GEOLOGY AND PUBLIC LANDS

CONFERENCE PROCEEDINGS AND FIELD GUIDE

EDITED BY
JANE ALEXANDER

GEOLOGICAL ASSOCIATION OF NEW JERSEY

XXIX ANNUAL CONFERENCE AND FIELDTRIP

OCTOBER 12 – 13, 2012

THE ENVIRONMENTAL CENTER AT LORD STIRLING PARK, THE GREAT SWAMP, NJ

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(COLLEGE OF STATEN ISLAND/CUNY)

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SCHEDULE

Friday, October 12th – Environmental Education Center – Lord Stirling Park, Basking Ridge, NJ

11:00-3:30	Registration
11:30-12:30	Teacher Workshop – New Jersey Field Site Information Exchange Jane Alexander, College of Staten Island/CUNY, and GANJ President
1:00-1:20	Welcoming Remarks Jane Alexander, College of Staten Island/CUNY, and GANJ President
1:20-1:40	Paleozoic Trace Fossils from the Delaware Water Gap National Recreational Area Robert Metz, Kean University
1:40-2:00	Geology along the Trail across Mount Paul Memorial County Park, Chester, New Jersey John Puffer, Rutgers University-Newark
2:00-2:20	Telling the Story of Glacial Lake Passaic Using Public Lands Scott Stanford, New Jersey Geologic Survey
2:20-2:40	Break
2:40-3:00	Geology of Round Valley Recreation Area: The Complex Interplay of Metamorphic, Magmatic and Tectonic Processes Richard Volkert, New Jersey Geological Survey
3:00-3:20	Soil Survey of Crooke's Point, Gateway National Recreation Area as a Guide for Vegetation Management Amanda Rollizo, College of Staten Island/CUNY
3:20-3:40	New Jersey's Public Lands: An Invaluable Resource for Geology Education Emma Rainforth, Ramapo College of New Jersey
3:45-4:45	<u>Keynote Speaker</u> Richard G. Lathrop, Jr., Rutgers University Dept. of Ecology, Evolution, and Natural Resources Director, Grant F. Walton Center for Remote Sensing & Spatial Analysis Editor of <u>The Highlands: Critical Resources, Treasured Landscapes</u>
6:00	Dinner and Business Meeting - Ridge Tavern, Basking Ridge, NJ

Saturday, October 13th - Field Trip

8:00 am - 5:00 pm. Meet at the Environmental Education Center

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INTRODUCTION

The 29th annual meeting of the Geological Association of New Jersey (GANJ) celebrates the public lands of New Jersey and the opportunities they offer for research and education. Public lands range from small local parks run at the town or county level to larger state parks and federally run recreation areas. All provide access to open space for the residents of the most densely populated state.

All of the speakers have utilized access to New Jersey's public lands in their research. Talks encompass all of the physiographic provinces, from the Ridge and Valley in the Delaware Water Gap National Recreation Area, through the Highlands at Round Valley State Park, the Piedmont at Mount Paul and the Coastal Plain at Gateway National Recreation Area. They provide an understanding of how geologic processes shaped the state from the earliest rocks, to the most recent glaciation and its impact on present day drainage. They show how parks and trails can be used to educate people about geology, and in turn, how geologic research can assist in preserving these lands for future generations.

The field stops will allow participants to experience New Jersey's geologic history from the Paleozoic rocks of Round Valley State Park to the present day processes at work in the wetlands of the Great Swamp. The trip will visit established state and county parks, but will also introduce participants to the recently acquired Mount Paul Preserve in Chester, which was only opened to the public this year.

PALEOZOIC TRACE FOSSILS FROM THE DELAWARE WATER GAP NATIONAL RECREATION AREA, NEW JERSEY AND PENNSYLVANIA

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Department of Geology and Meteorology, Kean University, Union, New Jersey 07083

Abstract

Marine strata representing deposits of the Devonian Oriskany Formation, Onondaga Formation and the Mahantango Formation at the Delaware Water Gap National Recreation Area have yielded a variety of trace fossils. Specimens include *Cruziana* isp., *Dactylophycus quadripartitum, Diplichnites* isp., *Helminthoidichnites tenuis, Helminthopsis hieroglyphica, Nereites missouriensis, Planolites annularis, Planolites beverleyensis, Protovirgularia dichotoma, Psammichnites* isp., *Skolithos linearis, Treptichnus bifurcus, Treptichnus pollardi,* and an escape structure. The traces were formed under a variety of environments ranging from a high energy, shallow marine; low-energy, well-oxygenated, moderately-deep marine; to that of hemipelagic marine deposits formed in an offshore setting. Trace-makers largely included annelids, several types of mollusks, trilobites, and arthropods.

Introduction

The Delaware Water Gap National Recreation Area (DWGNRA) was established by Congress in 1965, and is the largest recreation area in the eastern United States. Paralleling the Delaware River, it includes almost 70,000 pristine acres of land in northwestern New Jersey and northeastern Pennsylvania. A variety of activities can be enjoyed by a visitor, such as hiking, fishing, swimming, and camping among others. It is also home to extensive exposures of marine to non-marine Silurian and Devonian sedimentary rocks. Though Paleozoic strata within the boundaries of the Delaware Water Gap National Recreation Area of northwestern New Jersey and adjacent northeastern Pennsylvania have been widely studied, detailed and debated (e.g., Weller, 1900; Epstein et al., 1967; Spink, 1967; Epstein, 1970; Barnett, 1970; Epstein and Epstein, 1972), the presence of trace fossils were largely neglected (e.g., Martino and Zapecza, 1978). The author has attempted to fill in this gap through a series of field investigations of the Silurian and Devonian strata present. Summaries of such publications were presented in a previous paper of the Geological Association of New Jersey (Metz, 2007). The present article details additional field research (Metz, 2008a, 2008b, 2009) since that publication and provides a compilation of trace fossils within Devonian strata as well as their likely paleoenvironment.

Geologic Setting and Systematic Ichnology

The Oriskany Formation (Lower Devonian) consists of beach or barrier-bar deposits. The unit, up to 10 m thick, is exposed in a northeast-southwest trend along Wallpack Ridge in northwestern New Jersey, and thins to the northeast, eventually pinching out at Peters Valley (Drake et al, 1996). The strata in this area consist of orange-gray-to brownish-orange-weathering, light-to medium-gray, crossbedded, brachiopod-rich, carbonate-cemented quartz sandstone and quartz-pebble conglomerate (Epstein, 1970; Drake et al., 1996). Field investigation of the Oriskany Formation in the Lake Maskenozha Quadrangle, approximately 3 km southwest of Wallpack Center, revealed the presence of the trace fossils *Planolites beverleyensis*, *Skolithos linearis*, and a bivalve escape structure. At this location the strata consist of fossiliferous (brachiopods), light-to medium-gray calcareous sandstone, quartz-pebble to gravel conglomerate, and minor moderate yellowish to dark reddish brown to dark-yellowish orange weathering, thin beds and lenses of medium-to medium dark-gray siltstone, dark-gray chert, and limestone. Interestingly, detailed field investigations at other exposures of the Oriskany Formation within the DWGNRA failed to produce evidence of trace fossils. A low diversity and relatively high abundance of trace fossils (*Planolites* and *Skolithos*) were recovered from this site. Schlirf and Uchman (2005) indicated that the presence of these two trace fossils, by themselves, does not indicate a particular environment due to their facies-crossing habits. However, *Skolithos* has been most commonly recorded from shallow water high-energy environments. Furthermore, many of the specimens of *Skolithos* collected from calcareous sandstones are infilled with sediment coarser than the host rock, while several field examples display a sharp erosional contact between underlying calcareous sandstones exhibiting gravel-filled burrows of *Skolithos* derived from overlying conglomerates. Indeed, the coarser filling of the *Skolithos* burrows combined with the sedimentary structures, nature of and largely reworked condition of the fossil brachiopod fauna, as well as the escape structure, points to a high energy, shallow water setting subject to periodic inundation by coarser, land-derived sediment. In contrast, *Planolites*, though very abundant, is limited to thin, sporadic occurrences of bioturbated siltstone. In addition, compared to the vertical nature of the *Skolithos* tubes, *Planolites* occurs as an unlined, largely horizontal burrow. Field evidence also indicates that, at times, abundant *Planolites* and minor *Skolithos* coexisted. Thus, the shallow burrowing, deposit-feeding nature of *Planolites* suggests formation under somewhat more quiescent energy conditions, including limited land-derived coarser sediment inundation, whereby worm-like organisms took advantage and exploited organic-rich nutrients present in the finer-grained sediment. As such, *Skolithos* as well as *Planolites* displayed pioneering, opportunistic (r-selected) strategies of colonization (Miller and Johnson, 1981). In this case, typified by low diversity and relatively high density, organisms produced burrows in a dominantly high-stress environment over relatively differing periods of time, largely influenced by the changes in rates of terrigenous-derived sediment accumulation and varying energy conditions.

Ichnogenus *Planolites* Nicholson, 1873
***Planolites beverleyensis* (Billings, 1862)**
Figs. 1

Material. At least 175 specimens, with many more in the field. Figured specimen New Jersey State Museum 21860.

Description. Simple, mostly horizontal to sub-horizontal, straight to slightly curved, smooth, unlined burrows. Burrows are 2-6 mm in diameter, up to 60 mm in length, and preserved in convex hyporelief. Crossovers occur between individual specimens. Burrow fill is lighter in color and coarser than the host rock.

Remarks. *Planolites* differs from the similar appearing ichnotaxon *Palaeophycus* in lacking a wall lining and having burrow fill which differs from the host strata (Pemberton and Frey, 1982; Fillion and Pickerill, 1990; Keighley and Pickerill, 1995). Keighley and Pickerill (1997) addressed the various problems associated with differentiating between two of the ichnospecies of *Planolites*. *Planolites* occurs in marine (e.g., Moghadam and Paul, 2000), as well as no-marine (e.g., Kim et al., 2005) environments. *Planolites* has been depicted as representing the active backfilling of an interim burrow by a mobile deposit-feeding worm-like organism (Pemberton and Frey, 1982), while arthropods have been proposed for non-marine deposits (Buatois and Mángano, 1993).

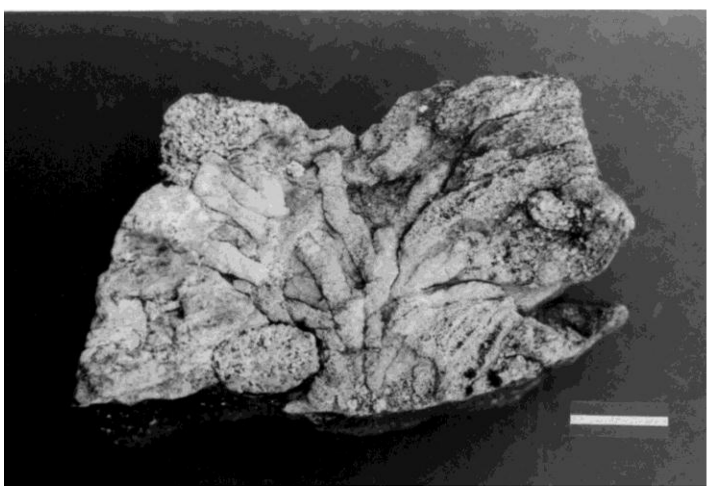

Figure 1. *Planolites beverleyensis*. Scale = 1 cm.

Ichnogenus *Skolithos* Halderman, 1840
***Skolithos linearis* Halderman, 1840**
Fig. 2

Material. Eleven slabs with at least 27 specimens, with more in the field. Figured specimen NJSM 21863.

Description. Straight to slightly inclined, unbranched, cylindrical to sub-cylindrical, vertical to steeply inclined burrows, mostly of constant width though some are slightly variable. Some specimens flare upwards into a funnel-shaped top. Sporadic specimens exhibit annulations. Burrow diameter 3-10 mm, maximum length observed up to 90 mm, limited by the thickness of the collected slabs; diameter of funnel 15-20 mm. Walls mostly distinct, rarely indistinct, mostly unlined, a few specimens are thinly lined with dark- or light-colored clayey material. Burrow fill structureless, with many specimens filled with coarser sediment than the surrounding matrix.

Remarks. Shape, ornamentation, burrow diameter, uniformity of burrow diameter, and distinctiveness of burrow walls are features used to differentiate ichnospecies of *Skolithos*. Following criteria established by Alpert (1974), as well as taxonomic comments by Fillion and Pickerill (1990), the specimens are assigned to *S. linearis*. The presence of straight, vertical burrows of greater burrow diameter, characterizes *S. linearis*. Several authors (e.g., Goodwin and Anderson, 1974; Dam, 1990) suggested that *Skolithos* resulted from erosion of the upper portion of *Monocraterion* tubes, and that both traces were produced by the same trace-maker (Goodwin and Anderson, 1974). Schlirf and Uchman (2005) provided a through in-depth comparison of *Skolithos* and *Monocraterion*. In doing so, they note that the debate over the taxonomy of *Skolithos* and *Monocraterion* largely revolves around material subsequently illustrated by Westergard (1931), who Schlirf and Uchman (2005) claim did not illustrate the original type material of Torell (1870), but what Westergard (1931) believed Torell may have intended. In addition, importantly, they state that the Lower Cambrian Mickwitzia sandstone is the one source for *Monocraterion tentaculatum*. However, only one specimen exhibiting "tentacles" (as indicated by its name) can be accounted for, while the few other specimens are funnel-shaped. Following a lengthy discussion, they note that *Skolithos* and *Monocraterion* (whatever its origin) have little in common. Thus, Schlirf and Uchman (2005) strongly recommended that specimens of *Skolithos*-like structures having funnel-shaped tops, not be referred to as *Monocraterion* since the taxonomic status of this specific structure remains problematic. Though most often indicative of shallow marine environments (Mángano and Buatois, 2004), *Skolithos* is a wide ranging form having also been found in deep marine (Crimes, 1977), and non-marine (e.g., Bromley and Asgaard, 1979; Buatois and Mángano, 2007) deposits.

Figure 2. ***Skolithos linearis,*** **central specimen exhibiting funnel-shaped top. Scale = 1 cm.**

Escape Structure
Fig. 3

Material. One slab with one specimen. Figured specimen NJSM 21864.

Description. Vertical to slightly curved burrow with well-defined , downward oriented deflections of adjacent sedimentary laminations. Diameter of burrow 15-20 mm, which expands into a triangular funnel-shaped opening at the very top approximately 35 mm across, preserved length 180 mm. Burrow is infilled with very coarse sand to gravel-sized sediment and lacks evidence of distinctive wall structure.

Remarks. Most authors have assigned escape structures as having been formed by bivalves (e.g., Pieńkowski, 1985), as well as polychaetes (e.g., Pickerill et al., 1977). The escape structure is preserved in a calcareous sandstone slab, the lower bedding surface of which is relatively rich in poorly preserved, largely abraded brachiopods. Interestingly, though brachiopods dominate the mega fauna, associated bivalves have also been documented from the Oriskany Formation (Albright, 1987). Thus, this structure was most likely produced by a bivalve, resulting from the pushing and pulling movements of the foot during upward escape activities (e.g., McCarthy, 1979).

Figure 3. Escape structure. Scale = 1 cm.

The Middle Devonian Onondaga Formation (formerly Buttermilk Falls Formation), as much as 82 m in thickness, consists of medium- to dark-gray, fossiliferous, clayey limestones and calcisilts which locally contain nodular and bedded black chert. The limestone weathers light- to medium-light-gray (Drake et al., 1996). In northwestern New Jersey, the formation is exposed in a northeast-southwest trend along the western side of Wallpack Ridge. Body fossils previously found include brachiopods, corals, cephalopods, and trilobites (Albright, 1987). Field investigation of clayey limestones and calcisilts of the Onondaga Formation in the Lake Maskenozha and Flatbrookville Quadrangles revealed the trace fossils *Nereites missouriensis* and *Psammichnites* isp. The Onondaga occurs in intermittent exposures along the shoreline of the Delaware River as well as Old Mine Road. Specimens of *Nereites* are best seen on grayish-orange, yellow-gray, and pale yellowish-brown weathering bedding surfaces along Old Mine Road. Investigation of well-exposed sections of the four members of the Onondaga Formation in Pennsylvania (see road log and stop descriptions in Inners and Fleeger, 2001), revealed that the trace fossil-bearing strata in New Jersey compare most favorably the Moorhouse Member (likely within the middle portion of this unit). This is the first reported occurrence of these trace fossils in the Delaware Water Gap National Recreation Area, and only a few exposures in New Jersey contained them. Ver Straeten (1996) provided an in-depth, detailed investigation of Onondaga strata, suggesting deeper water deposits for middle rocks of the Moorhouse Member (also see Epstein, 1970; Wolfe, 1977). A low diversity and moderate abundance of trace fossils were recovered from bioturbated sediments of the Onondaga Formation, distinguished by abundant *Nereites*. It would appear from the degree of bioturbation and the largely horizontal, sinuous movement the *Nereites* trace-maker, and the fine grain of the sediment, that an abundant food supply was available which the organism readily took advantage of so as to extract the maximum amount, and that there was likely a minimum of coarse sediment influx into the area. As such, both *Nereites* and *Psammichnites* indicate largely horizontal to slightly inclined feeding strategies occurring in relatively low-energy, well-oxygenated, moderately-deep marine waters with moderate sediment influx. As noted by MacEachern et al. (2007), most trace fossil suites assigned to the *Nereites* ichnofacies typically occur in deeper-water environments. Thus, though tempting to assign these ichnotaxa to a *Nereites* ichnofacies, they will be placed in an association, as Bromley (1996) noted the need for caution in labeling individual trace fossils (e.g., *Nereites*) for such assignment.

Ichnogenus *Nereites* MacLeay, 1839
***Nereites missouriensis* (Weller, 1899)**
Fig. 4

Material. At least 175 specimens with many more uncollected in the field. Figured specimen NJSM 21867.

Description. Mostly horizontal to sub-horizontal, unbranched, cylindrical to mostly ellipsoidal in cross section, winding to meandering burrows, preserved in hyporelief and epirelief. Trace diameter 2-7 mm, though highly variable within a single specimen, maximum preserved length 210 mm. Central backfilled tunnel consisting of well-spaced, transverse, scalariform, packeted menisci, having homogeneous fill similar to host rock (clayey limestone/calcisilt). Scalariform menisci 0.6-2.5 mm wide with distance between menisci 2-3 mm, typically less than, rarely equal to burrow diameter.

Remarks. Mángano et al. (2002a) noted that *Nereites missouriensis* typically occurs in monospecific assemblages. As such, interestingly in the Onondaga Formation strata, only a single specimen of *Psammichnites* occurs beside very abundant *Nereites*. *Nereites* has been interpreted to represent a grazing trace, combining the activities of feeding and locomotion, and likely produced by an infaunal worm-like form (e.g., Seilacher, 1983).

Fig. 4. ***Nereites missouriensis*****. Scale = 1 cm.**

chnogenus ***Psammichnites*****Torell, 1870**
Psammichnites **isp.**
Fig. 5

Material. One specimen. Figured specimen NJSM 21868.

Description. Straight to slightly curved, bilobed trail, 20-25 mm wide, length approximately 120 mm, preserved in epirelief. Surface of each lobe is elevated, somewhat rounded and smooth with sporadic, somewhat poorly preserved, closely spaced, fine, delicate transverse and less commonly obliquely oriented striae. Sporadic, longitudinally oriented fine ridges external of the bilobate structure are present, forming an acute angle. Median groove distinct, shallow, 2-3 mm wide. Whole trail divided up longitudinally into segments, 10-15 mm in length. There is minor evidence of somewhat arcuate backfill where the trail has collapsed.

Remarks. Overall, the generally poor preservation does not allow for ichnospecific assignment. Several investigators (e.g., Hofmann and Patel, 1989; Mángano et al., 2002b; Seilacher, 2007) have helped to clarify the internal structure and taxonomy of the complex morphology of this trace. Seilacher (2007) provides an excellent review of the many preservational variants of *Psammichnites*. Proposed trace-makers include crustaceans (Torell, 1870), worms (e.g., Matthew, 1890), mollusks (e.g., Glaessner, 1969), gastropods (Häntzschel, 1975), annelids (McIlroy and Heys, 1997), echiurans (Runnegar, 1982), halkieriids (Seilacher-Drexler and Seilacher, 1999), and a molluscan producer (Mángango et al., 2002b).

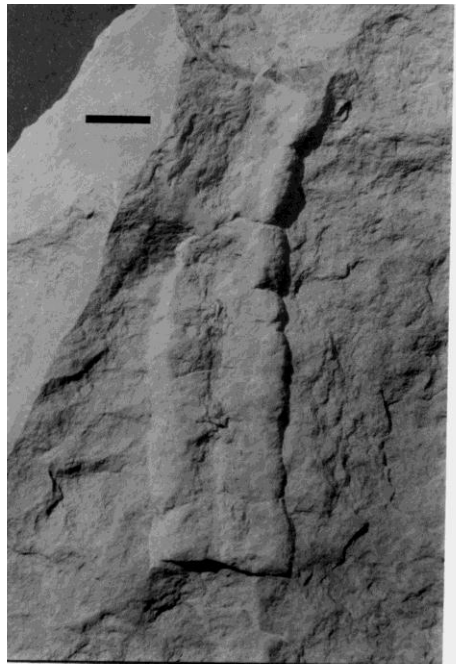

Figure 5. *Psammichnites* isp. Scale = 1 cm.

The Mahantango Formation (Upper Middle Devonian) is up to 400 meters thick and makes up the bulk of the Hamilton Group present in Pennsylvania and neighboring southeastern New York (Faill, 1985; Prave et al., 1996). It consists of predominantly fossiliferous mudstones and sandstones representing shallow marine deposits arranged in coarsening-upward cycles (Prave et al., 1996). The formation delineates prograding clastic wedges that were associated with the Devonian Acadian orogeny (e.g., Dennison and Hasson, 1976; Faill, 1985). Prave et al. (1996) concluded that cyclicity could be explained due to frequent progradation and retreat of what was basically a straight, tide-influenced shoreline onto a storm-dominated marine shelf. Thick deposits of the Mahantango Formation are present in a series of very steep, cliff exposures along Route 209 between Bushkill and Milford, Pennsylvania. Extensive fragmentation along bedding and cleavage intersections as well as weathering along vertical joint faces has resulted in an expansive apron of shale-chip rubble. As such, prospecting for trace fossils is typically limited to scattered small clasts consisting of an assortment of upper, middle, and lower portions of the steep exposure. However, recent rainstorms at that location resulted in slumping of large clasts that came from an isolated lowermost exposure of the Mahantango Formation not subject to mixing from above. Due to the distinctive freshness of the sheered face, the clasts were estimated to represent an approximate thickness of 6 m. The rocks are predominantly dark-gray mudstones, weathering varies from brownish-gray, olive-gray, pale-brown, and grayish red-purple. Detailed field investigation revealed the presence of the trace fossils *Cruziana* isp., *Dactylophycus quadripartitum*, *Diplichnites* isp., *Helminthoidichnites tenuis*, *Helminthopsis hieroglyphica*, *Planolites annularis*, *Planolites beverleyensis*, *Protovirgularia dichotoma*, *Treptichnus bifurcus*, and *Treptichnus pollardi*. Associated fossils include cephalopods and bivalves. Slattery (1993) provided a detailed description of a measured section which lies approximately 0.7 km north of the present site. Comparison to that of Slattery (1993) indicates that the deposits at the present site are most similar to a portion of Facies Association 1, dominated by locally fossiliferous, medium-to dark-gray mudstone. Slattery (1993) and Prave et al. (1996) interpret the mudstones of Facies Association 1 as that of hemipelagic deposits formed in an open marine setting (= offshore of MacEachern et al., 2007) below fair-weather weather base, reflecting relatively low energy levels. At the present site, the biogenic structures are dominantly horizontal traces of deposit-feeding organisms (e.g., *Planolites*, *Helminthopsis*, *Treptichnus*). As such, the overall diversity and characteristics of the dominantly horizontal, biogenic structures allow the trace fossil assemblage to be referred to the *Cruziana* ichnofacies (Frey and Seilacher, 1980).

Ichnogenus *Cruziana* d'Orbigny, 1842
***Cruziana* isp.**
Fig. 6

Material. Ten specimens. Figured specimen NJSM 21880

Description. Straight to slightly curved specimens preserved in convex hyporelief. Trace width 3-4 mm, and up to approximately 60 mm long, composed of two low-relief symmetrical lobes, each 0.75-1 mm wide, separated by a shallow median furrow. Transverse scratch marks are somewhat faint though discernible.

Remarks. Trilobites are commonly designated as the marine producers of *Cruziana* (Seilacher, 1970). Though *Cruziana* is typically a shallow marine form (e.g., Crimes et al., 1977; Mángano et al., 1996), it has also been reported from marginal marine (e.g., Buatois and Mángano, 1997), deep marine (e.g., Pickerill, 1995), and continental strata (e.g., Bromley and Asgaard, 1979).

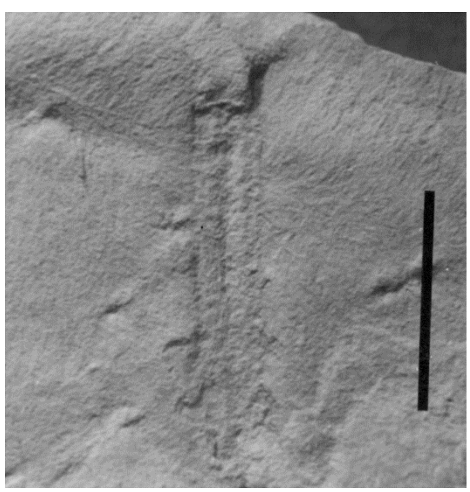

Figure 6. *Cruziana* isp. Scale = 1 cm.

Ichnogenus *Dactylophycus* Miller and Dyer, 1878
***Dactylophycus quadripartitum* Miller and Dyer, 1878**
Fig. 7

Material. Two specimens. Figured specimen NJSM 21882.

Description. Small, branching burrows which radiate outward in one direction in a flabellate manner. Four branches can be distinguished, most of which terminate in a point-like configuration. Individual branches have a diameter 1.5-2 mm, with maximum length up to 8 mm. Preserved in convex hyporelief. On one specimen, one of the branches exhibits delicate annulations on the lateral ridges which are transverse to a narrow medial furrow, imparting a bilobate appearance, while on a second branch these features are very faint and poorly preserved. The lack of annulations on the other branches is due to recent erosion. A second specimen possesses a faint median groove on two of the branches taking up more than half of the width, while a third branch exhibits faint annulations.

Remarks. *Dactylophycus* is a relatively rare trace fossil having been reported from Upper Ordovician shallow-water marine deposits in the Cincinnati area, Ohio (Miller and Dyer, 1878; Osgood, 1970), and Upper Cambrian? to Lower Ordovician lagoonal strata from Newfoundland, Canada (Fillion and Pickerill, 1990). Thus, to the author's knowledge, this represents the youngest occurrence of this trace fossil. Osgood (1970) did an excellent restudy of the type materials of Miller and Dyer (1878). Suzanne Constanza (Harvard University, Botanical Museum) was kind enough to send this author close-up photographs of the three slabs exhibiting *D. quadripartitum* (all numbered HBM 3174). Overall, as noted by Osgood (1970), the diameter of the branches (2-4 mm) as well as the maximum length (up to 15 cm) is greater than the present specimen. The same is true for specimens (2-2.3 mm wide, up to 10 mm long) reported by Fillion and Pickerill (1990). Though Osgood (1970) considered *Dactylophycus* a deposit feeder, both he and Fillion and Pickerill (1990) noted the absence of a main shaft. Interestingly, though one of the present specimens shows a hint of what could have been a portion of a shaft, there is no evidence of a master tunnel (cf. Osgood, 1970).

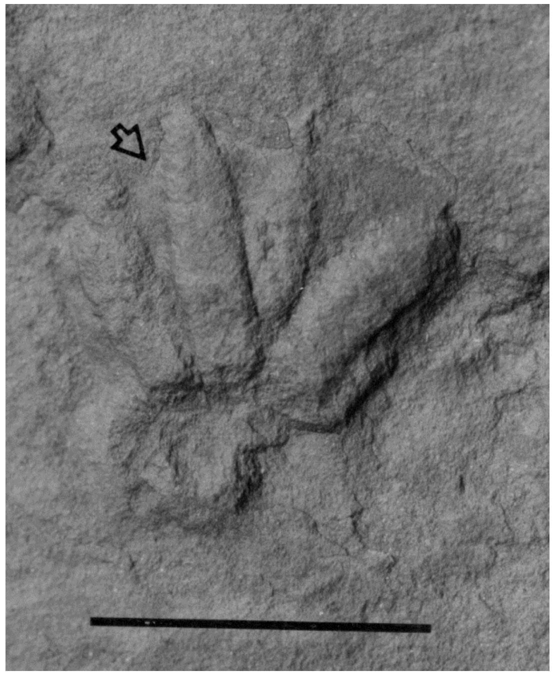

Figure 7. *Dactylophycus quadripartitum*. Note annulations (arrow) on ridge which are transverse to a median furrow. Scale = 1 cm.

Ichnogenus *Diplichnites* Dawson, 1873
***Diplichnites* isp.**
Fig. 8

Material. Several specimens on a single slab. Figured specimen NJSM 21883.

Description. Main trackway consists of two parallel rows of transverse ridges, lacking a median groove. Individual ridges are spindle-shaped, partially smooth but also exhibit striations which are parallel to the trackway axis. Series comprises 5 distinct, three additional poorly preserved imprints. Ridges are 4-7 mm long, 0.75-2 mm wide. Trackway length of series of 5 ridges is 15 mm, preserved width 7-15 mm, and they are spaced 2-2.5 mm apart. Other portions of the slab exhibit sporadic spindle-shaped ridges, but not occurring in a series.

Remarks. Keighley and Pickerill (1998) discussed the problems associated with both the initial introduction of the *Diplichnites* by Dawson (1873), as well as the subsequent adaptation for varied morphologies (see also Fillion and Pickerill, 1990). *Diplichnites* has been recorded in marine (e.g., Seilacher, 1955) as well as non-marine (Lucas et al., 2004) deposits. Due to the problems associated with the varied trackway morphologies linked to *Diplichnites*, assignment is best left in open ichnospecific nomenclature.

Figure 8. ***Diplichnites*** **isp. on same slab as** ***Dactylophycus*****. Scale = 1 cm.**

Ichnogenus ***Helminthoidichnites*** **Fitch, 1850**
Helminthoidichnites tenuis **Fitch, 1850**
Fig. 9

Material. Ten specimens. Figured specimen NJSM 21884.

Description. Small, simple, smooth, unbranched, straight to gently winding burrows preserved in hyporelief and epirelief. Burrow fill is structureless and similar to surrounding sediments. Width ranges from 0.5-1.5 mm, and is constant within individual specimens; maximum observed length 80 mm. Crossovers between different specimens are present.

Remarks. This trace is very similar to *Unisulcus minutes* reported by Hitchcock (1858). Interestingly, Jensen (1997) noted that theoretically, the distinction between *Helminthoidichnites* and *Planolites* is the lack of difference between the fill and surrounding sediment, a difficult assessment in burrows of very small size. This ichnogenus ranges in age from the Late Precambrian to the Cretaceous (e.g., Narbonne and Aitken, 1990), and has been reported from marine (e.g., Jensen, 1997) and non-marine (e.g., Buatois et al., 1997) deposits.

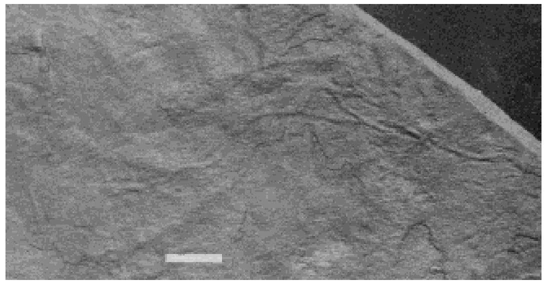

Figure 9. ***Helminthoidichnites tenuis*****. Scale = 1 cm.**

Ichnogenus *Helminthopsis* Heer, 1877
***Helminthopsis hieroglyphica* Heer in Maillard, 1887**
Fig. 10

Material. Three specimens. Figured specimen NJSM 21885.

Description. Horizontal, smooth, unbranched, unlined, irregularly meandering trails or burrows. Diameter varies from 1.0-1.5 mm, though is constant within each specimen; maximum preserved length of 65 mm. Preserved in negative epirelief and positive hyporelief. Specimens display straight segments separated by irregular meanders.

Remarks. The greater sinuousity and irregular meandering differentiates *Helminthopsis* from *Helminthoidichnites* (Hofmann and Patel, 1989). *Helminthopsis* is distinguished from *Gordia* by its lack of self-overcrossing and tendency to meander (Pickerill, 1981). *Helminthopsis* is interpreted to represent a grazing trail produced by a deposit-feeding organism (Buatois et al., 1997). In marine environments, proposed trace-makers include worm or worm-like forms (e.g., Chamberlain, 1971) and priapulans and polychaetes (Książkiewicz, 1977). Non-marine producers may include nematodes and arthropods (Mángano et al., 1996; Metz, 1996).

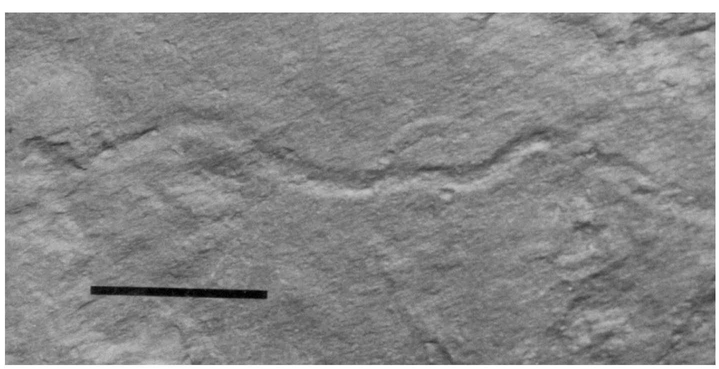

Figure 10. ***Helminthopsis hieroglyphica*****. Scale = 1 cm.**

Ichnogenus *Planolites* Nicholson, 1873
***Planolites annularis* (Walcott, 1890)**
Fig. 11

Material. One specimen. Figured specimen NJSM 21886.

Description. Straight, horizontal, unbranched, ellipsoidal, burrow exhibiting regularly spaced annulations. Burrow diameter 4 mm and constant, preserved length 25 mm. A burrow lining is not present, and the burrow fill is lighter in color and is of coarser grain size than that of the host sediment. Preserved in convex hyporelief.

Remarks. *Planolites* is distinguished from *Palaeophycus* by the lack of a wall lining and having burrow fill that differs from the host strata (Pemberton and Frey, 1982). It has been reported from marine (e.g., Heinberg and Birkelund, 1984) as well as non-marine (e.g., Pickerill, 1992) strata. The specimen of *P. annularis* is most comparable to a similar ichnotaxon in Osgood (1970, Plate 77, figure 3), as well as weathered specimens from the same location collected by the author (Metz, 1996). *Planolites* has been interpreted to represent active backfilling of a temporary burrow by a mobile deposit-feeding organism, such as polychaetes (Pemberton and Frey, 1982).

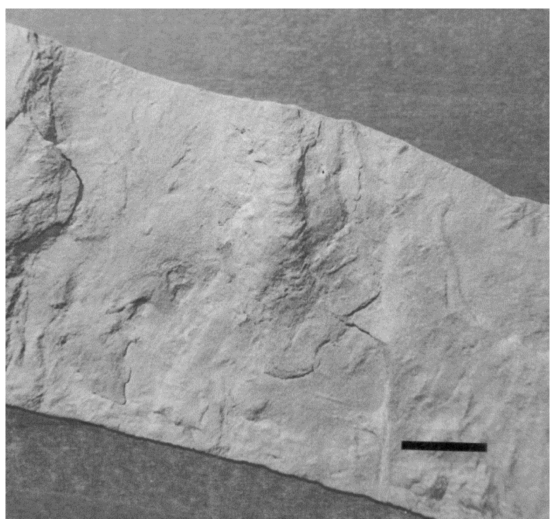

Figure 11. *Planolites annularis*. Scale = 1 cm.

Ichnogenus *Planolites* Nicholson, 1873
***Planolites beverleyensis* (Billings, 1862)**
Fig. 12

Material. Two specimens. Figured specimen NJSM 21887.

Description. Simple, horizontal, smooth-walled, circular to ellipsoidal unlined burrows. Diameter 3 mm, fairly constant within each specimen; maximum length observed 50 mm. Preserved in convex hyporelief.

Remarks. Refer to Fig. 1.

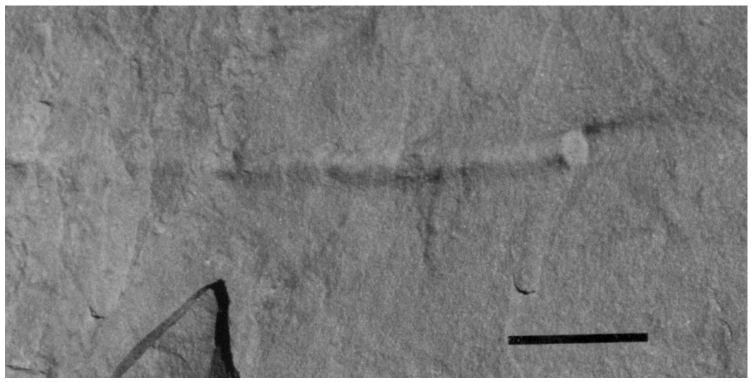

Figure 12. *Planolites beverleyensis*. Scale = 1 cm.

Ichnogenus *Protovirgularia* M'Coy, 1850
***Protovirgularia dichotoma* M'Coy, 1850**
Fig. 13

Material. One specimen. Figured specimen NJSM 21888.

Description. Straight, unbranched, horizontal trace, 30 mm in length, 3 mm in diameter, a portion of which exhibits a median ridge with paired, lateral, wedge-shaped, closely-spaced appendages. Lateral appendages are normal to median ridge, with length of 1 mm. Preserved in convex hyporelief.

Remarks. Recently, a number of researchers have provided excellent in-depth reviews of *Protovirgularia* (e.g., Han and Pickerill, 1994; Mángano et al., 2002a). Based on revisions by Han and Pickerill (1994), citing morphologic variation exhibited by specimens of *Protovirgularia*, and comments by other authors, the present specimen is assigned to *P. dichotoma*. Potential trace-makers proposed for *Protovirgularia* include annelids or arthropods (e.g., Greiner, 1972), scaphopods (Seilacher and Seilacher, 1994), and bivalves (e.g., Han and Pickerill, 1994; Seilacher and Seilacher, 1994). Since bivalves are present in the Mahantango Formation, they are the likely trace-maker responsible for *Protovirgularia dichotoma*.

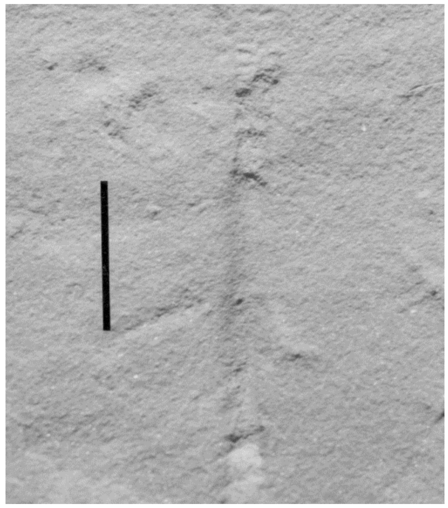

Figure 13. ***Protovirgularia dichotoma*****. Scale = 1 cm.**

Ichnogenus *Treptichnus* Miller, 1889
***Treptichnus bifurcus* Miller, 1889**
Fig. 14

Material. Eight specimens. Figured specimen NJSM 21890.

Description. Straight to curved horizontal burrows, 1-2 mm in diameter, individual segments 4-7 mm in length, with short projections 1-1.5 mm extending from junctures between longer segments. Several of the projections curve upward, others are swollen compared to the main burrow. Burrow fill is similar to the host rock. Preserved in convex hyporelief.

Remarks. Where the burrow segments of *T. bifurcus* are straight, projections occur on alternate sides, where curved, the projections occur on the outside of the curved portion. The projections represent bedding plane indications of vertical shafts (Maples and Archer, 1987; Buatois and Mángano, 1993). Several examples of *T. bifurcus* appear to grade into *T. pollardi*, and vice-versa. Proposed trace-makers for *Treptichnus* include vermiform animals and insect larvae (Buatois and Mángano, 1998). Most recordings of *Treptichnus* have been from non-marine (e.g., Uchman et al., 2004) environments, though marginal marine (e.g., Archer et al., 1995) to deep marine (Crimes et al., 1981) examples have also been noted.

Figure 14. ***Treptichnus bifurcus*****. Scale = 1 cm.**

Ichnogenus *Treptichnus* Miller, 1889
***Treptichnus pollardi* Buatois and Mángano, 1993**
Fig. 15

Material. At least forty-five specimens. Figured specimen NJSM 21891.

Description. Simples, zigzag, curved, or straight burrow segments possessing small pits, indication openings of vertical shafts, located either at junction between, or at some position within individual segments. Diameter of segments 1 mm, individual segments range in length from 4-8 mm. The maximum number of burrow segments is 7, most have less. Sporadic burrow segments have short projections at junctures, similar to *T. biburcus*. There are several examples of small pits as well as zigzag burrows lacking pits reflecting upper layer and lower layer morphology, respectively. Preserved in convex hyporelief and concave epirelief.

Remarks. The presence of pits and absence of projections distinguishes *Treptichnus pollardi* from *T. pollardi*. In the present material, the presence of pits lacking evidence of segments, as well as segments lacking pits compares quite favorably to that illustrated by Buatois and Mángano (1993, p. 219, figure 3B, D). As with *Treptichnus bifurcus*, most recordings of *T. pollardi* have come from non-marine environments (e.g., Buatois and Mángano, 1998).

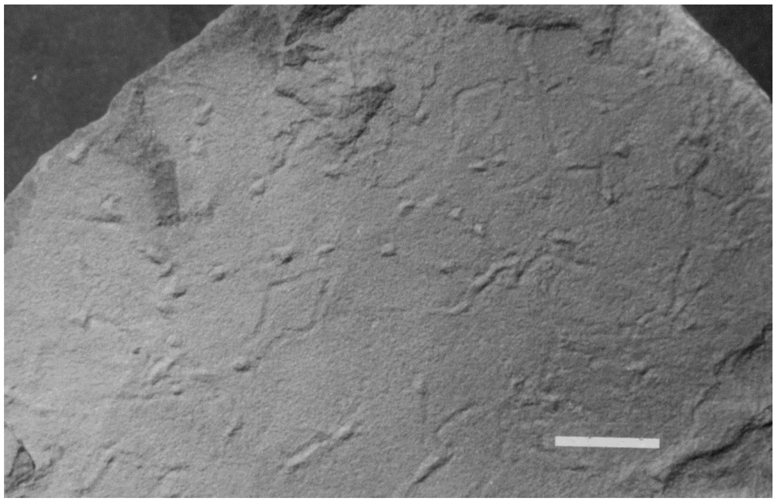

Figure 15. *Treptichnus pollardi*. Scale = 1 cm.

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GEOLOGY ALONG THE TRAIL ACROSS MOUNT PAUL MEMORIAL COUNTY PARK, CHESTER, NEW JERSEY

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Abstract

The Mt Paul Trail offers a rare opportunity to observe good exposures of a wide range of rock types (granite, gneiss, slate, conglomerate, and mylonite) of widely varying age (Precambrian, Ordovician, and Triassic) and a wide range of geologic structures (The Flemington Fault, The Peapack – Ralston Fault, The Peapack Klippe, and some plunging folds) along about 3 miles of gentle hill slope.

Introduction

The Mt Paul trail is a small part of the rather extensive trail system of Chester, New Jersey. The trail system is currently being upgraded and several trail plans are about to be implemented. The Conservation Resources Inc. serves as a consultant for this project. According to Conservation Resources "The Chester Trails Plan is a collaborative effort by Chester Township and Chester Borough to develop a plan for a trail ecosystem that will connect significant open space, environmental, cultural and historic features in the Chesters." This year (2012) Conservation Resourses published a 36 page plan on their web site that includes a "geology trail" with 47 points of interest. Several of these points will have QR codes on posts along the trail that will describe the rock exposures.

One of the more interesting of the Chester trails is a proposed "blue" or secondary spur of Patriots Path that will cross through the Mount Paul Memorial County Park. An important step in the development of the Mt. Paul trail was the recent purchase of the "Shale Pits Property" by the Schiff Natural Lands Trust. This purchase has connected State Route 24 (510) or Mendham Rd. W to Mt Paul Memorial Park. The Shale Pits include several acres of excellent exposure of Ordovician shale.

Geologic Setting

An easy way to think of the geologic development of the earth over the last couple of billion years is to think of two supercontinents (**Rodinia and Pangea**) that existed by themselves, one after the other, in a huge global ocean. **Rodinia** formed first during Precambrian time. Most of Chester formed during the assembly of **Rodinia** about one billion years ago. Then about 550 million years ago at the end of Precambrian time **Rodinia** broke up and slid away from a huge hot-spot to form smaller continents (Puffer, 2002). Sediments that eroded off of one of these continents (Laurentia) was deposited in the ocean as sand and mud about 480 million years ago during early Paleozoic time (specifically the Cambrian and Ordovician periods). Some of the mud lithified into the slates exposed in the quarry north of Mt. Paul. Then Laurentia and the other continents slowly drifted together and gradually merged into the second continent (**Pangea**) about 300 Million years ago. When **Pangea** started to break apart about 200 Million years ago, volcanic rocks extruded out of the fissures (Puffer et al. 2009); and rift valleys formed that later widened to form the Atlantic Ocean. Some of the sand that was deposited in one of the rift valleys lithified into the red sandstone found at the top of Mt. Paul. Therefore, Mt. Paul has participated in a large fraction of the geological development of the earth. It's actually that simple, although geologists are still working out the details.

We are, therefore, left with three kinds of rock exposed around Mt. Paul: 1. Late Precambrian granites and metamorphic rocks formed deep under Rodinian mountains; 2. Slates and sandstones formed from mud and sand deposited in Paleozoic oceans, and 3. Sandstone deposited in rift valleys formed during the break-up of Pangea.

1. Late Precambrian rocks

a. Losee Metamorphic Suite (Ylo) – is a volcanic rock that was metamorphosed about 1.1 billion years ago into gray gneiss composed of quartz and plagioclase feldspar (the two most abundant minerals in the earth's continental crust) together with minor quantities of biotite and magnetite. The volcanic rock was extruded onto a large island arc (a volcanic island chain such as Japan) and was metamorphosed during the assembly of Rodinia. The metamorphic activity was very intense and caused some of the rock to melt. The portion that melted became an igneous rock called a trondhjemite (Puffer and Volkert, 1991). The Losee is commonly veined with quartz and magnetite. Where the magnetite content exceeds about 20 percent it was mined as an iron ore at several locations throughout the Chester area (Puffer, 2001).

b. Byram Intrusive Suite (Ybh) – is a pink to tan granite that intruded into the Losee Metamorphic Suite about 0.8 billion years ago. The granite is composed of microcline feldspar, quartz, plagioclase feldspar, and hornblende with minor magnetite. A few abandoned magnetite (iron ore) mines in the Chester area are also found in the Byram granite. Most of the Chester area is underlain by Byram Granite and Losee Gneiss.

2. Paleozoic Rocks

a. Jutland klippe upper unit B (Ojtb) – is red and green shale that has been slightly metamophosed into a slate. The rock was deposited as shallow marine mud in an Ordovician ocean about 480 to

490 million years ago. The rock is interbedded with dolomite, siltstone, sandstone, and conglomerate. The rock contains conodont, graptolite, and brachiopod fossils (Volkert et al. 1990) supporting a marine environment. These Ordovician rocks are locally confined to a structure known as a Klippe which is an isolated block of rocks separated from underlying rocks by a fault (see accompanying cross section).

b. Beekmantown Group – are gray Ordovician dolomites and limestones (Obu and OBl) that contain conodont fossils (the teeth of extinct marine ell-like animals).

c. Allentown Dolomite – is a gray dolomitic mudstone (OCa) deposited in shallow water as indicated by mud cracks and algal stromatolites. The rock is of late Cambrian to early Ordovician age (about 490 million years).

d. Leithsville Formation – is gray dolomite (Cl) containing Archaeocyathid fossils (a horn coral animal that lived in shallow warm coral reefs). The rock was deposited during the late Cambrian Period (about 500 million years).

e. Hardyston Quartzite- is a Cambrian sandstone (Ch) that contains trilobite fossils.

3. Triassic rocks

A layer of dark red Quartz-pebble conglomerate that was deposited during the late Traissic period (about 200 to 210 million years ago) is exposed at the top of Mt. Paul. The rock is sandstone containing abundant pebbles composed of quartz. The red color is due to iron oxide (rust). The rock layers are not exposed at the surface but provide evidence of their existence as common pebbles and cobbles in the soil. The red Triassic sandstones of New Jersey contain common dinosaur and alligator footprint fossils together with mud-crack, raindrop, and ripple-mark providing evidence of nonmarine deposition. Some vegetation is also preserved in the sandstone as thin layers of coal-like material. Most of the Triassic dinosaurs including those that lived in the Chester area were small, about the size of a person. Most of these dinosaurs became extinct during the eruption of volcanic basalt flows that suddenly covered eastern North America, eastern South America, northwestern Africa, and southern Europe at the beginning of the Jurassic period (Puffer et al., 2009). But a few species survived and evolved into larger Jurassic and Cretaceous dinosaurs.

The Mt. Paul Trail

From the parking area about 1/8th mile south of Rt. 24 hike west into the abandoned shale/slate quarry area. The quarrying activity has resulted in excellent exposures of the Ordovician Jutland klippe upper unit B. The area has been mapped by Volkert et al, (1990) who have measured the strike and dip at several locations around the quarry. The dip measurements average about 70^o to the south-east and strike about 45^o to the north-east. The beds have formed folds that plunge at low angles (about 6 to 16 degrees to the northeast and the southwest. The composition of the red slate and the green slate is the same except for the oxidation state of the iron in the red slate that has oxidized to rust.

Continuing south and west on the trial very little Ordovician rock is exposed. However, as the summit of Mt Paul is approached within about ¼ mile in any direction pebbles and cobbles of red conglomerate are exposed in gullies and in eroded soil exposures. This red Triassic rock is about 300 million years younger than the Ordovician rocks. The pebbles in the conglomerate are composed of quartz and are cemented together with silt enriched in red iron.

As the trail from the summit descends toward the southwest more Ordovician rock is encountered until Gladstone Brook is reached. Gladstone Brook marks the fault contact with Byram granite that is nicely exposed on the valley wall to the west. The fault along the brook is a major New Jersey fault known as the Flemington Fault. The younger Ordovician rocks on the east side dropped down while the much older Byram Granite on the west side lifted up as indicated by the arrows on the cross section along the line A'-A. Therefore, the relative motion makes this portion of the fault a normal-type fault. The contact at Gladstone Brook is a contact between rock deposited about 480 million years ago and granite intruded about 800 million years ago. The granite and the adjacent Losee Gneiss were therefore uplifted several miles from their original sites of intrusion or metamorphism to reach their current position. This also means that the several miles of rock overlying the Byram and Losee have eroded away resulting in countless cubic miles of sediments, most of which were deposited as sand and mud in shallow marine environments.

It can also be observed that the Byram Granite has experienced the effects of crushing and grinding during movement along the Flemington Fault and has become finer grained and foliated. The sheared, fine-grained portion of the granite has become a rock type known as mylonite.

Acknowledgments

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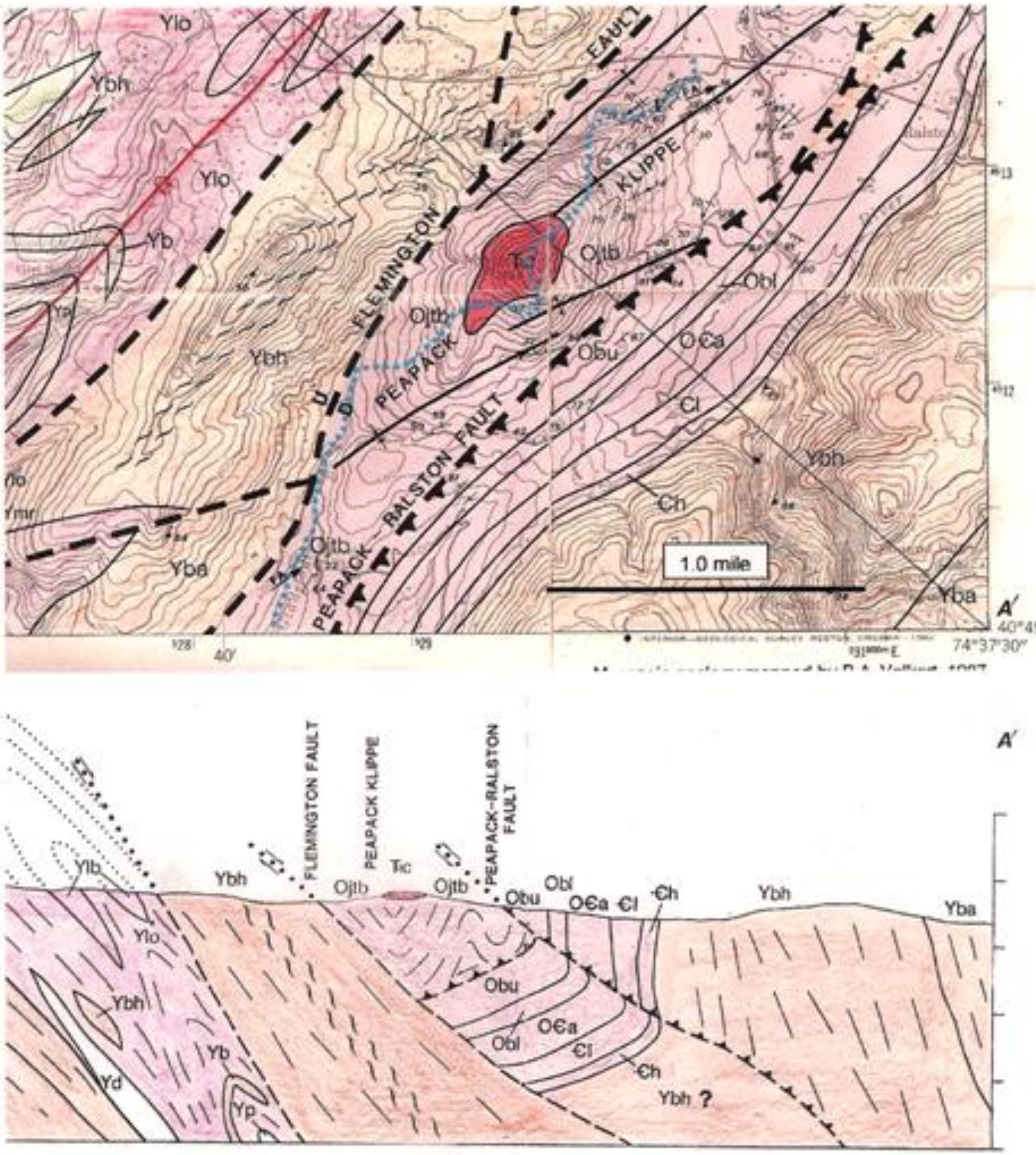

Geologic map and cross section after Volkert et al. (1990). Mt. Paul is located near the center of the Peapack Klippe. The peak of Mt. Paul is Triassic sandstone (TrC) mapped in dark red and the base is Ordovician slate (Otjb) mapped in light red. The Precambrian Byram Granite (Ybh) west of the Flemington Fault is mapped in orange and the Precambrian Losee Gneiss (Ylo) is mapped in pink. The trail mapped as a blue line.

TELLING THE STORY OF GLACIAL LAKE PASSAIC USING PUBLIC LANDS

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Introduction

Lake Passaic, the largest glacial lake entirely within New Jersey, formerly filled the central Passaic River basin between Second Watchung Mountain and the Highland Escarpment. The history of the lake includes preglacial river drainage unlike that of today, early glacial lakes during at least two pre-Wisconsinan glaciations, three glacial lake stages during the late Wisconsinan glaciation, and three postglacial lakes. Although the basin of glacial Lake Passaic today is largely suburban and urban, many of the places important to an understanding of the history of the lake are within municipal, county, or federal parks. Spillways and sluiceways for the three late Wisconsinan glacial lake stages can still be seen in their natural states. Depositional features such as ice-contact deltas and fans, and varves, can be observed in a few spots, and several overlooks give a sense of scale of the lake. Additionally, the glacial history of the lake is part of the back story of current environmental issues in the basin, including flooding, water supply, and land use, providing a conceptual link between ongoing social concerns and our local geologic heritage.

This paper summarizes the history of the lake, explains the stratigraphy of the glacial aquifers in the lake basin, and describes nine sites on public land in the basin that illustrate key features and sediments of the lake. The site descriptions can serve as a self-guided field trip for individuals or small groups. Parts of this discussion are from a 2010 New York State Geological Association field trip (Stanford, 2010a). Additional sources of information on Lake Passaic include Cook (1880), who named the lake and described shoreline features; Salisbury (1893), Salisbury and Kummel (1894, 1895), Kummel (1895), and Salisbury (1902), who mapped deposits and shoreline features in detail, described the glacial and postglacial lake stages, and delineated preglacial and glacial drainage routes; and recent surficial geologic mapping at 1:100,000 scale (Stone and others, 2002) and 1:24,000 scale (for example, Stanford, 2003, 2007a, 2007b; quadrangle maps covering the entire basin are available as pdfs at www.njgeology.org/pricelst/geolmapquad.htm), which refine the history and timing of preglacial drainage, glacial and postglacial lake stages, ice-margin positions and ice-flow directions. The maps also use well, boring, and geophysical data to map subsurface stratigraphy and bedrock-surface topography. A simple account of the lake history is provided in the summer 2007 edition of the N. J. Geological Survey newsletter (www.njgeology.org/enviroed/newsletter/v3n2.pdf).

History of Lake Passaic

The arcuate trace of the west-dipping cuesta ridges of the Watchung Mountains, formed on resistant basalt flows, reflects the canoe-shaped geometry of the Watchung syncline. The Ramapo Fault on its west side brings Proterozoic gneiss against the syncline. Erosion of the soft shale and sandstone enclosed between the basalts and the gneiss excavated a 30 mile long by 10 mile wide basin between Second Watchung Mountain and the Highland Front. This basin is punctured by only two sets of gaps: the paired gaps at Little Falls and Paterson and at Short Hills and Millburn. These gaps were cut by the Hudson River in the Pliocene (Johnson, 1931; Stanford, 2010b). An early Pleistocene glaciation, which occurred sometime between 2.5 Ma (Ma=million years ago) and 800 ka (ka=thousand years ago), blocked these gaps, forming a glacial lake (Lake Watchung) in the southernmost part of the basin (fig. 1A). Deeply weathered and eroded deltaic and lacustrine-fan sand and gravel in the Bernardsville area are the only surviving evidence for this lake. The height of the Bernardsville delta indicates that this lake rose to an elevation of more than 450 feet, about 100 feet higher than the highest late Wisconsinan stage. This higher level occurred because the early Pleistocene glacier advanced further south than the later glaciers and so blocked the gap at Moggy Hollow, which remained open during the later glaciations.

The early Pleistocene glacier also diverted the Hudson from its course through the Watchungs. A new local drainage network, including the lower Passaic and Raritan rivers, was established after retreat of the glacier (fig. 1B). The Passaic exited the basin through the Short Hills-Millburn pair of gaps. During the early and middle Pleistocene, between 2 Ma and 150 ka, this new drainage incised valleys into shale bedrock between the basalt ridges, and eroded away most of the early Pleistocene glacial deposits.

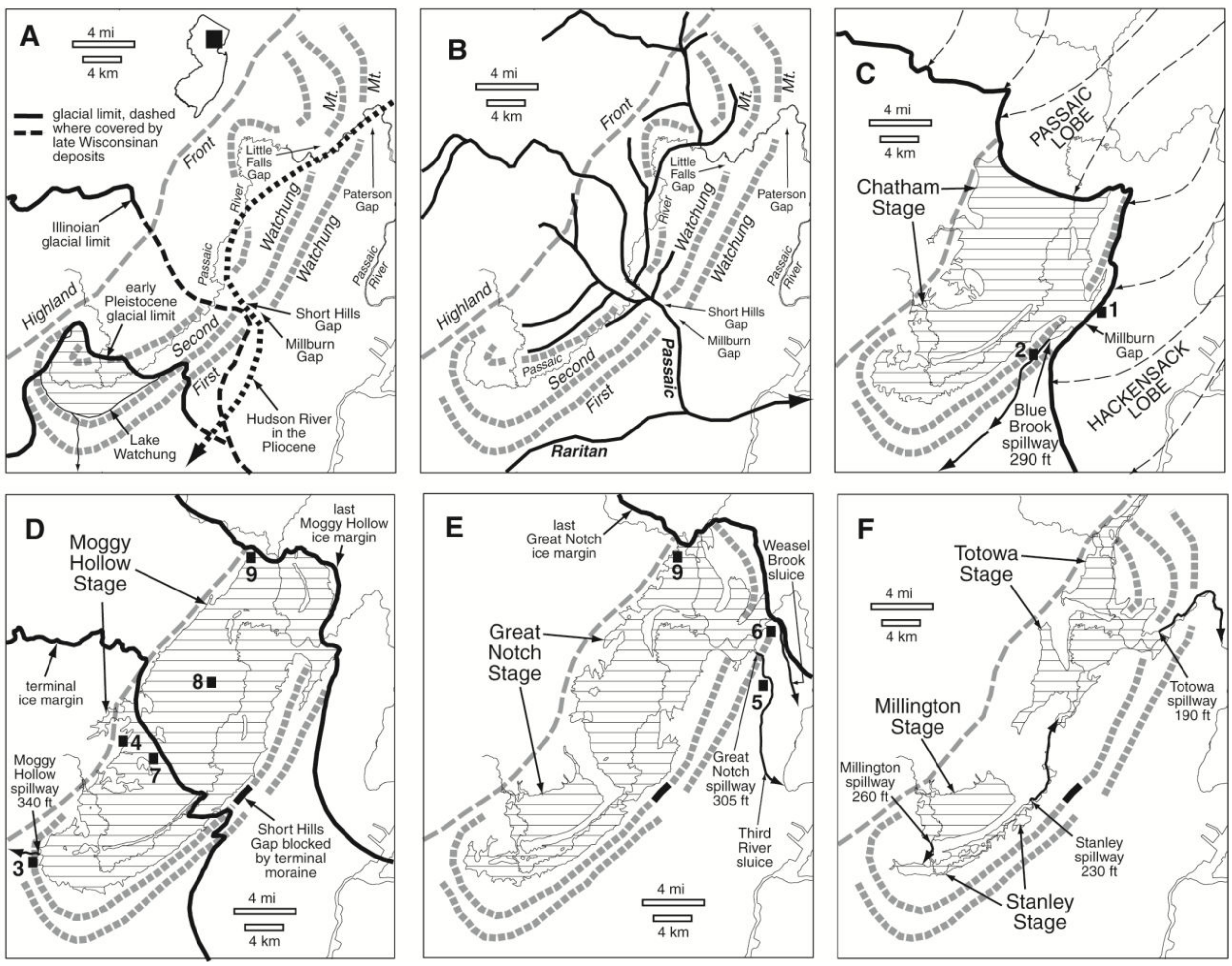


Figure 3. History of Lake Passaic. A. Pliocene route of Hudson River, limits of early Pleistocene and Illinoian glaciation (dashed where covered by late Wisconsinan deposits), and extent of early Pleistocene Lake Watchung. Inset shows location. B. Fluvial drainage after retreat of early Pleistocene glacier. C. Advancing Hackensack Lobe of late Wisconsinan glacier blocks Millburn Gap, establishing the Chatham stage. D. Continued advance of Hackensack Lobe to terminal position blocks Short Hills Gap, establishing the Moggy Hollow stage. Moraine fills Short Hills Gap, allowing Moggy Hollow stage to expand northward. Ruling shows its maximum extent, just before uncovering of Great Notch. E. Maximum extent of Great Notch stage. Uncovering of Great Notch released 2.5 mi³ of water down the Third River sluice. Uncovering of Paterson Gap drained the Great Notch stage and released 1.2 mi³ of water down the Weasel Brook sluice. E. Postglacial stages. Numbered boxes show educational sites.

The next glacier to enter the Passaic basin is probably of Illinoian age (*Illinoian* is the North American stage term for the next-to-last continental glaciation), which reached its maximum position at about 150 ka. Illinoian glacial deposits in and south of the Short Hills gap indicate that this glacier sealed the gap and thereby created an Illinoian version of Lake Passaic, probably similar to the highest (Moggy Hollow) late Wisconsinan stage. Illinoian deltaic and lacustrine-fan sand and gravel, and lake-bottom silt and clay, occur beneath late Wisconsinan deposits in the central and southern sections of the basin ("overramp zone' on fig. 2), where they fill the incised pre-Illinoian valleys in shale and also extend over the low interfluves between the valleys. The Illinoian fill in Short Hills Gap rises to about 200 feet in elevation (section AA', fig. 2), so during Illinoian retreat a lake was maintained at that level until the gaps at Little Falls and Paterson were uncovered. This lake was much shallower than the late Wisconsinan recessional stages and limited the vertical accretion of Illinoian valley fill sediments north of Short Hills Gap. Erosion of the Illinoian valley fill during the interval between Illinoian retreat and arrival of late Wisconsinan ice (150 to 25 ka) was minimal, because basalt in Little Falls Gap at an elevation of 180 feet established a high base level for the basin, limiting fluvial incision.

The late Wisconsinan (*Wisconsinan* is the North American stage term for the most recent continental glaciation; the late Wisconsinan is the last phase of this glaciation, which reached its maximum extent about 25 ka) history of the lake includes three glacial lake stages (fig. 1C, D, E) and three postglacial lakes (fig. 1F). The advancing late Wisconsinan glacier in the Newark Basin included two lobes of ice: the Hackensack and Passaic. The Hackensack Lobe filled the broad lowland between First Watchung Mountain and the Palisades and was unimpeded by topography. It therefore flowed further south than the Passaic Lobe, which was slowed as it advanced over the Watchung Mountains. When the advancing Hackensack Lobe blocked Millburn Gap (Site 1), the Chatham stage of the lake formed (fig. 1C), controlled by a spillway at the head of the Blue Brook valley at an elevation of 290 feet (Site 2). Filling of this lake stage buoyed the Passaic Lobe ice and allowed the Passaic Lobe to ramp over rather than erode pre-existing sediments as it advanced to the terminal position. These pre-existing sediments included Illinoian deposits and delta, fan, and lake-bottom sediments laid down in the Chatham stage in front of advancing ice (fig. 2, sections AA', BB'). An earlier, shallow, advance-phase lake probably formed when the Hackensack Lobe blocked Paterson Gap. At this time the Passaic River was ponded to the level of the Illinoian fill in Short Hills Gap. However, since this fill was only about 20 feet (6 m) higher than the Little Falls base level, the lake was shallow, accumulated little sediment, and did not buoy the Passaic Lobe.

Continued advance of the Hackensack Lobe blocked the Blue Brook valley and then sealed Short Hills Gap. This caused the Chatham stage to rise 50 feet to the Moggy Hollow stage (fig. 1D), which was controlled by a spillway into the Raritan basin across Second Watchung Mountain near Far Hills (Site 3). The 50-foot rise in lake level further buoyed the Passaic lobe, allowing it to ramp over the back end of a large fan-delta complex built into the lake at the terminal position. Till of the terminal moraine was deposited along the back end of this large delta deposit (fig. 2, sections AA', BB').

Deposition of till of the terminal moraine in Short Hills Gap filled the gap to an elevation of more than 400 feet, about 50 feet higher than the Moggy Hollow stage, allowing this lake to expand northward as the ice front retreated. When Great Notch, a gap through First Watchung Mountain east of Little Falls, was uncovered, the Moggy Hollow stage dropped 80 feet to the Great Notch stage (fig. 1E). This drop released about 2.5 mi^3 of water down the Third River sluice (Site 5) downhill from Great Notch into the lower Passaic. The configuration of the glacier margin at this time ("last Moggy Hollow ice margin" on fig. 1D) is fixed by the last ice-contact deltas deposited in the Moggy Hollow stage: one along the Highland Front near Riverdale (Site 9) and one in the north end of the Preakness valley, where the ice front was lodged along the crest of Second Watchung Mountain.

Further retreat of the Hackensack Lobe uncovered the north end of First Mountain at Paterson Gap (Site 6), allowing the Great Notch stage to drain down the Weasel Brook sluice. This flood released about 1.2 mi^3 into the lower Passaic valley. Again, the position of the glacier margin at this time is fixed by Great Notch-stage deltas, including a delta at Riverdale (Site 9) and one at the north end of the Preakness valley. Both of these deltas are reworked from adjacent Moggy Hollow deltas by meltwater draining from local lakes adjacent to Lake Passaic, held in by ice deployed as shown in fig. 1E.

After the Great Notch stage drained, sediment dams held in three postglacial lakes. At Totowa, a large lacustrine fan blocked the Passaic valley in a narrow reach downstream from Little Falls (along Riverview Drive at Laurel Grove Cemetery, where I-80 crosses the river) forming the dam for the Totowa stage. This lake extended up the Pompton and Ramapo valleys, where several large ice-contact deltas were deposited in it, including the Pompton plain, which nearly filled the northern bay of the Totowa lake, and a large delta which fills the Ramapo valley further north. Sand terraces were deposited by the Rockaway, Whippany, and Passaic rivers in the western and southern bays of the lake.

The terminal moraine formed a dam across the Passaic at Stanley (in the Passaic River Parkway, a Union County park, just upstream from Mount Vernon Avenue in Summit) forming the Stanley stage. The Passaic and Dead rivers deposited terraces of silt and fine sand in this shallow lake. The Great Swamp basin north of Lake Stanley was dammed by the terminal moraine-delta complex to the northeast, which blocked the former preglacial Passaic valley at Chatham. Postglacial Lake Millington filled this basin and was controlled by a spillway at Millington in a gap on Long Hill, the basalt ridge that forms the southern rim of the basin. The postglacial Passaic River exits the Great Swamp basin by way of this gap. Erosion of the spillway here formed the Millington gorge of the Passaic River, which is partly preserved as open space on the east side of Pond Hill Road, south of South Maple Avenue, in Bernards Township.

Loantaka and Great Brooks eroded sand from the delta fronting the terminal moraine to the north of the Great Swamp and deposited it in terraces in the northeast end of Lake Millington. These sand terraces overlie lake clay in the Great Swamp. After Lake Millington drained the terraces were incised and channeled by Loantaka and Great Brooks. Permafrost developed, forming lenses of ground ice within the terrace sand. When these lenses melted they left shallow basins in the terraces. Basins may also have formed from wind erosion of fine sand and silt on unvegetated terrace surfaces. The channeling and basin development created an intricate landform mosaic of sandy islands, marshes, and swamps, which today contribute to the ecologic diversity of the swamp. These features are traversed by trails in the wilderness area of the Great Swamp National Wildlife Refuge, the Great Swamp Outdoor Education Center of the Morris County Park Commission in Chatham Township, Loantaka Brook Reservation in Harding and Chatham Townships, and the Environmental Education Center at Lord Stirling Park of the Somerset County Park Commission in Bernards Township.

Each of the postglacial lakes was controlled by spillways on erodible material (sand and gravel for Totowa, till for Stanley, and weathered and fractured basalt for Millington). The spillways were gradually lowered by erosion and the lakes eventually drained to leave the broad floodplains and marshes that now occupy their floors. Today, these floodplains and marshes are valuable open spaces that provide thousands of acres of flood storage and wildlife habitat in an otherwise fully built environment.

Chronology

Radiocarbon dates and varve counts provide age control for the glacial and postglacial events. Late Wisconsinan ice arrived at the terminal moraine by 21 to 20.5 ka (dates in radiocarbon years), based on radiocarbon dates of concretions in varves in the Great Swamp (Reimer, 1984; Stone and others, 2002), and on radiocarbon dates of organic materials above and below late Wisconsinan till in New Jersey and Long Island (summarized in Stanford, 2010a). Glacial varves counted from split-spoon samples obtained from borings in the Great Swamp indicate that ice stood at the terminal moraine for about 750 years (Reimer, 1984). The Great Swamp basin received sediment while ice stood at the terminal moraine but was sealed by the moraine from further glacial sedimentation once the ice margin retreated from the moraine. Complete varve counts have not been made for the lake basins north of the moraine but correlation to the varve and radiocarbon chronology in the Hackensack and Hudson valleys suggests that the Great Notch stage had drained by about 19.5 ka. This timing leaves at most about 1000 years for deposition of the glacial recessional deposits north of the moraine, with perhaps a few hundred additional years for deposition of the Pompton Plain delta in the postglacial Totowa stage, which was glacially sourced. Lake-bottom sediment in the northern basin of the lake is as much as 200 feet thick (fig. 2, section BB', north end), giving a minimum accumulation rate of 2.5 inches/year for these deposits.

The postglacial lakes lasted longer. All three postglacial lakes formed at around 19.5 ka, when the Great Notch stage drained. Lake Millington persisted to at least 14 ka, based on a concretion date from silt and clay deposited in the lake (Reimer, 1984), but had drained by 10 ka, based on peat dated to 9.6 ka (E.W. B. Southgate, personal communication) and pine, spruce, and birch pollen in peat (Meyerson, 1970) filling ground-ice basins on the stream terraces laid down as the lake lowered and drained. The Stanley and Totowa stages are not dated but probably drained earlier than Lake Millington because their dams were on erodible sediment rather than fractured basalt.

Hydrogeology of the Lake Passaic Basin

The valley-fill deposits of the Lake Passaic basin produce over 15,000 million gallons per year and are the most productive glacial aquifers in New Jersey (Hoffman and Quinlan, 1994). They are the principal or sole water source for several municipal systems and so are classified by the EPA as a sole-source aquifer. The "overramp zone" (fig. 2), where Illinoian and Chatham-stage lacustrine sand and gravel are thick, extensive, and fill the buried pre-Illinoian fluvial valleys, is the most productive sector of the aquifer system (see distribution of production wells on fig. 2). The Illinoian and Chatham-stage sands and gravels are overlain by low-permeability silt and clay lake-bottom deposits (Site 7) and till. Till matrix is chiefly silty sand to sandy silt but till is of lower permeability than its grain size alone would indicate because it has been highly consolidated in most places by the weight of overlying ice. The till and lake-bottom deposits act as confining or semi-confining layers. Some wells bored through these sediments into underlying sand and gravel flowed at the surface when first drilled, indicating artesian conditions, although pumping in recent decades has greatly lowered the piezometric surface and wells no longer flow (Meisler, 1976; Hoffman and Quinlan, 1994). Gradients on the piezometric surface in the overramped sand and gravel indicate flow toward Short Hills Gap, along the original pre-Illinoian fluvial valley gradient, although there is much perturbation of the natural gradient by pumping-induced cones of depression. Recharge to the confined, overramped sand and gravel is by infiltration of precipitation on the delta fronting the terminal moraine, which connects to the overramped sand and gravel in the subsurface (fig. 2, section BB'), by stream loss from the Passaic and Whippany rivers where they flow across the delta-moraine complex, and from vertical leakage of surface water through the lake-bottom sediment and till capping the overramped sand and gravel at and north of the moraine.

Neither Illinoian nor late Wisconsinan ice advanced south of the moraine complex (fig. 2, late Wisconsinan limit), and pre-Illinoian deposits in this area are thin, patchy erosional remnants largely above the level of the valley fill. Thus, no sand and gravel occurs in the valley fills south of the moraine complex. These fills instead consist of late Wisconsinan lake clays and silts deposited atop Illinoian lake clays and silts. These deposits confine the underlying bedrock but provide no water themselves.

North of the overramp zone, ice of the Passaic Lobe was not buoyed by Chatham-stage lake water, since Millburn Gap was not yet blocked by Hackensack ice. The Passaic Lobe thus eroded rather than overran any Illinoian stratified deposits on the landscape during advance. In fact, ice of the Passaic Lobe was so erosive in this sector that it scoured overdeepenings as much as 300 feet deep into shale bedrock in the lowlands between the basalt ridges (fig. 2, late Wisconsinan glacial overdeepenings). The valley fill in this northern sector of Lake Passaic thus consists exclusively of recessional lacustrine sediments on a basal till on bedrock, with no buried pre-advance materials (fig. 2, section BB', north end). The recessional lacustrine deposits are chiefly silt and clay lake-bottom sediments but in places, particularly in the valley fill along the Highland Front, permeable lacustrine-fan sand and gravel underlies the lake-bottom material. These fan deposits supply several municipal well fields in this area (Site 9). The fan gravels were deposited at the mouths of subglacial tunnels as the ice front retreated, and were later buried by lake-bottom sediment that accumulated vertically as fine sediment settled out of the lake. In places the fans rise above the lake-bottom fill, or were deposited on high areas of the lake bed, and so are exposed as knolls and ridges (Site 8). In places along the Highland Front they built up far enough to reach the lake surface, where they prograded into the lake to form deltas. Other deltas along the Highland Front were deposited where ice-lateral meltwater channels draining down the Front entered the lake. The outcropping deltas and fans act as recharge conduits for the buried fans, where the deposits are physically connected.

Shallow-water delta sands were deposited in the postglacial lakes by meteoric or meltwater streams, especially in the Totowa stage. They lie above the thick deposits of lake clay laid down in the Moggy Hollow and Great Notch stages (fig. 2, north end of BB', and fig. 11, section AA'). These sands may be unconfined aquifers if sufficiently thick, particularly adjacent to rivers, which are in hydraulic connection with the sands. However, no large-capacity wells tap these deposits in the Lake Passaic basin.

Flooding is another hydrologic consequence of the glacial history of the basin. The northward reroute of the Passaic through the gaps at Little Falls and Paterson after Short Hills Gap was filled in the Illinoian and late Wisconsinan glaciations had the effect of lengthening and flattening the course of the river compared to the pre-Illinoian route. The pre-Illinoian baselevel in Short Hills Gap is at an elevation of 70 feet; the postglacial baselevel at Little Falls is at an elevation of 180 feet. Gradients of the Passaic and Pompton Rivers across the basin are 1.5 and 0.8 feet/mile, respectively, for 40 and 12 miles, respectively, upstream of Little Falls today, compared to 15 and 6.5 feet/mile for 12 and 20 miles, respectively, to the same origin points upstream of Short Hills Gap in the pre-Illinoian valleys. For most of the flat reaches above Little Falls the Passaic and Pompton flow on low-permeability lake clay, which prevents infiltration of floodwater. The flat gradient and impermeable substrate cause floodwater to back up upstream of Little Falls. Urbanization since the mid-20th century has worsened flooding by increasing the area of impervious surface in the basin and thereby adding more storm runoff to streams. Damaging floods along the Passaic and Pompton occurred in 1810, 1865, 1878, 1882, 1886, 1896, 1902, 1903, 1920, 1936, 1945, 1951, 1952, 1968, 1984, 1999, 2007, 2010, and 2011 (March and August) (www.nj.gov/dep/passaicriver/). This recurring hazard has led to a long and continuing history of flood-control plans, levee construction, and property buy-outs.

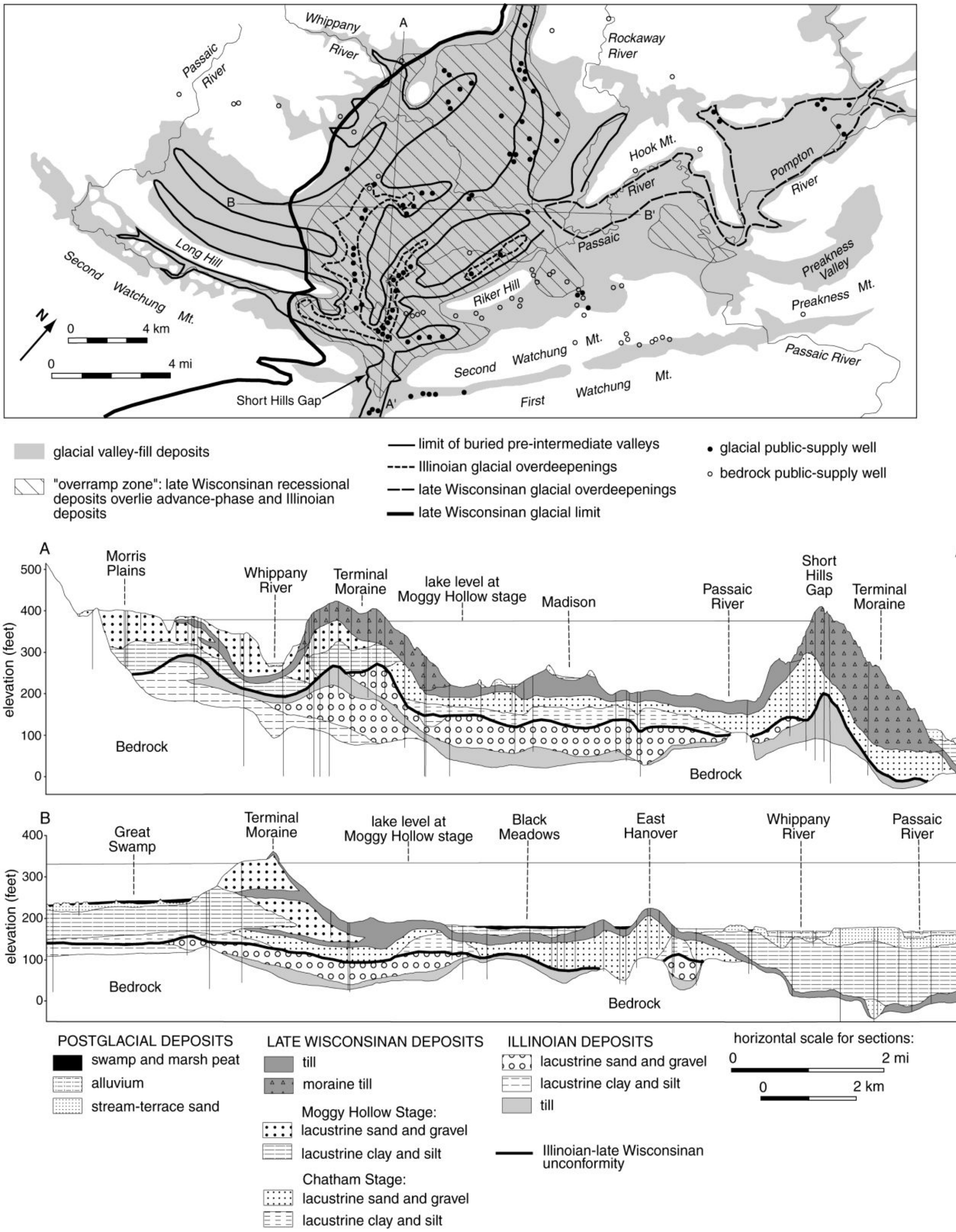


Figure 4. Stratigraphy and geomorphology of the valley-fill aquifer system in the Lake Passaic basin.

Selected Educational Sites

Site 1: Millburn and Short Hills Gaps (Washington Rock overlook, South Mountain Reservation, Millburn). This view is best seen from the top of a former quarry at 40°43.779'; 74°18.186', which is on the south side of a yellow-blazed trail leading west from a concrete viewing platform at 40°43.717'; 74°18.062'. The platform provides an alternate, though more restricted view, of Millburn Gap. These gaps, originally cut by the Hudson River in the Pliocene (fig. 1A), later served as the exit for the Passaic River through the Watchungs until the Illinoian glacier partly filled Short Hills Gap with moraine deposits, diverting the river northward through the Little Falls and Paterson gaps. During late Wisconsinan glacial advance the Hackensack Lobe, moving southward down the broad lowland to the east (left) blocked Millburn Gap, forming the Chatham stage in the Passaic basin west of the gap (to the right) (fig. 3). The Chatham stage was controlled by a spillway into Blue Brook in the valley between First and Second Mountains. Continued advance across the Blue Brook valley and onto Second Mountain closed this spillway and caused the lake to rise 50 feet to the Moggy Hollow level. Over the approximately 750 years that ice stood at the terminal moraine, Short Hills Gap was filled with till, allowing the Moggy Hollow stage to expand northward as the ice front retreated.

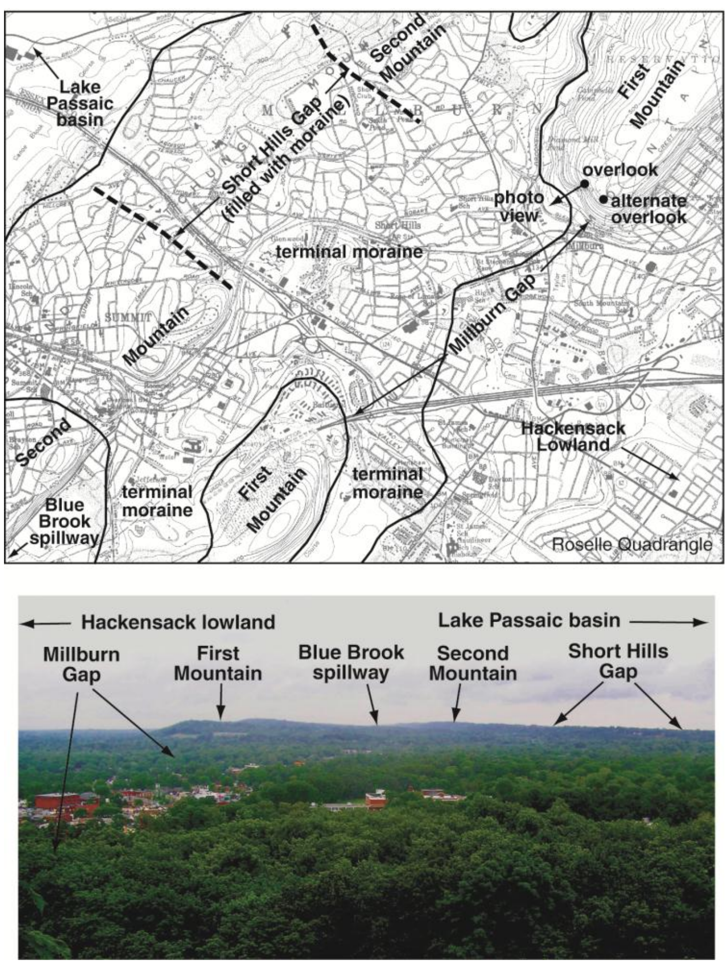


Figure 3. Map and photo of Millburn and Short Hills gaps.

Site 2: Blue Brook Sluiceway (Deserted village of Feltville, Watchung Reservation, Berkeley Heights). With Millburn Gap closed, water in the basin was ponded to the level of the Blue Brook spillway to form the Chatham stage of Lake Passaic. The spillway was on the divide at the head of Blue Brook between First and Second Mountains at an elevation of about 290 feet (fig. 4). The spillway area was later covered by outwash laid down after the glacier advanced across the valley. The channel cut by the outflow is well preserved in Watchung Reservation, particularly near the deserted village of Feltville. This channel, and the channels at Moggy Hollow and Great Notch (sites 3 and 5), carried, at the very least, the full discharge of the Passaic River that currently passes the Great Falls at Paterson. Adding to this meteoric discharge is the volume of glacial meltwater generated from the ice sheet within the Passaic basin. The lower trail along the north bank of the sluiceway east of the village provides views across the channel, for example, from the old shale quarry at 40°40.970'; 74°23.029'. Narrow side ravines along the south bank at 40°40.708'; 74°23.210' and 40°40.838'; 74°23.029' were cut by tributaries after the main channel was downcut by the outflow. Upstream of Feltville, Lake Surprise is an artificial pond built in the sluiceway. Downstream from Feltville the sluiceway broadens a bit before narrowing and dropping over a small waterfall on basalt bedrock downstream of Seeleys Pond. Material eroded from the sluiceway was flushed downvalley (past the present location of US Route 22), where it was later buried by gravels of the Plainfield outwash plain. Discharge through the sluice can be estimated from channel dimensions and slope using the Manning equation (Chorley and others, 1984). At its narrowest point at Feltville, the channel is about 200 feet wide, with a slope from the spillway to the baselevel on basalt at Seeleys Pond of 0.006. For a flow depth of 20 feet, and a Manning roughness value of 0.045, which is typical of channels in bedrock, the channel discharges 65,000 ft^3/s, at a velocity of 16 ft/s. For a flow depth of 10 feet, the channel discharges 22,000 ft^3/s, at a velocity of 10 ft/s. The 1903 flood of record for the Passaic basin discharged 34,000 ft^3/s at Great Falls in Paterson. A 15 to 20-foot deep flow through the channel at Feltville could thus accommodate a large surplus of glacial meltwater in addition to flood-stage meteoric basin discharge.

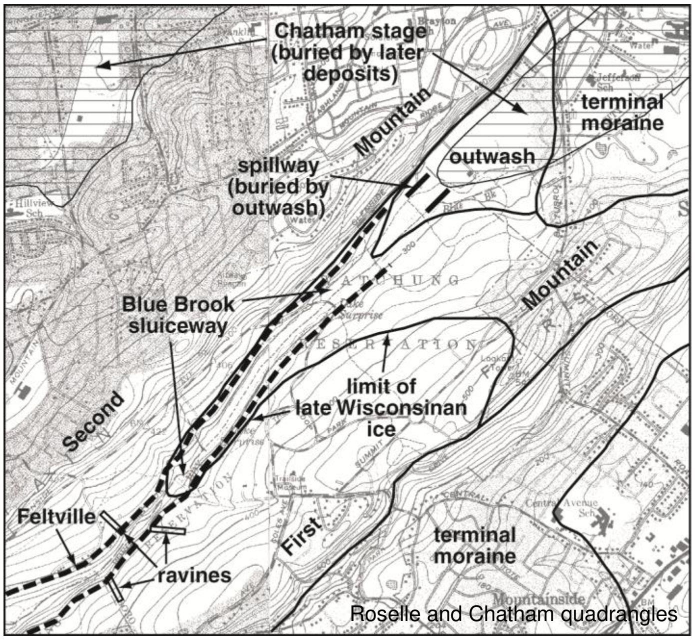


Figure 6. Map of Blue Brook sluiceway.

Site 3: Moggy Hollow Sluiceway (Leonard J. Buck Garden, Far Hills). With Short Hills Gap (and Great Notch and Paterson Gap) closed, the low point on Second Mountain is at the head of Moggy Hollow near Far Hills. The spillway itself is crossed by Liberty Corner Road (Route

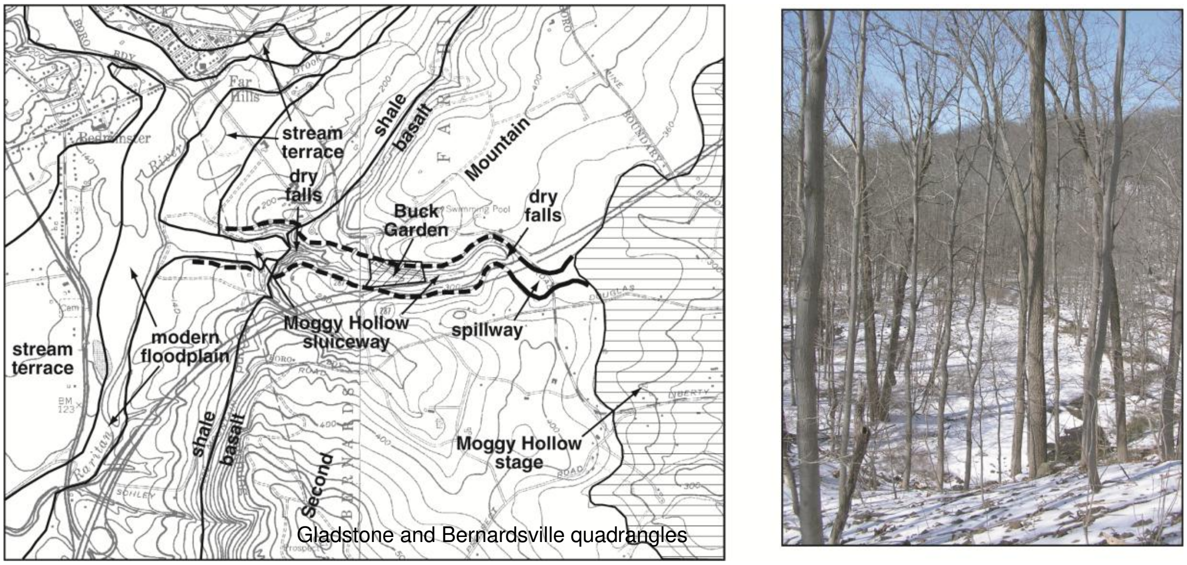


Figure 7. Map of Moggy Hollow sluiceway and photo of upper dry falls.

512) just south of I-287 (at 40°40.326'; 74°36.651'), and is marked by a marsh on the east side of the road. Moggy Hollow is the sluiceway leading 1.5 miles down the west slope of Second Mountain from the spillway to the North Branch of the Raritan River (fig. 5). Leonard J. Buck Garden (40°40.390'; 74°37.427') is a 33-acre formal rock garden run by the Somerset County Park Commission. It is situated within the channel. The basalt outcrops and slopes that formed cascades and rapids within the channel are now design elements of the garden. Between the garden and the spillway, a wild segment of the hollow is preserved as open space owned by the Raritan Headwaters Association. This segment of the channel (accessible from a small gravel pull-off on the south side of Route 512 just north of I-287) contains a 20-foot dry waterfall (see photo, fig. 5), a wetland in a former plunge-pool downstream from the fall, and a rocky bed of angular basalt cobbles on the channel floor. The sluice downstream to the west of Buck Garden is on private property and is not accessible, but contains another dry waterfall on basalt, broadens after crossing onto shale bedrock and terminates at the North Branch of the Raritan River. The size of the sluiceway (20 to 60 feet deep, 400 to 600 feet wide, about 1 mile long in basalt) indicates that much basalt was eroded by the outflow, but there is no basalt-rich gravel fan at the mouth of the hollow or in the fluvial terraces along the Raritan. This anomaly may indicate that the basalt gravel was eroded by the Raritan and buried by non-basalt-rich gravel after flow through the hollow ceased. Another possibility is that much of the channel incison occurred during the Illinoian and early Pleistocene glaciations, when Moggy Hollow also functioned as a spillway, and that the resulting gravel fan has been eroded and weathered away.

Discharge through the hollow, as at the Blue Brook sluice, can be estimated from channel dimensions and slope using the Manning equation. The channel averages 200 feet wide with a flow depth of about 10 feet and has a slope from the spillway to the base of the lower falls of 0.04. These dimensions, using the 0.045 roughness value for bedrock channels, indicate that Moggy

Hollow would discharge 54,000 ft^3/s of water at a velocity of 27 ft/s. Like the 20-foot deep flow at Feltville, a 10-foot deep flow through Moggy Hollow thus could accommodate a large surplus of glacial meltwater on top of flood-stage meteoric basin discharge. Moggy Hollow is significantly steeper than the other Lake Passaic sluiceways, which is why it is deeper and more prominent as a landform than the others. It also operated for a longer period, from the time of arrival of the glacier at the terminal moraine at about 21 ka to the time the ice uncovered Great Notch at perhaps 20 ka or slightly later. This chronology indicates about 1000 years of operation for Moggy Hollow, compared to at most a few hundred years for the Blue Brook and Third River sluices.

Site 4: Overlook of the Moggy Hollow Stage (Fort Nonsense, Morristown National Historic Park, Morristown). Fort Nonsense is on the Highland Front, which forms the west wall of the lake basin. The view to the east from the fort (40°47.636'; 74°29.273') over Morristown shows the width and depth of the lake (fig. 6). Morristown is built on a delta laid down in the Moggy Hollow stage and so marks the lake surface. The highest points on the terminal moraine just back of Morristown rose above the lake, forming an island in the Moggy Hollow stage (fig. 1D). The moraine and delta later emerged as a peninsula extending to Long Hill at Chatham when the lake dropped to the Great Notch stage (fig. 1E). Second Mountain on the horizon was the eastern shore of the lake. The moraine forms a continuous ridge across the basin from Morristown to Chatham, well above the extensive swamps and marshes on the former lake bottom to the north and south. The moraine thus provides a transportation corridor from New York City westward across New Jersey that was followed by both colonial-era roads and, later, by the Morris and Essex Railroad and its successors the Delaware, Lackawanna, and Western Railroad, and New Jersey Transit. This strategic position was one reason Washington decided to encamp his army here during the winters of 1776-1777 and 1779-1780 while the British occupied New York.

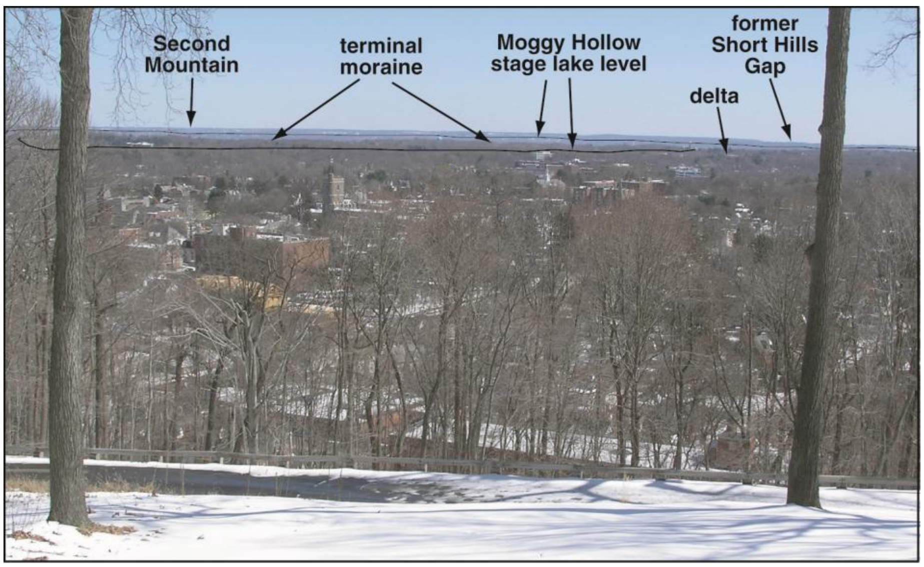


Figure 8. View to east from Fort Nonsense.

Site 5: Third River Sluiceway (Alonzo Bonsal Wildlife Preserve, Montclair). With Short Hills Gap filled with till, the Moggy Hollow stage expanded northward in the basin until Great

Notch, a gap in First Mountain, was uncovered. This allowed the lake to drop 80 feet to the Great Notch stage. The outflow from this drop consisted of an initial sudden release of 2.5 mi³ of water, followed by a steady discharge equal to that formerly draining through Moggy Hollow. These flows cut a channel into the till-over-sandstone upland southeast of Great Notch that is now occupied by the upper reach of Third River (fig. 7). This channel is not as steep as the Moggy Hollow and Blue Brook sluiceways, and so is not as deeply incised, and most of it is urbanized, but a segment can still be seen in its natural state in the Alonzo Bonsal Wildlife Preserve in Montclair. A footbridge across Third River at 40°50.964'; 74°11.272' is within the channel, with the south bank adjacent. A footpath through woods along the north bank provides views across the channel (for example, at 40°51.059'; 74°11.335' and 40°50.999'; 74°11.242'), particularly when leaves are off the trees. Downstream, to the east of Broad Street in Bloomfield, the channel leads out onto the Brookdale outwash plain, which continues about 4 miles down the Third River valley to downtown Bloomfield. Some of the outwash gravel in the plain is from erosion of the sluice. A smaller channel branches from the main sluiceway near Grove Street, Montclair, and descends to the Brookdale plain through Yantacaw Brook Park in Montclair. It may have been cut by an overflow during the initial flood surge.

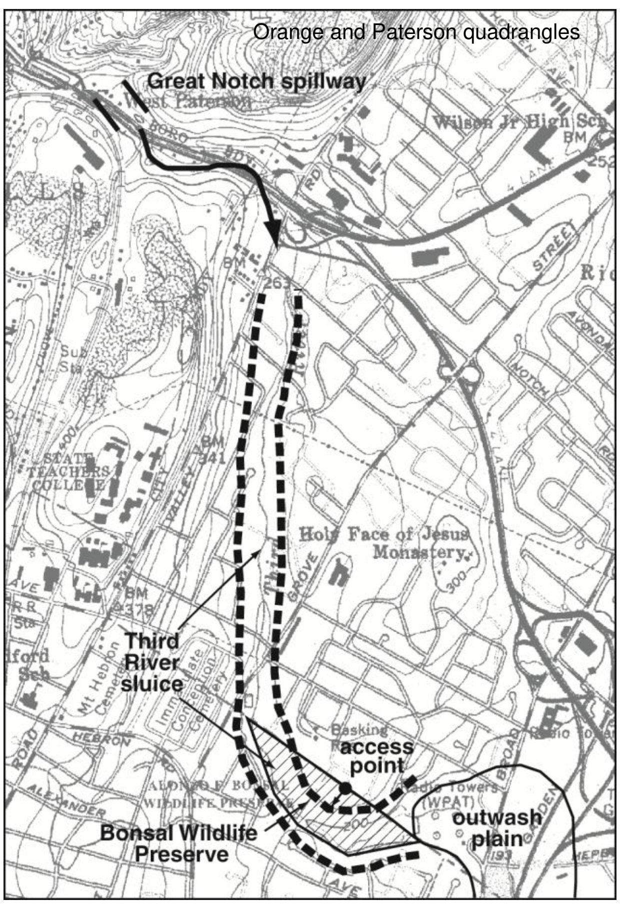


Figure 9. Map of Third River sluiceway.

Bankfull discharge through the sluice at the Bonsal Preserve, using the Manning equation with a channel width of 300 feet, a flow depth of 20 feet, a slope of 0.009 from the spillway to the outwash plain, and a roughness of 0.045, is 125,000 ft³/s. At this rate, the initial 2.5 mi³ release would take about a month to drain through the sluice and would do so with a velocity of about 20 ft/s. More rapid drainage would require overflow through the Yantacaw channel. The post-release steady outflow could easily be accommodated by the Third River sluice alone. A half-bankfull channel geometry (200 feet wide with a 10 foot flow depth) discharges 27,000 ft³/s, which was probably sufficient for non-peak meteoric and meltwater outflow.

Site 6: Weasel Brook Sluiceway (overlook of Paterson, Garrett Mountain Reservation, Paterson). The Great Notch stage drained when the north end of Garrett Mountain (a segment of First Mountain that forms the south abutment of Paterson Gap) was deglaciated. This drainage released 1.2 cubic miles of water to cut the Weasel Brook sluice along the base of First Mountain south of the gap (fig. 8). The release point is at the base of the cliff below the overlook (40°54.362'; 74°10.488'). The sluiceway, which now contains NJ Route 19 and the former Erie and Delaware, Lackawanna, and Western railroads, is entirely urbanized but can still be seen as a broad valley up to 50 feet deep between Broad Street and Hazel Street in Clifton, and is crossed by bridges on the Garden State Parkway and US Route 46. The sluice geometry is somewhat unusual. It is steep and narrow at the head, where it was walled, and perhaps floored, by ice on the east side, and broadens and flattens south of Broad Street, where it runs out across a till-on-sandstone upland. Downstream of US Route 46 the sluice branches into two subvalleys. The westerly branch leads to the head of the Brookdale outwash plain in the Third River valley. The easterly branch leads into several small ice-dammed lakes on the west side of the lower Passaic valley. These subvalleys may have conducted outflow simultaneously or sequentially from west to east as the ice margin retreated to the northeast.

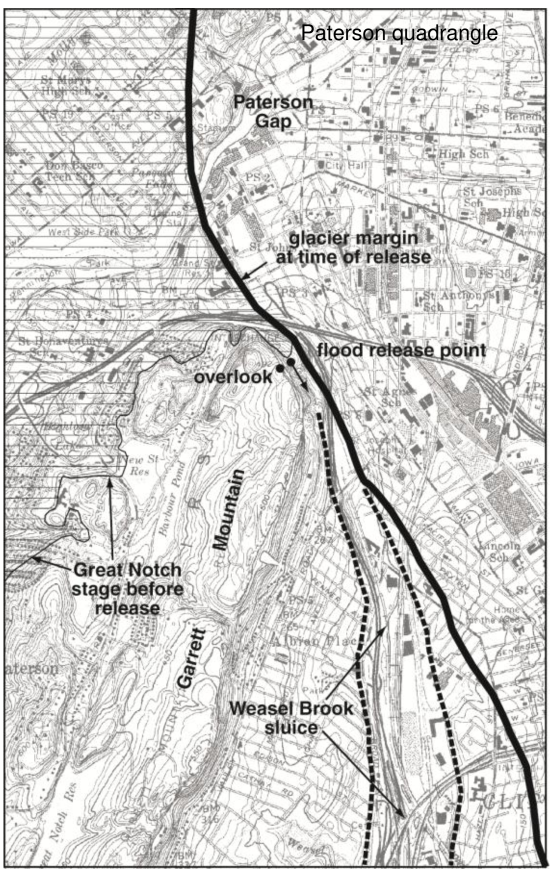
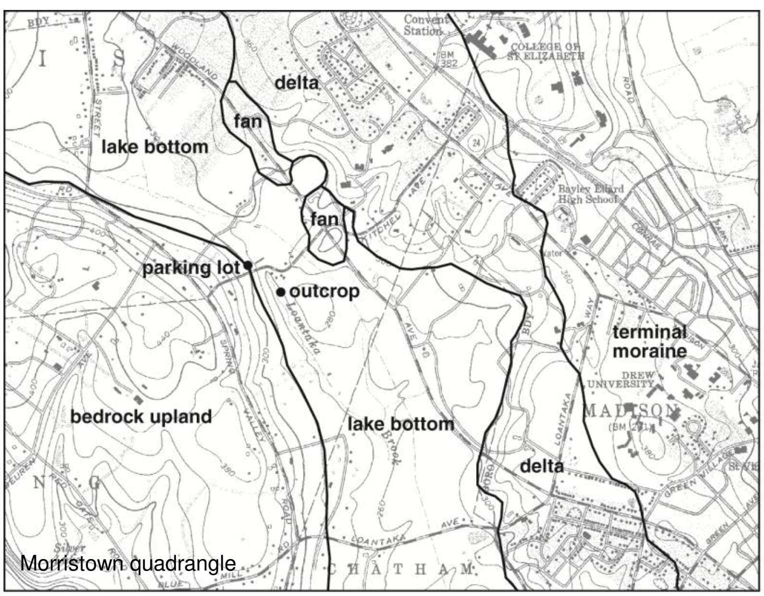


Figure 10. Map of Weasel Brook sluiceway.

Discharge through the main sluice, again calculated using the Manning equation and based on a 10-foot flow depth through the 1000-foot wide channel north of the Garden State Parkway, with a gradient of 0.015 from the release point to the point where the channel branches at US Route 46, and a roughness of 0.045, was 180,000 ft^3/s. At this rate the 1.2 mi^3 release would drain in about 11 days, at a velocity of 18 ft/s. There was little time for a steady outflow following the release in this sluice because retreat of the glacier margin from the release point immediately opened progressively lower drainage routes for the Passaic River through Paterson Gap. The Passaic eventually stabilized in the current route over Great Falls once the gap was fully deglaciated.

Site 7: Lake-bottom deposits (Loantaka Brook Reservation, Harding Township). Silt and clay were laid down on the lake floor by density underflows in quiet water away from the ice margin. These deposits are as much as 250 feet thick and underlie most of the low-lying terrain in the basin. They are rarely exposed because they are commonly covered by postglacial alluvial and wetland sediments. One good outcrop is along Loantaka Brook (40°46.003'; 74°27.247') just east of the paved bike path to the south of Kitchell Road (fig. 9). The outcrop is on the east bank of the brook but is reached from the west side by crossing a wet floodplain and the creek itself. It may inaccessible during high water. The quality of the outcrop will vary depending on how recently the bank has been eroded. This exposure is a natural streambank cut into the apron at the foot of the delta deposited in the Moggy Hollow stage while ice stood at the terminal moraine. The varves were laid down in about 100 feet of water depth. They consist of light-brown clay beds up to 0.5-inch thick alternating with grayer, laminated silt and very fine sand beds up to 4 inches thick. The clay beds were laid down in winter, when the lake was iced over, water was still, and there was little sediment coming into the lake. Under these quiet conditions, clay could settle onto the lake floor. The silt and sand were laid down in summer, when the water was more turbulent, keeping the clay in suspension, and much sediment came into the lake from the melting glacier. Together, each clay-silt couplet, known as a varve, represents a year of deposition. Variations in couplet thickness reflect changes in melt rate and sediment input from year-to-year and can be used to correlate varve records from site to site and lake to lake, potentially providing a detailed timescale for tracking glacial retreat. This has been successfully accomplished for the Connecticut River valley in New England (Ridge, 2008).



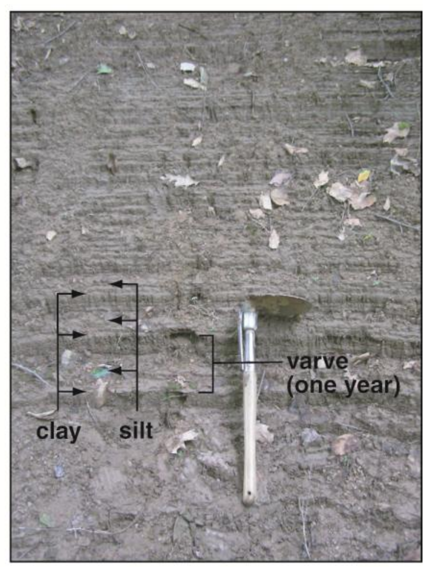


Figure 11. Map and photo of lake-bottom deposits along Loantaka Brook.

Site 8: Lacustrine fan (Old Troy Park, Parsippany). Sand and gravel was deposited in deltas and fans in the lake. Deltas are bodies of sand and gravel built out into the lake from the glacier margin with flat tops that mark the lake level. They include a large, continuous delta along the front of the terminal moraine between Chatham and Morristown, and smaller deltas along the east and west shore of the lake north of the moraine. Fans are knolls and ridges of sand and gravel laid down on the lake floor at the mouths of sediment-laden subglacial tunnel channels discharging into the lake under high water pressure at the front of the glacier. Deltas and fans were extensively mined in the basin for construction material but all the pits are now closed and most have been developed as residential or commercial properties. A lacustrine fan in its natural state is located in Old Troy Park (fig. 10). The yellow trail from the parking lot on Reynolds Avenue leads to the crest of a northwest-southeast trending till ridge at 40°50.371'; 74°24.141', which marks a recessional position of the glacier margin. Just to the east, across West Brook, the yellow trail comes to a gravel pit in the lacustrine fan at 40°50.288'; 74°24.041', which exposes pebble-to-cobble gravel and sand. The yellow trail continues in a loop along the fan, which here is a ridge elongated along the former glacier margin, on the same trend as the till ridge. The southwest-facing slopes of the till ridge and the fan are steeper than the northeast slopes, suggesting the ice overrode and smoothed the northeast slopes after the till and fan were deposited. Another possibility is that these slopes were confined beneath the front edge of the glacier during deposition. The lake was about 110 feet deep above the crest of the fan, which helped buoy the front edge of the ice off the lake floor, opening some space beneath the margin for deposition.

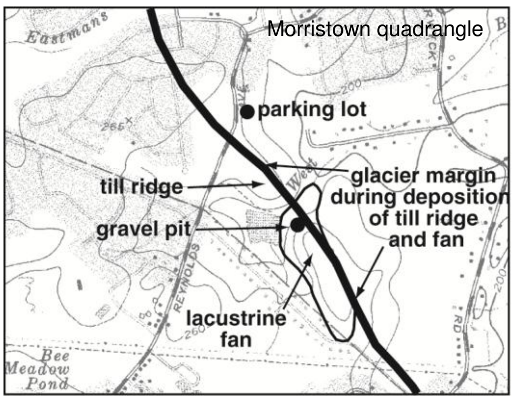


Figure 12. Map of lacustrine fan at Old Troy Park.

Site 9: Delta-fan complex (Foothills Park, Pequannock Township). Foothills Park and the office complex to the east over to NJ Route 23 are within a former gravel pit. This pit was dug into a lacustrine fan deposited in the Moggy Hollow stage of Lake Passaic. The fan formerly rose as much as 50 feet above the present land surface. The excavated part of the fan is the top of a deposit more than 200 feet thick (fig. 11, section AA'), extending down to a thin basal layer of till resting on bedrock. The pit exposures, for example, bare areas and trail beds in the southwest corner of the park lawn at 40°58.756'; 74°18.828', and logs of test and production wells adjacent to and within the former pit, show that the fan is pebble-to-cobble gravel and sand with some interbeds of finer sediment, for its entire thickness. The north wall of the pit is dug into a Great Notch stage delta with a top surface at 320 feet, visible as the treeline on the hill across the park to the north. The pit wall here shows delta foreset sand overlying pebble-to-cobble fan gravel, indicating that the delta prograded over the older fan deposit. Fan deposits laid down in the Moggy Hollow stage also crop out north of the delta, and a small Moggy Hollow delta borders the Great Notch delta to the west along I-287 (fig. 11), indicating that the Great Notch delta was not deposited in contact with the glacier. Rather, as the ice front retreated northward from the fan complex toward the present position of NJ 23, an outlet channel for Lake Butler, a local lake that occupied a north-draining valley just to the west of, and higher than, Lake Passaic, was opened at the 500-foot elevation on the hill to the west of the fan complex (arrowed line on fig. 11). At the same time, Hackensack lobe ice on the east side of the Lake Passaic basin uncovered Great Notch, and Lake Passaic lowered 80 feet from the Moggy Hollow to the Great Notch stage. This allowed the Butler outflow to erode the north end of the Moggy Hollow delta and adjoining fans and redeposit the sediment in the Great Notch stage delta. These relationships fix the geometry of the last Moggy Hollow ice margin (fig. 1D) between Great Notch and the Highlands.

The thick, permeable fan deposits here are a prolific aquifer. Production wells for Pequannock Township (circled dots on fig. 11) are screened between 120 and 200 feet in depth and, when first drilled, yielded between 700 and 800 gpm. The fan deposits here comprise the full thickness of the valley fill and so are unconfined, but fan gravels are more commonly buried, and confined, by lake clays. Such buried, confined fan gravels are the principal valley-fill aquifers in New Jersey.

Of particular note here is the beautiful glacially polished and striated gneiss outcrop at 40°58.810’; 74°18.814’. The rock is interlayered white, buff, and light green quartz-plagioclase gneiss and dark gray amphibolite of Middle Proterozoic age. The gneiss is ductilely deformed because it is very close to the Ramapo Fault, the bounding normal fault on the west edge of the Newark Basin (Volkert, 2010). The striations show ice flow to the S10°W, with less prominent, later striations showing flow to the S0° to S5°E. Ice here was moving out of the Wanaque valley in the Highlands to the north and arrived slightly before the main southwest-flowing Passaic Lobe entered the lowland. This northerly ice flow accounts for the absence of red sandstone and basalt from the fan gravel exposed here, which is about 60% gneiss and 40% Paleozoic sandstone and conglomerate from northerly sources.

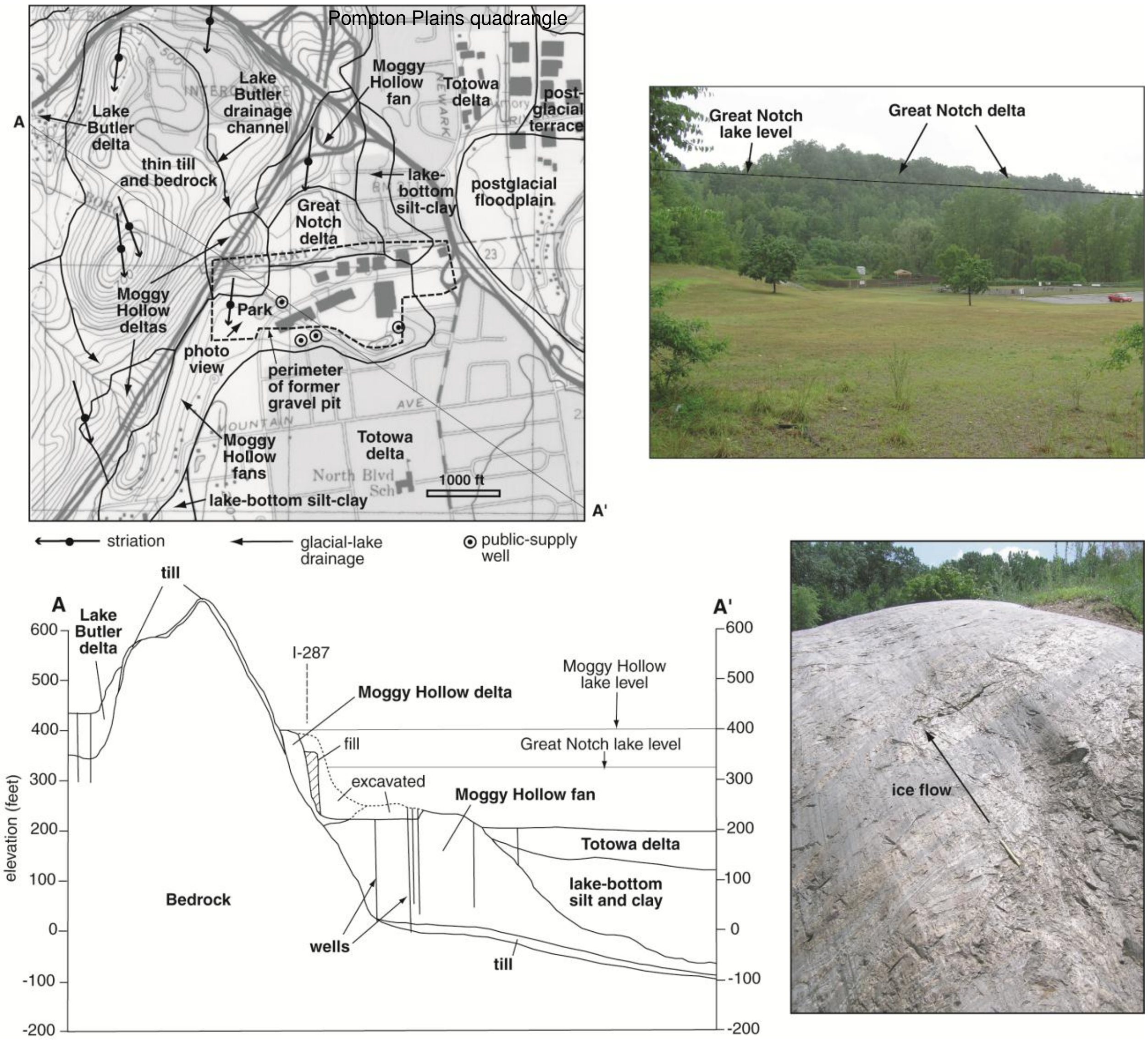


Figure 13. Map and section of fan-delta complex, Foothills Park, Pequannock. Photos show Great Notch delta and striated gneiss outcrop in park.

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GEOLOGY OF ROUND VALLEY RECREATION AREA: THE COMPLEX INTERPLAY OF METAMORPHIC, MAGMATIC AND TECTONIC PROCESSES

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Introduction

Round Valley Recreation Area, jointly managed by the New Jersey Department of Environmental Protection and the New Jersey Water Supply Authority, was officially opened to the public in 1977. Located in north-central New Jersey in Hunterdon County, it occupies 5,300 acres of scenic, low rolling hills and a prominent, horseshoe-shaped ridge known as Cushetunk Mountain (Fig. 1). Round Valley derives its name from the naturally occurring shape of the valley (now beneath the reservoir) that is rimmed on the north, south, and east by Cushetunk Mountain. The name Round Valley has been used for this distinctive geographic feature since before 1836, when Cushetunk Mountain was known either as Round Valley Mountain or Pickel's Mountain.

At the heart of Round Valley Recreation Area, and the reason for its development, is Round Valley Reservoir. Situated within the Raritan River watershed, Round Valley was first proposed as a site for a water storage reservoir in 1945 to help meet the growing need for a stable water supply in northern New Jersey. The project was begun in 1959 and completed in 1965. The South Dam was constructed across Prescott Brook at the southwestern end of Cushetunk Mountain and the North Dam and Dike across South Branch Rockaway Creek at its northwestern end (Fig. 2). The construction was followed by gradual flooding of the interior valley to an elevation of approximately 385 feet above sea level.

Cushetunk Mountain provides a natural barrier for most of Round Valley Reservoir because its ridge top elevations range from 440 to 834 feet above sea level, while elevations on the valley floor average 280 feet. With a width of 1.5 miles and length of 2.6 miles, the reservoir has a surface area of 2,350 acres. It is the second largest freshwater body in New Jersey, surpassed only by Lake Hopatcong and the second deepest, surpassed only by Merrill Creek Reservoir. Round Valley Reservoir is 160 feet at its deepest point, and with a storage capacity of 55 billion gallons (Buss, 1990), it is the largest reservoir in New Jersey. Typically, it provides 70 million gallons of water per day.

Geology of Round Valley Recreation Area

Mostly in the Piedmont province, Round Valley Recreation Area straddles the boundary separating the Highlands and Piedmont physiographic provinces. Only the extreme western end of the reservoir, including the boat launch areas, beach complex, Pine Tree Trail area, and Park office, south to the South Dam, is located in the Highlands. This area is underlain by Proterozoic and Paleozoic rocks that constitute the most complex geology of the area. There, bedrock forms a disrupted mosaic of fault blocks underlain by Proterozoic granite and gneiss that are in some places overlain by, and other places in fault contact with, equally disrupted bodies of Paleozoic rocks (Fig. 3). In general, Proterozoic and Paleozoic rocks at Round Valley Recreation Area occur in north-south or northwest-southeast-trending blocks.

Mesoproterozoic Geology

Proterozoic rocks at Round Valley Recreation Area consist mainly of microperthite alaskite, amphibolite and quartz-plagioclase gneiss. Most of the Proterozoic rocks display a penetrative metamorphic (crystallization) foliation formed during their deformation to granulite-facies conditions between 1045 and 1024 Ma (Volkert et al., 2010) during the Ottawan orogeny, the collision between eastern Laurentia (proto North America) and Amazonia (proto South America) that formed the supercontinent Rodinia. Temperature estimates for this high-grade metamorphism are 769^{o}C based on calcite-graphite geothermometry (Peck et al., 2006) and pressure estimates are 500 to 550 MPa (~5.5 Kb) based on hornblende-plagioclase geobarometry (Volkert, 2004).

Microperthite alaskite has a fairly uniform texture and grain size. It is a medium-grained, massive, pale pinkish-white, moderately foliated granitoid rock composed primarily of microcline microperthite and quartz (Fig. 4A) and, by definition, containing less than 5 percent mafic minerals. Alaskite at Round Valley is part of the Byram Intrusive Suite, which along with the comagmatic Lake Hopatcong Intrusive Suite, forms the Vernon Supersuite that underlies roughly 50 percent of the Highlands. Byram granite dated elsewhere in the Highlands has yielded sensitive high-resolution ion microprobe (SHRIMP) U-Pb zircon ages of 1184 to 1182 Ma (Volkert et al., 2010). Alaskite forms the northern part of the prominent ridge that rises more than 100 feet above the main reservoir between the Dike and South Dam.

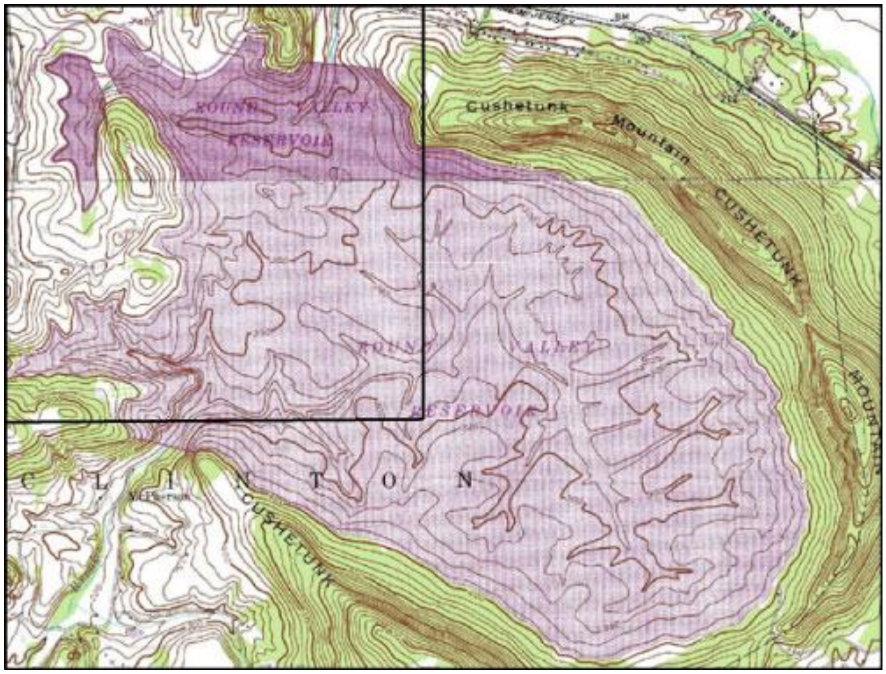

Figure 1. Topography of Round Valley Recreation Area composited from U.S.G.S quadrangle maps of Flemington (bottom) and Califon (top). Cushetunk Mountain forms a horseshoe-shaped ridge that partially encloses Round Valley, now covered by the reservoir. Area of figure 2 is outlined by box in upper left.

Figure 2. Location of the North Dam, Dike and South Dam at Round Valley Reservoir. Figure modified from Google Earth.

Because of the texture and the absence of extensive fracturing of the alaskite, it is fairly resistant to erosion. However, the effects of physical weathering are exceptionally well displayed among these outcrops. As the rock slowly disintegrates in place, it first forms rock fragments and, with continued weathering, gravel and coarse sand. Much of the unconsolidated terrace found along the western edge of the reservoir is made up of such fragments from the underlying alaskite.

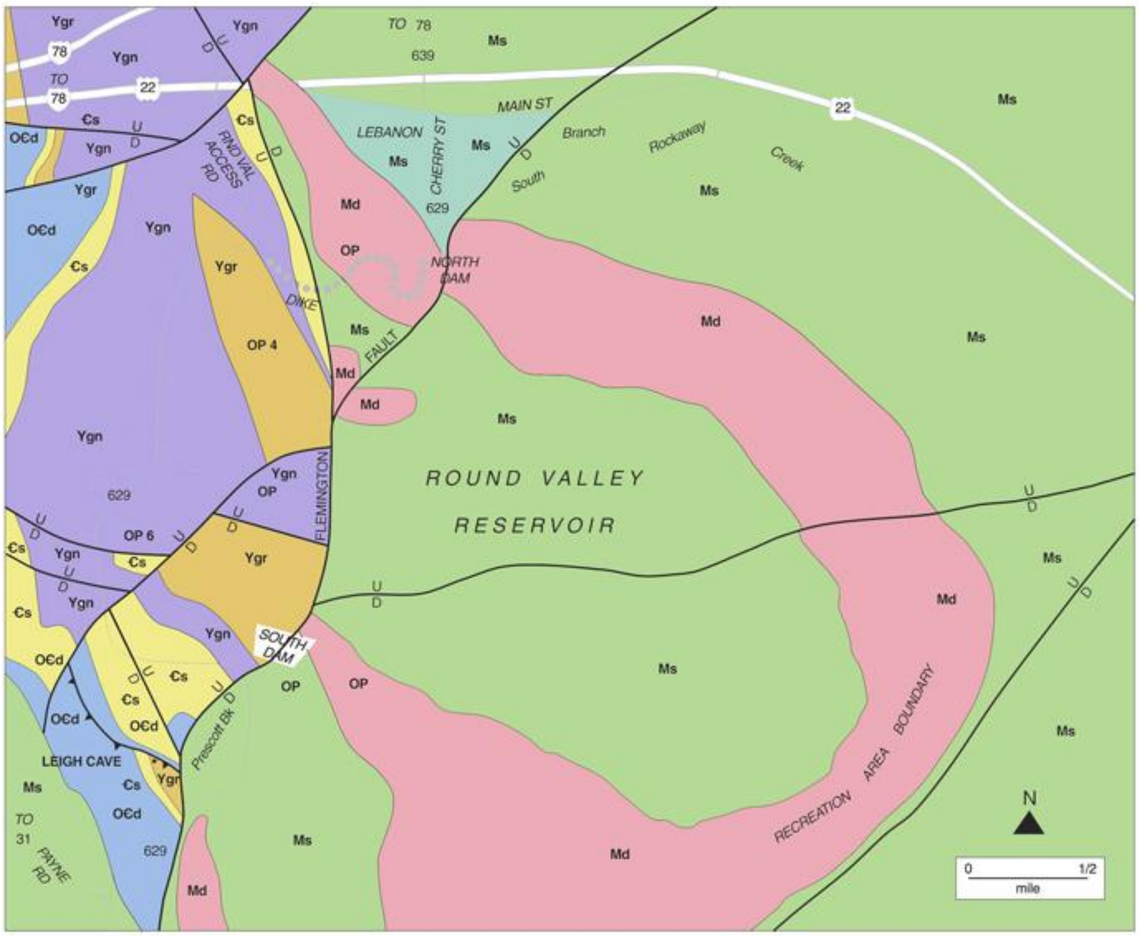

Figure 3. Simplified geologic map of Round Valley Reservoir and the surrounding area. Symbols are: Md, Jurassic diabase; Ms, Jurassic and Triassic sedimentary rocks; O€d, Ordovician and Cambrian dolomite; €s, Cambrian sandstone; Ygr, Proterozoic granite; Ygn, Proterozoic gneiss; U and D, upthrown and downthrown sides, respectively, on faults (bold lines).

Amphibolite (Fig. 4B) crops out at the reservoir's edge east and west of the southern boat launch area and along the lower part of adjacent hillslopes. It is a black or grayish-black, medium-grained, foliated gneiss composed of hornblende and plagioclase ± clinopyroxene and magnetite. The protolith was a fairly mafic rock that likely was basalt. A few outcrops display subhedral phenocrysts of plagioclase as long as 0.5 inch that may be a relict feature inherited from the protolith. Amphibolite here is part of the Losee Suite, an assemblage of calc-alkaline rocks formed in a continental-margin magmatic arc environment along the Laurentian margin (Volkert, 2004). Rocks of the Losee Suite yield SHRIMP U-Pb zircon ages of 1282 to 1248 Ma (Volkert et al., 2010). Amphibolite is the principal host rock for magnetite deposits at Round Valley Reservoir (Large mine) and elsewhere in the area.

Minor amounts of pale pink to white-weathering, medium-grained, foliated gneiss composed of plagioclase and quartz (Fig. 4C), also part of the Losee Suite, is conformably interlayered with amphibolite. Quartz-plagioclase gneiss at Round Valley was formed from either a volcanic protolith (dacite) or a plutonic one (tonalite).

Very locally, amphibolite outcrops are intruded discordantly by white, coarse-grained, unfoliated pegmatite (Fig. 4D) composed of plagioclase, amphibole, clinopyroxene, magnetite, and trace amounts of biotite and chlorite. Textural evidence indicates they were generated through partial melting of amphibolite. Widespread pegmatites elsewhere in the Highlands yield U-Pb zircon ages of 1004 to 986 Ma (Volkert et al., 2005).

Paleozoic Geology

Paleozoic rocks at Round Valley Recreation Area are mainly Cambrian in age and they include Hardyston Quartzite (Fig. 5A) and Leithsville Formation (Fig. 5B). These sedimentary formations were deposited along a passive margin during breakup of the supercontinent Rodinia. The Early Cambrian Hardyston documents initial fluvial sedimentation across the eroded Proterozoic surface and the subsequent drowning of the continental margin during a marine transgression. The overlying dolomite of the Leithsville Formation marks the stabilization of a carbonate platform. West of Round Valley, near Clinton, Ordovician dolomite crops out and farther west, near Jutland, a marine sequence of Ordovician shale interbedded with minor sandstone and limestone is exposed. This ocean basin was destroyed during continental collisions involving eastern North America about 450 million years ago (Taconic orogeny) and 290 million years ago (Alleghanian orogeny). In New Jersey these events renewed uplift and mountain formation in the Highlands.

Mesozoic Geology

Mesozoic rocks at Round Valley are Triassic and Jurassic in age. They underlie the Piedmont province, which in the area of Round Valley is sometimes called the Hunterdon Plateau. The Piedmont is located south and east of the Highlands and is an area of gently rolling hills and lowlands with minor ridges and uplands. At the start of the Mesozoic, the supercontinent of Pangea separated and the Earth's crust in what was to become New Jersey was pulled apart along the Ramapo fault and other border fault segments, creating a down-dropped block east of the Highlands forming the Newark basin. The uplifted Highlands began to erode, and provided sediment from Proterozoic and Paleozoic rocks that filled the Newark basin. Mafic magma rose through rift-related extensional fractures in the crust, extruding onto the surface as basalt flows, or emplaced into sediments as diabase dikes, sills and other intrusive bodies.

Most of Round Valley Recreation Area is underlain by Mesozoic sedimentary and igneous rocks. Sedimentary rocks are mainly reddish-brown, medium- to thin-bedded siltstone and mudstone of the Passaic Formation (Fig. 6A) that crop out in the valley beneath the reservoir, as well as the lowland areas north, east, and south of Cushetunk Mountain. Conglomeratic facies of the Passaic Formation (Fig. 6B) is preserved along the Flemington Fault north of the reservoir. These rocks are interbedded reddish-brown cobble to pebble conglomerate, pebbly sandstone and sandstone. Clasts are commonly matrix-supported, subangular to subrounded, locally imbricated, hematite-stained quartz and quartzite in a matrix of coarse- to medium-grained sandstone.

Igneous rocks at Round Valley are mainly diabase (Fig. 6C) of Jurassic age that underlies all of Cushetunk Mountain, the hill west of the reservoir's North Dam, and a low hill to the south which is currently under water. Diabase is medium grained and greenish gray where fresh, but weathers light tannish-gray and locally is yellowish-brown stained from the oxidation of iron in the rock. Diabase is composed mainly of clinopyroxene, plagioclase and opaque iron oxides. The tight interlocking texture of the mineral grains (Fig. 6D) and the relatively fine grain size of the rock combine to give diabase a strength and durability that resists erosion, enabling it to form uplands and steep ridges. Because of these qualities, it was used extensively for riprap during construction of the reservoir and Round Valley Recreation Area.

Geochemical analyses of Cushetunk Mountain diabase indicate that it is chemically similar to other diabase from the west-central Newark basin, as well as to the Palisades sill (Puffer and Lechler, 1979; Houghton et al., 1992).

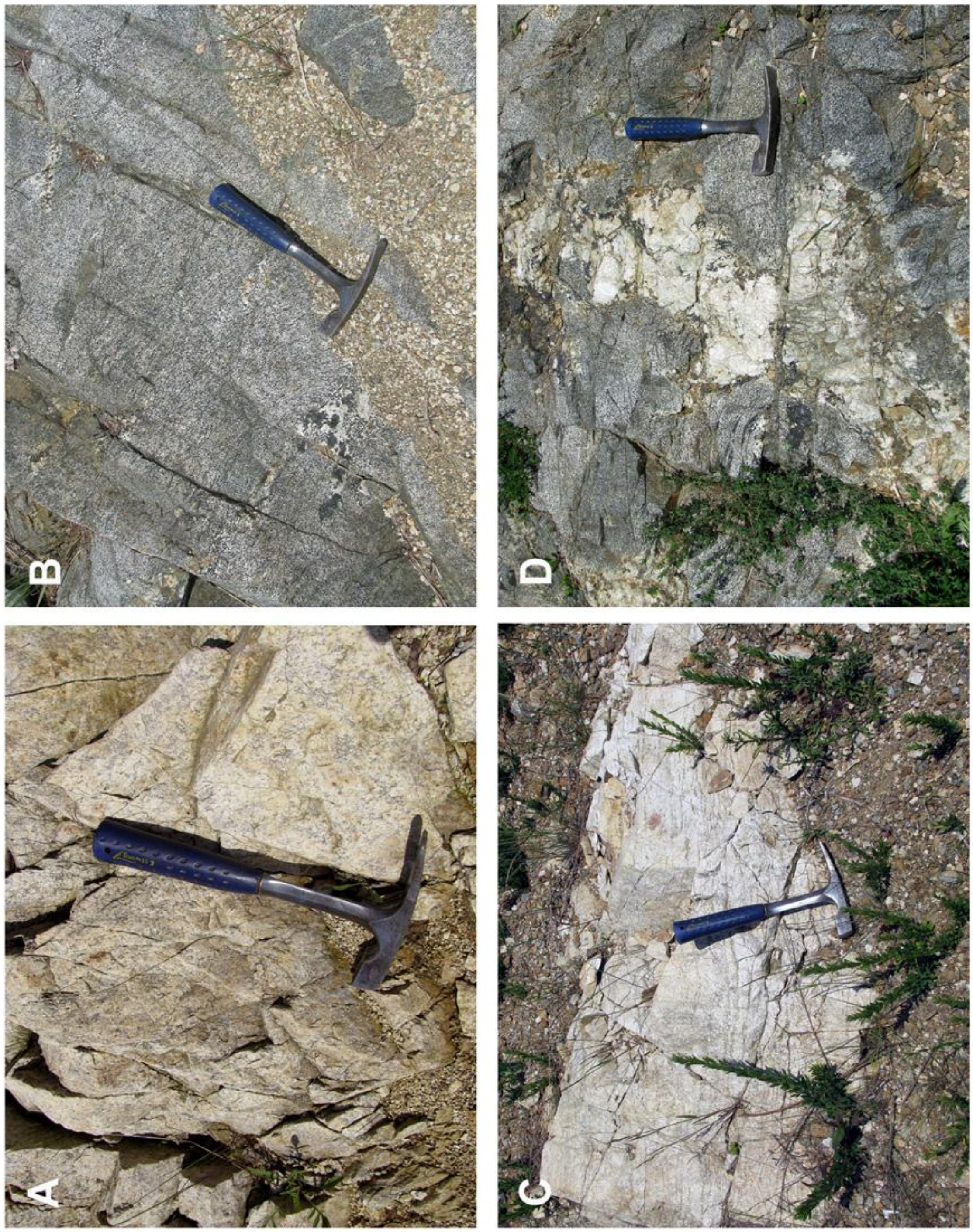

Figure 4. Representative Proterozoic rocks at Round Valley Recreation Area. (A) Medium-grained, massive microperthite alaskite of the Byram Intrusive Suite. (B) Foliated hornblende-plagioclase amphibolite. (C) Medium-grained, foliated quartz-plagioclase gneiss of the Losee Suite. (D) Unfoliated pegmatite composed of plagioclase, hornblende and magnetite intruding amphibolite.

Cenozoic and Recent Geology

From the end of the Mesozoic through the Cenozoic Era, the predominant geologic process occurring was erosion. Most areas of the region were again reduced to a nearly flat surface. The present landscape has since evolved slowly, largely through stream erosion. Streams have cut deeply into the Paleozoic and Mesozoic sedimentary rocks leaving the more resistant Proterozoic granites and gneisses and Jurassic diabase rising above them as ridges. The landscape continues to undergo gradual change as sediments eroded from rocks of the region are transported by local tributaries to the Raritan and Delaware Rivers.

At several places along the edge of the reservoir, the erosional development of low shoreline bluffs with steep faces and talus at the slope base is visible. Physical weakening of the bedrock from faulting, and repeated seasonal freezing and thawing cycles along fractures, causes the rock to weather and break apart easily. Water and wave action continue to erode these low bluffs and claim still more of the shoreline.

Bedrock Structure

Proterozoic Foliation

Crystallization foliation is the parallel alignment of mineral grains (Figs. 4B, 4C) resulting from compressional stresses during the Ottawan orogeny that deformed the rocks while they underwent high-grade metamorphism at 1045 Ma. It defines the strike of the Proterozoic bedrock at Round Valley Recreation Area and elsewhere in the Highlands. Northwest of the reservoir, foliation is relatively uniform and displays an average strike of N20^{o}W. The dip ranges from 31^{o} to 70^{o} northeast. West and southwest of the reservoir, foliation is more varied due to folding of the rocks during the Proterozoic, as well as by the effects of faulting. Most foliations strike northeast or northwest and dip moderately to steeply east at 36^{o} to 81^{o}, except along the southern shoreline, in the hinge area of folds, where foliations strike N.72^{o}E. to N87^{o}W. and dip north at 33^{o} to 77^{o}. The plunge of all mineral lineations on foliation surfaces averages 31^{o} N34^{o}E., parallel to the axes of minor folds in outcrop.

Paleozoic Bedding

Despite dissection of the Paleozoic rocks into numerous structural blocks, the strike and dip of bedding is fairly uniform. West of Round Valley Reservoir, most beds in the Hardyston and Leithsville strike between N17^{o}E. and N16^{o}W. and dip west at 20^{o} to 72^{o} (Volkert, 1989; Herman et al., 1992).

Mesozoic Bedding

The strike of bedding in the Passaic Formation is fairly uniform. In general, beds within Round Valley have an average strike of N70^{o}E. and dip north at 21^{o} to 54^{o} (Kümmel, 1900). Beds

Figure 5. Paleozoic rocks at Round Valley Recreation Area. (A) Dip slope of thin-bedded, feldspathic Hardyston Quartzite. (B) Medium-bedded dolomite overlain by shaly dolomite of the Leithsville Formation.

north and south of Cushetunk Mountain have a strike and dip similar to beds in Round Valley and, therefore, are discordant to the diabase intrusion. This effectively rules out any possibility that the geometry of Cushetunk Mountain is related to folding.

Cushetunk Diabase

The diabase intrusion that underlies Cushetunk Mountain forms a horseshoe-shaped ridge, the axis of which strikes about N60^{o}W and forms a low to moderate angle with bedding. The discordant relationship between the diabase and Passaic Formation at Round Valley has long been known (e.g., Bayley et al., 1914). Based on structure contouring of the stratigraphy, Houghton et al. (1992) calculated that the southern part of the Cushetunk Mountain intrusion is nearly concordant with the Passaic Formation and was emplaced into the sediments about 9,000 feet beneath their contact with the Orange (A) Dip slope of thin-bedded, fissile siltstone and mudstone of the Passaic Formation. (B) Quartzite clast conglomerate of the Passaic Formation. (C) Medium-grained, massive Cushetunk diabase. (D) Close-up of diabase showing dense, massive texture.

Mountain Basalt, while the northern part of the intrusion is markedly discordant. As indicated above, the shape of the intrusion is not related to folding because bedding does not display structural concordance to diabase on the northern limb, or in what would be the hinge of the fold to the southeast. Nor was the diabase intruded as a sill because of its highly discordant northern contact against the Passaic Formation.

Figure 6. Representative Mesozoic rocks that crop out north and east of Round Valley Recreation Area.

Faults and Fractures

The various faults that deform the rocks at Round Valley Recreation Area formed in response to multiple tectonic events that range from the Proterozoic through the Mesozoic. The Flemington Fault (Fig. 3) is a major structural feature that separates Proterozoic and Paleozoic rocks on the footwall from Mesozoic rocks on the hanging wall. At Round Valley and to the south, the fault strikes about N.10°E. and dips 50° southeast (Herman et al., 1992). Dominant movement was oblique normal with a right-lateral strike-slip component (Houghton et al., 1990). To the north, now beneath the reservoir, the fault bifurcates (Volkert, 1989; Drake et al., 1996). A north-striking western segment of the fault extends beneath the Dike, and an eastern segment that strikes about N.40°E. extends beneath the North Dam (Fig. 3). Other faults cutting diabase at the North Dam are clearly Mesozoic. They have an average strike and movement sense of N.55°W. (dip-slip normal), N.55°E. (left-lateral strike-slip), and N.05°E. (right-lateral strike-slip), and steep to vertical dips (Fig. 7), orientations that similar to some of the fracture groups measured in the Proterozoic rocks.

Proterozoic rocks at the southwest part of Round Valley Recreation Area locally display ductile deformation zones that range in width from 6 to 10 feet and that strike N.60°E. to N.60°W. and dip north, subparallel to crystallization foliation. High strain in these zones was accommodated by less competent layers in the amphibolite that contain biotite. Ductile features include shear band foliation, pinch-and-swell structures, boudinage, and shear folds (Fig. 8A). These structures were formed under fairly high metamorphic grade and, therefore, likely originated during the Proterozoic. Ductile features display what appears to be a dominantly right-lateral shear sense.

Proterozoic rocks at Round Valley display a pervasive brittle deformation fabric that overprints the ductile fabric. Some faults preserve evidence for cataclastic flow of fine-grained fault material and small breccia fragments (Fig. 8B). Other brittle deformation features include breccia, fracture cleavage, epidote-coated slickensides, and fault gouge. Proterozoic rocks along the reservoir are cut by a series of parallel brittle faults that strike N.55°E. to N.65°E. and dip subvertically. They are characterized by fracture cleavage spaced an inch or less apart (Fig. 8C), cataclasis and zones of anastomosing splays as much as two feet wide.

Breccia zones strike N.80°E. to N.50°E., subparallel to the strike of crystallization foliation. They are as much as four feet wide and contain subangular to subrounded clasts of amphibolite in a matrix of fine-grained cataclasite composed mainly of retrogressed amphibolite (Fig. 8D). Breccia clasts are well foliated, and some preserve isoclinal folds, indicating that brecciation took place in the upper crust when the rocks were cold, subsequent to Proterozoic uplift and cooling. Bedrock was broken up in situ, likely through hydraulic fracturing, and the clasts were then transported laterally. In some instances they can be reconnected to less deformed amphibolite along adjacent boundaries. Overall, clasts show little evidence of rotation and maintain conformity with foliation in the less deformed host. Breccia zones are cut by high-angle fractures that strike about N.40°W., N.25°W. and N.80°W.

Similar breccia zones developed in Proterozoic rocks elsewhere in the Highlands are associated with major structural features such as the Reservoir, East and Tranquility faults, all of which have undergone reactivation as compressional structures during Paleozoic orogenesis, which may have been when the breccia zones were formed. The breccia zone developed along the East Fault in Sussex County is interpreted as having formed during the Taconic orogeny (Gillespie, 1993).

Dominant fractures measured in Proterozoic outcrops define groups that have an average orientation and movement sense of N.55^{o}E. 74^{o} SE to 71^{o} NW (high-angle oblique normal, or high-angle oblique reverse), N.05^{o}E. 73^{o} NW (dip slip normal or reverse), N.45^{o}W. 74^{o} SW (high-angle oblique normal or reverse), N.40^{o}W. 70^{o} NE (left-lateral strike-slip),

Figure 7. High-angle faults striking N.05^{o}E. and N.55^{o}W. that deform Mesozoic diabase at the North Dam.

and N.80^{o}W. 59^{o} NE (left-lateral strike-slip). Cross-cutting relationships suggest the following sequence of formation for the groups: N.55^{o}E., N.05^{o}E., N.45^{o}W., and N.80^{o}W. While this relationship holds generally, there are some exceptions. The N.55^{o}E. group may have formed during the Proterozoic Ottawan orogeny and then undergone reactivation during subsequent tectonic events, each of which imparted a different movement sense. Fractures of this group display strong evidence of having formed earlier than other brittle deformation features, and they are consistently cut by fractures that strike N.35°W. to N.55°W. (Fig. 9). The N.35°W. to N.55°W. group is very similar to the strike of the thrust fault at Leigh cave that places Proterozoic rocks over Cambrian Leithsville Dolomite, and so this group may have formed, or at least been reactivated, during the Taconic orogeny.

Most of the faults at Round Valley Recreation Area are steeply-dipping to vertical but an exception is the thrust fault at Leigh cave (see description below). The presence of this fault has been known for some time and is mentioned briefly in Bayley et al. (1914). Proterozoic rocks on the west side of the fault were transported northeastward over the Paleozoic rocks. The thrust fault strikes about N.40°W. and dips southwest at 30° to 40° (Fig. 10).

Leigh Cave

Leigh cave, approximately 0.5 miles southwest of the reservoir's South Dam, helps to emphasize the geologic complexity of the area. The cave is probably the largest in New Jersey in total underground volume, and second or third largest in total length, containing more than 800 feet of passages (Dalton, 1976). Leigh cave is unique because the floor and walls are dolomite of the Leithsville Formation, whereas the ceiling is Proterozoic granite that was thrust over the dolomite during the Paleozoic. Because water dripping through fractures in the roof of the cave flows through insoluble Proterozoic rock instead of soluble limestone or dolomite, characteristic forms such as stalagmites, stalactites, and flowstone are lacking. The orientation of the cave passages is structurally controlled and they follow the strike of bedding in the dolomite, the thrust fault, and dominant joints and fractures in the bedrock (Dalton, 1976).

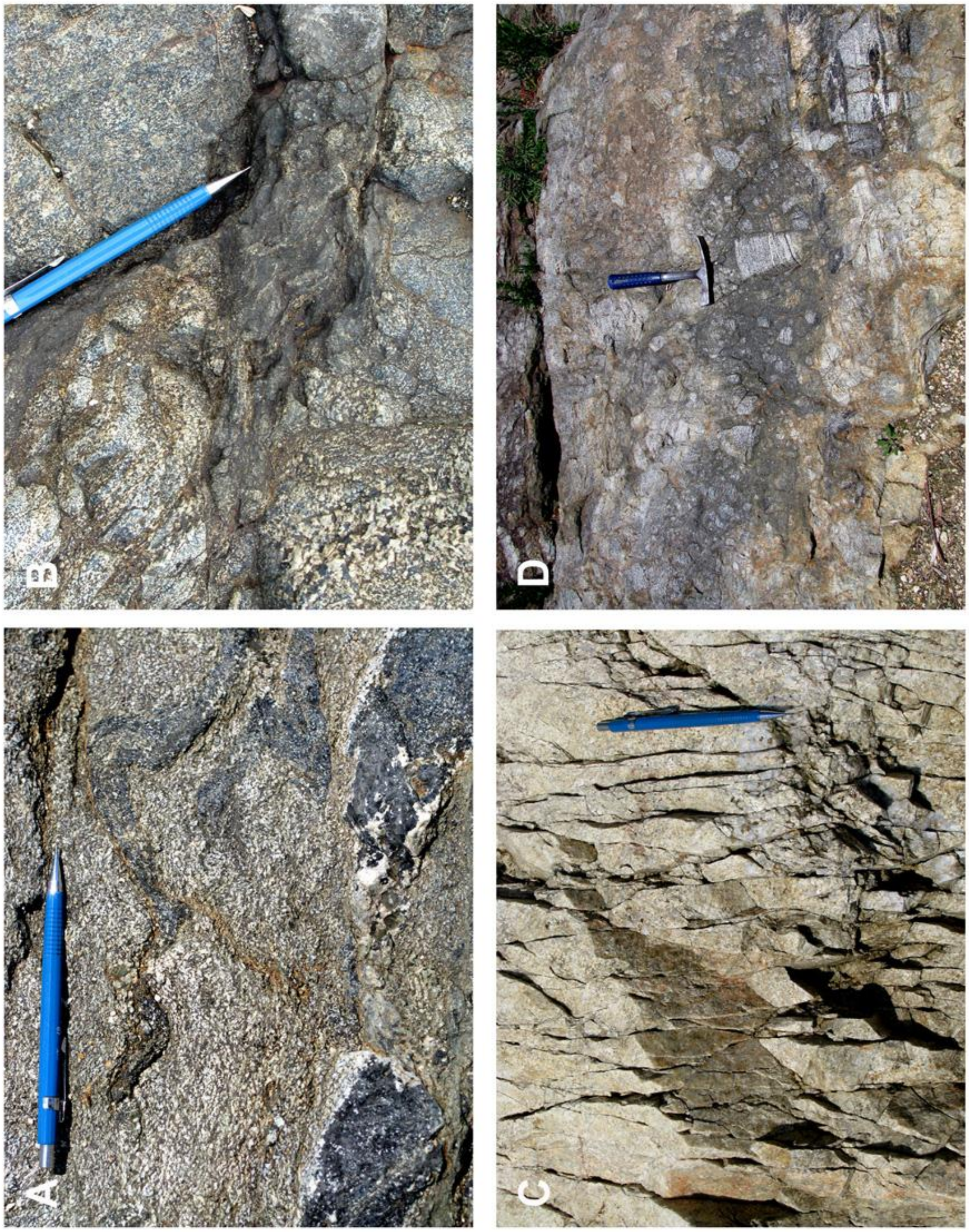

Figure 8. Examples of deformation features in Proterozoic rocks at Round Valley Recreation Area. (A) Ductile fabric consisting of shear folds, boudinage and shear band foliation in amphibolite. (B) Cataclastic flow of comminuted material and breccia fragments in brittle fault striking about N.55°E. (C) Fracture cleavage striking N.55°E. and spaced an inch or less apart in quartz-plagioclase gneiss. (D) Breccia zone about four feet wide composed of subangular to subrounded clasts of amphibolite in a matrix of cataclastic amphibolite. Breccia zone strikes N.50°E. to N.80°E., subparallel to crystallization foliation in the host.

Figure 9. Early brittle fault that strikes N.55^{o}E. and dips steeply SE to vertical cut by fault striking about N.60^{o}W. Outcrop of amphibolite along south shoreline of reservoir. (see text for discussion).

Figure 10. Exposure of the thrust fault at Leigh cave viewed above ground and looking north. Proterozoic granite (top of photograph) was transported northeast onto dolomite of the Leithsville Formation. Direction of thrusting is toward the upper right of photograph. Clipboard is on the fault surface.

ECONOMIC GEOLOGY AND RESOURCES

Hunterdon County has a rich and varied history, but people are often unaware of the role geology has played in the development of the Round Valley area. Indigenous peoples no doubt settled in Round Valley to take advantage of the sheltered protection offered by Cushetunk Mountain. Later, between 1710 and 1966, a thriving iron ore (mostly magnetite) industry existed throughout northern New Jersey which, during the early 19th century, led the nation in iron ore production. During the Revolutionary War, some of this ore was forged into weaponry for the Continental Army. Over 400 mines and prospects throughout northern New Jersey were worked, producing an estimated total yield of 50 million tons of iron ore. One of these, the Large mine (known also as the Lebanon mine), is located along the southwestern edge of Round Valley Reservoir. The mine intermittently produced iron ore from magnetite from about 1872 to 1875, and again from 1879 to 1880, and possibly until 1899. The mine was named for John K. Large of Whitehouse who opened the mine in 1872 on the property of D. K. Hoffman. The mine had a single shaft 95 feet deep and of uncertain length that worked a vein of ore 6 to 8 feet wide. Although the ore was rich, containing between 41 and 45 percent iron, its high sulfur content of 2.7 to 3.6 percent (Bayley, 1910) made it less suitable, and therefore less valuable, for use in the steel industry. The exact quantity of ore removed from the mine is unknown, although historical records show that in 1899 as much as 1,200 tons of iron ore was present on the dumps.

In addition to the Large mine, iron ore and minor amounts of graphite hosted by Proterozoic gneiss were mined a few miles to the north at Annandale, High Bridge and Cokesbury. Manganese ore hosted by Paleozoic rocks was mined a few miles to the west near Clinton, and copper ore hosted by Mesozoic sandstone was mined farther south near Flemington. Some of the local bedrock was also quarried for crushed stone. Quarries were developed in Proterozoic rocks at Glen Gardner and between High Bridge and Califon, in Paleozoic rocks west of Round Valley and near Clinton, in Mesozoic sedimentary rocks northeast of Lebanon, in Mesozoic diabase south of Round Valley Reservoir, and in Mesozoic basalt near Oldwick.

Agriculture is an important industry in the region, a direct result of geologic processes that created the rich, fertile soil covering much of the area around the reservoir. Additionally, farmers used limestone from local quarries to lime their fields. Some limestone was also mixed with iron ore for use as a flux during the roasting of ore in regional furnaces and forges.

Geology continues to play a role in the area's development by providing the unique conditions that make Round Valley Reservoir possible. In addition, ground water in fractured bedrock provides an important source of water for domestic, industrial and public supply wells.

ENGINEERING GEOLOGY

Of particular interest to any discussion of Round Valley Recreation Area is the geology of the reservoir's North Dam, Dike and South Dam. In the 1950's, extensive rock-core drilling was undertaken to locate faults and determine engineering suitability of the soil and bedrock for construction of the reservoir. Erosion of fractured bedrock adjacent to the faults resulted in the formation of gaps and low spots on the ground surface where Prescott Brook, South Branch Rockaway Creek, and the drainage north of the Dike flow through the prominent gaps. In order to impound water in the reservoir, dams and a dike were required to close the gaps and low spots. Bedrock fractures and openings were sealed with grout to strengthen the rock and prevent leakage before construction of the dams or the dike.

The North Dam is an earthfill structure 1,500 feet long and 135 feet high (Buss, 1990). Its construction is essentially the same as that of both the Dike and the South Dam. In all, approximately 370,000 cubic yards of impermeable clay was used for the cores of these structures. Approximately 3,700,000 cubic yards of soil and rock cover was also required for construction of the two dams and the dike (State of New Jersey, 1958). All of these geologic materials were obtained on site. This included clay, soil and rock fill, and crushed rock used as riprap. Drilling across the North Dam revealed this area to be entirely underlain by Jurassic diabase (State of New Jersey, 1958). Near the center of the dam drilling encountered an east-dipping segment of the Flemington Fault (Fig. 3), which follows the same northerly trend here as South Branch Rockaway Creek.

The Dike is a curved, earthfill structure 2,350 feet long and 80 feet high (Buss, 1990) designed to impound water across a natural low spot at the north end of the reservoir. It is constructed of an impermeable core of clay 30 feet wide, and is covered with fill of Mesozoic soil and sedimentary rock excavated from the floor of Round Valley. The outermost layer is an apron, or wall, of riprap that consists of boulder-sized pieces of diabase. Drilling across the area presently underlain by the Dike reveals even more complex geology than that at either dam. The Dike is underlain (from east to west) by Jurassic diabase and highly weathered sedimentary rock, Cambrian Hardyston Quartzite, and Proterozoic gneiss (State of New Jersey, 1958). Drilling at the west-central part of the Dike encountered a north-trending, east-dipping segment of the Flemington fault that separates the gneiss and Hardyston on the west from Mesozoic rocks on the east (Fig. 3).

The South Dam is an earthfill structure 1,400 feet long and 185 feet high (Buss, 1990). Drilling across the South Dam showed that the eastern half of this area is underlain by Jurassic diabase, while the western half is Proterozoic granite (Fig. 3). They are separated by a segment of the Flemington Fault, which was encountered during drilling near where the center of the South Dam now is. At the South Dam the fault roughly follows the same trend as Prescott Brook.

To insure that the reservoir remains full, water must sometimes be pumped 3.3 miles uphill from South Branch Raritan River through a 108-inch diameter pipeline because no streams flow into the reservoir. At other times water is released from Round Valley Reservoir to recharge Prescott Brook, South Branch Rockaway Creek, or South Branch Raritan River.

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SOIL SURVEY OF CROOKE'S POINT, GATEWAY NATIONAL RECREATION AREA AS A GUIDE FOR VEGETATION MANAGEMENT

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Introduction

Crookes Point is a 28-acre peninsula located within Gateway National Recreation Area in Staten Island, NY (Figure 1). Presently managed by the National Park Service, Crookes Point is dominated by highly dense invasive and non-native plant species. To accomplish the restoration of ecological functions in a sustainable manner, the development of a site-specific vegetation community management plan was required. The goal of this plan and the intent of the plantings would be to make Crookes Point as sustainable as possible by reestablishing a Mid-Atlantic Coastal, Back Dune, or Shrub-Scrub ecosystem through the removal of existing invasive vegetation and replanting of native species. To learn more about the current environmental conditions of the site, natural resource managers expressed the need for a detailed soil survey to gain further insight regarding nutrient availability, topography, and other factors which affect plant growth. With more information known, the optimal native plant species for the site could be selected and utilized at the site to ensure target future ecological conditions are met. The selections based upon this information would also ensure that any plantings that are to occur at the site will have a greater chance of success. Recently, removal of invasive species by physical and chemical methods has occurred, with native plantings set to begin in the Fall of 2013.

Figure 1: Map of Gateway National Recreation Area including Great Kills Park in Staten Island, NY

The purpose of this project was to establish the biological, chemical, and physical parameters of Crookes Point which are essential in the development and establishment of a successful vegetation management plan. The main outcome of this investigation is a soil characterization identifying morphological features relative to climate moisture retention and structure in addition to a basic chemical characterization (pH, nitrogen, phosphorus, potassium, calcium, magnesium, sulfate, manganese, iron, chloride, aluminum) of the soil relative to fertility. The natural dynamics and physical features existent in barrier island-type ecosystems were also studied. Coastal upland plant communities at Fire Island National Seashore in New York served as a representative of typical ecological communities in the mid-Atlantic region. Observations in conjunction with literature research were used to determine which vegetation species are optimal in these areas and which vegetation community structure (forest, shrub-scrub, etc.) would be best suited for the study area. Vegetation community structure dynamics were taken into consideration to determine which species should be planted to create the target-ecosystem type over time. Based on these parameters, a list of appropriate plant species for Crookes Point was generated and emphasis given to plant species which may attract wildlife necessary for sustaining a healthy ecosystem (i.e. pollinators) or for increased recreational opportunities (i.e. bird watching). Results were presented to the Natural Resource Management Division of Gateway National Recreation Area to be used for the native planting at Crookes Point.

Study Design

Crookes Point has been chosen as the next site for planting through the *MillionTreesNYC* initiative. The project is set to occur over the next several years, with the initial stages of the project beginning in Spring 2011, planting by Fall 2013, and monitoring and maintenance of the site until at least December 2015. The invasive species removal and tree planting at Crookes Point will occur in phases. Eight sections were initially proposed for restoration by New York City Department of Parks and Recreation (Figure 2). In Early January 2012, Area 1 was selected to begin treatment as part of a pilot project. Continuation of restoration efforts of other areas in Crookes Point will be contingent upon the success of this initiative.

Figure 2: Proposed multi-year upland restoration project area at Crookes Point as defined by New York City Department of Parks and Recreation Natural Resources Group (2010).

All field work for this proposed study was conducted in agreement with National Park Service research and collection regulations. Four project areas within the eight initial sections identified by *MillionTreesNYC* project coordinators were selected for soil sampling. The areas were selected based on restoration priority and differences in vegetation present. The exact location of the pit within the chosen "project area" was selected randomly (Figure 3). If necessary, dense invasive vegetation was cleared to create an access point to each area.

One test pit approximately 5ft × 5ft × 5ft was dug in each of the 4 project area locations in Crookes Point. Full pedon descriptions were completed in accordance to NRCS standards. This included notation of landscape features including elevation, aspect, slope, drainage, flooding, ponding, land cover, erosion, and runoff. Soil descriptions included notation of matrix color (using *Munsell* color chart), texture, rock fragments, structure, consistence, mottles/redoximorphic features, surface features, and roots. Soil horizons were identified and measured in the field and samples from each horizon were collected for further analysis.

Figure 3: Locations of the four sampling locations (CP1, CP2, CP3, CP4) within the Crookes Point project area. Sampling occurred between October 2010 and November 2011.

Summary of Soil Properties, Nutrient Content and Observed Vegetation

Soil Properties

Soil pedon description CP-1 (Figure 4)

Location Description: Forested area approximately 150’ from “open” herbaceous-dominated area and about 75’ from Orange Trail, in the southeast quadrant of Proposed Reforestation Area #3

Latitude: 40°31’55.96 N
Longitude: 74°08’12.40 W

Figure 4: Photograph of soil pit CP-1

Altitude: 10 feet

Slope Characteristics:
Slope: 0-3 percent
Complexity: Non-complex

Physiography:
Local: Fill
Major: Human-made land
Landuse: Parkland/Wildlife
Moisture Regime: Udic moisture regime
Precipitation: 40 to 50 inches per year

Natural Drainage Class: Well drained
Permeability: Rapid
Flooding: None
Ponding: None
Parent Material: Sandy dredge materials

Soil pedon description CP-2 (Figure 5)

Location Description: Vineland located approximately 50’ from trail in the southwest quadrant of Proposed Reforestation Area #1

Latitude: 40°31’59.91 N
Longitude: 74°08’07.30 W

Figure 5: Photograph of soil pit CP-2

Altitude: 9 feet
Slope Characteristics:
Slope: 0-3 percent
Complexity: Non-complex

Physiography:
Local: Fill
Major: Human-made land
Landuse: Parkland/Wildlife
Moisture Regime: Udic moisture regime
Precipitation: 40 to 50 inches per year

Natural Drainage Class: Well drained
Permeability: Rapid
Flooding: None
Ponding: None
Parent Material: Sandy dredge materials

Soil pedon description CP-3 (Figure 6)

Location Description: Woodland located in the northeast quadrant of Proposed Reforestation Area #4

Latitude: 40°32’01.61 N
Longitude: 74°08’19.37 W

Altitude: 10 feet
Slope Characteristics:
Slope: 0-3 percent
Complexity: Simple

Physiography:
Local: Fill
Major: Human-made land
Microfeature: Small swales
Landuse: Parkland/Wildlife
Moisture Regime: Udic moisture regime
Precipitation: 40 to 50 inches per year

Natural Drainage Class: Well drained
Permeability: Rapid
Flooding: None
Ponding: None
Parent Material: Sandy dredge materials

Figure 6: Photograph of soil pit CP-3

Soil pedon description CP-4 (Figure 7)

Location Description: Northeastern quadrant of Proposed Reforestation Area #6 in a stand of *Phragmites* adjacent to the gravel parking lot near Great Kills Harbor

Latitude: 40°32’06.00 N
Longitude: 74°08’03.90 W

Figure 7: Photograph of soil pit CP-4

Altitude: 3 feet
Slope Characteristics:
Slope: 0-3 percent
Complexity: Non-complex

Physiography:
Local: Fill
Major: Human-made land
Landuse: Parkland/Wildlife
Moisture Regime: Udic moisture regime
Precipitation: 40 to 50 inches per year

Natural Drainage Class: Well drained
Permeability: Rapid
Flooding: None
Ponding: None
Parent Material: Sandy dredge materials

Nutrient Content

The major essential nutrient elements supplied through the soil are nitrogen, phosphorus, potassium, calcium, magnesium, and sulfur (Tucker, 2009). Minerals are released from the decomposition of parent material, decomposition of organic matter, and decomposition of the soil from flood waters supplying nutrients for plant uptake.

The average pH of the soil in all four test locations was 6.3 and is within the range where plants grow well (between 5.0 and 8.5). Soils at Crookes Point could be generally characterized as “Slightly Acidic”. On average, had the most acidic soils and CP2 soils appeared to be most alkaline, however the most alkaline soil collected from Crookes Point was located in CP4-C3 with a pH of 7.9.

The ammonia and nitrite nitrogen concentrations at Crookes Point appear to be very low, with most of the nitrogen in the form of nitrate nitrogen. This is most likely due to the characteristics of the soil at Crookes Point: sandy and well-drained. The highest concentration of nitrogen was found in the A horizon of CP1 (75 ppm) in the form of nitrate nitrogen, suggesting that nitrification is occurring rapidly here. It is possible that the relatively non-diverse plant species which are present at Crookes Point do not promote the diverse existence of microorganisms and thus a high level of ammonia nitrogen is not being produced. In addition, the generally low organic content and few clay minerals observed in the soils at Crookes Point can also result in the decreased accumulation and retention of nitrogen.

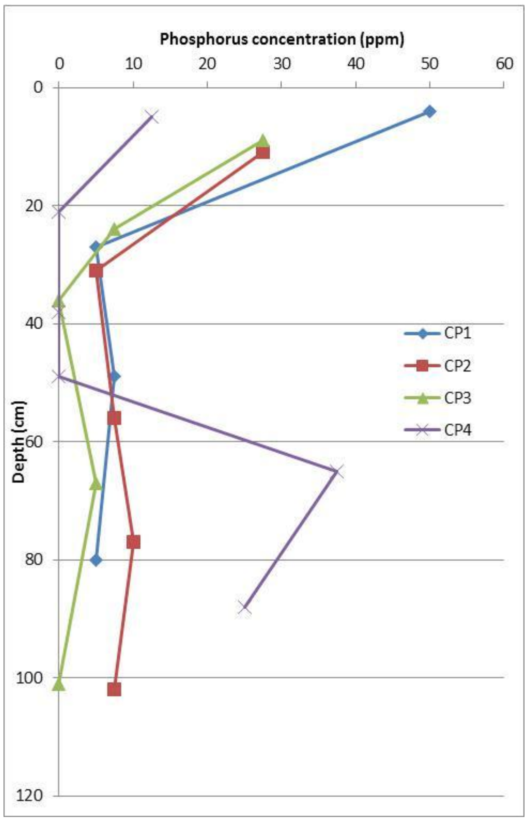

Figure 8: Graph of phosphorus concentration v depth at each location

An optimal level of phosphorus in sandy soil should be at least 25 ppm and more available phosphorus than this is desirable and beneficial (Tucker, 2009). At Crookes Point, all A horizons had available phosphorus concentrations higher than 25 ppm except for the A horizon of CP4 which had a concentration of 12.5 ppm. This decrease in available phosphorus may be due to the high demand of the nutrient from the large stand of *Phragmites* present at the site. Phosphorus concentrations were found to taper off with soil depth for all locations except for CP4 which had 37.5 ppm of available phosphorus in the C4 horizon and 25 ppm in the C5 horizon (concentrations higher than that in the A horizon). The high concentrations found in the C4 and C5 horizons may be due to calcium phosphate leaching from the C3 layer which had a large amount of oyster shells and calciferous matter in the fill material.

Potassium concentrations were low in most samples analyzed due to the sandy nature of the soil. The higher concentrations were found in the A horizons which is due to the increased weathering and erosion which occurs here. The A horizon of CP4 had the lowest level of available potassium in the soil (below 50 ppm) which may be due to the large stand of *Phragmites* present at the site resulting in increased plant uptake of this nutrient and decreased weathering and erosion. Most of the potassium at Crookes Point cannot be stored in the soil and leaches through the profile which explains why concentrations of available potassium were below detection limits for the majority of the samples.

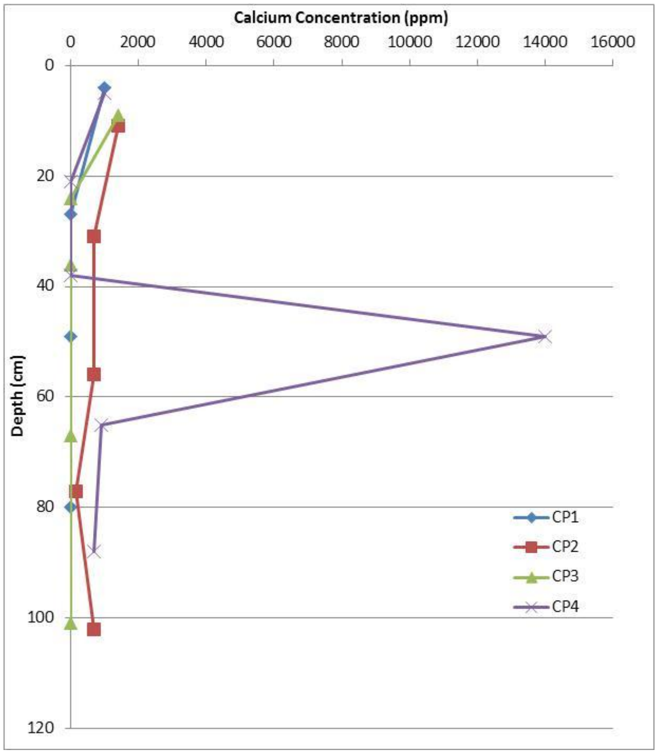

Figure 9: Graph of calcium concentration v depth at each location

Sandy soils normally contain less calcium than clay or organic soils and should give readings of approximately 500ppm for healthy plant growth (Tucker, 2009). At Crookes Point, the A horizons of all sites had concentrations at or above 1000ppm (Figure 9). Overall, CP4 had the highest concentrations of replaceable calcium throughout the profile with the largest concentration (≥14,000 ppm) found in CP4-C3. This soil horizon was predominately composed of oyster shells and other shell fragments. Leaching of calcium is visible as the concentration visible tapers off in CP4-C4 and CP4-C5. All samples from CP2, excluding the sample from the C3 horizon, had concentrations over 700ppm as well. Test results confirm and supplement soil acidity readings. Soil from CP1 and CP3 tended to be slightly more acidic which may indicate that the active calcium has been replaced by hydrogen or other ions (Troeh & Thompson, 2005). In addition, the pH of soil samples below CP4-C2 were noticeably more alkaline in comparison to soil samples taken at and above the horizon.

Most concentrations of magnesium were “low” (approximately 10 ppm), but in the A horizon of CP1 and the C3 horizon of CP3, magnesium concentrations were “medium to high” (resulting concentration value of approximately 52.5 ppm). The A horizon of CP3 and CP4 have a “Low to Medium” concentration of approximately 17.5 ppm. Magnesium deficiencies at CP2 may inhibit the growth of vegetation at this site. Plantings may be more successful at CP1 and CP3 where magnesium is less of a limiting factor.

In this study, results of the sulfate test were interpreted as concentrations at or below 50ppm, which is the lowest detectable limit. It is possible that results of the sulfate analysis are possibly inconclusive as there was no observed turbidity change or precipitate formed. However, as sulfate leaches easily, this figure may be accurate since the soil at Crookes Point is well drained.

Any positive test reading, even a very low reading, generally indicates the presence of sufficient available manganese to meet plant requirements (Tucker, 2009). However, soils analyzed from Crookes Point resulted in a negative test reading which may suggest a manganese deficiency.

It is possible that manganese is not at its soluble form or leached rapidly at Crookes Point since soils, on average, were slightly acidic.

While most soils contain abundant iron, only a fraction is soluble and readily available to the growing plant as iron readily forms insoluble complexes with carbonates, phosphates, and hydroxides in the soil solution (Tucker, 2009). Like manganese, this is particularly true in neutral or alkaline soils. Acid soils contain higher levels of available iron (Troeh & Thompson, 2005), which is evident when comparing Crookes Point soils. The largest concentration of iron was found in CP4 with 62.5 ppm in the C2 horizon and 50 ppm in C1. When comparing to pH, CP4-C2 and CP4-C1 were found to be the most acidic soils out of all Crookes Point samples. Looking at the profile on a whole, CP1 which is slightly more acidic than CP2 and CP3, has marginally more ferric iron available to plants, although it is still very low.

Active aluminum concentrations in Crookes Point soils varied, ranging from "Very Low" to greater than "Very High". CP2 appears to have a preferable concentration of active aluminum as each sample had a concentration less than 5 ppm. CP1 had soil which ranged from "Very Low to Low" in the A horizon, "Very High" throughout the rest of the profile and greater than "Very High" in the clay sample. A high test result indicates an undesirable acid soil and may have unfavorable conditions for planting. Plants which normally thrive on acid soils may fail on a soil with a high active aluminum test reading, much like that found at CP1 and CP4 (Tucker, 2009).

Observed Vegetation

CP1 – Herbaceous cover dominated by Japanese honeysuckle (*Lonicera japonica*), poison ivy (*Toxicodendron radicans*), Virginia creeper (*Parthenocissus quinquefolia*), multiflora rose (*Rosa multiflora*), oriental bittersweet (*Celastrus orbiculatus*), bristly dewberry (*Rubus hispidus*), and white poplar (*Populus* sp.) (Figure 10).

Figure 10: Herbaceous cover dominated by Japanese honeysuckle, poison ivy, Virginia creeper multiflora rose, oriental bittersweet, bristly dewberry, and aspen species in CP1.

CP2 – Vineland dominated by multiflora rose (*Rosa multiflora*) and Japanese honeysuckle (*Lonicera japonica*) (Figure 11).

Figure 11: Vineland dominated by multiflora rose and Japanese honeysuckle in CP2.

CP3 – Woodland dominated by black cherry (*Prunus serotina*), poison ivy (*Toxicodendron radicans*), and oriental bittersweet (*Calastrus orbiculatus*) (Figure 12).

Figure 12: Woodland dominated by black cherry, poison ivy, and oriental bittersweet at location CP3.

Figure 13: A large stand of *Phragmites* dominates the area at sampling location CP4.

CP4 – Herbaceous cover dominated by large stand of *Phragmites* sp. (Figure 12), multiflora rose (*Rosa multiflora*), and locust sp. (Figure 13).

Recommendations for Vegetation Planting

The goal of intended plantings at Crookes Point is to make the area as sustainable as possible by reestablishing a Mid-Atlantic Coastal, Back Dune, or Shrub-Scrub ecosystem through the removal of existing invasive vegetation and replanting of native species. Native species currently found in the inland maritime forest at Crookes Point include white poplar (*Populus grandidentata*), black cherry (*Prunus serotina*), eastern red cedar (*Juniperus virginiana*), sweet gum (*Liquidambar styraciflua*), paper birch (*Betula papyrifera*), white pine (*Pinus strobus*) and pitch pine (*Pinus* rigida). Native vines such as Virginia creeper (*Parthenocissus quinquefolia*) and poison ivy (*Toxicodendron radicans*) are also present. Shrubs of northern bayberry (*Myrica pensylvanica*) and winged sumac (*Rhus copallina*) are found closer to the shoreline. These species should be considered for planting at the site since they are known to historically exist at Crookes Point. Gateway National Recreation Area's herbarium database (National Park Service 1997) provides essential historical information regarding the additional herbaceous species present in the park. This useful tool is especially beneficial when selecting species to be planted in the shrub-scrub or herbaceous layers.

The maritime holly forests of Fire Island National Seashore can be seen as the ultimate goal for this site, as it is a stable successional maritime forest comprised of numerous native species. This rare ecological community is only found behind well-established sand dunes along the Atlantic coast from New Jersey to Massachusetts and is protected because of its rarity (Art, 1976). The maritime holly forest of Fire Island National Seashore is dominated by American holly (*Ilex opaca*), as well as black cherry (*Prunus serotina*), gray birch (*Betula populifolia*), post oak

(*Quercus stellata*), sassafras (*Sassafras albidum*), and pitch pine (*Pinus rigida*). Vines such as wild grape (*Vitis* sp.), roundleaf greenbriar (*Smilax rotundifolia*), Virginia creeper (*Parthenocissus quinquefolia*), and poison ivy (*Toxicodendron radicans*) are common and provide cover and fruit for birds and other wildlife. In addition to saplings of the dominant tree species, the shrub layer at Fire Island National Seashore includes northern bayberry (*Myrica pensylvanica*), highbush blueberry (*Vaccinium corymbosum*), and serviceberry (*Amelanchier canadensis*). Historically, wild sarsaparilla (*Aralia nudicaulis*), starry false solomons seal (*Smilacina* stellata) and pink ladies slipper (*Cypripedium acaule*) dominated the herbaceous layer, but browsing has reduced their prevalence and it is now characterized by Pennsylvania sedge (*Carex* pensylvanica), starflower (*Trientalis* borealis), and Canada mayflower (*Maianthemum* canadense) among others (Forrester et al., 2008).

When selecting species to plant at Crookes Point, the native species found in Great Kills should be given precedent, with weight also given to the native species found at Fire Island National Seashore. An extensive species list of native plants in Fire Island National Seashore can be used as an additional guide. Comparing these species with the "PLANTS Database," eastern red cedar (*Juniperus virginiana*), paper birch (*Betula papyrifera*), northern bayberry (*Myrica pensylvanica*), winged sumac (*Rhus copallina*), highbush blueberry (*Vaccinium corymbosum*), and serviceberry (*Amelanchier canadensis*) appeared to be the best suited for the soil texture and pH found at Crookes Point. Planting of these and other well-suited species should allow for the development of a successional maritime forest behind the dunes and a diverse shrub-scrub layer closer to the water's edge, resulting in increased ecological diversity and sustainability at Crookes Point.

Concerns regarding Herbicide use

Three herbicides have been proposed by the NYC Department of Parks & Recreation as possibilities for use in the removal of vegetation at Crookes Point:

Garlon 4 Ultra (EPA Reg No 62719-527)
Accord XRT II (EPA Reg No 62719-556)
Accord Concentrate (EPA Reg No 62719-324)

This assessment is primarily concerned with possible contamination of groundwater at Crookes Point, related to the soil composition and the shallow nature of the water table, due to Crookes Point having the hydrology of a typical barrier spit or island. There is not enough data available either for the chemicals in these herbicides, or for the soils and groundwater at Crookes Point for detailed modeling of the system. However, broad conclusions can be drawn based on literature studies of similar environments.

Garlon 4 Ultra

The active ingredient in Garlon 4 Ultra is tryclopyr butoxyethyl ester, sometimes referred to as tryclopyr BEE. It is used as an herbicide for broadleaf plants, but does not kill grasses and conifers. The Specimen Label from Dow states that "The use of this chemical in areas where soils are permeable, particularly where the water table is shallow, may result in groundwater contamination." (Dow AgroSciences LLC, 2007). The data behind this recommendation are summarized in a comprehensive literature review by Marin Municipal Water District (Kegley et al., 2010).

Tryclopyr BEE degrades in both soil and water to form tryclopyr acid, with a half-life of a few hours to half a day, and this degrades further to form 3,5,6-trichloro-2-pyridinol (TCP) and 3,5,6-trichloro-2-methoxypyridine (TMP), with a much longer half-life which is strongly dependent on environmental conditions. Tryclopyr acid is relatively mobile, as it is soluble and is not strongly adsorbed to soil, whereas Tryclopyr BEE and TCP are more strongly adsorbed to soils. However, studies indicate that there is much poorer adsorption to soils where porosity is high and organic matter is low, which is exactly the situation at Crookes Point (Kegley et al., 2010). This, combined with a relatively shallow water table leads to the possibility of extensive leaching of the herbicide and its initial breakdown products to the groundwater.

Accord XRT II and Accord Concentrate

The active ingredient in both Accord XRT II and Accord Concentrate is glyphosate isopropylamine (IPA) salt. Although glyphosate breaks down more slowly (half-life of 2 – 197 days in soil and from a few days to 91 days in water) it is considered to have moderate to no toxicity to fish and aquatic species, and under most conditions it adsorbs strongly to soils and will not be leached into the groundwater (Miller et al., 2010).

Unlike most other herbicide chemicals, glyphosate sorbs to mineral surfaces in the soil, rather than to organic matter. It is therefore the mineral composition of the soil, in particular the presence of iron and aluminum oxides, that controls the mobility of glyphosate in soil and its potential for leaching to the groundwater (Borggaard and Gimsing, 2008). Preliminary studies of the soils at Crookes Point indicate that they have much lower levels of iron and aluminum than average soils, and therefore there would be much less adsorption of the glyphosate to the soil than in 'typical' application areas. Further analysis of the soils would be necessary to check whether the herbicide would be taken up by the unusual soils at Crookes Point. If the herbicide is not strongly adsorbed to the soil, then the porous nature of the soil combined with the shallow water table would lead to the contamination of groundwater.

Conclusions

The purpose of this research was to establish the biological, chemical, and physical parameters of Crookes Point to develop a successful vegetation management plan and to restore Crookes Point to a healthy and sustainable ecosystem through removal of invasive species and replanting of native vegetation.

Soil within the 28-acre site was characterized through the completion of four pedon descriptions which identified key morphological features and defined the soil structure. Samples which were collected underwent pH and nutrient analysis in the laboratory to aid in the selection of appropriate plant species for reforestation efforts. Baseline vegetation surveys at the time of sampling in addition to vegetation data compiled from Gateway National Recreation Area's Herbarium offered additional insight regarding changes to the species composition at Crookes Point over time. Historical research documented the geological changes to Crookes Point over the last century and determined the source of the soil's parent material (fill) when the area was converted into a city park. Healthy ecosystems such as the holly forest of Fire Island National Seashore served as representatives of typical ecological communities in the mid-Atlantic region and provided additional information when selecting native species.

Based on this research, it was determined that species such as eastern red cedar (*Juniperus virginiana*), paper birch (*Betula papyrifera*), northern bayberry (*Myrica pensylvanica*), winged sumac (*Rhus copallina*), highbush blueberry (*Vaccinium corymbosum*), and serviceberry (*Amelanchier canadensis*) when planted would be well suited for the area and help form a Mid-Atlantic back dune coastal ecosystem. Recently, removal of invasive species at Crookes Point by physical and chemical methods has occurred, with native plantings set to begin in the Fall of 2013. If these and similar species are utilized, with proper monitoring and stewardship, Crookes Point can provide a richer habitat for migrating birds, pollinators, resident mammals and wildlife, and be restored to its full ecological potential.

Acknowledgements

This research was supported by the *Gateway Research Learning Center Fellowship Program* through the Natural Resource Management Division at Gateway National Recreation Area, and a grant from *Eastern National.* My sincere thanks and appreciation to Natural Resource Management Division Chief Douglas A. Adamo and Biological Science Technician William S. Parker for their phenomenal guidance and much needed assistance in the field.

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THE HIGHLANDS: CRITICAL RESOURCES, TREASURED LANDSCAPES

Richard G. Lathrop Jr.

The Highlands, a 1,000,000+ hectare area of rugged uplands, stretch from northeastern Pennsylvania, across northern New Jersey, southern New York into northwestern Connecticut. An outlier of the New England physiographic province, the Highlands serve as a natural geographic boundary delimiting the northern edge of the New York City and Philadelphia metropolitan regions. Adjacent to some of the most densely populated areas in the United States and undergoing increasing pressure from sprawling development, the Highlands region presents a compelling case study in how humans in concert with environmental forces have shaped and continue to shape a landscape. Equally compelling are the myriad efforts that have been undertaken through watershed protection, open-space preservation, and bioregional conservation planning to conserve the essential elements of that landscape. Simultaneously there has been a push to forge a broader regional multistate identity that spans more parochial interests.

In this light, the book "*The Highlands: Critical Resources, Treasured Landscapes"*(published by Rutgers University Press, 2011) was written to provide a comprehensive look at the natural and cultural history of this vital region that serves as the Mid-Atlantic region's back yard. The volume was edited by Dr. Richard Lathrop, Professor of Ecology, Evolution and Natural Resources and Director of the Walton Center for Remote Sensing & Spatial Analysis at Rutgers University, who has been intimately involved in Highlands' environmental research and policy for over 10 years. The chapters were written by leading researchers and specialists at a level that is accessible to students, teachers, policy makers and interested laypersons.

So where exactly are the Highlands? Encompassing a four-state region that extends from Pennsylvania, in the south, northeast up through New Jersey, New York, and into Connecticut, the Highlands come under many different local names. Physiographically, the Highlands are generally defined by the extent of their underlying Precambrian crystalline bedrock and its upland terrain. Administratively, the federal Highlands Conservation Act in 2004 takes a more expansive view by including the entirety of the area of any municipality that falls partly within the Highlands ecoregion, thereby extending the boundary well beyond the ecoregion proper and encompassing nearly 3.5 million acres (see Figure 1).

One of the most aesthetically appealing aspects of the Highlands is the rocks: big, bold, substantial rocks—from the occasional sheer rock cliff with a skirt of tumbled talus at its base to the stray glacial "erratic" boulder the size of a small house. As described in chapter 1 (co-authored by Alexander Gates and David Valentino), the Highlands are composed of billion-year-old bedrock that formed the roots of the ancient Grenville Mountains. Over the eons, the Highlands rose and fell only to rise again, and the bedrock subsequently folded and transformed under the tremendous pressures of colliding continental plates. The native bedrock metamorphosed into various forms of gneiss, a crystalline rock that is highly resistant to weathering and erosion. The shearing forces of the colliding plates also faulted and fractured the bedrock, creating conduits for mineral-rich fluids that cooled to form veins rich in iron. These deposits of iron served as the impetus for much of the early industrial development of the Highlands, playing a vital role in the rise of the United States as an industrial powerhouse in the nineteenth century.

Tens of thousands of years earlier, during what is known as the Wisconsin glacial episode, a continental-scale ice sheet bulldozed southward across the Highlands, stopping in New Jersey. The resulting terminal moraine forms a boundary separating the northern glaciated and the southern unglaciated Highlands. Chapter 2 (authored by Scott Stanford) recounts how, north of the moraine, glaciers sculpted the landscape, gouging valleys while scraping ridgetops bare. In the glacially gouged low spots, water pooled to form lakes and ponds. The moving ice and meltwater transported, sifted, and deposited sand and gravel in beds tens of feet thick in some Highlands valleys. The resulting valley-fill aquifers serve as a big sponge, holding millions of gallons of potable water (a vital resource for Highlands residents).

Along with geology and topography, a location's soils are a major determinant of possible human use of the land. Chapter 3 (co-authored by John Tedrow and Richard Shaw) discusses the many factors that determine a location's soil, attributes of the soil, and how these characteristics vary across the Highlands landscape. On the glacially scoured uplands, the soils are often shallow and rocky; generally unsuitable for agriculture, such areas remain or have returned to forest. In low spots on the landscape, fine-grained clays have been deposited, forming an impervious layer that impedes the downward infiltration of water. Under these conditions, the higher water tables and waterlogged soils promote wetlands vegetation. South of the moraine, the valleys are broader, and the soils, often formed from calcareous limestone, are deeper and mellower, making for rich farmland. The diversity of the Highlands forest and wetland ecosystems is a reflection of the interplay between the bedrock geology, topography, glacial history, and soils.

Clean and abundant water is a resource that we often take for granted in the humid northeastern United States. Turn on the tap, and as long as drinkable water flows out, that is the end of story for many of us. However the backstory behind water is much more complex and much more interesting. Indeed, many consider water as <u>the</u> critical resource in the present-day Highlands; many of the fights over land are in reality fights over water. While iron may have been the key Highlands commodity in the nineteenth century, water has become the key commodity of the twentieth century and will likely remain so in the twenty-first century.

Chapter 4 (co-authored by Otto Zapecza, Donald Rice and Vincent DePaul) discusses the importance of the hydrological cycle in determining the quality and quality of both the surface and groundwater in the Highlands. The region's abundant precipitation either infiltrates the surface to become groundwater or runs off across the surface to a neighboring stream. Whereas water can infiltrate between the grains of porous sedimentary rocks, water has a hard time infiltrating through the tightly knit crystalline matrix of the gneiss bedrock so characteristic of the Highlands. Instead of moving through the rock matrix (and consequently being filtered in transit), water moves through the open conduits formed by the many fractures. As a result, these so-called fractured rock aquifers do not store large amounts of water, and the groundwater that is there can be easily compromised. Where they occur, valley-fill aquifers are sources of large amounts of readily available water, but this groundwater is also easily polluted. One of the most important twists about the Highlands is that most of the surface water supplies stored in the region's many reservoirs are exported beyond the Highlands proper to serve the citizens of the broader mid-Atlantic and southern New England metropolitan region. Residents living within the Highlands rely primarily on groundwater pumped from wells.

The connection between human alteration of the landscape (through development and agriculture) and the consequent degradation of the region's water quality is a critical issue. As discussed in chapter 5 (authored by Daniel VanAbs), the mid-1800s brought increasing recognition that water-borne disease constituted a major public health challenge and that protecting watersheds protected public health. Metropolitan governments outside the Highlands looked to the Highlands as the ideal nearby location to acquire watershed land and to situate reservoirs. A complex system of reservoirs and interconnections to move water around has evolved over the past century and a half. While the Highlands produce millions of gallons of water per day, water supplies have been highly taxed during periods of prolonged drought, leading to shortages and severe water use restrictions. More recently, the concept of water rights has expanded to ensure sufficient releases of water to sustain downstream aquatic ecosystems. With the increasing competition for the remaining open lands, there are fewer prospects of building new reservoirs, thus engendering a greater emphasis on management and protection of existing supplies.

Forests are central to the character of the Highlands landscape. As highlighted throughout this section, natural ecosystems are dynamic, continually responding to varying environmental conditions or discrete disturbances; change is a constant. Chapter 6 (authored by Emily Southgate) discusses how the Highlands' forests were heavily exploited in the eighteenth and nineteenth centuries as a homegrown energy source in the form of charcoal to power the iron industry or as fuelwood to heat homes. Most of the region was repeatedly cut over so there is little or no original "virgin" forest to be found. As recounted in chapter 7 (authored by William Schuster), when given a reprieve from the wholesale cutting in the late 1800s, these forests quickly regenerated to again cover the Highlands in a thick mantle of upland forest dominated by a mixture of broad-leaved deciduous trees. Chapter 8 (authored by Joan Ehrenfeld) explores the region's wetlands and explains how swamps, bogs, and herbaceous marshes add to the landscape's habitat and biotic diversity.

As noted in chapters 9 (co-authored by Gerry Moore and Steven Glenn) and 10 (authored by Elizabeth Johnson), the Highlands are home to a rich diversity of plant and animal life, but very few endemic species are unique to the Highlands. In fact, the Highlands forest is quite similar in structure and composition to forests across the entire sweep of the Appalachians from Georgia to Maine. What makes the Highlands forests different and of special value is their proximity to the teeming human population centers of the Hartford to New York City to Philadelphia region. These renewed forests and wetlands are increasingly valued for the "ecosystem services" they provide, such as their ability to serve as natural water and air filtration systems or their role in storing carbon. Not to be forgotten or belittled—from a selfish human point of view—the Highlands forests, the wetlands, and the resident plant and wildlife provide aesthetic values enhancing the quality of life for the human residents as well. While demonstrating great resiliency in the past, the Highlands forests, wetlands, and native biota are under assault from a host of new forces: habitat conversion, imbalances in wildlife populations, invasive plants, exotic pests, point/nonpoint-source water pollution, atmospheric pollutant deposition and global climate change.

When considering the lush forested vistas of the present-day Highlands, it is easy to overlook this region's previous incarnation as a center of early-American industry. Chapter 11 (co-authored by Theodore Kury and Peter Wacker) lays out the history of the iron industry and the Highlands' role in helping to forge America's rise as an industrial giant. The iron industry also helped shape the development of the region's rail and canal infrastructure of the nineteenth century. While most of the ironworks are long gone, the Morris Canal defunct, and the minor rail lines abandoned, the iron industry and associated transportation networks helped set the template of later urban development in the twentieth century. Not to be forgotten is the agricultural heritage of the Highlands, which was regionally significant in the nineteenth century. Chapter 12 (authored by Richard Lathrop) recounts how farming across the Highlands quickly evolved from an initial subsistence phase to a commercial phase oriented to supporting the rapidly growing urban centers of the mid-Atlantic and southern New England. By the mid-1800s agriculture in the region was on the wane, afflicted by, among other troubles, increased competition from the newly settled American Midwest. While much diminished, this agricultural heritage still lives on in some of the more fertile Highlands valleys, especially in the southern Highlands of New Jersey and Pennsylvania.

As northeastern US urban populations expanded, enhanced transportation and greater amounts of leisure time enabled a return to the land—this time for recreation. The Highlands saw many "firsts" in the development of American outdoor recreation—early resort development, leadership in the state parks movement, completion of the first leg of the Appalachian Trail (chapter 13 authored by Daniel Chazin). The mid- to late twentieth century saw the beginnings of a major transformation of the Highlands. Post–World War II population growth, increasing affluence, and the birth of the state and interstate highway system unleashed a massive wave of sprawling suburbanization. Concern about the loss of the Highlands' wild and scenic character,, in particular the Hudson Highlands and Storm King as iconic American grandeur, spurred the growth of the environmental movement and the push for more stringent land-use planning.

As discussed in chapter 14 (co-authored by Robert Pirani, Thomas Gilbert and Corey Piasecki), the region serves as a test bed for multistate cooperation in the form of the four-state Highlands Conservation Area as well as state-level planning as mandated by the New Jersey Water Protection and Planning Act. As made clear in chapter 15 (coauthored by Richard Lathrop, Mary Tyrrell and Myrna Hall), a failure by these aforementioned conservation initiatives and a "business as usual" continuation of existing development trends will lead to further reductions in forests and farmlands and a reduction in watershed integrity. A future vision for the Highlands is laid out, where large tracts of unfragmented forest help to sustain watershed integrity, uncluttered vistas, and thriving communities within a working landscape.

It is the authors' hope that this book will help to promote understanding of the Highlands as a coherent region with a unique identity, to explain the significance of its resources as critical to the mid-Atlantic and southern New England states and beyond, and to underline why this landscape is treasured by so many.

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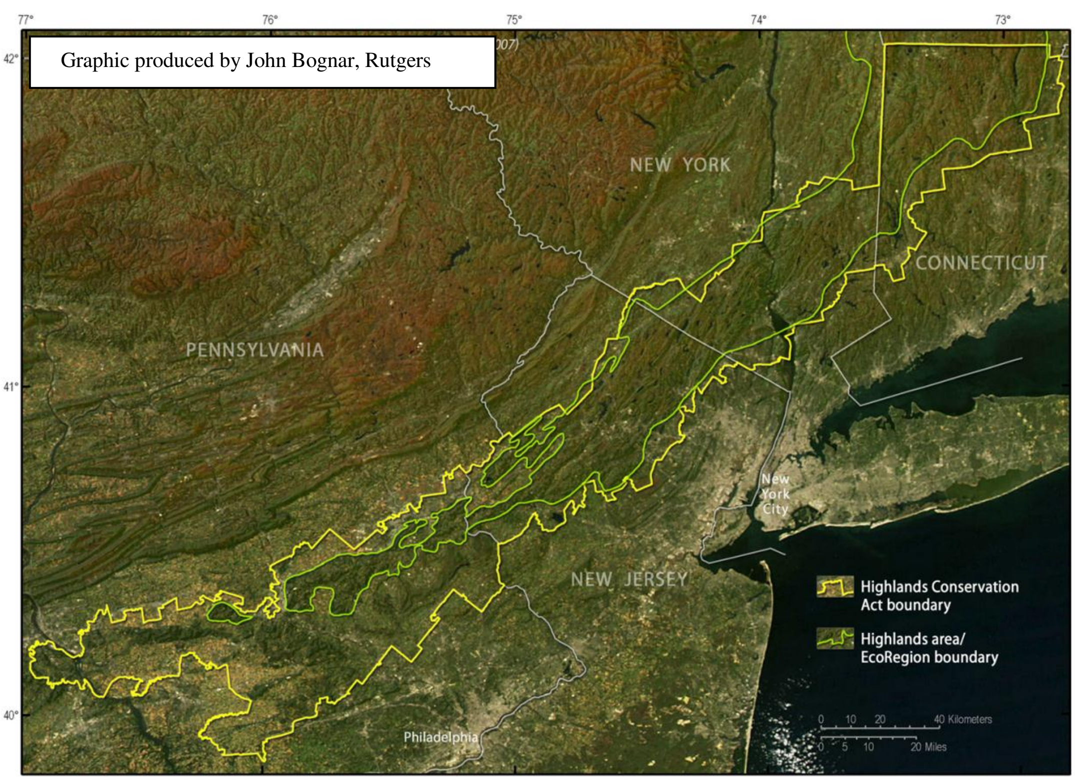


MODIS (Moderate Resolution Imaging Spectroradiometer) Aqua satellite image, dated October 11, 2008, of the greater 4-state Highlands region. Image source: NASA; Conservation Act boundary source: U.S. Forest Service; Ecoregion source: Commission for Environmental Cooperation, courtesy of the US Forest Service.

Figure 1. Satellite image map of the 4-state Highlands region.

FIELD TRIP

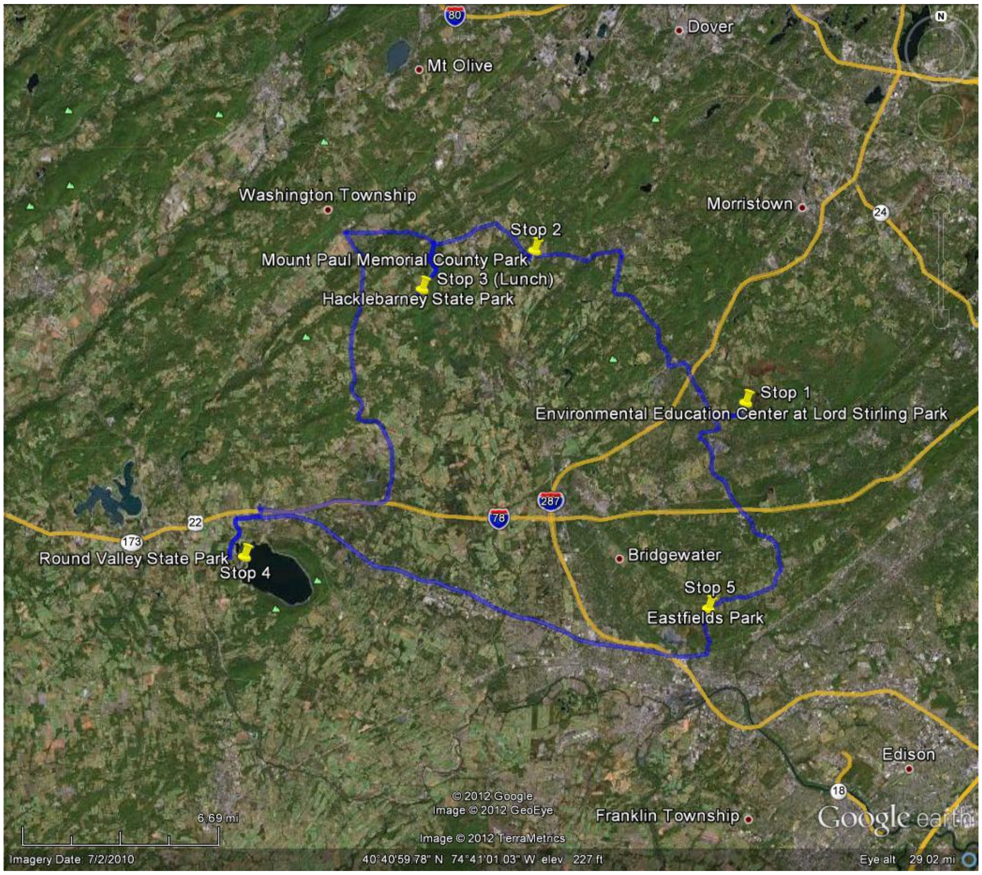


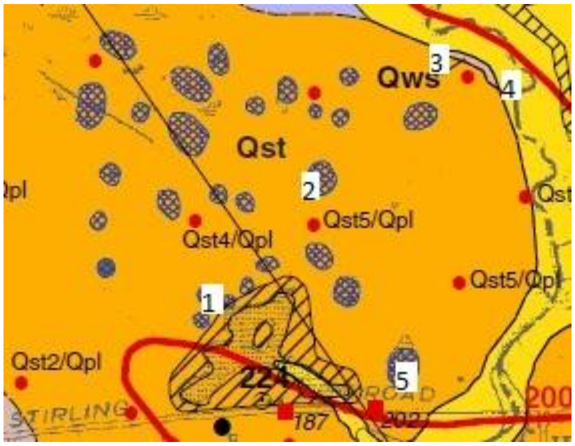
Figure 1: Map of stops to be visited

Stop 1 Environmental Center at Lord Stirling Park

Lat 40°41'42.28"N Lon 74°31'14.94"W

Jane Alexander, College of Staten Island/CUNY

The Environmental Center at Lord Stirling Park provides access to trails at the western end of the Great Swamp. The swamp has formed on the former site of glacial Lake Passaic, and is important as a unique ecological environment and because of its impact on flooding on the Passaic River. We will examine the surface drainage of the Great Swamp and how this has been influenced by past geologic events.



Qws	Weathered Shale
Qal	Alluvium
Qst	Stream Terrace Deposits
	Artificial Fill
	Thermokarst Basin

Figure 2: Surficial Geology of the Great Swamp. (From: OFM 74, Surficial Geology of the Bernardsville Quadrangle, Morris and Somerset Counties, New Jersey, Stanford, Scott D., 2008, scale 1 to 24,000)

Figure 3: Trail map of Lord Stirling Park showing points of interest.

1. Bullfrog Pond

Bullfrog Pond is a permanent pond located in a thermokarst basin (Figure 2 and 3). These shallow basins were formed from melting ground ice as the glacier retreated. Water levels in this pond vary seasonally depending on rainfall, as evidenced by the bordering plants that are adapted to the wetland environment.

Figure 4: Thermokarst basin.

2. Wooded Swamp Area

Several shallow thermokarst basins in the wooded area of the swamp are flooded seasonally, although do not hold permanent ponds. The water is stagnant and anaerobic, leading to the preservation of fallen leaves and the buildup of thin peat deposits (Figure 4).

3. Lenape Meadow

Lenape Meadow has a notably different ecology to the surrounding swamp area. It is located on an outcrop of weathered shale bedrock, rather than the think alluvial deposits typical of the Great Swamp (Figure 2). The marginally higher altitude and good drainage allow for the establishment of meadow plants. The meadow is, however, artificially maintained by annual burning, to prevent the encroachment of forest trees.

4. East Observation Tower on the Passaic River

This observation tower affords a view of a section of the Passaic River that is frequently flooded (Figure 3). This flood plain – and the storage capacity of the Great Swamp wetland as a whole – reduces flooding downstream on the Passaic River, by releasing water from heavy rainfall over a longer period of time. The tower also provides an overview of a large meander on the river, which makes it an excellent location to bring students learning about stream processes.

5. Lily Pad Pond

This is one of the larger, permanent ponds found in a thermokarst basin (Figure 2 and 5). Unlike Bullfrog Pond, the water in this pond is connected to the Passaic River, and as it fills up during heavy rainfall, the excess water drains. Consequently, there is little change in the water level, and the plant life reflects this.

Figure 5: Lily Pad Pond

Depart Environmental Education Center at Lord Stirling Park for Stop 2 (11.1 mi - about 25 mins)
1. Head west on Lord Stirling Rd toward Riverside Dr 1.0 mi
2. Turn right onto Somerset 657/S Maple Ave 1.0 mi
3. Turn left onto E Oak St 331 ft
4. Turn right onto N Finley Ave/Somerset County Road 613 1.1 mi
5. Turn left onto Morristown Rd 0.6 mi
6. Slight right onto Olcott Square 164 ft
7. Slight right onto Anderson Rd/Anderson Hill Rd, Continue to follow Anderson Hill Rd 1.0 mi
8. Continue onto County Road 525/Mendham Rd, Continue to follow County Road 525 2.5 mi
9. Turn right onto County Road 525/Hilltop Rd 0.5 mi
10. Slight left to stay on County Road 525/Hilltop Rd 1.0 mi
11. Turn left onto County Road 510/W Main St/New Jersey 24, Continue to follow County Road 510/New Jersey 24 2.2 mi
12. Turn left onto Mount Paul Place 0.2 mi

Arrive at Mount Paul Memorial County Park on right

Stop 2 Mount Paul Memorial County Park

Lat 40°46'12.38"N Lon 74°39'31.23"W

The Mt. Paul Trail

John H. Puffer, Rutgers University, Newark.

From the parking area about 1/8th mile south of Rt. 24 hike west into the abandoned shale/slate quarry area. The quarrying activity has resulted in excellent exposures of the Ordovician Jutland klippe upper unit B. The area has been mapped by Volkert et al, (1990) who have measured the strike and dip at several locations around the quarry. The dip measurements average about 70^{o} to the south-east and strike about 45^{o} to the north-east. The beds have formed folds that plunge at low angles (about 6 to 16 degrees to the northeast and the southwest. The composition of the red slate and the green slate is the same except for the oxidation state of the iron in the red slate that has oxidized to rust.

Continuing south and west on the trial very little Ordovician rock is exposed. However, as the summit of Mt Paul is approached within about ¼ mile in any direction pebbles and cobbles of red conglomerate are exposed in gullies and in eroded soil exposures. This red Triassic rock is about 300 million years younger than the Ordovician rocks. The pebbles in the conglomerate are composed of quartz and are cemented together with silt enriched in red iron.

As the trail from the summit descends toward the southwest more Ordovician rock is encountered until Gladstone Brook is reached. Gladstone Brook marks the fault contact with Byram granite that is nicely exposed on the valley wall to the west. The fault along the brook is a major New Jersey fault known as the Flemington Fault. The younger Ordovician rocks on the east side dropped down while the much older Byram Granite on the west side lifted up as indicated by the arrows on the cross section along the line A'-A. Therefore, the relative motion makes this portion of the fault a normal-type fault. The contact at Gladstone Brook is a contact between rock deposited about 480 million years ago and granite intruded about 800 million years ago. The granite and the adjacent Losee Gneiss were therefore uplifted several miles from their original sites of intrusion or metamorphism to reach their current position. This also means that the several miles of rock overlying the Byram and Losee have eroded away resulting in countless cubic miles of sediments, most of which were deposited as sand and mud in shallow marine environments.

It can also be observed that the Byram Granite has experienced the effects of crushing and grinding during movement along the Flemington Fault and has become finer grained and foliated. The sheared, fine-grained portion of the granite has become a rock type known as mylonite.

Reference cited

Volkert, R.A., Markewicz, F.J., and Drake, A.A., 1990, Bedrock geologic map of the Chester Quadrangle, Morris County, New Jersey. Geologic Map Series 90-1 New Jersey Geological Survey.

Depart Mount Paul Memorial County Park for Stop 3 (8.0 mi - about 18 mins)
1. Head north on Mount Paul Place toward County Road 510/New Jersey 24 0.2 mi
2. Turn left onto County Road 510/New Jersey 24 2.6 mi
3. Turn left onto County Rd 513 S/Main St/New Jersey 24, Continue to follow County Rd 513 S/ New Jersey 24 2.3 mi
4. Turn left onto Parker Rd 92 ft
5. Turn left onto State Park Rd 1.8 mi
6. Turn right to stay on State Park Rd 0.2 mi
7. Continue onto Hacklebarney Rd 0.4 mi
8. Turn left 0.3 mi
Arrive at Hacklebarney State Park

Stop 3 Hacklebarney State Park

Lat 40°45'3.88"N Lon 74°43'54.74"W

This location has facilities suitable for a lunch stop, although there will be no guided tour. The banks of the Black River have exposures of the Losee Gneiss and Byram Granite, and the river itself flows over large glacial boulders (Figure 6).

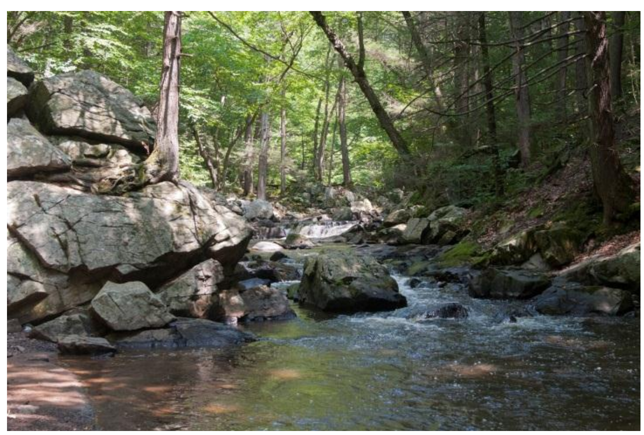


Figure 6: The Black River, Hacklebarney State Park

Depart Hacklebarney State Park for Stop 4 (23.5 mi - about 44 mins)
1. Head west toward Hacklebarney Rd 0.3 mi
2. Turn right onto Hacklebarney Rd 0.4 mi
3. Continue onto State Park Rd 0.2 mi
4. Turn left to stay on State Park Rd 1.8 mi
5. Turn right onto Parker Rd 92 ft
6. Turn left onto County Rd 513 S/New Jersey 24, Continue to follow County Rd 513 S 3.1 mi
7. Turn left onto Fairmount Rd 3.6 mi
8. Continue onto County Rd 517/Old Turnpike Rd 5.8 mi
9. Continue onto County Rd 523/Oldwick Rd 1.0 mi
10. Take the ramp onto I-78 W 4.3 mi
11. Take exit 20A toward Lebanon 0.2 mi
12. Merge onto Cokesbury Rd 0.3 mi
13. Turn right onto US-22 W 0.8 mi
14. Turn left onto Round Valley Access Rd 0.8 mi
15. Continue onto Lebanon - Stanton Rd/Round Valley Access Red, continue to follow Lebanon - Stanton Rd 0.4 mi
16. Continue onto Co Rd 629/Stanton Lebanon Rd 0.4 mi

Arrive at Round Valley State Park

Stop 4 Round Valley State Park

Lat 40°37'8.01"N Lon 74°50'51.47"W

Mesoproterozoic rocks along the west shoreline of Round Valley Reservoir.

Rich Volkert, New Jersey Geological and Water Survey

Intermittent outcrops along the shoreline of the reservoir, east of the boat launch area, display a variety of interesting petrological and structural features that range in age from Proterozoic to Mesozoic. The challenge in deciphering these features is “seeing through” the pervasive overprinting deformation. The outcrops at this stop are on the footwall of the Flemington Fault, a major regional structure that separates Proterozoic and Paleozoic rocks of the Highlands from down-dropped Mesozoic rocks of the Piedmont on the hanging wall. The fault, located about 500 feet east of the shoreline, strikes N.10^{o}E. and dips about 50^{o} SE. Bedrock geologic relationships are shown on the maps below (Figs. 7 and 8).

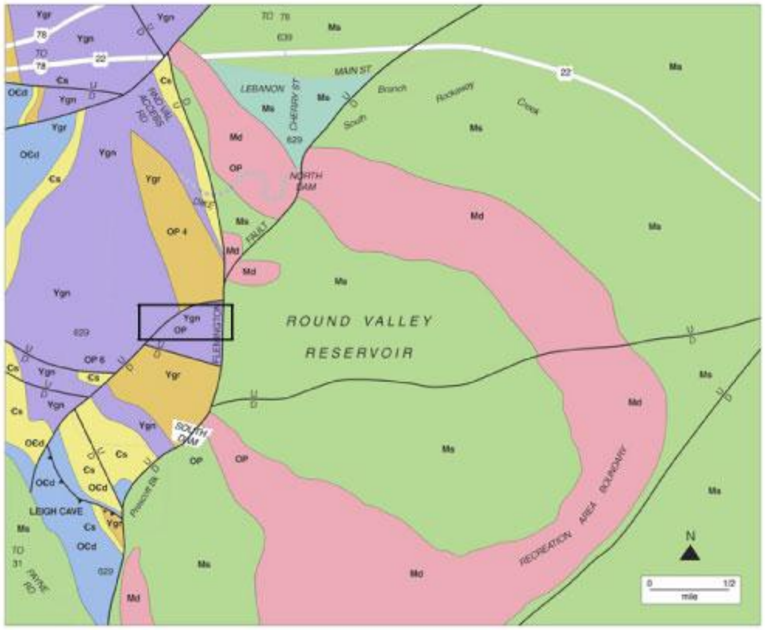


Figure 7. Simplified geologic map of Round Valley Recreation Area. Small box outlines area shown in figure 2. Md, Jurassic diabase (pink); Ms, Jurassic and Triassic sedimentary rocks (green); O€d, Ordovician and Cambrian dolomite (blue); €s, Cambrian sandstone (yellow); Ygr, Proterozoic granite (gold); Ygn, Proterozoic gneiss (purple). U and D are upthrown and downthrown sides, respectively, on faults (bold lines).

Most of the gneiss at this stop is salt-and-pepper weathering, black to grayish-black, medium-grained, foliated amphibolite composed of hornblende and plagioclase ± clinopyroxene and magnetite (Fig. 9A). The rock displays a penetrative metamorphic foliation that has obliterated most primary igneous textures of the basaltic protolith. However, a few outcrops display subhedral phenocrysts of plagioclase as long as 0.5 inch (Fig. 9B) that may be a relict feature inherited from the protolith. Some amphibolite outcrops contain abundant magnetite as disseminations, lenses and thin layers that are concordant to foliation and locally folded. The magnetite may be a more distal part of the thicker ore body hosted by amphibolite that was mined 2,500 to the west at the Large (Lebanon) mine, located on the north side of the reservoir.

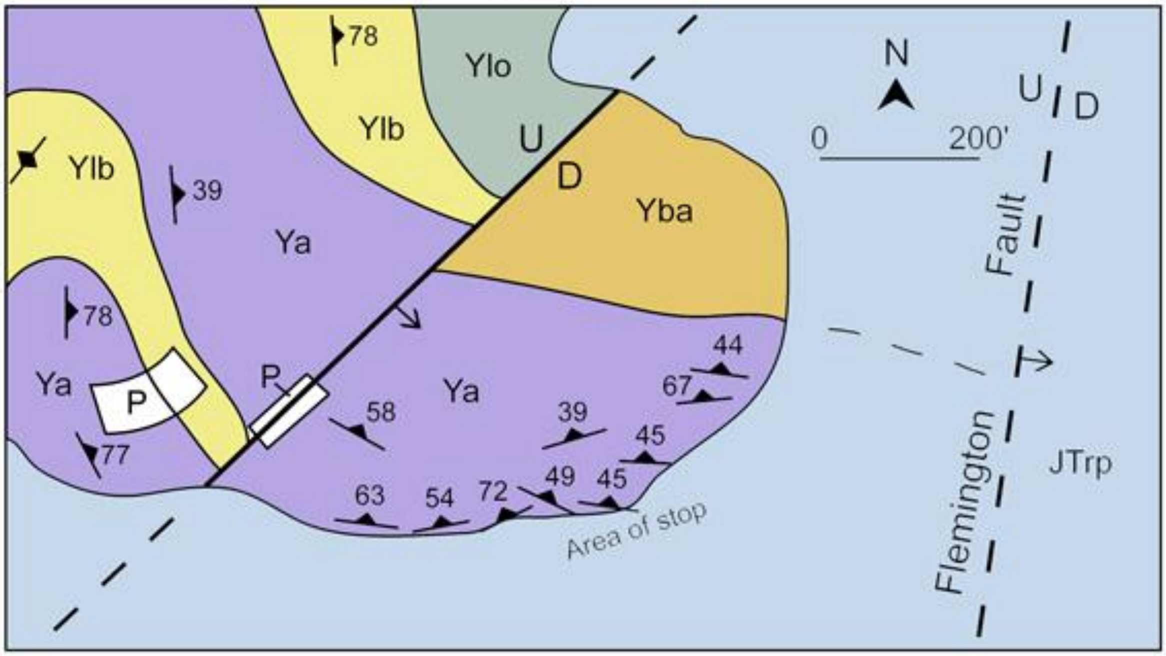


Figure 8. Simplified geologic map of the area of this stop. Symbols are: Ya, amphibolite; Ylo, quartz-plagioclase gneiss; Ylb, biotite-quartz-plagioclase gneiss; Yba, microperthite alaskite; and P, parking lot. U and D are upthrown and downthrown sides, respectively, on faults. Arrows show direction of dip of faults.

Very locally, amphibolite outcrops are intruded discordantly by white, coarse-grained, unfoliated pegmatites (Fig. 9C) composed of plagioclase, amphibole, clinopyroxene, magnetite, and trace amounts of biotite and chlorite. Most of the larger pegmatites along the shoreline are one to two feet wide and range in strike from N.15°E. to N.10°W. Note the rotation of the crystallization foliation along pegmatite contacts, suggesting that dilation of the amphibolite by a shear couple with left-lateral movement sense created space for the intrusion. Widespread pegmatites elsewhere in the Highlands yield U-Pb zircon ages of 1004 to 986 Ma (Volkert et al., 2005). Pegmatites at Round Valley are postorogenic and, therefore, comparable in age.

Amphibolite outcrops here also contain thin, conformable layers of coarse-grained partial melt generated in situ that is mineralogically similar to, and has gradational contacts with, the parent amphibolite. Some partial melt layers were subsequently folded along with the enclosing foliation (Fig. 3A) during high-grade metamorphism at ~1045 Ma, indicating they are older than the unfoliated pegmatite intrusions.

Intermittent outcrops of white or pale pink-weathering, medium-grained, foliated quartz-plagioclase gneiss ± biotite and hornblende (Fig. 9D) form layers as much as several feet thick that are in conformable contact with amphibolite. Similar quartz-plagioclase gneiss in the Highlands is part of the Losee Suite, an assemblage of calc-alkaline rocks formed in a continental-margin magmatic arc at 1282 to 1248 Ma (Volkert et al., 2010). Quartz-plagioclase gneiss at Round Valley may have formed from a volcanic protolith (dacite) or a plutonic one (tonalite). Geochemical analyses are lacking from these outcrops and bedrock exposure is insufficient to permit a definite interpretation. Regardless, the field relationships provide an important temporal constraint that indicates amphibolite here is also part of the Losee Suite.

Crystallization foliation in the Proterozoic rocks (Figs. 9A and 9D) is an inherited feature formed from west-directed compressional stress during the Ottawan orogeny. The timing of its formation has been dated by U-Pb zircon geochronology at 1045 to 1024 Ma (Volkert et al., 2010). Along the shoreline, foliation strikes mainly N.72^{o}E. to N.86^{o}W. (Fig. 8) and the dip ranges from 33^{o} to 77^{o} north. Outcrops at this stop are on the hanging wall of a northeast-striking normal fault that has detached the limbs of a synform, now preserved on the footwall, from the hinge of the fold on the hanging wall. Foliations in outcrops to the immediate north strike north to northwest and dip east, defining the northwest overturned limbs of the synform. The hinge of a complimentary antiform is mapped to the west, just north of the parking area (Fig. 8). Outcrops along the shoreline are deformed by faulting and fracturing and foliations locally diverge from the ranges given above. Mineral lineations on foliation surfaces plunge northeast and average 31^{o} N34^{o}E., parallel to the axes of minor folds observed in outcrop.

Proterozoic rocks locally display ductile deformation zones that strike N.60^{o}W. to N.80^{o}W. and dip north. Ductile features include shear band foliation oriented parallel to crystallization foliation, pinch-and-swell structures, boudinage, and shear folds (Fig. 10A). Because these deformation features reflect a higher temperature origin than the brittle features, they were probably formed during the Proterozoic.

Brittle deformation features include breccia zones as much as four feet wide (Fig. 10B), fracture cleavage (Fig. 10C), fault zones as much as two feet wide composed of cataclasis, fractures and slickensides coated by epidote or chlorite (Fig. 10D), and gouge. The strike of the breccia zones is somewhat variable but remains subparallel to the strike of crystallization foliation. The zones are composed of subangular to subrounded clasts of amphibolite in a matrix of cataclastic and retrogressed amphibolite. Breccia clasts are well foliated and some preserve isoclinal folds (Fig. 10B), providing an important constraint on the timing of brecciation as being post-Proterozoic and likely related to Paleozoic orogenesis. During brecciation, bedrock was broken up in situ, largely through hydraulic fracturing, and the clasts were then transported laterally. In some instances they can be reconnected to less deformed amphibolite along adjacent boundaries. The breccia zones are cut by fractures that strike N.35^{o}W. to N.65^{o}W. Breccia zones developed in Proterozoic rocks elsewhere in the Highlands are similarly confined to the footwall of major structural features such as the Reservoir, East and Tranquility faults, all of which have undergone reactivation during the Paleozoic.

Outcrops along the shoreline of the reservoir are deformed by pervasive brittle fractures that define dominant groups having an average orientation and movement sense of N.55^{o}E. 74^o SE to 71^o NW (high-angle oblique normal, or high-angle oblique reverse), N.05^{o}E. 73^o NW (dip-slip normal or reverse), N.45^{o}W. 74^o SW (high-angle oblique normal or reverse), N.40^{o}W. 70^o NE (left-lateral strike-slip), and N.80^{o}W. 59^o NE (left-lateral strike-slip). Cross-cutting relationships suggest a sequence of formation for the groups of N.55^{o}E., N.05^{o}E., N.45^{o}W., and N.80^{o}W. Assigning an age to each of these groups is hindered by the fact they show evidence for overprinting movement indicators resulting from multiple reactivations. The N.55^{o}E. group may have formed during the Proterozoic and then undergone reactivation during the Paleozoic and Mesozoic. The N.44^{o}W. group is very similar to the strike of the dominant joint set in Proterozoic rocks that ranges from N.40^{o}W. to N.60^{o}W. throughout the Highlands (Volkert, 1996). It is also similar to the strike of the thrust fault at Leigh cave that places Proterozoic rocks over Cambrian Leithsville Dolomite, suggesting this fracture group may also have originated during the Proterozoic and then undergone reactivation during the Taconic orogeny. Faults cutting diabase at the North Dam are clearly Mesozoic. They have strikes of about N.55^{o}W., N.55^{o}E. and N.10^{o}E. that overlap some of the dominant fractures measured in the Proterozoic rocks at Round Valley.

Looking south across the reservoir, toward the South Dam, the low slope on the right is underlain by Proterozoic and Paleozoic rocks, and to the left of the dam is the southern tip of Cushetunk Mountain, underlain by Jurassic diabase. The Flemington Fault extends through the midpoint of the dam. North of this stop the fault bifurcates beneath the reservoir. A western segment extends through the midpoint of the Dike, and an eastern segment extends through the western part of the North Dam (Fig. 7).

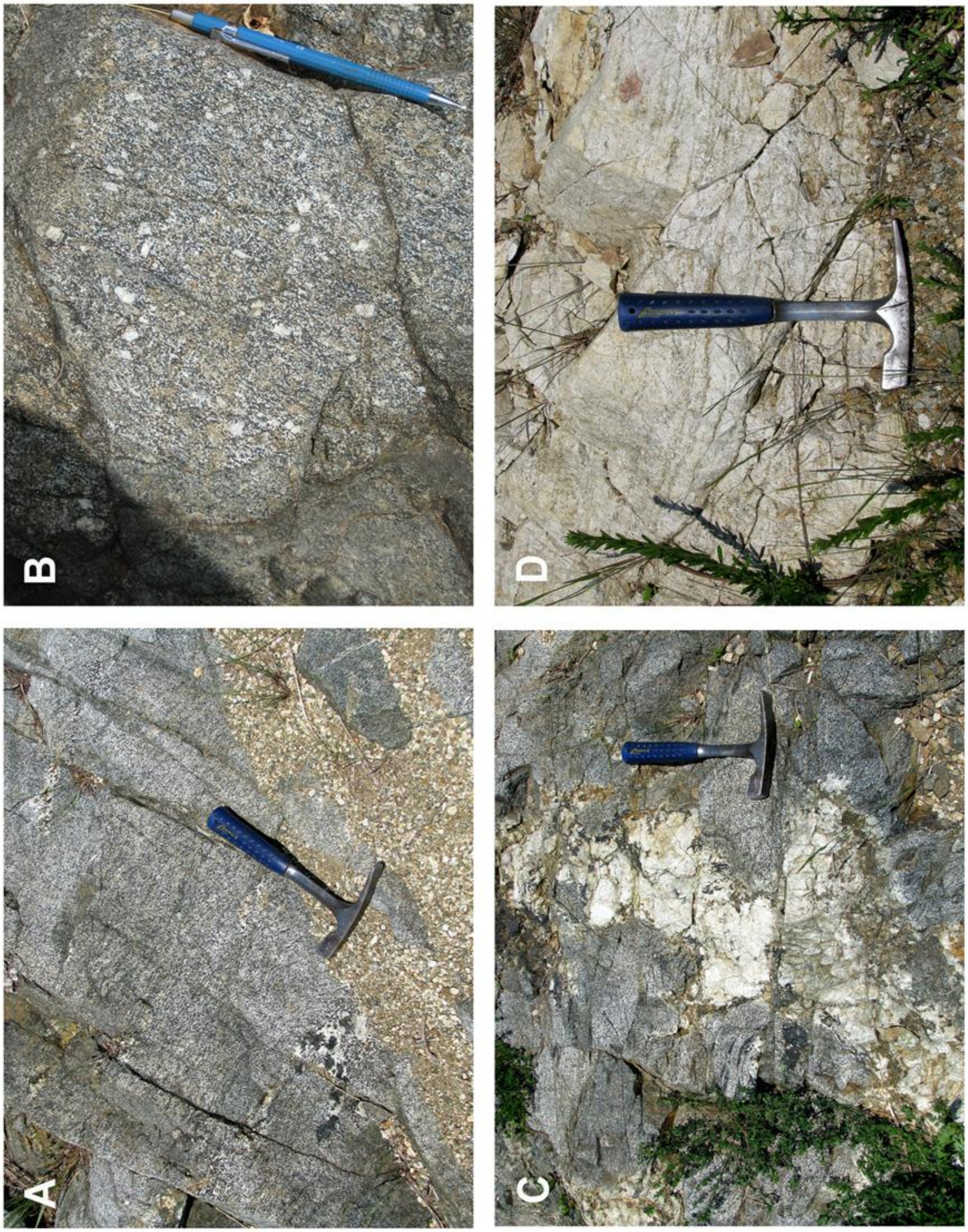


Figure 9. Proterozoic lithologies exposed along the west side of Round Valley Reservoir. (A) Medium-grained, foliated amphibolite. Note fold defined by thin layer of partial melt left of hammer head. (B) Amphibolite with subhedral phenocrysts of plagioclase. (C) Unfoliated pegmatite composed of hornblende, plagioclase and magnetite discordantly intruding amphibolite. (D) White weathering, medium-grained quartz-plagioclase gneiss in conformable contact with amphibolite.

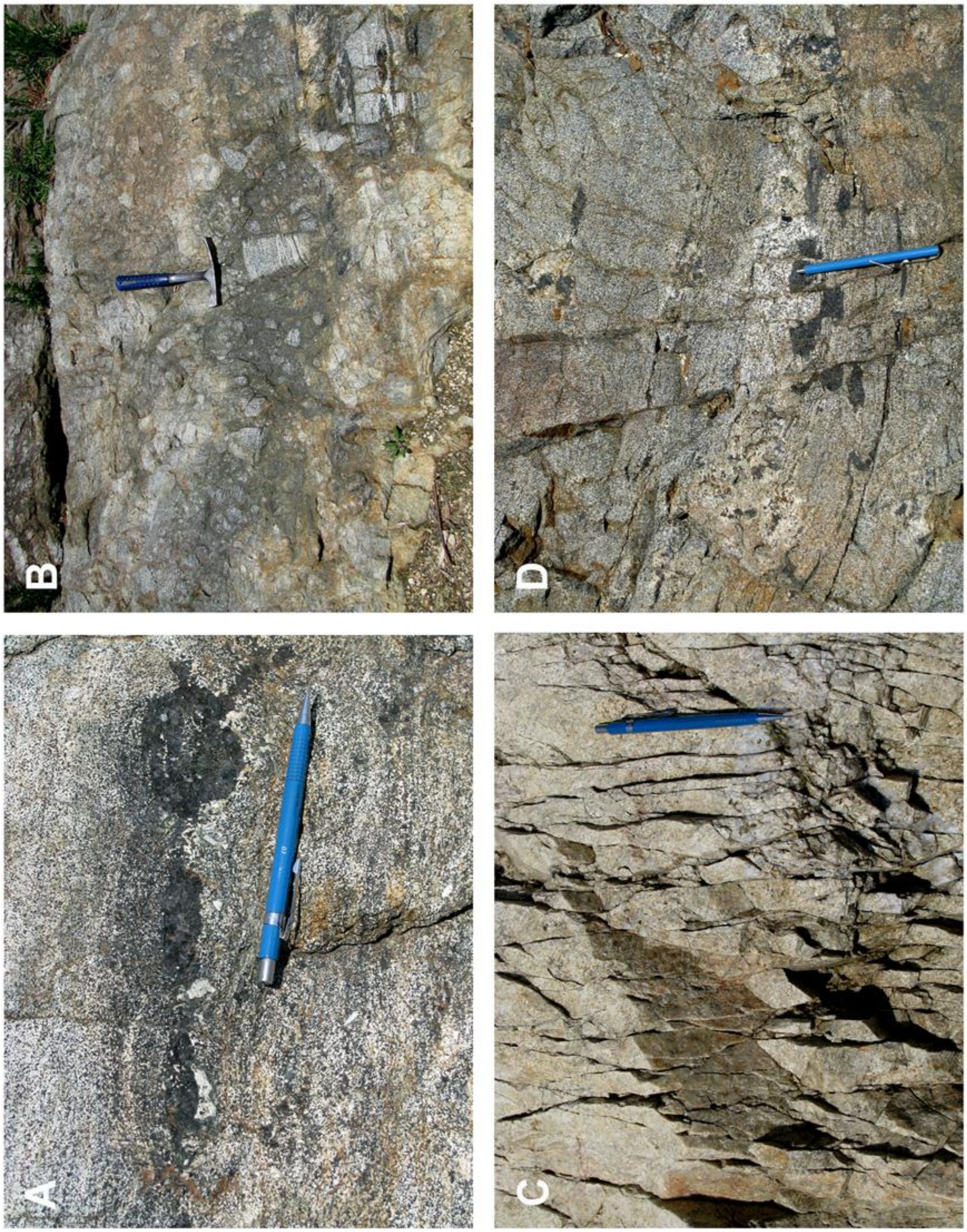


Figure 10. Some representative deformation features in Proterozoic outcrops at this stop. (A) Boudins formed by extension of competent mafic layer in less competent groundmass of foliated amphibolite. (B) Breccia zone striking about N.55^{o}E. in amphibolite. (C) Fracture cleavage oriented N.55^{o}E. in quartz-plagioclase gneiss. (D) Offset of magnetite lenses and layers along steeply-dipping fractures striking N.35^{o}W. to N.25^{o}E.

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Volkert, R.A., Aleinikoff, J.N., and Fanning, C.M., 2010, Tectonic, magmatic and metamorphic history of the New Jersey Highlands: New insights from SHRIMP U-Pb geochronology, *in* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M., eds., From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region: Geological Society of America Memoir 206, p. 307-346.

Volkert, R.A., Zartman, R.E., and Moore, P.B., 2005, U-Pb zircon geochronology of Mesoproterozoic postorogenic rocks and implications for post-Ottawan magmatism and metallogenesis, New Jersey Highlands and contiguous areas, USA: Precambrian Research, v. 139, p. 1-19.

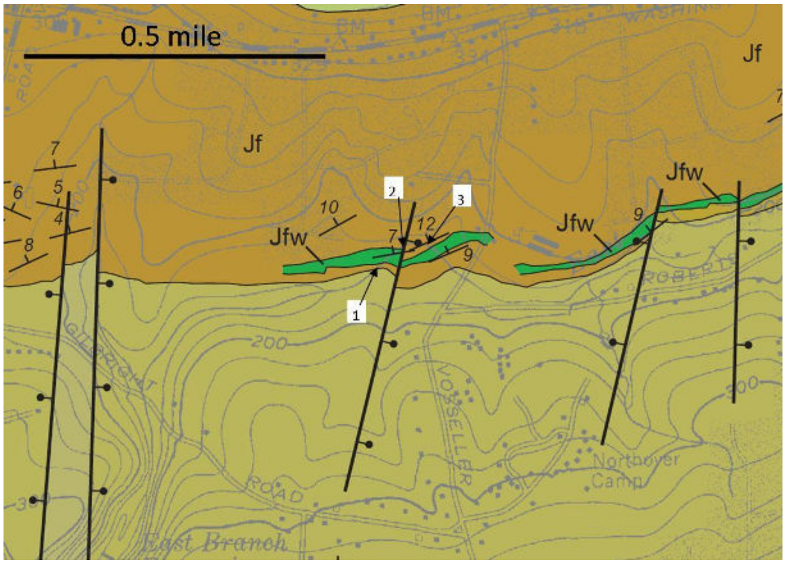
Depart Round Valley State Park for Stop 5 (20.1 mi - about 27 mins)
1. Head north on Co Rd 629/Stanton Lebanon Rd toward Pine Tree Dr 0.4 mi
2. Continue onto Lebanon - Stanton Rd 0.4 mi
3. Continue onto Round Valley Access Rd 0.7 mi
4. Turn right onto US-22 E 17.1 mi
5. Take the Vosseller Ave exit toward Bound Brook 397 ft
6. Turn left onto Vosseller Ave 1.3 mi
Arrive at Eastfields Park

Stop 5 Eastfields Park

Lat 40°35'35.63"N Lon 74°32'43.03"W

Jane Alexander, College of Staten Island/CUNY
Emma Rainforth, Ramapo College of New Jersey

The Blue Brook provides an excellent outcrop of the contact between the Orange Mountain Basalt and the overlying Feltville Formation, including a condensed section of the Washington Valley Member (Olsen, 1980).



Jf	Feltville Formation
Jfw	Washington Valley Member
Jo	Orange Mountain Basalt

Figure 11: Bedrock Geology of Eastfields Park (From: OFM 89, Bedrock Geologic Map of the Bound Brook Quadrangle, Somerset and Middlesex Counties, New Jersey, Volkert, Richard A. and Monteverde, Donald H., 2011, scale 1 to 24,000)

On entering the park, walk north to the stream, and then follow it to the west for about 800 ft to the top of the Orange Mountain Basalt (location 1 in Figure 11). The surface of the basalt is vesicular in places, and the tops of columnar structures are visible in the stream bed. The basalt is overlain by about 10 cm of conglomerate containing basalt clasts at the base of the Feltville Formation.

Figure 12: Outcrop of Feltville Formation along the Blue Brook.

The Feltville Formation immediately above the basalt has limited outcrop in the stream bed, but further east along the stream it is well exposed in a low cliff (locations 2 and 3 in Figure 11 and Figure 12). It consists of interbedded reddish-brown mudstone, siltstone and fine sandstone. The Washington Member bed comprises carbonaceous limestone and calcareous siltstone and shale, along with the red beds. At this location, soft sediment deformation prior to lithification has resulted in blocks of the limestone floating within the mudstone (Olsen, 2010).

At location 2 (Figure 11) there is a break in the cliff resulting in a small gully that marks the location of the fault shown on the map. There is a series of similar faults in this area, all north trending and of Mesozoic age (Volkert and Monteverde, 2011). There is significant fracturing in the sedimentary rocks on either side of the gully, with joints dipping at 85° with a strike of N. 20° E, similar to the average trend for the area (Volkert and Monteverde, 2011). A second set of joints is present in some places with a dip of 63° and the same strike.

References Cited

Olsen, P. E., 1980, The Latest Triassic and Early Jurassic formations of the Newark Basin (Eastern North American Newark Supergroup): Stratigraphy, structure and correlation: New Jersey Academy of Science Bulletin, v. 25, no. 2, p. 25-51.

Olsen, P. E., 2010, Fossil Great Lakes of the Newark Supergroup – 30 years later. *in* Benimoff A. I. (ed.) New York State Geological Association 82nd Meeting Field Trip Guide Book. P. 101-162.

Volkert R. A. and Monteverde D. H., 2011, Bedrock Geologic Map of the Bound Brook Quadrangle, Somerset and Middlesex Counties, New Jersey. New Jersey Geological Survey Open-file map OFM 89.

Depart Eastfields Park for return to Environmental Education Center at Lord Stirling Park (11.5 mi - about 24 mins)

1. Head northeast on Vosseller Ave toward Perrine Rd 0.5 mi
2. Turn right onto Washington Valley Rd 2.8 mi
3. Continue onto Mountain Blvd 0.2 mi
4. Turn left onto Mt Bethel Rd 1.9 mi
5. Continue onto King George Rd 2.0 mi
6. Slight left to stay on King George Rd 164 ft
7. Continue straight onto County Route 512/Valley Rd 0.5 mi
8. Continue onto Stonehouse Rd 1.3 mi
9. Turn right onto S Finley Ave 0.2 mi
10. Take the 1st right onto Cross Rd 0.6 mi
11. Turn left onto Somerset 657/S Maple Ave 0.5 mi
12. Take the 2nd right onto Lord Stirling Rd 1.0 mi

Arrive at Environmental Education Center at Lord Stirling Park