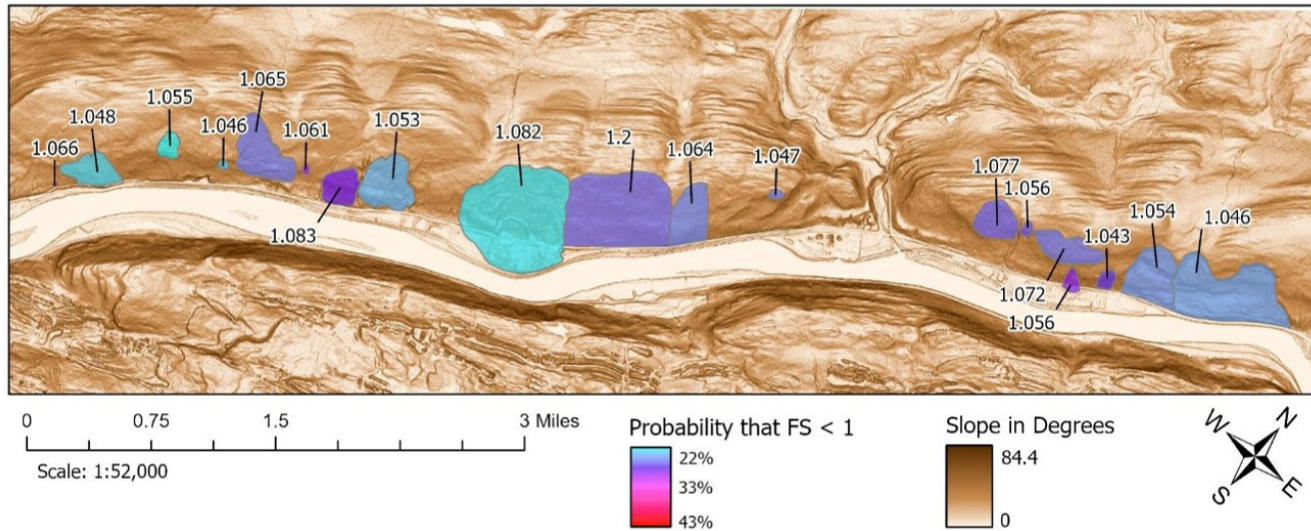


Figure 4.9. Factor of Safety Probability Map. Median FS value and probability of unstable slopes ($FS < 1$) are shown for each landslide at Shickshinny Mountain.

present-day stability of the slope. To address this, the slip-surface geometry interpreted from the ERT and MASW surveys was incorporated into probabilistic slope stability models to estimate the factor of safety (FoS) for each mapped landslide. Unlike traditional deterministic analyses that assume a single value for each parameter, this approach incorporated uncertainty in key geotechnical and hydrologic properties—including rock density, cohesion, friction angle, groundwater conditions, and slip-surface geometry—through Monte Carlo simulation. The resulting distributions of factor of safety indicate that many of the Shickshinny Mountain landslides exist in a state of marginal stability (**Figure 4.9**), with relatively modest increases in groundwater levels or pore-water pressure capable of reducing stability and promoting renewed movement. These findings are consistent with the dendrologic record, which suggests episodic reactivation during periods of elevated moisture rather than continuous slope movement.

Question 3. What are the potential consequences of future slope failure to nearby infrastructure,

the Susquehanna River, and surrounding communities?

Evaluating landslide hazards requires consideration of both the likelihood of failure and its consequences. Hydrologic simulations indicate that renewed failure of the largest landslide could partially, or wholly, obstruct the Susquehanna River, resulting in upstream inundation of floodplains and nearby communities within approximately 24 hours. Transportation analyses further demonstrate that blockage of U.S. Route 11 would substantially disrupt regional travel, increasing travel times by approximately 20–25 minutes and limiting access for emergency response, commercial transportation, and local residents. When considered together with the evidence for continued landslide activity and the large volume of material available for failure, these results demonstrate that Shickshinny Mountain presents a hazard extending well beyond the slope itself, emphasizing the importance of continued monitoring and risk-informed management.

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Return to Wilkes Barre Inn and Suites. End of Day 1

Day 2

Start: 08:00 AM : Wilkes-Barre Inn and Suites, 800 Kidder Street, Wilkes-Barre PA 18702
(41.25107, -75.84039)



Stop 5a: Luzerne Outcrop

(41.28796, -75.90527) Exit 6 of Route 309 North.

This stop will include an overview of the Luzerne outcrop stratigraphic section along the exit ramp from Route 309.



Stop 5b: Toby Creek Collapse and Remediation

(41.28796, -75.90379) Luzerne Lumber Company

Presentation on the April 2024 Toby Creek subsidence, subsequent solutions, and continued efforts.



Stop 5c: Toby Creek Geophysical and Hydrology Investigation

(41.28212, -75.89855)

Presentation on geophysical investigations (electrical and gravity), and likely reasons for subsidence development at Toby Creek. Presentation of maps of faults and existing sinkholes/subsidences.

Stop 5 – The Toby Creek Subsidence

Bobak Karimi¹

¹ Biology and Earth Systems Sciences, Wilkes University, Wilkes-Barre, PA, USA

This location includes three separate stops that will be visited by the entire group at the same time. There are short hikes along the paved (gravel), flat path of the Back Mountain Trail.

Introduction

The Wyoming Valley of northeastern Pennsylvania bears the legacy of more than a century of underground anthracite coal mining. Beneath much of the valley lies an extensive network of abandoned mine workings that continues to influence groundwater flow, surface drainage, and ground stability long after mining ceased. The interaction between natural geologic processes and these abandoned excavations has produced a variety of geologic hazards, including mine subsidence, stream loss, mine pool development, and localized flooding. Understanding these hazards is critical for managing land use and water resources throughout the region.

On April 14, 2024, these interactions became dramatically evident when a portion of Toby Creek collapsed behind Luzerne Lumber Company (41.288011°, -75.903346°) in Luzerne Borough, Luzerne County, Pennsylvania (**Figure 5.1**). The collapse produced a large whirlpool that rapidly diverted streamflow into abandoned underground mine workings, leaving the downstream channel temporarily dry (**Figure 5.2**) and prompting an emergency response by local agencies and the Pennsylvania Department of Environmental Protection (DEP, 2024; Mocarsky, 2024). Investigation of the collapse determined that the stream had breached shallow abandoned workings of the Mill Hollow Colliery within the Bottom Ross coal seam, approximately 18 m (60 ft) beneath the streambed (DEP, 2024). Although emergency remediation (**Figure 5.3**) restored surface flow, continued stream loss following the repair demonstrated that groundwater remained hydraulically connected to the underlying mine system (DEP, 2024).

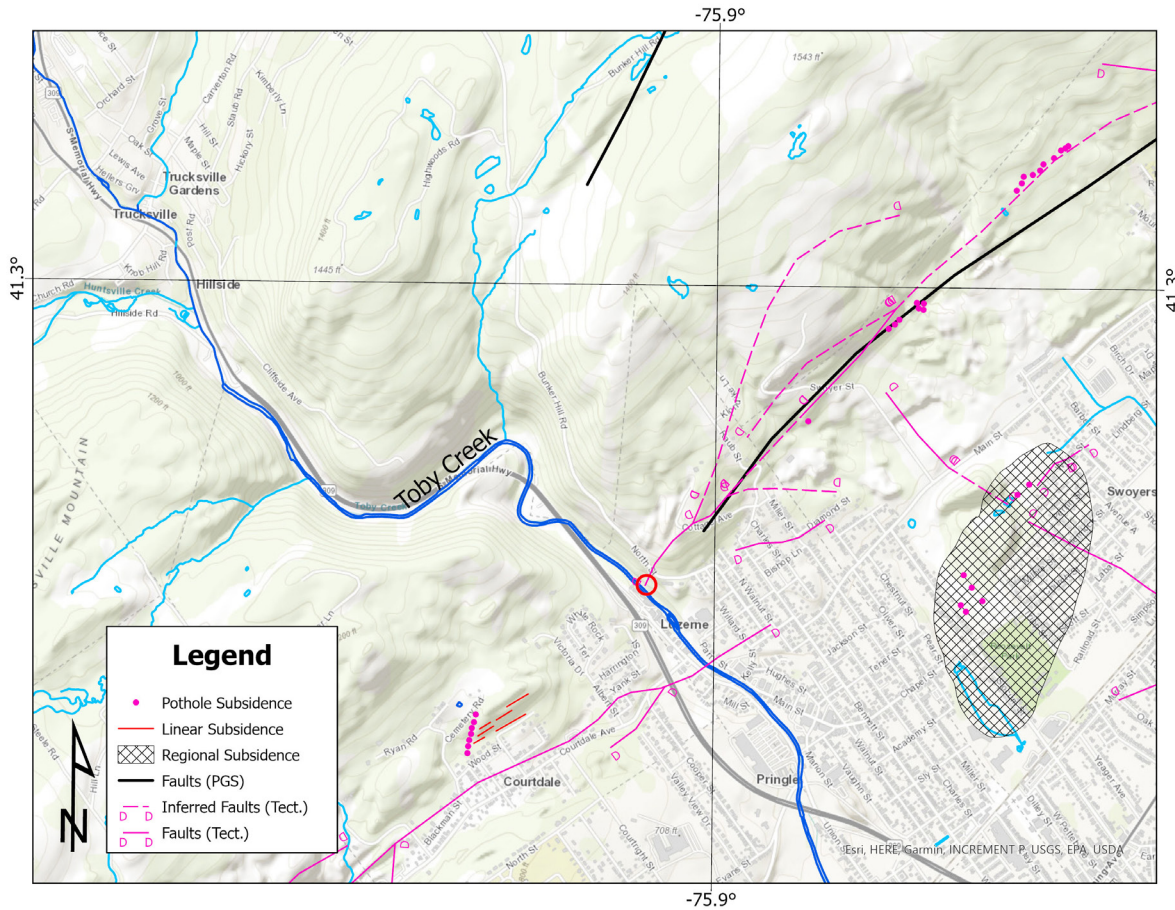


Figure 5.1. Map of Toby Creek with additional data on historical subsidences (pothole, linear, and regional) and faults from various historical datasets. The open red circle represents the location of the subsidence.



Figure 5.2. Mine subsidence and emergency remediation at Toby Creek on April 14, 2024. Note: Screenshot captured from the WNEP-TV news video Mine subsidence in Luzerne County (Kohut & Strub, 2024).



Figure 5.3. Photograph of Toby Creek mine subsidence work. Note: Photograph by J. Ardan, 2024. From “Water flow restored to Toby Creek in Luzerne after mine subsidence,” by S. Mocarsky, 2024, *Wilkes-Barre Citizens’ Voice*, <https://www.citizensvoice.com/2024/04/15/water-flow-restored-to-toby-creek-in-luzerne-after-mine-subsidence/>. Copyright 2026 by Wilkes-Barre Citizens’ Voice.

The Toby Creek subsidence illustrates how abandoned mine workings, fractured bedrock, and groundwater flow can interact to produce sudden and unexpected ground failure. While the collapse itself occurred within seconds, the conditions leading to failure developed over decades through the combined influence of natural geologic structure and historic mining. Consequently, the site provides an exceptional opportunity to examine the relationships among bedrock geology, anthropogenic modification of the subsurface, groundwater movement, and geologic hazards.

This chapter follows the progression of the 2024 investigation through three field stops. Stop 5A introduces the regional geologic and structural framework exposed at the Luzerne Outcrop and examines the geologic controls on groundwater movement. Stop 5B overlooks the site of the 2024 collapse and discusses the sequence of events, mining history, and factors that contributed to the subsidence. Stop 5C concludes at Toby Creek, where streamflow measurements, electrical resistivity surveying, and gravity surveys were used to investigate subsurface groundwater pathways and evaluate the potential for future subsidence.

Geologic Setting

Toby Creek drains approximately 94.5 km² (36.5 mi²) of the Back Mountain region before flowing southeast through Luzerne Borough and joining the North Branch Susquehanna River near Edwardsville (Hughes, 2020). Originating in the glaciated uplands northwest of the Wyoming Valley, the creek traverses a variety of geologic and land-use settings before entering the Northern Anthracite Coal Field. Unlike many watersheds within the Wyoming Valley, Toby Creek has historically maintained relatively good water quality and has not experienced the widespread impacts of abandoned mine drainage. Nevertheless, portions of the lower watershed remain susceptible to stream loss and hydrologic alteration where the creek intersects abandoned underground mine workings and structurally deformed bedrock (Hughes, 2020; Eastern Pennsylvania Coalition for Abandoned Mine Reclamation [EPCAMR], n.d.).

The field excursion is located near the southwestern margin of the Northern Anthracite Coal Field, where Pennsylvanian-age coal-bearing strata have been folded, faulted, and extensively mined for more than a century. The Luzerne Outcrop (Stop 5A) exposes three formations representative of the regional stratigraphic succession (**Figure 5.4**). From oldest to youngest, these include the Mauch Chunk Formation, Pottsville Formation, and Llewellyn Formation. Together, these units record the

transition from Late Mississippian terrestrial sedimentation to Pennsylvanian fluvial and coal-forming environments that characterize the anthracite region. The stratigraphic relationships shown in **Figure 5.4** provide the geologic framework for understanding the structural setting, distribution of coal seams, and hydrogeologic conditions examined throughout this field trip.

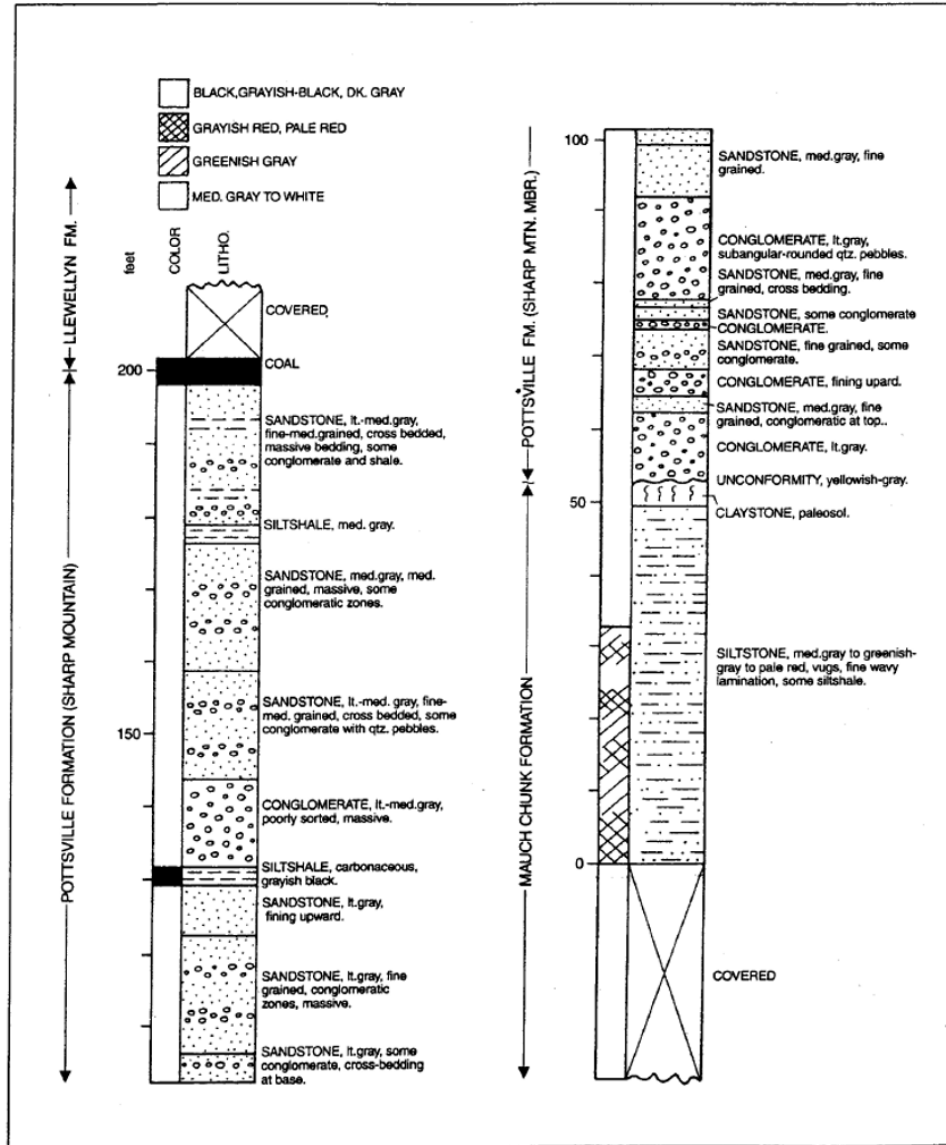


Figure 5.4. Stratigraphic column of the rock units exposed at the Luzerne Outcrop (Stop 5A). From oldest to youngest, the outcrop exposes the Mauch Chunk Formation, Pottsville Formation, and Llewellyn Formation. These units represent the regional stratigraphic framework of the Northern Anthracite Coal Field and provide the geologic context for the structural features, coal-bearing strata, and groundwater pathways discussed throughout this chapter. Adapted from Fig. 2 in Edmunds & Eggleston (1989). Note: Image taken from the 1997 guidebook for the Field Conference of Pennsylvania Geologists.

Regional deformation associated with Appalachian mountain building produced northeast-trending folds and numerous faults that strongly influence the modern landscape. At the Luzerne Outcrop, bedding dips moderately to the southeast and is displaced by several dip-slip faults. These structures, together with abundant joints and fractures, increase the secondary permeability of the bedrock and provide preferential pathways for groundwater movement. Consequently, groundwater flow is governed not only by lithology but also by the structural architecture of the bedrock.

Historic underground mining further modified this naturally fractured system. Room-and-pillar extraction within the Mill Hollow Colliery created an interconnected network of mine workings beneath portions of Toby Creek. At the 2024 subsidence site, abandoned workings within the Bottom Ross coal seam occur approximately 18 m (60 ft) beneath the modern stream channel (DEP, 2024). The interaction between fractured bedrock, fault-controlled groundwater flow, and abandoned mine voids created the hydrogeologic conditions that ultimately contributed to the April 2024 stream capture and subsidence event.

The three field stops presented in this chapter examine these relationships at progressively smaller scales. Stop 5A introduces the regional geologic framework exposed at the Luzerne Outcrop. Stop 5B examines the 2024 subsidence from an overlook that places the event within its geomorphic and mining context. Stop 5C concludes at Toby Creek, where streamflow measurements and geophysical investigations provide insight into the subsurface pathways responsible for stream loss and continued groundwater movement.

Field Stops

Stop #5A – Geologic Framework of the Toby Creek Subsidence Area

Leader: Bobak Karimi (Wilkes University)

The Luzerne Outcrop provides an excellent exposure of the stratigraphy and structural geology that underlie the Toby Creek watershed and the southwestern margin of the Northern Anthracite Coal Field. The outcrop illustrates many of the geologic features that govern groundwater movement beneath Toby Creek. Bedding orientation, lithologic contrasts, joints, and faults exposed here collectively influence the migration of groundwater toward the valley floor and ultimately into the abandoned underground mine workings responsible for the 2024 stream capture.

The exposure contains three formations representative of the regional stratigraphic succession (**Figure 5.2**). From oldest to youngest, these include the Mauch Chunk Formation, Pottsville Formation, and Llewellyn Formation. Together, these units document the transition from Late Mississippian terrestrial deposition to Pennsylvanian fluvial and coal-forming environments that characterize the Northern Anthracite Coal Field.

Stratigraphy

The lowest exposed unit is the **Mauch Chunk Formation**, which consists primarily of red shale, siltstone, and fine-grained sandstone. These rocks were deposited in fluvial and floodplain environments during the Late Mississippian and represent conditions that predate the widespread coal-forming swamps of the Pennsylvanian. The comparatively fine-grained lithologies weather readily and commonly produce gentler slopes than the overlying formations.

The **Pottsville Formation** conformably overlies the Mauch Chunk Formation and is composed predominantly of resistant quartz-rich sandstone and conglomerate with subordinate shale and thin

coal horizons. The increased abundance of coarse clastic sediments reflects deposition by high-energy fluvial systems during the early Pennsylvanian. Throughout northeastern Pennsylvania, the Pottsville Formation commonly forms prominent ridges and cliffs because of its resistance to weathering.

The uppermost unit exposed at the outcrop is the **Llewellyn Formation**, the principal coal-bearing formation of the Northern Anthracite Coal Field. Cyclic sequences of sandstone, shale, siltstone, claystone, and numerous anthracite coal seams record repeated changes in depositional environments associated with Pennsylvanian coastal plain and peat-forming swamp systems. Historic underground mining throughout the Wyoming Valley, including the Mill Hollow Colliery, targeted coal seams within this formation.

Structural Geology

The Luzerne Outcrop clearly demonstrates the structural deformation that accompanied Appalachian mountain building. Bedding strikes approximately northeast and dips moderately to the southeast, consistent with the regional structural grain of the Northern Anthracite Coal Field. This orientation reflects folding associated with the Appalachian orogeny and controls both the geometry of the coal-bearing strata and the movement of groundwater through the bedrock.

Several dip-slip faults offset the stratigraphic succession and are readily recognized by displaced bedding contacts, and smaller structural features within the exposure (**Figure 5.5**). When projected southwest across U.S. Route 309, this fault system is spatially coincident with the location of the April 2024 Toby Creek subsidence (**Figure 5.1**). Although the fault cannot be continuously traced beneath the valley floor, its alignment with the collapse suggests that structural discontinuities may have influenced groundwater movement and hydraulic connectivity between the surface and the underlying abandoned mine workings.

Figure 5.5. The Luzerne Outcrop showing formation contacts (red solid lines), bedding (red dashed lines), and some of the major faults (yellow solid lines).



Hydrogeologic Significance

The lithologic and structural relationships observed at the Luzerne Outcrop provide an important framework for understanding the 2024 Toby Creek subsidence. Groundwater movement within fractured sedimentary rocks is governed less by the primary porosity of the rock matrix than by secondary permeability associated with joints, bedding-plane partings, and faults. These discontinuities create preferential pathways that allow precipitation and stream water to migrate downward through the bedrock.

The spatial coincidence between the projected fault system exposed at the Luzerne Outcrop and the 2024 subsidence suggests that fault-controlled permeability may have focused groundwater flow toward the abandoned workings beneath Toby Creek. Rather than acting independently, fractured bedrock and legacy mine voids likely function as an integrated hydrogeologic system in which structural discontinuities convey water to the underlying mine network. Concentrated groundwater flow along these pathways may promote internal erosion, sediment transport, and progressive weakening of the overlying materials, increasing the likelihood of localized subsidence where shallow mine workings are present.

Although additional subsurface investigation would be required to confirm the continuity and hydraulic behavior of the fault at depth, the exposed structures at the Luzerne Outcrop provide a compelling geologic analogue for understanding the processes responsible for the 2024 stream capture.

Stop #5B – The 2024 Toby Creek Subsidence: Causes, Impacts, and Mine Controls

Leader: Bobak Karimi (Wilkes University), David McGovern (DEP-BAMR)

From this overlook, participants can view Toby Creek, the location of the April 2024 subsidence, and the Luzerne Outcrop visited during Stop 5A (**Figure 5.1**). This elevated perspective places the collapse within its broader geologic setting and illustrates the relationship between the mountain front, the valley floor, and the historic underground mine workings beneath Toby Creek.

Perhaps the most significant observation from this location is the alignment between the structural features exposed at Stop 5A and the location of the subsidence. The dip-slip fault system observed within the Luzerne Outcrop projects southwest across U.S. Route 309 toward the collapse site. Although the continuity of the fault beneath the valley floor cannot be directly observed, its projection suggests that the same structural framework extends beneath Toby Creek and may influence groundwater movement through the subsurface.

Structural & Mining Controls

Now that participants have observed the outcrop, the relationship between geology and the 2024 subsidence becomes more apparent. The southeast-dipping strata and associated dip-slip faults exposed at Stop 5A increase the secondary permeability of the bedrock by creating interconnected pathways for groundwater movement. Beneath the valley floor, these natural structural features intersect abandoned room-and-pillar workings of the Mill Hollow Colliery, producing a hydrogeologic system in which groundwater can move between fractured bedrock and underground mine voids.

The coincidence between the projected fault system and the location of the 2024 collapse suggests that groundwater flow may be structurally focused before entering the abandoned mine network. Although mining created the void into which Toby Creek drained, the structural architecture of the bedrock may have influenced where groundwater was concentrated and why the collapse developed at this particular location.

Rather than viewing the subsidence as solely a mining-related feature, it is more appropriately interpreted as the product of interacting natural and anthropogenic processes.

Landscape Interpretation

This overlook also highlights the geomorphic relationship between the mountain front and Toby Creek. Surface runoff originating on the uplands is directed toward the valley through fractured bedrock before entering Toby Creek. Where faults and fractures intersect shallow underground mine workings, groundwater and surface water become hydraulically connected, allowing streamflow to enter the abandoned mine network.

Although the stream capture itself is no longer visible, the landscape viewed from this stop illustrates how geologic structure, topography, and historic mining combine to control groundwater movement beneath the valley.

The structural relationships observed at Stops 5A and 5B provide a conceptual model for groundwater movement beneath Toby Creek. However, the abandoned mine workings and groundwater pathways responsible for the 2024 subsidence cannot be directly observed from the surface. At Stop 5C, streamflow measurements, electrical resistivity tomography, and gravity surveys are used to investigate these concealed subsurface conditions and evaluate the potential for future subsidence.

Stop #5C – Investigating the Subsurface: Streamflow, Electrical Resistivity, and Gravity Surveys

Leader: Bobak Karimi (Wilkes University), Anneke Bouma, Joey Earl, Emma Klusmeyer

The previous two stops established the geologic framework of the Toby Creek subsidence and demonstrated how bedrock structure and historic underground mining combine to influence groundwater movement. However, the processes responsible for the April 2024 collapse occur almost entirely beneath the ground surface and cannot be directly observed. At this stop, geophysical investigations and streamflow measurements are used to evaluate the hidden hydrogeologic system beneath Toby Creek and assess the potential for future subsidence.

Multiple complementary methods were employed to characterize groundwater movement and subsurface conditions. Streamflow measurements quantified the magnitude of water loss within Toby Creek, electrical resistivity tomography (ERT) imaged resistivity variations in subsurface moisture or lithology, and gravity surveys identified density variations potentially associated with abandoned mine workings. Together, these datasets provide a more complete understanding of the hydrogeologic system than any individual method alone.

How Does Toby Creek Compare to Other Wyoming Valley Streams?

The 2024 subsidence raised an important question: is stream loss at Toby Creek an isolated occurrence, or is it representative of a broader hydrogeologic pattern throughout the Wyoming Valley? To place Toby Creek in a regional context, stream discharge measurements were collected at multiple locations along nine streams throughout the Wyoming Valley (**Figure 5.6**). Survey sites were selected to evaluate changes in discharge upstream and downstream of mapped faults, abandoned underground mine workings, or both. Streams were grouped into three categories: (1) streams without known intersections with mapped faults or mine workings, (2) streams intersecting mapped faults only, and (3) streams intersecting both faults and abandoned mine workings (**Table 5.1**).

Table 5.1. Percent stream loss measured along surveyed streams categorized by the presence of mapped faults and abandoned underground mine workings.

Possible Stream Loss Causes and Percent Change in Discharge					
No Cause/No Loss		Fault		Mine/Fault	
Abraham's Creek*	21%	Laurel Run	-42%	Mill Creek	-100%
Harvey's Creek	37%	Spring Brook**	-40%	Toby's Creek	-19%
Hunlock Creek	36%			Nanticoke Creek	-66%
				Solomon Creek	-36%
*Need additional downstream points.					
**Several inferred faults nearby that may intersect w/ Spring Brook.					

Streams lacking obvious structural or mining influences gained flow between measurement locations due to groundwater inputs and tributary inflows. In contrast, streams intersecting mapped faults commonly experienced measurable stream loss, averaging approximately 40% between survey locations. The greatest losses occurred where streams crossed both mapped faults and abandoned underground mine workings, with measured losses ranging from 19% to complete stream capture (100%).

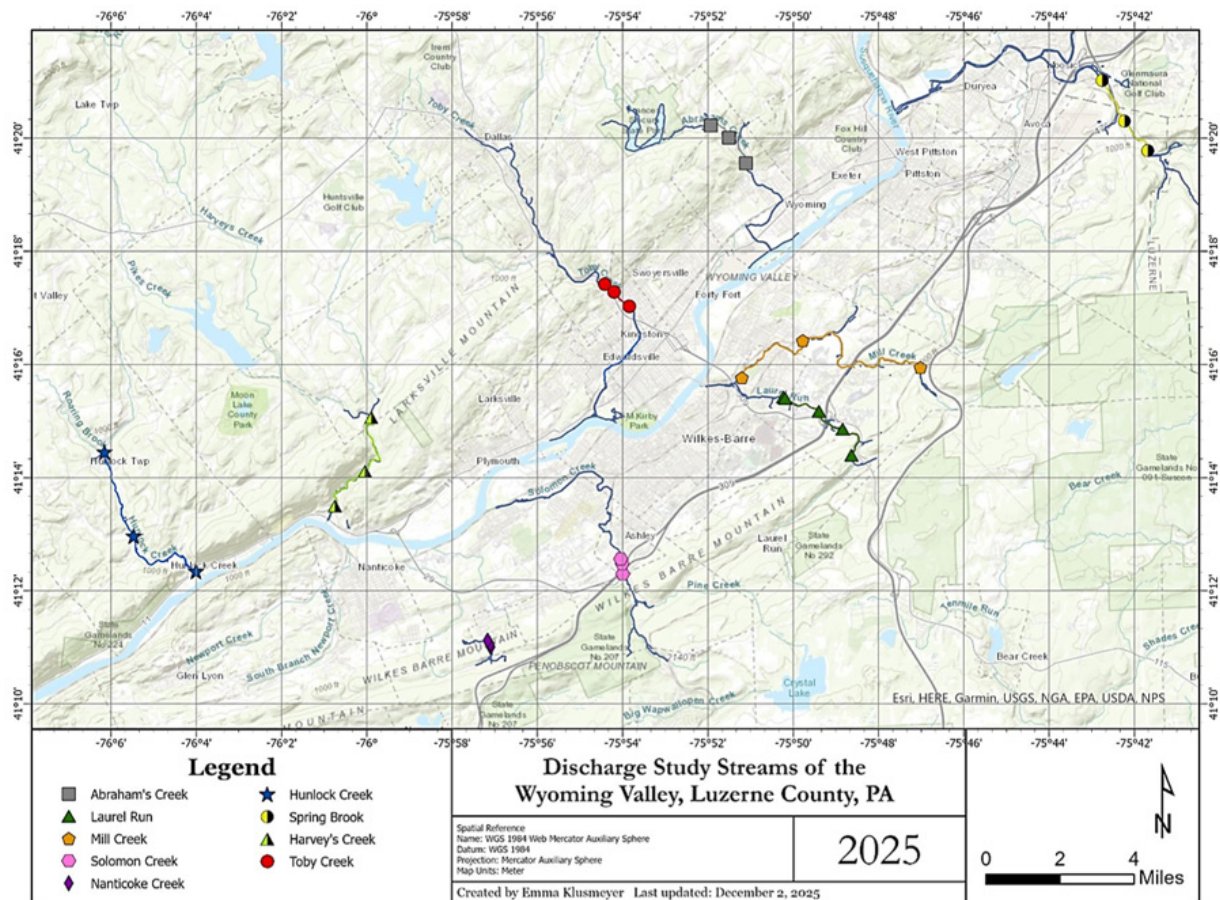


Figure 5.6. Map showing streamflow measurement locations throughout the Wyoming Valley.

Several streams illustrate the range of observed responses. Mill Creek exhibited complete loss of streamflow where it intersected abandoned mine workings and structural features, while Nanticoke Creek and Solomon Creek experienced discharge reductions of approximately 66% and 36%, respectively. Following emergency remediation, Toby Creek exhibited an overall discharge reduction of approximately 19% across the study reach. However, discharge measurements collected immediately upstream and downstream of the 2024 subsidence documented a localized stream loss of approximately 36%. Streamflow partially recovered farther downstream through inflow from tributaries, stormwater infrastructure, and other anthropogenic drainage inputs, resulting in a net reduction substantially smaller than the localized loss associated with the collapse.

These observations suggest that faults alone may promote groundwater infiltration by increasing secondary permeability within the bedrock. However, where faults intersect abandoned underground mine workings, stream losses become substantially greater, indicating that mine voids provide an efficient subsurface conduit capable of capturing and transporting surface water. Within this regional framework, Toby Creek behaves consistently with other streams affected by both structural discontinuities and legacy mining. The 2024 subsidence therefore represents a localized manifestation of a broader hydrogeologic process occurring throughout the Northern Anthracite Coal Field.

Where Does the Water Go?

Although streamflow measurements demonstrate that water is lost from Toby Creek, they do not reveal the pathways by which it enters the subsurface. To better understand groundwater movement following the 2024 subsidence, electrical resistivity surveys were conducted along two transects within the study area (**Figure 5.7**). One survey was completed parallel to Toby Creek and centered on the subsidence site, while a second was acquired along the base of the Luzerne Outcrop. Together, these profiles provide insight into the structural controls governing groundwater movement between the mountain front and the abandoned underground mine workings.

The resulting electrical resistivity models reveal distinct zones of reduced resistivity that coincide with mapped fault systems and major fracture zones. Rather than occurring randomly within the subsurface, these conductive features follow structurally controlled pathways that are consistent with groundwater infiltration through fractured bedrock. Several of the interpreted flow paths are aligned with faults that are subparallel to bedding, while others appear to exploit lithologic contacts between formations where permeability contrasts may further promote groundwater movement (**Figure 5.7**).

The two survey lines provide complementary perspectives on the same hydrogeologic system. The profile collected along Toby Creek illustrates groundwater movement beneath the stream and through the vicinity of the 2024 subsidence. In contrast, the profile collected along the base of the Luzerne Outcrop demonstrates that the structural features observed at Stop 5A continue to influence groundwater flow beyond the exposed bedrock. Together, these surveys suggest that groundwater infiltrates along structurally controlled pathways extending from the mountain front toward the valley floor, where the pathways intersect abandoned mine workings beneath Toby Creek.

These observations support the conceptual model developed throughout the field excursion. The dip-slip faults observed at the Luzerne Outcrop (Stop 5A) project toward the subsidence site (Stop 5B), and the resistivity models indicate that these same structural discontinuities likely function as preferential groundwater pathways. Rather than flowing uniformly through the bedrock, groundwater appears to exploit existing faults, fractures, and lithologic contacts before entering the abandoned mine network. This interaction between natural geologic structure and legacy underground mining provides a plausible explanation for the localized stream loss observed during the 2024 subsidence.

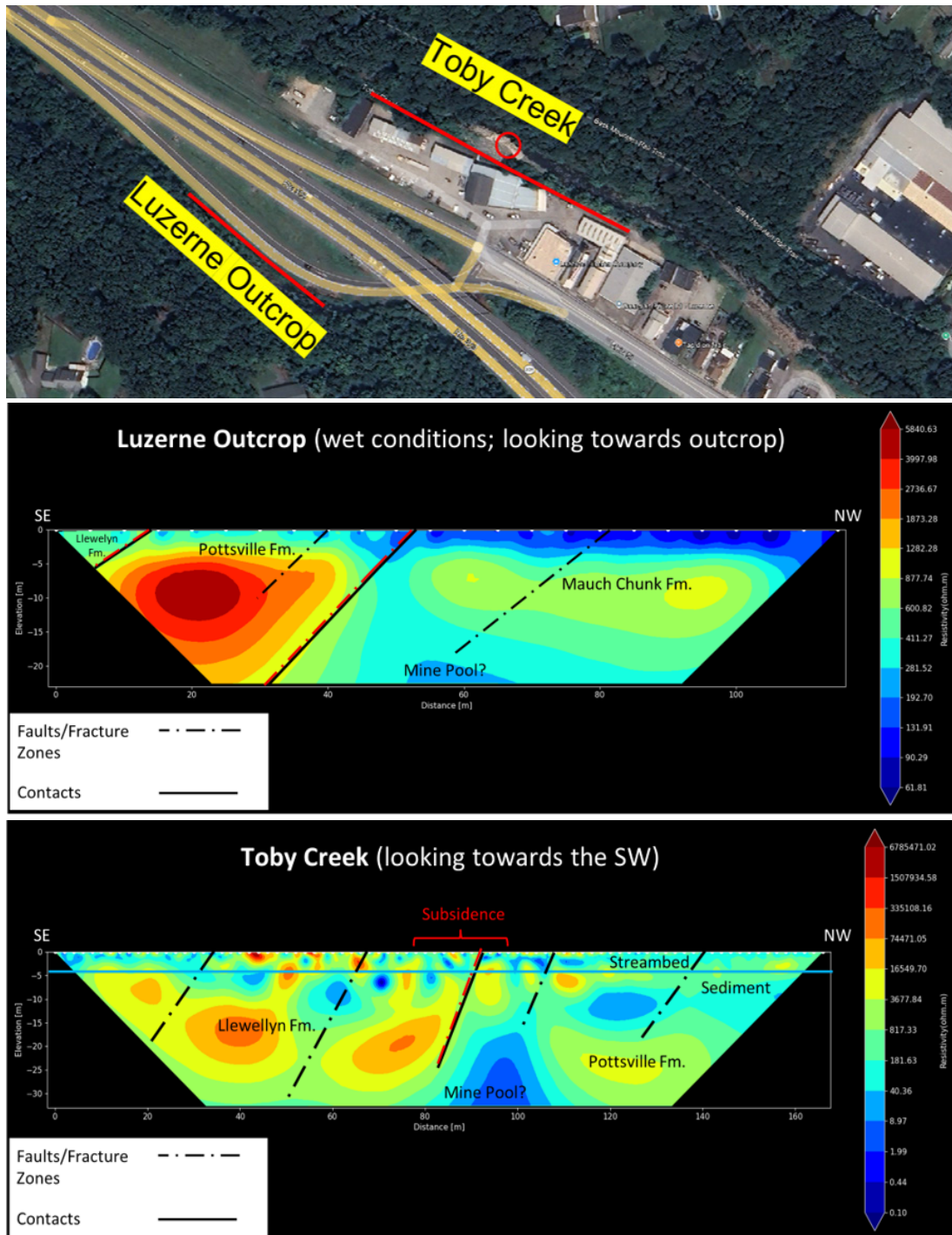


Figure 5.7. Electrical survey profiles and tomographic images. (A) Map of the two electrical resistivity survey lines conducted. Below are representative ERT profiles interpreted alongside mapped geology and the projected fault systems. (B) Line parallel to the Luzerne Outcrop and (C) paralleling Toby Creek.

Evaluating Future Subsidence Potential

Although the electrical resistivity surveys provide evidence for structurally controlled groundwater pathways, they do not directly evaluate the distribution of subsurface density variations that may be associated with abandoned mine workings or weakened overburden. To complement the resistivity investigation, gravity measurements were collected throughout the study area, including along the two electrical resistivity survey lines (**Figure 5.8**).

Gravity profiles constructed along both survey transects are consistent with the interpretations developed from the resistivity data. Variations in the gravity response correspond with major structural features and zones interpreted as groundwater pathways, providing an independent line of evidence that supports the conceptual model of groundwater migration through fractured bedrock toward the abandoned mine

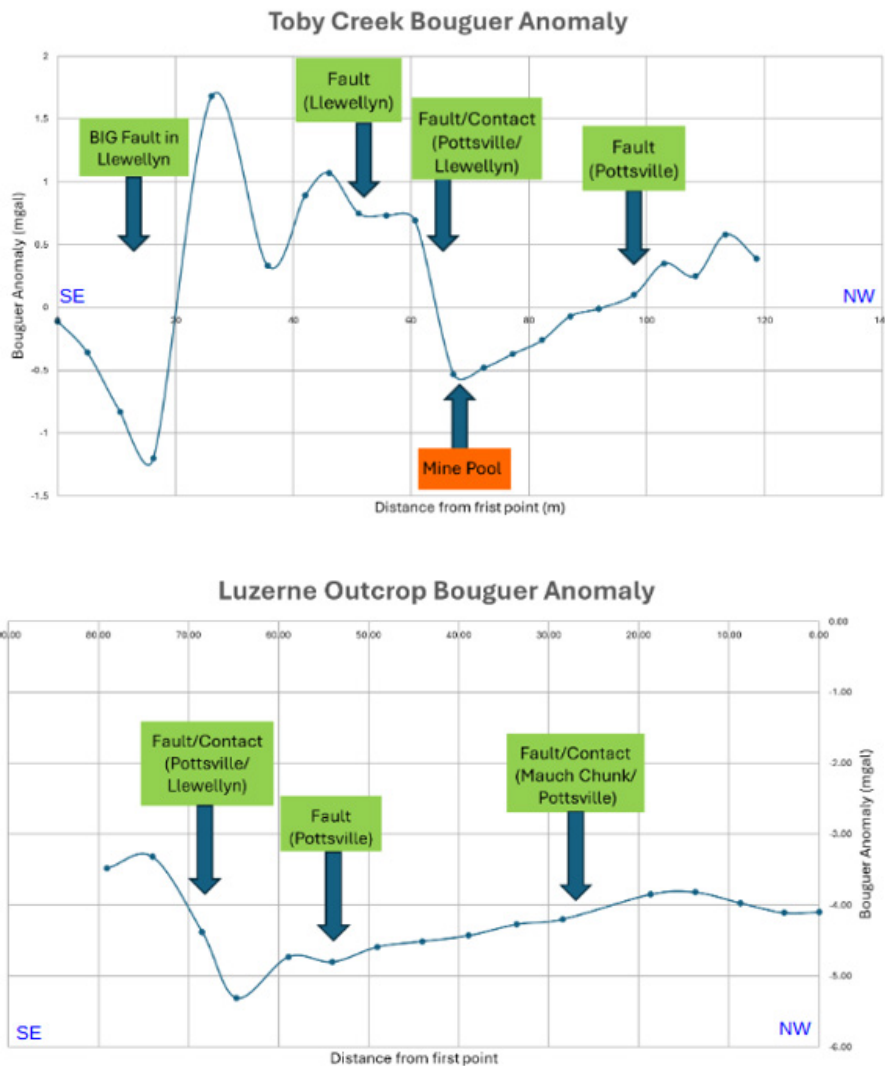


Figure 5.8. Gravity profiles collected along the two electrical resistivity survey lines showing the correspondence between gravity anomalies and interpreted structural features identified from the resistivity models.

system. The close agreement between the gravity and resistivity datasets increases confidence that the observed geophysical anomalies reflect meaningful subsurface features rather than artifacts associated with a single survey method.

At present, however, the gravity survey remains incomplete. Although the existing profiles provide valuable constraints along the surveyed transects, additional gravity measurements are needed to improve spatial coverage between profiles and better define the lateral extent of the interpreted subsurface features. As the dataset expands, gravity modeling will help refine the geometry of low-density zones that may be associated with abandoned mine workings, collapsed material, or other subsurface conditions influencing groundwater flow and ground stability.

While the current gravity dataset does not independently identify areas of future subsidence, it corroborates the structural and hydrogeologic interpretations derived from the electrical resistivity surveys. Collectively, the geophysical investigations suggest that groundwater movement beneath Toby Creek is strongly influenced by the interaction of faults, fractures, lithologic contacts, and abandoned underground mine workings. Continued integration of gravity data with electrical resistivity and streamflow measurements will improve understanding of the subsurface architecture and aid in identifying areas where similar conditions may increase susceptibility to future subsidence.

Integrated Interpretation

The three field stops presented in this chapter illustrate how geologic observations collected at different scales can be integrated to better understand the 2024 Toby Creek subsidence. The Luzerne Outcrop (Stop 5A) establishes the regional geologic framework, where southeast-dipping Pennsylvanian strata are disrupted by several dip-slip faults and fracture systems that increase the secondary permeability of the bedrock. From the overlook at Stop 5B, these same structural features can be projected across the landscape toward the location of the 2024 collapse, suggesting that the structural framework observed at the outcrop extends beneath the valley floor.

Regional streamflow measurements demonstrate that Toby Creek is not an isolated example of stream loss within the Wyoming Valley. Streams intersecting mapped faults commonly experience measurable reductions in discharge, while the greatest losses occur where faults coincide with abandoned underground mine workings. These observations indicate that geologic structure alone can promote groundwater infiltration, but the combination of faults and legacy mining substantially enhances the transfer of surface water into the subsurface.

Electrical resistivity surveys further refine this conceptual model by identifying conductive zones that are consistent with groundwater movement along structurally controlled pathways. The resistivity profiles suggest that groundwater preferentially exploits faults, fractures, and lithologic contacts before entering the abandoned mine system beneath Toby Creek. Gravity profiles collected along the resistivity transects independently support these interpretations, although additional gravity measurements are needed to better constrain the geometry and extent of the subsurface features responsible for groundwater movement.

Collectively, these observations indicate that the 2024 Toby Creek subsidence cannot be attributed solely to abandoned mine workings or solely to natural geologic structure. Rather, the collapse represents the interaction of fractured and faulted bedrock, shallow room-and-pillar mine workings, and groundwater flow within a complex hydrogeologic system. Although developed in response to a single subsidence

event, the investigative framework presented here—including regional streamflow comparisons, structural geologic mapping, and integrated geophysical surveys—provides a transferable approach for evaluating stream loss, groundwater pathways, and subsidence hazards throughout the anthracite region.

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Stop 6: Pocono Raceway - Surficial Geological Mapping of the Blakeslee Quadrangle

(41.05301, -75.51573) Pocono Race Track, Long Pond PA

Discussion of the Surficial Geological Map of the Blakeslee 7.5-Minute Quadrangle. Presentation in the Main Grand Stand overlooking the terminal moraine complex of the late-Wisconsinan advance



Pocono Raceway

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This location includes lunch and a presentation at the 300-level (Terrace Club) of the Main Grain Stand at the Pocono Raceway. The stop involves a strenuous walk to the upper part of the Main Grain Stand to observe the raceway and surrounding area.

Introduction

The Pocono Raceway is located in the village of Long Pond in northeastern Pennsylvania (**Figure 1**). The site is located within the Blakeslee 7.5-Minute quadrangle, which straddles the late-Wisconsinan age terminal moraine complex of the Laurentide Ice Sheet. The raceway and surrounding grounds are located south of the moraine complex, in a low relief landscape mapped as Illinoian age till (Sevon & Braun, 1997). An updated surficial geology map of the Blakeslee 7.5-Minute quadrangle was recently created by Wilkes University faculty (Matt Finkenbinder and Bobak Karimi) and students (Sam Winter and Faith Morcombe) and funded by a U.S. Geological Survey EDMAP grant (Winter et al., 2025). The field stop will include presentation of the map and supplemental datasets including grain size analysis of coarse grained (sand and gravel) glacial deposits and radiometric dating of select glacial deposits. The Main Grain Stand at the Pocono Raceway will serve as our vantage point to observe the surrounding landscape and consider the evidence for ice sheet glaciation of the region (**Figure 2**). In addition, Pocono Raceway staff will provide an overview of the development of the track facility and a history of track paving.

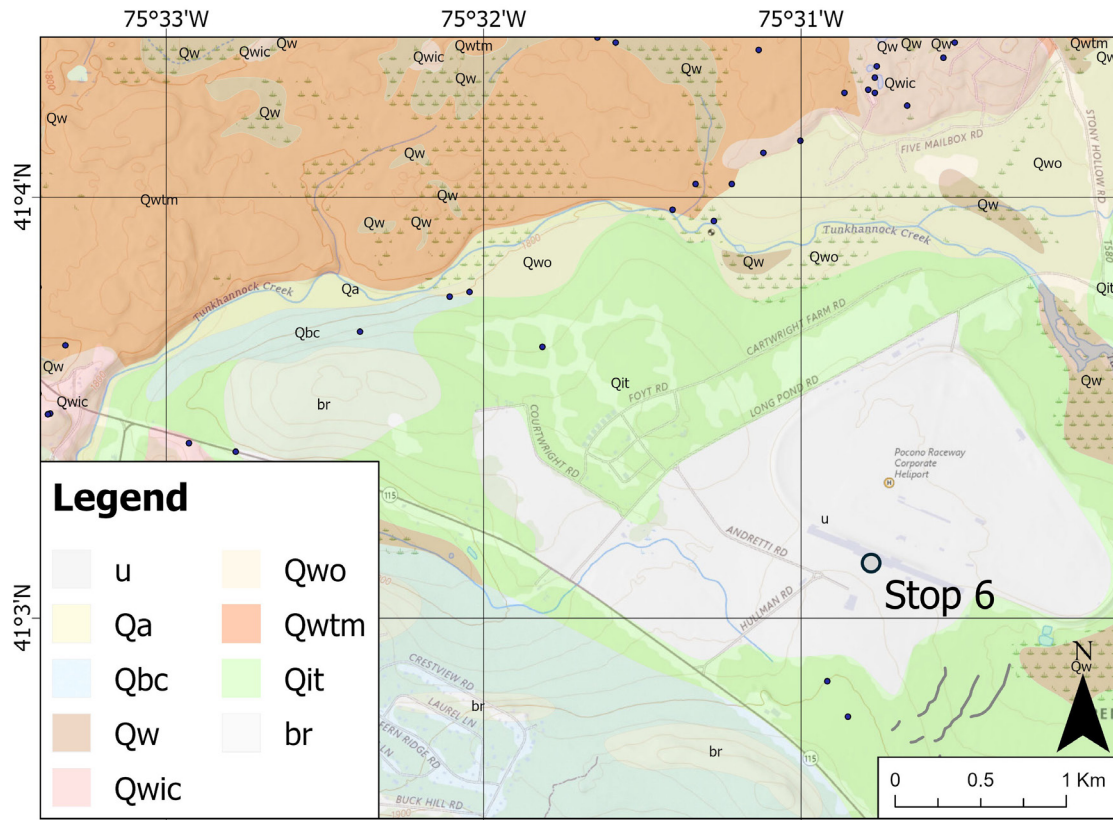


Figure 1. Surficial geologic map of the Pocono Raceway (Stop 6) and surrounding area showing the distribution of glacial sediments and landforms. Surficial geology units include urban land (u), alluvium (Qa), boulder colluvium (Qbc), wetlands (Qw), ice-contact stratified sand and gravel (Qwic), outwash (Qwo), Wisconsinan end moraine (Qwtm), Illinois till (Qit), and bedrock (br) (Winter et al., 2025). Black dots are field check and/or sample collection locations. The gray lines south of the Pocono Raceway represent gelifluction lobes.

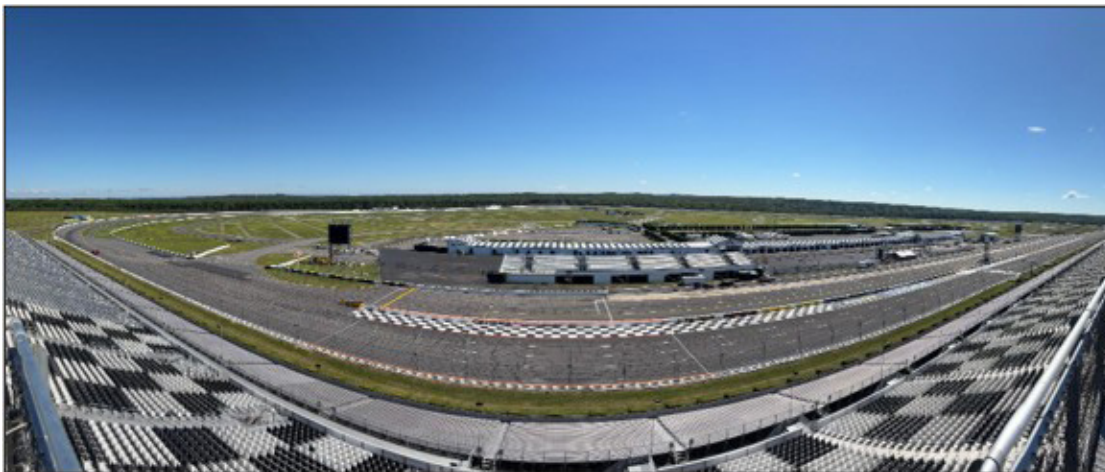


Figure 2. Panoramic photograph of the Pocono Raceway from the Spotters Stand of the Main Grain Stand looking across the frontstretch. The Fern Ridge, or late-Wisconsinan age terminal moraine complex, defines the maximum extent of ice sheet glaciation in the region and is visible on the opposite side of the track.

Field Stop

Stop #6 – The Pocono Raceway Main Grain Stand (41.0525° N, 75.5136° W)

Surficial Geology Mapping of the Blakeslee 7.5-Minute Quadrangle (Winter et al., 2025)

Leader: Matt Finkenbinder, Sam Winter, and Faith Morcombe (Wilkes University)

The stop provides an unprecedented view of the Pocono Raceway and surrounding landscape (**Figure 2**). The vantage point offers a spectacular view of the Fern Ridge, or late-Wisconsinan age terminal moraine complex, to the north and beyond the Tunnel Turn. The moraine complex is forested and largely undeveloped, owing to its hummocky topography, till sedimentology, and abundance of ponds and wetlands, which makes it rather unsuitable for large-scale development or for agricultural use. Areas located to the south, including the area surrounding the raceway facility, are underlain by older glacial sediments from a more extensive episode of ice sheet glaciation. The Blakeslee 7.5-Minute quadrangle has been the focus of recent surficial geology mapping efforts by Wilkes University faculty and students (Winter et al., 2025), and the raceway serves as a venue to present this research.

The Blakeslee 7.5-Minute Quadrangle straddles a portion of the terminal moraine complex of the late-Wisconsinan age (17,000 to 22,000 years old) (Crowl & Sevon, 1980) advance and contains a well-preserved geomorphic record of the advance and retreat of the Laurentide Ice Sheet from the region. Previous mapping work in the study area began with Sevon (1975), who mapped the distribution of major landforms and sediment deposits at 1:24,000 scale on the basis of field mapping and 7.5' topographic map/aerial photography interpretation in the early 1970's. Sevon (1975) mapped the late-Wisconsinan terminal moraine primarily on the basis of hummocky, knob and kettle topography. Subsequently, Braun (2019) updated and revised the surficial geological map of the Blakeslee Quadrangle at 1:24,000 scale in the late 1990's on the basis of additional field mapping (completed in 1997), 7.5-Minute topographic map/aerial photography interpretation, and by using county soil surveys to map sand and gravel deposits. Braun (2019) later reviewed lidar hillshade maps (acquired in 2005-2007) and aerial photographs, and concluded that areas of moraine topography, bedrock ledges, and smooth slopes underlain by boulder colluvium could be quite accurately mapped, however no field checking was completed for verification. Additional changes to map unit contacts were made prior to submission of the draft map to the Pennsylvania Geological Survey (PaGS) in 2019.

While field-based confirmations were conducted, they were few in number (Sevon, 1975) or not included (Braun, 2019) in products by these prior mappers. As a result, the existing maps lack sufficient metadata regarding field check locations and the confidence of mapped deposits and landforms (i.e. existing maps are not U.S. Geological Survey Geologic Map Schema or GeMS compliant). Furthermore, none of the previous studies generated age constraints on the glacial landforms and/or sediments using absolute dating techniques, and therefore the age of the late-Wisconsinan terminal moraine complex and the timing of ice sheet deglaciation is unknown. The overarching purpose and justification to create a new surficial geological map of the Blakeslee 7.5-Minute Quadrangle was motivated by 1) rising population in northeast Pennsylvania over the last several decades, 2) limitations and errors found in previously published maps which limit their reliability, 3) the need to fill a critical spatial data gap of surficial geological materials mapped using a consistent set of map units and landforms, 4) the availability of recently acquired, high-resolution geospatial datasets including Quality Level 2 lidar, and 5) the general lack of absolute age control of glacial deposits in northeast Pennsylvania.

Geospatial Analysis Methods: The surficial geological map was initially generated using geospatial analysis of high-resolution topographic datasets (and derivatives) and aerial photography, using previously published surficial maps of the study area by Sevon (1975) and Braun (2019) as a guide. A standard set of surficial units and names was used for mapping and to ensure regional consistency (Johnson & Finkenbinder, 2023). The surficial units include till moraine, till, glacial outwash, ice-contact stratified drift, alluvium, boulder colluvium, swamp/wetland/peat, open water, bedrock, landslides, and artificial fill. Sub-meter resolution Quality Level 2 (QL2) lidar topographic data was used to generate primary base maps and datasets (contours, hillshade, digital slope model, and surface roughness). Aerial photography, acquired by the National Agriculture Imagery Program in 2022, was used as another primary basemap.

The high-resolution basemaps and additional datasets were used to aid in visualization of the land surface for mapping. The Late-Wisconsinan age terminal moraine complex and associated deposits were initially mapped on the basis of hummocky, knob and kettle topography from the hillshade map. Bedrock exposures were mapped using a combination of aerial photographs and the DSM base map by notion areas of high local relief, which is related to steep cliffs formed on resistant Devonian age sandstone and/or conglomerate ledges. Modern floodplains and alluvium, open water, swamps, and peat bogs were mapped using aerial photographs and hillshade maps. Late-Wisconsinan age till deposits, which likely blanket most of the landscape north of the terminal moraine between bedrock exposures and valley bottoms, were subsequently mapped. Coarse grained glacial deposits (sand and gravel) were mapped using parent material data from county soil survey maps (Lipscomb, 1981; Fisher et al., 1962), morphostratigraphic relationships interpreted from hillshade maps, field observations, and geologic data from the Pennsylvania Groundwater Information System. Illinoian age glacial deposits were mapped on the basis of landscape position and subdued, hummocky topography from lidar hillshade maps. Landslides were mapped from hillshade maps based on prominent headscarps, hummocky toe slope deposits and/or solifluction lobate form through geospatial analysis of topographic datasets (Karimi et al., 2019).

Field Mapping Methods: Field mapping and checking of map units involved observations of surficial landforms, sediment deposits, and bedrock exposures for confirmation or re-interpretation. Field observations were recorded in hardbound notebooks and included the following: coordinate location, description of the landscape/landforms, sediment description (if relevant), and any sample collection data. Field observations were used to confirm the validity of mapped units or to make re-interpretations and adjustments.

Grain Size Analysis Methods: Nineteen (19) sand and gravel samples were collected from ice-contact stratified drift (Qwic) deposits for quantitative grain size analysis via sieving. The samples were oven-dried prior to analysis and then sieved at 1 ϕ intervals from -6 ϕ (64 mm) to 5 ϕ (32 μ m) using a Ro-Tap RX29 sieve shaker. Grain size distribution data were analyzed, and the individual weight percent and cumulative weight percent were calculated. The graphical plotting method was used to calculate percentile grain size values and the grain size statistical parameters of mean grain size and standard deviation of grain size (sorting) were calculated (Folk and Ward, 1957). Individual weight percent data were summed to calculate the percent gravel, sand, and combined silt and clay grain size fractions.

Geochronology Methods: Three (3) rock samples were collected from the uppermost surface of erratic boulders perched atop the late Wisconsinan age terminal moraine complex for 10-Beryllium cosmogenic exposure age dating. The boulder samples were all located in stable positions on isolated ridges within the moraine complex. Samples from the upper 1-2 cm of the top of the boulders were collected using a hammer and chisel or rock saw. Topographic shielding data were collected every 20 degrees using a Brunton compass. The samples were processed and measured at the Purdue Rare Isotope Measurement (PRIME) Laboratory at Purdue University. 10-Beryllium ages were calculated using the exposure age calculator formerly known as CRONUS Earth, code version 3, the global production rate calibration dataset (Borchers et al., 2016), and the 'LSDn' scaling scheme (Lifton et al., 2014). Exposure ages are reported in years (yr) prior to the year of collection (CE 2024). Shielding parameters were calculated from field measurements taken with a clinometer and using the CRONUS topographic shielding calculator (Balco et al., 2008). No adjustments were made for snow cover or weathering.

Three (3) samples were isolated from lacustrine sediments cores from Lost Lakes bog (41.08280° N, 75.48574° W), a kettle pond complex situated on the terminal moraine complex, for radiocarbon dating. The 14-Carbon samples were isolated from organic rich sediments stratigraphically above glacial sediments and therefore provide a minimum limiting age for the terminal moraine. The samples were chemically pre-treated at Wilkes University using a standard Acid-Base-Acid wash and rinse to neutral pH (Abbott and Stafford, 1996). Subsequently, the samples were combusted to CO₂ gas, converted to filamentous graphite, pressed in Aluminum targets, and measured at the National Ocean Sciences Accelerator Mass Spectrometry Facility at the Woods Hole Oceanographic Institution. Radiocarbon ages were calibrated to calendar years before present (cal yr BP) using the IntCAL20 calibration curve (Reimer et al., 2020) and the CALIB version 8.2 radiocarbon calibration program (Stuiver and Reimer, 1993).

Mapping Results

The study area contains evidence of at least two episodes of Laurentide Ice Sheet advance and retreat from the map area. The deposits are associated with the last ice age (i.e. Wisconsinan) and an earlier more extensive episode of glaciation, which has been attributed to the Illinoian (Braun, 2019). The late Wisconsinan terminal moraine is approximately 1 km in width (from north to south) and spans across the central portion of the map area trending east to west around Interstate-80 (**Figure 3**). The moraine landform is hummocky, contains multiple push ridges, has scattered boulders perched atop various topographic positions (**Figure 4E**), is poorly drained, and largely consists of glacial till. The Wisconsinan till sedimentology is highly variable in terms of grain size but generally consists of a very poorly sorted sediment with a clay to silt matrix including larger pebbles and cobbles. A significant portion of the moraine landform consists of ice-contact stratified drift (**Figure 4C, 4D**) and ice-contact delta deposits of very poorly to moderately sorted sands and gravels (**Figure 3**). Areas located north of the terminal moraine complex largely consist of Wisconsinan age till and bedrock exposures. Ice-contact sand and gravel deposits are mapped along small tributaries to the Tobyhanna Creek, and modern alluvial deposits are mapped in valley bottoms.

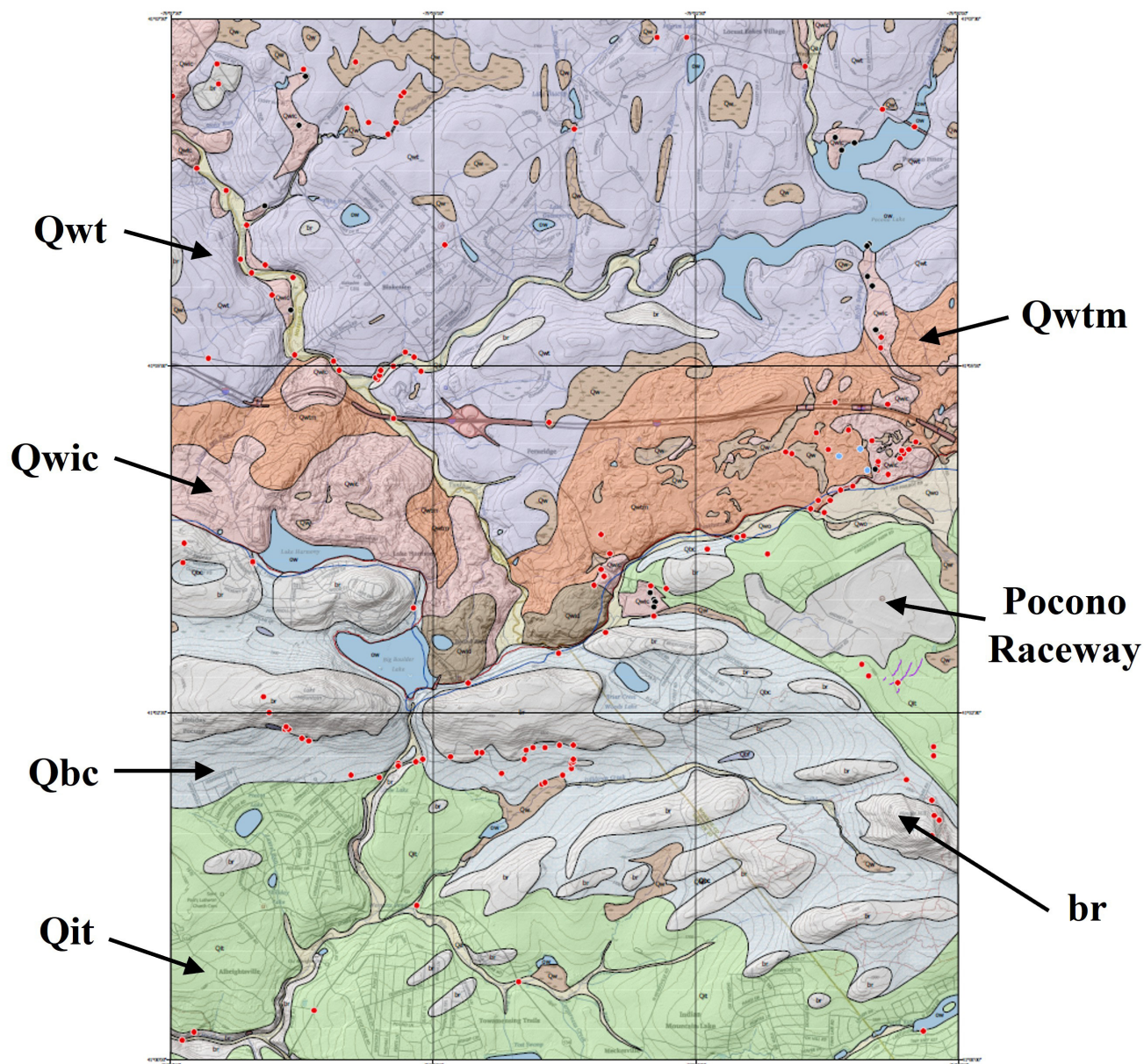


Figure 3. Surficial geology map of the Blakeslee 7.5-Minute Quadrangle with annotation of major map units and landforms (Winter et al., 2025). The labeled surficial geology units include Wisconsinan till (Qwt), ice-contact stratified sand and gravel (Qwic), boulder colluvium (Qbc), Wisconsinan end moraine (Qwtm), Illinois till (Qit), and bedrock (br). Black circles are field check locations, red circles are sand and gravel sample collection sites, and blue circles are boulder sample collection sites.



Figure 4. Photographs of select map units and landforms including A) Illinoian till exposure at the Long Pond Barrens, B) boulder colluvium along the southern hillslope of Lake Mountain, C) sand and gravel quarry along Route 118, D) exposure of sand and gravel deposit along the Route 118 quarry, E) erratic boulder on the terminal moraine sampled for cosmogenic exposure dating, and F) Sam Winter, Aidan Hogan, and Faith Morcombe at Lost Lakes bog collecting sediment core samples.

The area located immediately south of the late Wisconsinan age terminal moraine largely consists of boulder colluvium (**Figure 3, 4B**). Three boulder fields are mapped along the slopes of Lake Mountain and consist of large boulders (> 1 meter intermediate axis dimension) derived from the local late-Devonian age bedrock consisting of sandstones and conglomerates from the Catskill Formation. A relatively small pocket of glacial outwash and modern alluvium is mapped along the Tunkhannock Creek, while the areas further south consist of bedrock exposures or Illinoian age (consistent with the interpretation of Braun, 2019) glacial till (**Figure 3**). The Illinoian age glacial till displays a rusty orangish-red color (**Figure 4A**), suggesting a greater degree of chemical weathering compared to the more recently deposited Wisconsinan till, and pebbles and cobbles have thicker weathering rinds. The Pocono Raceway track and surrounding area is mapped as urban land, given the amount of land disturbance and impervious surface which precludes characterization of map units (**Figure 3**). Several gelifluction lobes are mapped south of the Pocono Raceway track, and a single landslide is mapped along the banks of Mud Run in the southwestern portion of the map area.

Grain size analysis of sand and gravel samples from ice-contact stratified drift (Qwic) deposits were very poorly sorted to moderately sorted (**Figure 5**). The mean grain size of samples ranged from fine sand (0.15 mm) to pebble (12.4 mm), while a total of 14 samples had a majority of sand sized clasts and 5 samples had a majority of gravel sized clasts (**Figure 5**). Most samples had a relatively small proportion of fine-grained sediments (silt and clay) (**Figure 6**). Overall, the textural properties of the sand and gravel samples are consistent with a moderate degree of fluvial (glacial meltwater) re-working of glacial sediments in ice-proximal environments.

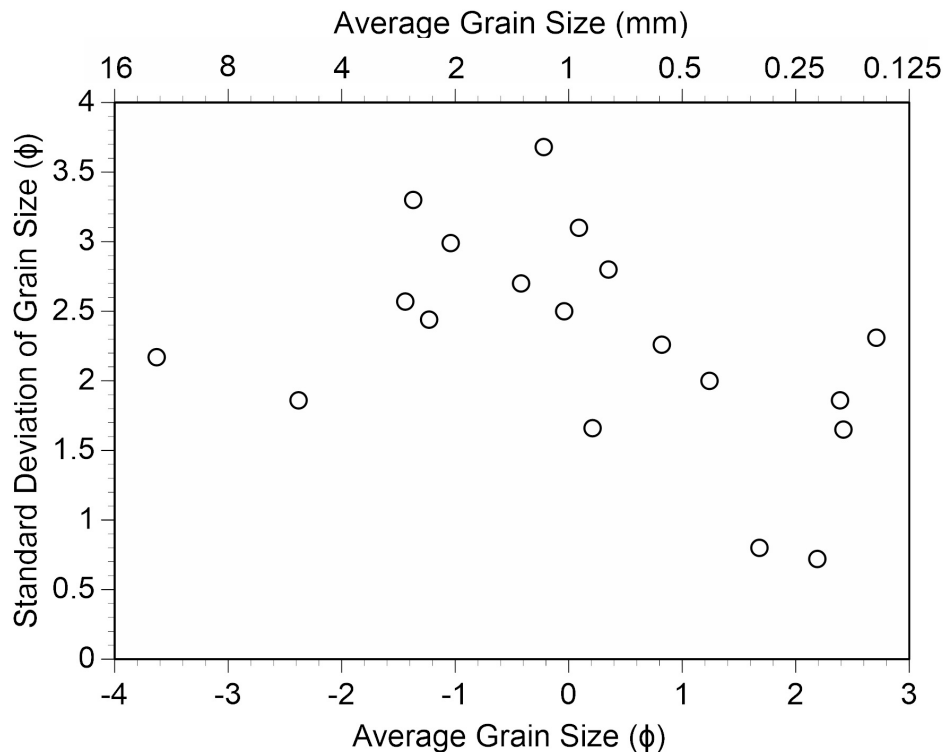
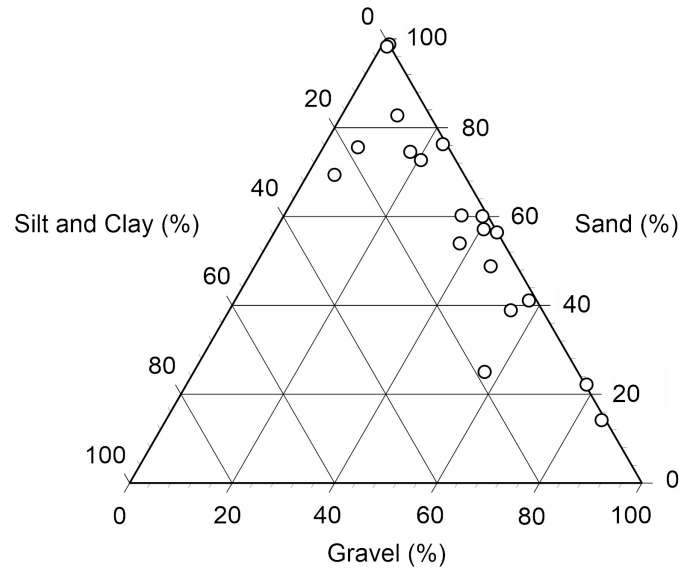


Figure 5. Biplot of the average of grain size (ϕ units bottom axis, mm units top axis) vs. standard deviation of grain size (ϕ) (sorting) of ice-contact stratified drift deposits (Qwic) from the Blakeslee 7.5-Minute quadrangle (Winter et al., 2025).

Figure 6. Ternary diagram of ice-contact stratified drift deposits from the Blakeslee 7.5-Minute quadrangle showing the relative abundance of gravel, sand, and fine-grained sediments (silt and clay) expressed as percents (Winter et al., 2025).



The 10-Beryllium exposure age of the terminal moraine boulder samples are $13,190 \pm 1140$ yr (sample MT-24-02), $21,420 \pm 1660$ yr (sample MT-24-05), and $14,180 \pm 1200$ yr (sample MT-24-06). The exposure age of sample MT-24-05 fits within the time range of the maximum advance of global ice sheets during the late Wisconsinan or Last Glacial Maximum (Clark et al., 2009) (**Figure 7**) and is broadly consistent with published 10-Beryllium exposure age chronologies from New Jersey (Corbett et al., 2017) and southern New England (Balco et al., 2002) including Long Island (Balter-Kennedy et al., 2024). The exposure age of samples MT-24-02 and MT-24-06 are discordant with regional chronologies defining moraine stabilization and ice retreat and appear anomalously young. The exact reason for this discrepancy is unclear, although it may reflect myriad factors including 1) landform instability and rolling or rotation of the boulders after deposition, 2) sampling of relatively small boulders with only ~1 meter of relief above the ground surface, or 3) sampling of weathered boulders with enhanced surface erosion. As a result, we conclude the exposure age of the late Wisconsinan age terminal moraine complex in the map area is best constrained by sample MT-24-05 and indicates moraine stabilization by $21,420 \pm 1660$ yr (**Figure 7**).

The basal calibrated (median) radiocarbon age isolated from post-glacial sediments, stratigraphically above basal minerogenic sediments, from Lost Lakes bog (**Figure 4F**) is ~10,650 calendar years before CE 1950 (cal yr BP) (~10,430 to ~11,070 cal yr BP range) (**Figure 7**). The radiocarbon age is substantially younger than regional estimates defining the age of ice retreat and deglaciation from lakes and bogs in the region (Finkenbinder et al., 2026; Peteet et al., 2012). The apparent young ages may reflect delayed stabilization of the terminal moraine complex due to the presence of persistent buried ice in an ice cored moraine (e.g. Prince et al., 2024), and thus a substantial lag in kettle basin formation. Alternatively, the young ages may reflect a period of non-deposition or erosion of previously deposited sediments in the Lost Lakes bog basin after deglaciation and kettle formation, due to very cold periglacial climates (Ciolkosz et al., 1986) or dry late Pleistocene to early Holocene climates in the region (Shuman & Burrell, 2017).

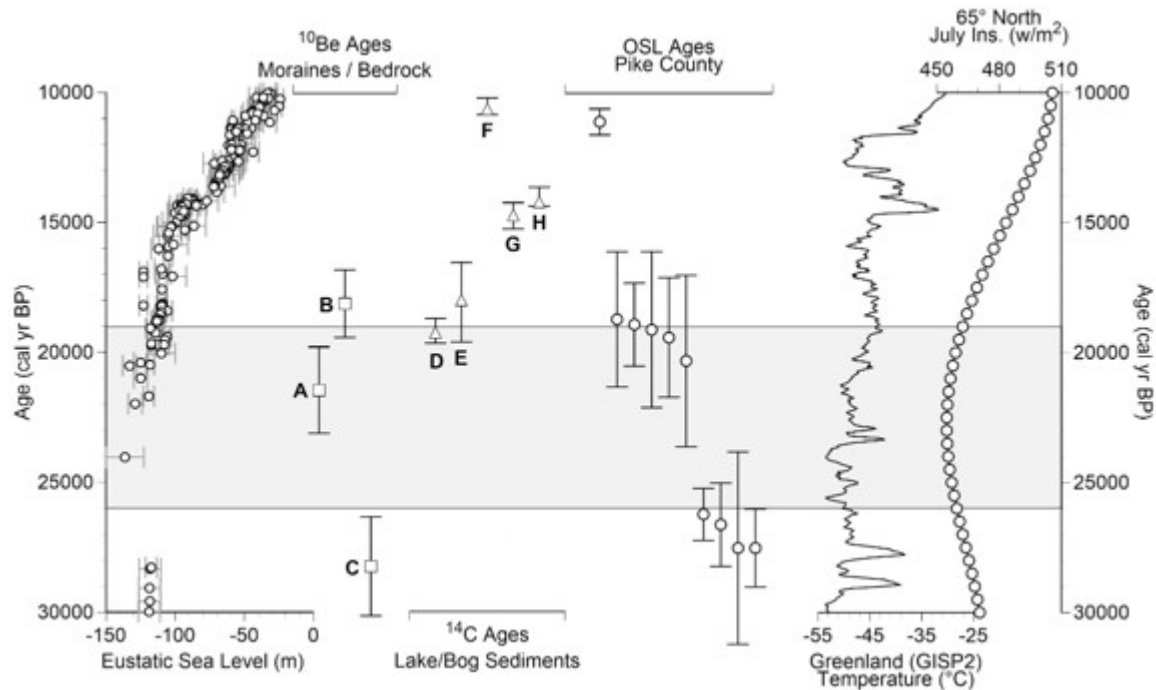


Figure 7. Comparison of Last Glacial Maximum proxy records and the culmination age of glacial moraines and age of deglaciation from regional lake and bog records, including eustatic sea level (Lambeck et al., 2014); ^{10}Be cosmogenic exposure dates (black square symbols) from (A) the Pocono Lake moraine at the Maple Tract Preserve (Winter et al., 2025), (B) the Pocono Lake moraine at State Game Lands 318, and (C) the Seven Tubs Recreation Area; Basal radiocarbon ages (black open triangle symbols) from lakes including (D) Nuangola Lake, Pennsylvania (Finkenbinder et al., 2026), (E) Cranberry Pond, Pennsylvania (Finkenbinder et al., 2026), (F) Lost Lakes bog, Pennsylvania (Winter et al., 2025), Halfmoon Lake, Pennsylvania (Cotter et al., 1986), and Tannersville Bog, Pennsylvania (Peteet et al., 2012); Optically Stimulated Luminescence (OSL) ages on glacial sand and gravel deposits from Pike County, Pennsylvania (Johnson and Finkenbinder, 2023); Greenland Ice Sheet Project 2 (GISP2) temperature (Alley, 2000); and 65° North July insolation (Laskar et al., 2004). The gray bar depicts the time range for the Last Glacial Maximum (Clark et al., 2009).

History of The Pocono Raceway

Leaders: Ben May and George Ewald (Pocono Raceway)

The Pocono Raceway traces its origins to the early 1960s, when developers began planning a major motor sports facility in the Pocono Mountains. Construction faced numerous delays, but the project eventually moved forward under the leadership of Dr. Joseph and Dr. Rose Mattioli. In 1968, the facility opened with a paved three-quarter-mile oval, while work continued on a larger track that would become the raceway's signature feature. The 2.5-mile length tri-oval shape track (**Figure 8**) paved with asphalt was completed in 1971. The track was designed with three distinct turns, each with varying banking and radius, inspired by famous racetracks around the country. Turn 1 has a banking of 14° and radius of 675 feet and was modeled after the former Trenton Speedway. Turn 2 (also known as the Tunnel Turn) has a banking of 8° and radius of 750 feet and was modeled after the Indianapolis Motor Speedway. Turn 3 has a banking of 6° and radius of 800 feet and was modeled after the Milwaukee Mile raceway. The front straightaway is 3,740 feet in length and is often cited as the longest frontstretch on any North American speedway.

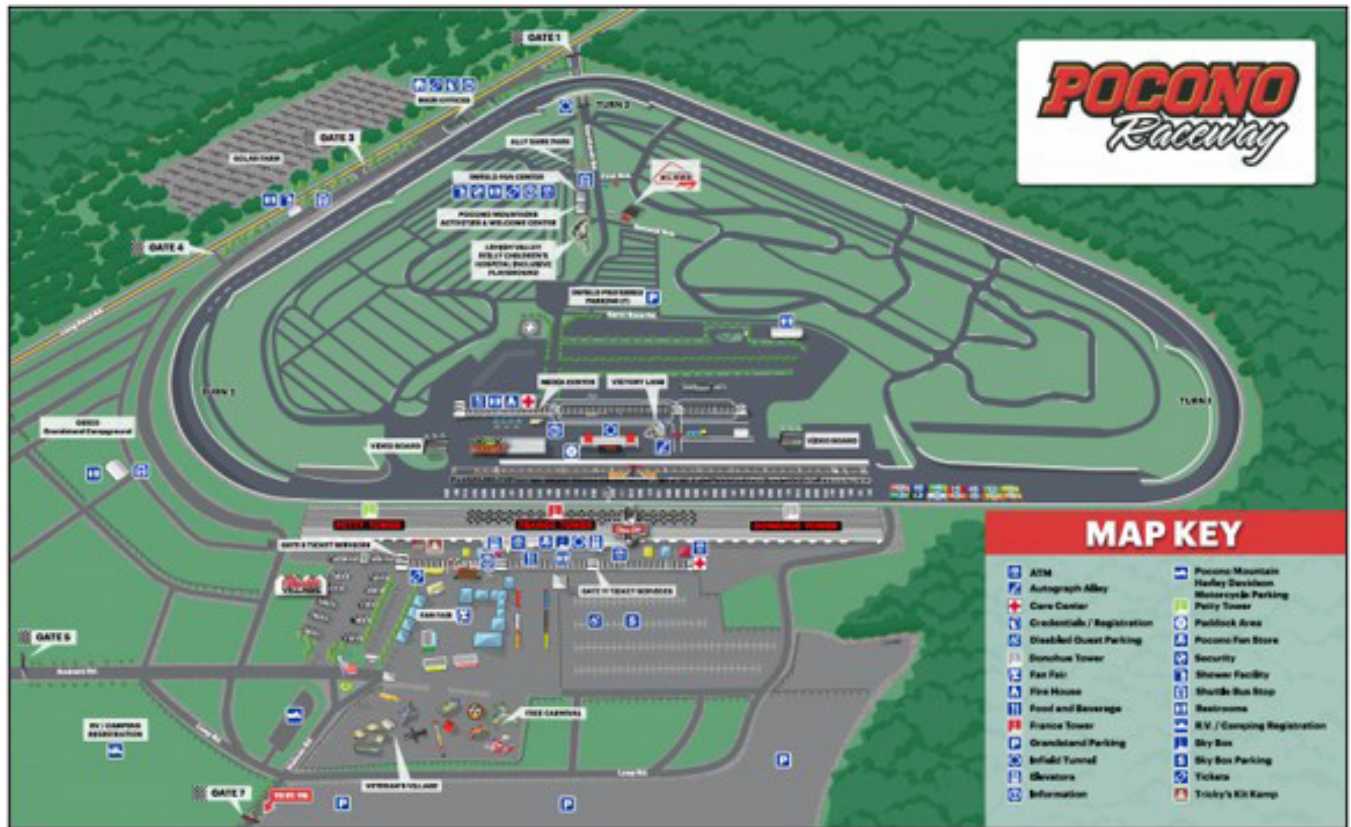


Figure 8. Pocono Raceway facility map (<https://www.poconoraceway.com/facility-map/>).

Since opening, Pocono Raceway has hosted a wide variety of major racing events. The first headline event on the triangular course was a 500-mile United States Auto Club (USAC) Indy car race in 1971, won by Mark Donohue. NASCAR arrived in 1974, when Richard Petty won the inaugural Pocono 500, helping establish the track as a regular stop on the stock car racing calendar. Over the years, the raceway has welcomed IndyCar and the NASCAR Cup Series, O'Reilly Series, and Craftsman Truck Series, along with numerous other racing events. Additional information about the history of the Pocono Raceway is available on the official track website (<https://www.poconoraceway.com/50-years-of-the-tricky-triangle/>). Local journalist Zach Sturniolo wrote a 5-Part Through the Decades Series in 2021 that highlights some of the big moments from each decade at Pocono Raceway.

Pocono Raceway Paving History

The raceway's surface and design have played an important role in its history. The original triangle was completed and paved in time for its first major race in 1971. Over the 1970s through early 1990s, the track surface aged and weathered over time. Exposure to harsh northeastern winters contributed to cracking, wear, and unevenness, making resurfacing an eventual necessity. In 1995, the raceway received a full repave by stripping the upper 2-3 inches of the wearing surface and re-paving with asphalt. The new asphalt extended the usable life of the track and helped maintain competitive racing conditions while supporting increasing speeds and evolving racing technology. More recently in 2011-2012, the track underwent another complete repaving, only the third full paving project in its history. The paving project involved removal of the existing worn asphalt down to the base layer (dirt foundation) of the track (**Figure 9A**) and installation of a new asphalt surface across the entire 2.5-

mile track (**Figure 9B-F**) using a specialized asphalt mixture designed for durability and gradual wear. The track base (soil) was tilled and mixed with 8 percent Portland cement and then rolled out to 12 inches thick. Subsequently, a 3-inch-thick sub-base of limestone aggregate was placed on top of the track base, and then a 2-inch-thick wearing course made of an asphalt-slag mixture was rolled on the surface.

The asphalt mixture used to pave the raceway incorporates Bethlehem Steel blast furnace slag (**Figure 10**) as aggregate, which has been used for all three paving projects at Pocono Raceway. The unique mix wears in a round pattern, compared to a diamond shape seen at most asphalt racetracks, and allows for a longer wear and tear lifetime period for asphalt. Haynes and Kibblehouse (H&K Group, Inc.) were contracted to supply the specialty asphalt mixture for the 1995 and 2011-2012 repave projects. The asphalt mixture consists of two different size fractions of blast furnace slag at different proportions (30% small pebbles and 40% granules to coarse sand) combined with silica sand (30% granule to coarse sand) and bitumen (AC-30 viscosity grade) (**Figure 10**). The repave project used 111,871 square yards of the asphalt-slag mixture (or 6,215 cubic yards) to completely refinish the wearing surface of the track. Pocono Raceway anticipates the current surface will last another 2-3 years before requiring another repave to upgrade the surface for continued use for NASCAR and other performance racing events.



Figure 9. Photographs of the 2011-2012 repave project at the Pocono Raceway including a) removal from the existing wearing surface down to the dirt foundation and b-f) re-pave using the blast furnace slag asphalt mixture. Source: Pocono Raceway Facebook page Track Improvement photo album.

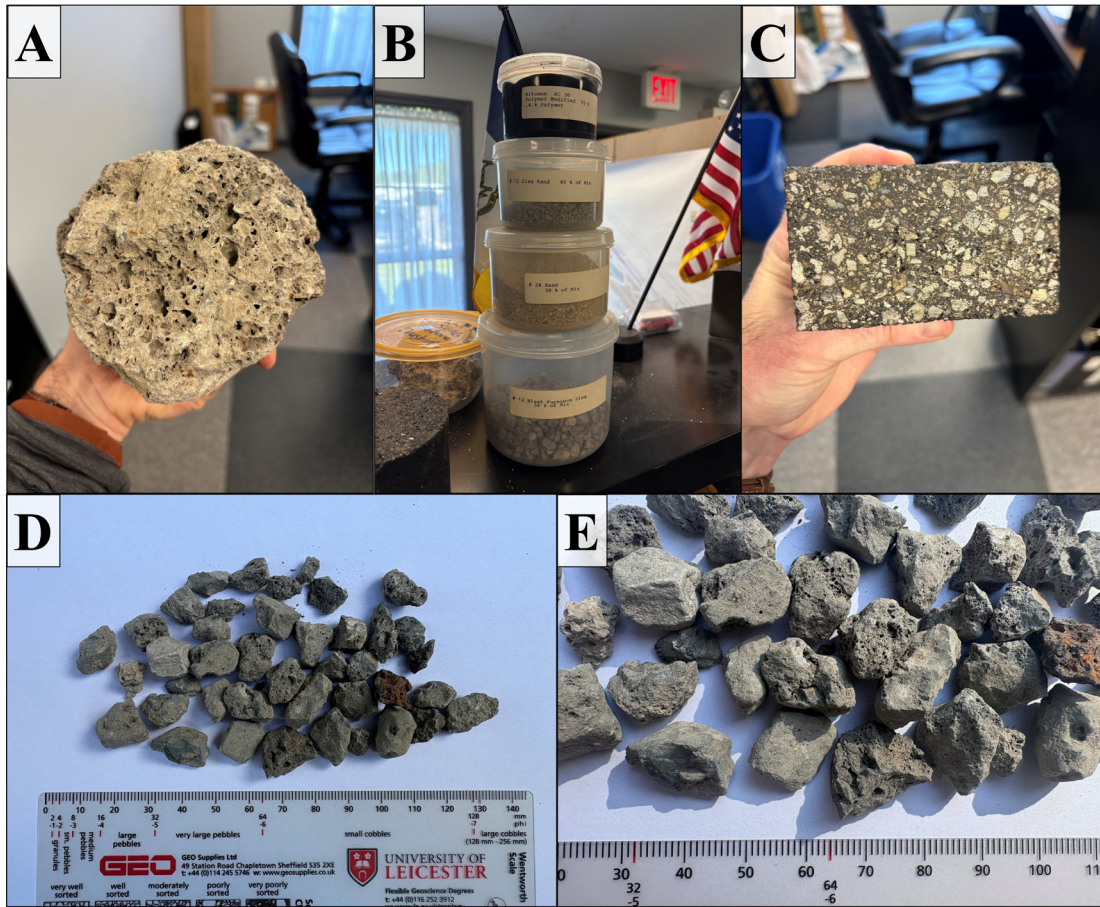


Figure 10. Photographs of Pocono Raceway paving materials including a) a Bethlehem Steel blast furnace slag cobble, b) various sizes of blast furnace slag aggregate and asphalt bitumen used for the 2011-2012 track repave, c) a representative core sample of the uppermost paving surface collected after paving, d) small pebble size fragments of slag displaying a vesicular texture, and e) zoom in view of the same pebble size fragments.

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Stop 7: Pocono Lake Moraine - Surficial geological mapping and exposure dating of moraine boulders

(41.09786, -75.49293) Stoney Hollow Road, Lost Lakes Trailhead

The field stop will examine the morphology and 10-Be exposure ages from the late-Wisconsinan terminal moraine complex.

Stop 7 – Pocono Lake Moraine Mapping and Exposure Dating

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This location involves a 1.0-mile-long (round trip) hike along the Lost Lakes Trail in State Game Lands 318. The trail is uneven and the terrain is low to moderate in relief.

Introduction

The Pocono Lake terminal moraine (informal name) represents the maximum advance of the late-Wisconsinan advance of the Laurentide Ice Sheet (**Figure 1**). The moraine landform is arguably the largest and best-developed and preserved section of the ice margin in northeast Pennsylvania. The moraine trends east-west along and north of Interstate-80 across the Blakeslee and Pocono Pines 7.5-minute topographic quadrangles. The landform displays classic moraine topography and sedimentology – it is ~1 km in extent (from north to south) and is characterized as hummocky, knob and kettle topography with poor internal drainage and includes till and ice-contact stratified drift deposits (Winter et al., 2025; Winter et al., in prep). The field stop includes presentation of a recently updated and published surficial geology map of the site by Wilkes University faculty and students (Sam Winter, Aidan Hogan, and Joseph Kozlowski), a demonstration of sampling of erratic boulders on the moraine, and an overview of cosmogenic exposure dating of the boulders and comparison and synthesis of the

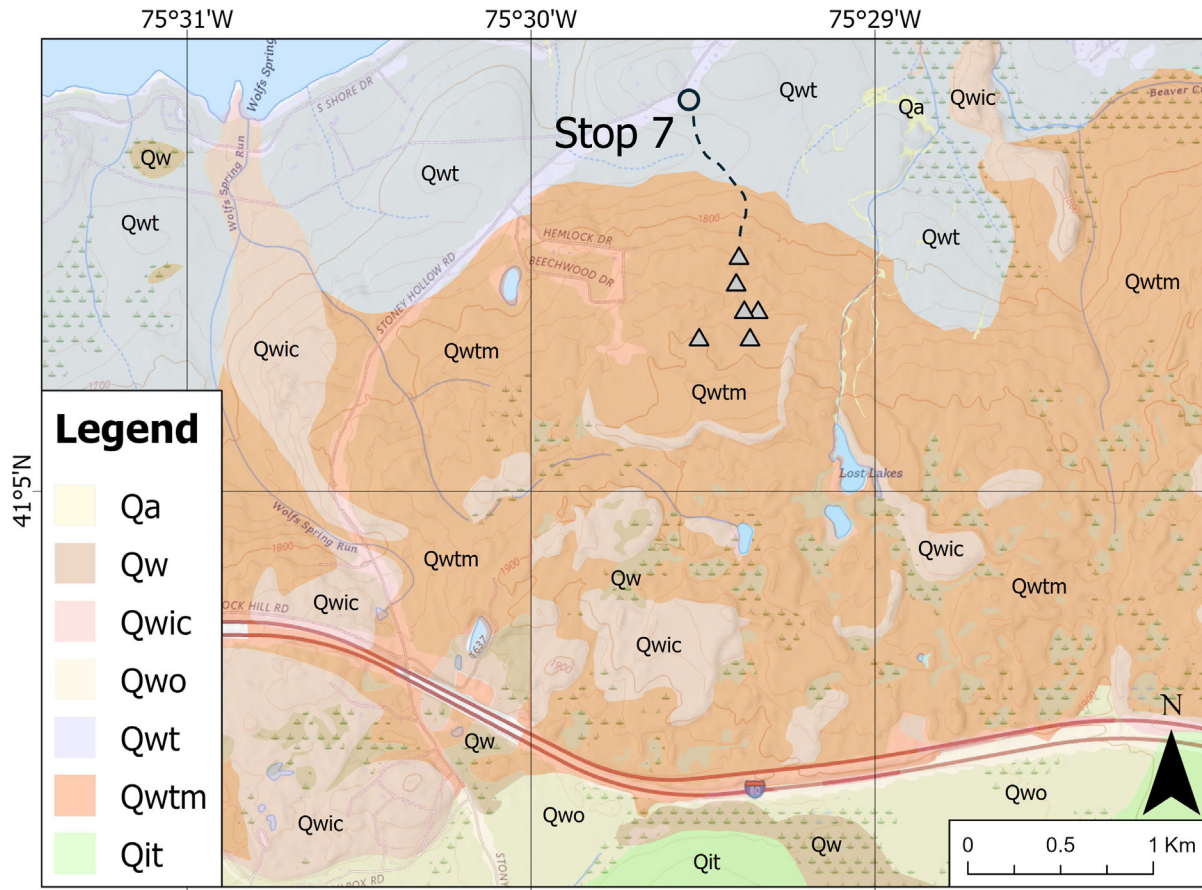


Figure 1. Surficial geologic map of the Pocono Lake terminal moraine and surrounding area showing the late-Wisconsinan ice margin and location of erratic boulders sampled for cosmogenic exposure dating (gray triangle symbol). The dashed line represents the 0.5-mile length path from the Lost Lakes Trailhead to the field site. Surficial geology units include alluvium (Qa), wetlands (Qw), ice-contact stratified sand and gravel (Qwic), outwash (Qwo), Wisconsinan till (Qwt), Wisconsinan till moraine (Qwtm), and Illinois till (Qit) (Winter et al., 2025; Winter et al., in prep).

ages with other regional datasets.

Surficial Geology Mapping of the Pocono Pines 7.5-Minute Quadrangle

The Pocono Pines 7.5-Minute Quadrangle straddles a portion of the late-Wisconsinan age terminal moraine complex of the Laurentide Ice Sheet (17,000 to 22,000 years old) (Crowl & Sevon, 1980) and contains an excellent geomorphic record of ice sheet margin landforms and deposits. Previous mapping work in the study area began with Berg et al (1977), who mapped the distribution of major landforms and sediment deposits at 1:24,000 scale on the basis of field mapping and 7.5' topographic map/aerial photography interpretation in the early 1970's. Berg et al (1977) mapped the late-Wisconsinan terminal moraine primarily on the basis of hummocky, knob and kettle topography. Subsequently, Braun (2019) updated and revised a portion of the surficial geological map of the Pocono Pines Quadrangle at 1:24,000 scale in the late 1990's on the basis of additional field mapping (completed in 1988, 1989, and 1997), 7.5-Minute topographic map/aerial photography interpretation,

and by using county soil surveys to map sand and gravel deposits. All areas located south and beyond the maximum extent of the late-Wisconsinan terminal moraine complex were not revised. Braun (2019) later reviewed lidar hillshade maps and aerial photographs (acquired in 2005-2007) and concluded that areas of moraine topography, bedrock ledges, and smooth slopes underlain by boulder colluvium could be quite accurately mapped, however no additional field verification was completed. Additional changes to map unit contacts were made prior to submission of the draft map to the Pennsylvania Geological Survey (PaGS) in 2019.

While field-based confirmations were conducted, they were few in number (Berg et al., 1977) or not included (Braun, 2019) in products by these prior mappers. As a result, the existing maps lack sufficient metadata regarding field check locations and the confidence of mapped deposits and landforms (i.e. existing maps are not U.S. Geological Survey Geologic Map Schema or GeMS compliant). Furthermore, none of the previous studies generated age constraints on the glacial landforms and/or sediments using absolute dating techniques, and therefore the age of the late-Wisconsinan terminal moraine complex and the timing of ice sheet deglaciation is unknown. As a result, Wilkes University undergraduate level Geology and Environmental Science students Sam Winter, Aidan Hogan, and Joseph Kozlowski completed a capstone Senior Project and thesis to create an updated GeMS compliant surficial geology map of the Pocono Pines 7.5-Minute Quadrangle (Winter et al., in prep).

Geospatial Analysis and Field Mapping Methods: The surficial geological map was generated using the same methods for the Blakeslee 7.5-Minute quadrangle at Stop 6 (Winter et al., 2025). using previously published surficial maps of the study area by Berg et al. (1977) and Braun (2019) as a guide. A standard set of surficial units and names from the adjacent Blakeslee 7.5-Minute quadrangle (Winter et al., 2025) was used for mapping and to ensure regional consistency. Field mapping and checking of map units involved observations of surficial landforms, sediment deposits, and bedrock exposures for confirmation or re-interpretation.

Mapping Results

The study area contains evidence of at least two episodes of Laurentide Ice Sheet advance and retreat from the map area. The deposits are associated with the last ice age (i.e. Wisconsinan) and an earlier more extensive episode of glaciation, which has been attributed to the Illinoian glacial advance (Berg et al. 1977). The late Wisconsinan terminal moraine is approximately 1 km in width (from north to south) and spans across the central portion of the map area trending east to west around Interstate-80 (**Figure 2**). In the eastern portion of the map area, the terminal moraine straddles the hillslopes of Camelback Mountain, while the mountain summit remained ice free. The moraine landform is hummocky, contains multiple push ridges, has scattered boulders perched atop various topographic positions (**Figure 3C**), is poorly drained, and largely consists of glacial till. The till sedimentology is highly variable in terms

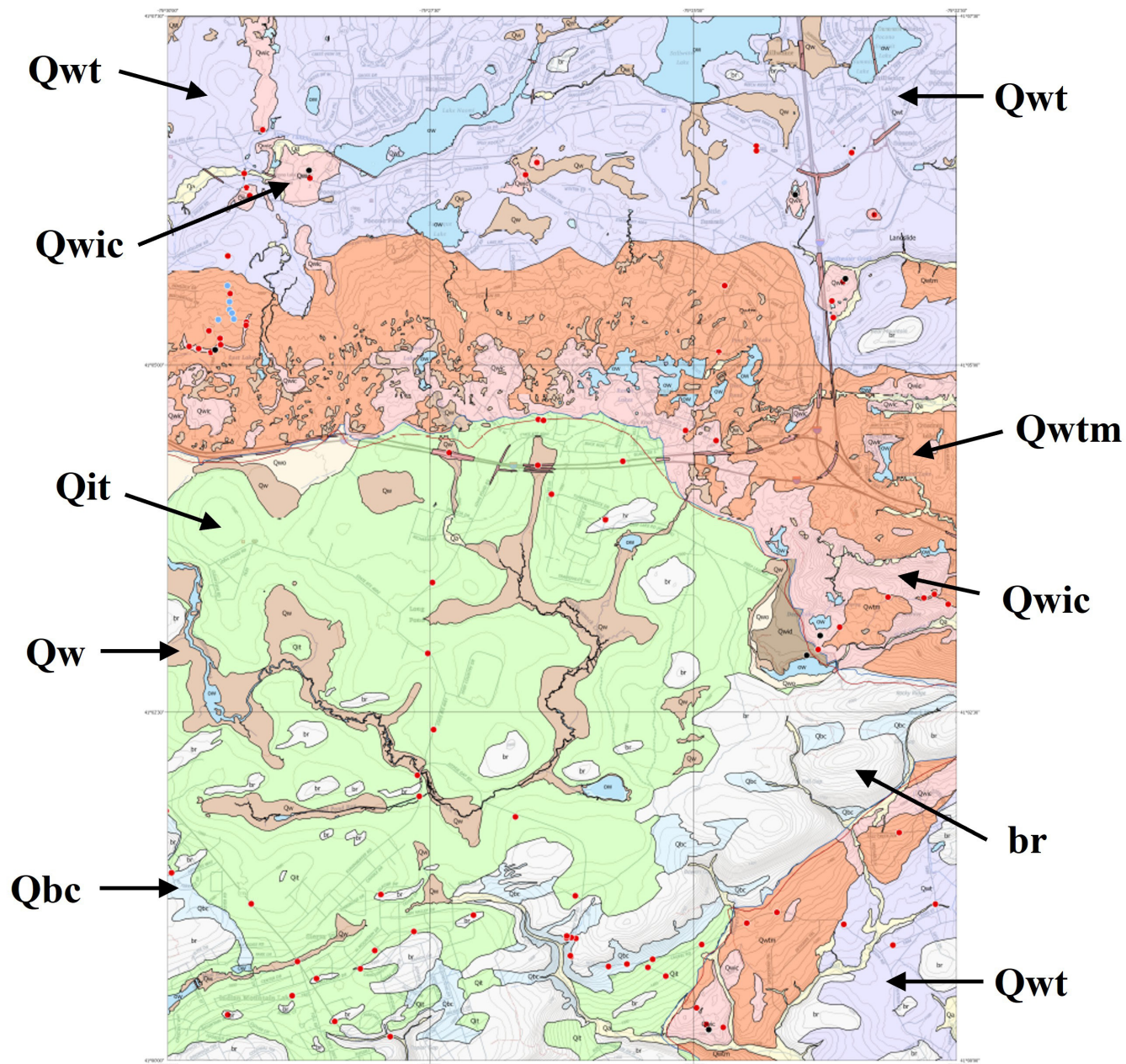


Figure 2. Preliminary surficial geology map of the Pocono Pines 7.5-Minute Quadrangle with annotation of major map units and landforms (Winter et al., in prep). The labeled surficial geology units include Wisconsinan till (Qwt), wetlands (Qw), ice-contact stratified sand and gravel (Qwic), boulder colluvium (Qbc), Wisconsinan till moraine (Qwtm), Illinois till (Qit), and bedrock (br).

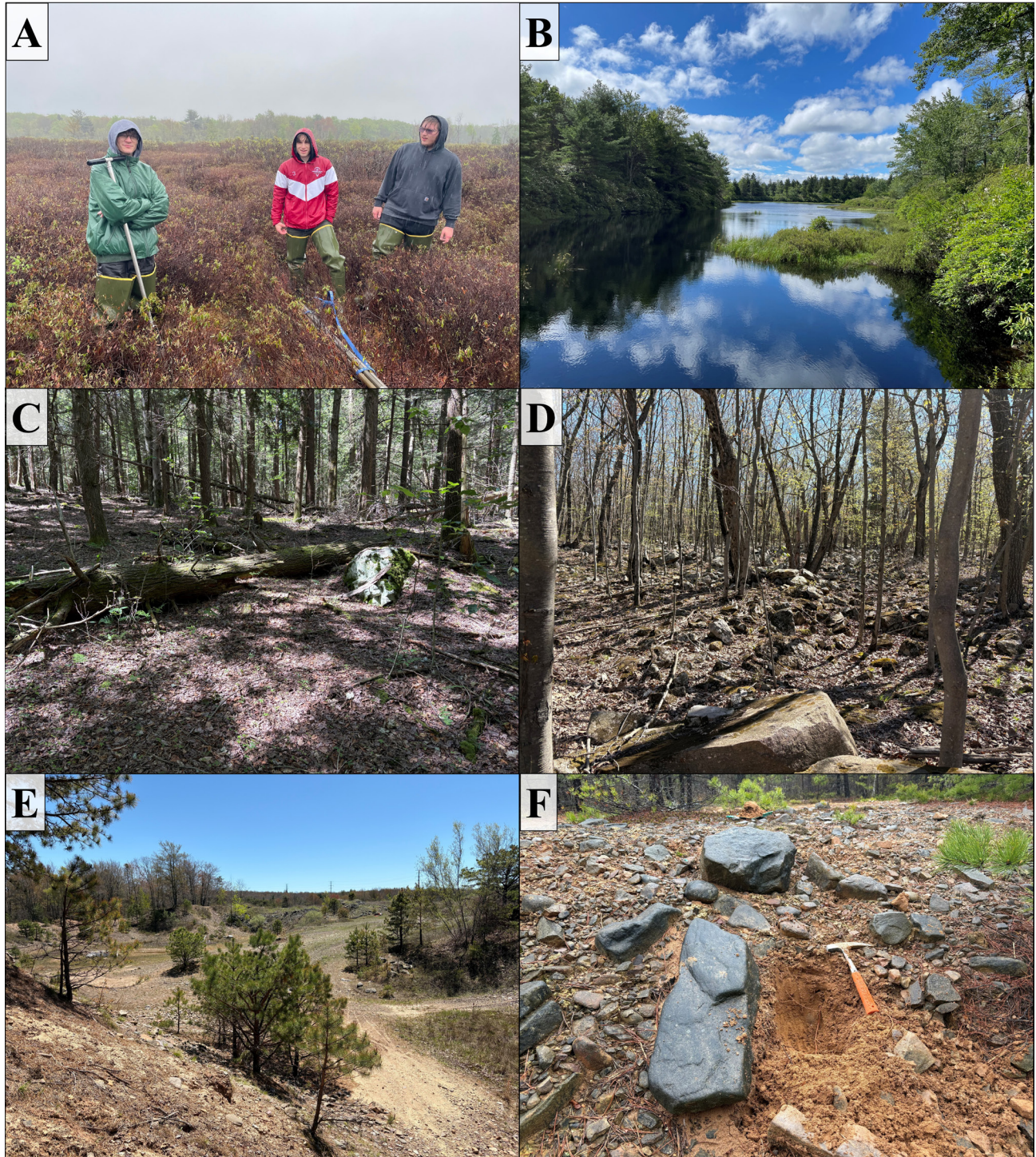


Figure 3. Photographs of select map units and landforms including A) Hauser Mire wetland (Qw), B) Lost Lakes (open water), C) Wisconsinan till moraine (Qwtm) with erratic boulder PLM-02 sampled for cosmogenic exposure dating at State Game Lands 318, D) boulder colluvium along the southeastern hillslope of Camelback Mountain in State Game Lands 38, E) ice-contact stratified sand and gravel (Qwic) quarry located immediately south of the Kalahari Waterpark Resort and Convention Center, and F) pavement exposure of ice-contact stratified sand and gravel (Qwic) deposit south of the intersection of Lond Pond Road and Route 940 in Pocono Summit.

of grain size but generally consists of a very poorly sorted sediment with a clay to silt matrix including larger pebbles and cobbles. A significant portion of the moraine landform consists of wetlands (**Figure 3B**) and ice-contact stratified drift of very poorly to moderately sorted sand and gravel (**Figure 2**).

Areas located north of the terminal moraine complex largely consist of Wisconsinan age till and bedrock exposures. Ice-contact sand and gravel deposits are mapped as a north-south trending esker complex east of Pocono Lake and along the outer edge of the terminal moraine complex (**Figure 3E-F**), and modern alluvial deposits are mapped in valley bottoms. Boulder colluvium deposits (**Figure 3D**) are mapped along steeper hillslopes down-slope of shallow bedrock along Camelback Mountain. The area located south of the late Wisconsinan age terminal moraine largely consists of bedrock exposures, Illinoian age glacial till, and wetlands. The Illinoian age glacial till displays a rusty orangish-red color, suggesting a greater degree of chemical weathering compared to the more recently deposited Wisconsinan till, and pebbles and cobbles have thicker weathering rinds. Two small landslides are mapped in low-order colluvial valleys (hollows) in the eastern portion of the map area (Karimi et al., 2019). Wetlands (**Figure 3A**) and open water associated with Long Pond are mapped in the west, southwest portion of the map area.

Field Stop

Stop #7 – Pocono Lake Moraine Exposure Dating (41.0909° N, 75.4901° W)

Leaders: Tim Lane (Aarhus University), Kathryn Adamson (Manchester Metropolitan University), and Matt Finkenbinder (Wilkes University)

This field stop includes a ~1.0-mile (round trip) hike along the Lost Lakes Trail in State Game Lands 318 that traverses the late-Wisconsinan age terminal moraine complex of the Laurentide Ice Sheet. The purpose of the stop is to observe the moraine landform, and its impressive topography, and to carry out a demonstration of collecting rock samples from erratic boulders for exposure dating (**Figure 4**). A presentation of preliminary ^{10}Be exposure ages from six (6) moraine boulders and comparison with



Figure 4. Photographs of select erratic boulders located on the Pocono Lake terminal moraine sampled for cosmogenic exposure dating. The upper 2 cm of boulders located in stable topographic positions were sampled using a handheld rock saw and rock hammer with chisel.

other regional datasets including basal radiocarbon ages from Nuangola Lake (Stop 1), Cranberry Pond (Finkenbinder et al., 2026), and other proximal records along with other exposure ages from the region (Stop 3), will be discussed (**Figure 5**).

An introduction to the cosmogenic nuclide method was presented at Stop 3 and the same methodology has been applied at the Pocono Lake moraine, however on erratic boulders (**Figure 4**) as opposed to bedrock (similar to the work discussed at Stop 6). The same geological and methodological assumptions and uncertainties exist for samples taken from boulders, but more emphasis is placed on the post-depositional history of the moraine or feature being dated. As explained previously, constraining the timing of LIS retreat is important for understanding ice sheet behavior during deglaciation. On and close to the LGM terminal moraine, cosmogenic nuclide and organic ^{14}C ages disagree by several thousand years (Peteet et al., 2012; Halsted et al., 2024). A compilation of cosmogenic nuclide, ^{14}C , and varve ages by Halsted et al (2024) suggests initial LIS recession occurred from 24,000 - 20,000 years. They suggest that the difference in age estimates is largely due to (1) inherited ^{10}Be from pre-LGM exposure which adds 2,000 to 6,000 years to exposure ages, and (2) an 4,000 to 8,000 year lag between ice retreat and organic material accumulation in lake basins. Specifically, exposure dating of the LGM terminal moraine to the east in Massachusetts and

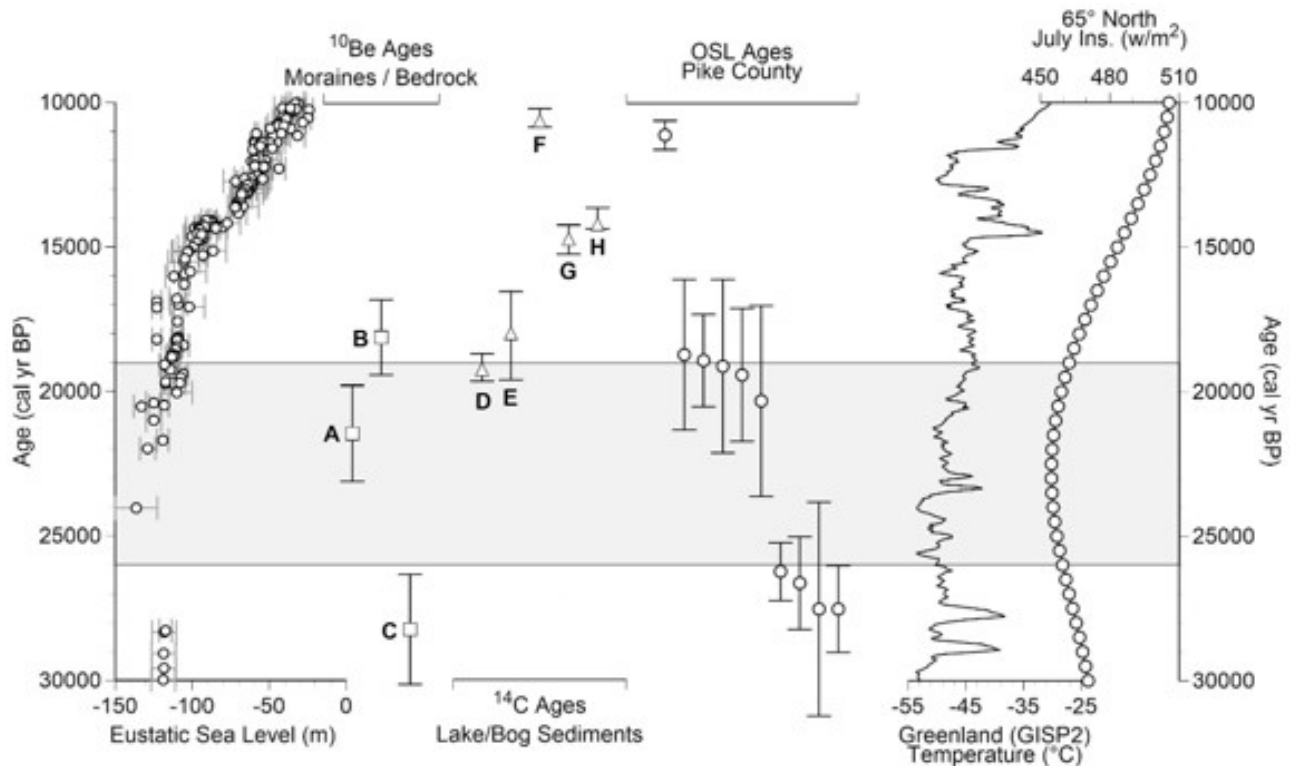


Figure 5. Comparison of Last Glacial Maximum proxy records and the culmination age of glacial moraines and age of deglaciation from regional lake and bog records, including eustatic sea level (Lambeck et al., 2014); ^{10}Be cosmogenic exposure dates (black square symbols) from (A) the Pocono Lake moraine at the Maple Tract Preserve (Winter et al., 2025), (B) the Pocono Lake moraine at State Game Lands 318, and (C) the Seven Tubs Recreation Area; Basal radiocarbon ages (black open triangle symbols) from lakes including (D) Nuangola Lake, Pennsylvania (Finkenbinder et al., 2026), (E) Cranberry Pond, Pennsylvania (Finkenbinder et al., 2026), (F) Lost Lakes bog, Pennsylvania (Winter et al., 2025), Halfmoon Lake, Pennsylvania (Cotter et al., 1986), and Tannersville Bog, Pennsylvania (Peteet et al., 2012); Optically Stimulated Luminescence (OSL) ages on glacial sand and gravel deposits from Pike County, Pennsylvania (Johnson and Finkenbinder, 2023); Greenland Ice Sheet Project 2 (GISP2) temperature (Alley, 2000); and 65° North July insolation (Laskar et al., 2004). The gray bar depicts the time range for the Last Glacial Maximum (Clark et al., 2009).

New Jersey produce moraine ages of $27,500 \pm 2200$ yrs (Balco et al. 2002) to $25,100 \pm 2,100$ years (Corbett et al. 2017), respectively, similar to other exposure ages on recessional moraine boulders in Connecticut and New England (Balco and Schaefer, 2006; Balco et al., 2009).

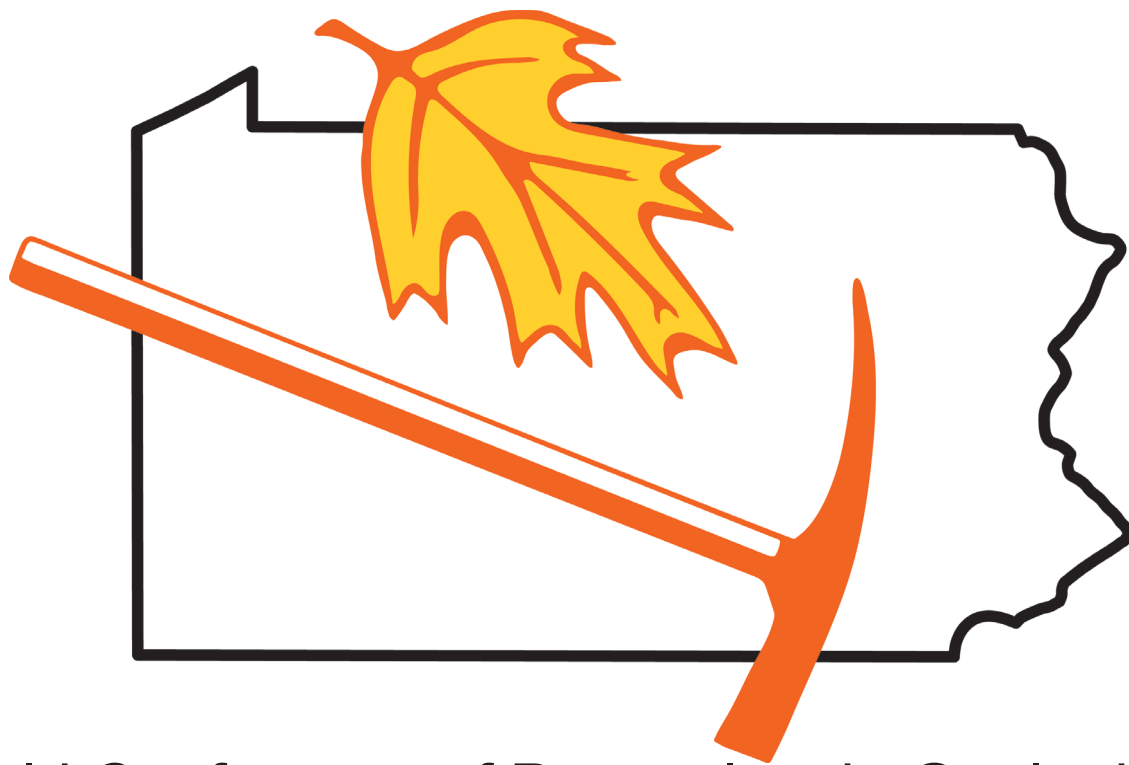
We sampled boulders from the Pocono Lake Moraine complex to investigate if this pattern extends to the continuation of the LGM moraine found in the region, and if the 2,000 to 6,000 years of inheritance hypothesized by Halsted et al (2024) could be constrained at this site. Six samples were taken from quartz arenite boulders (Pocono Formation) perched atop ridge crests along the large LGM moraine complex (**Figure 4**). This moraine is a large, composite feature, but samples were taken from a small area of the moraine to ensure similar depositional and post-depositional history. Samples were taken from erratic boulders which were judged to have been deposited subglacially, and thus not contain any inherited nuclides. The boulders were edge rounded in places, and raised above the surrounding terrain, to ensure minimal postglacial sediment cover or vegetation growth. Samples were measured at the National Environmental Isotope Facility, UK and exposure ages were calculated using the global production rate (Borchers et al., 2016) and the LSDn time-dependent scaling framework (Lifton et al., 2014) in years (yr) prior to sampling (2018 CE). The six samples have exposure ages ranging from 15,000 to 21,000 yr with a mean exposure age of 17,500 yr. However, based on the minimum limiting basal age from Nuangola which requires deglaciation prior to $\sim 19,260$ cal yr BP (Stop 1; Finkenbinder et al., 2026), we can remove younger outliers, leaving a mean age for retreat from the LGM moraine of $\sim 19,300$ yr (**Figure 5**).

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Return to Wilkes Barre Inn and Suites. End of Day 2



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