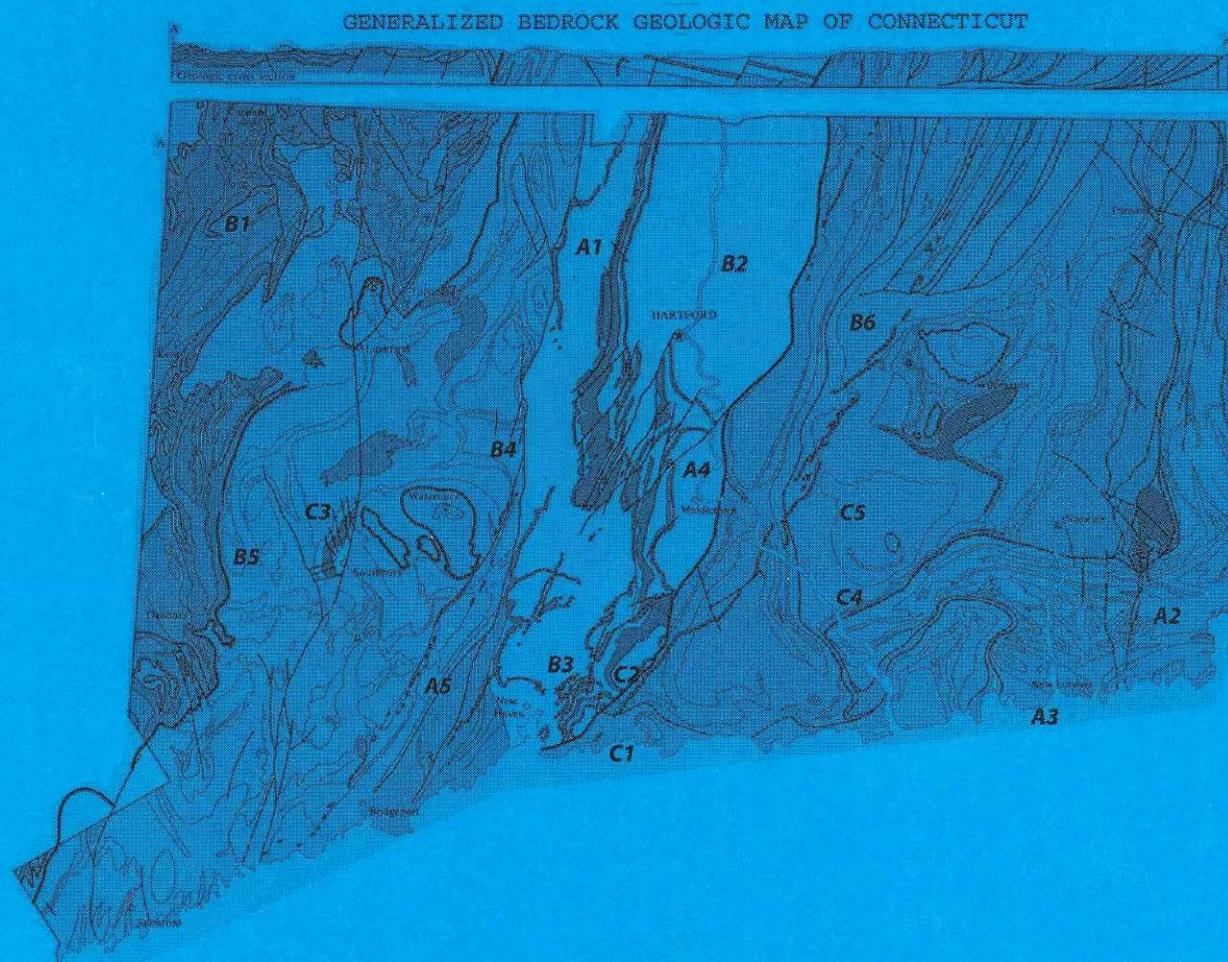


# 97<sup>th</sup> Annual Meeting of the New England Intercollegiate Geological Conference

## GUIDEBOOK FOR FIELD TRIPS IN CONNECTICUT



*Hosted by*  
Department of Geology and Geophysics, Yale University  
New Haven, Connecticut  
September 30, October 1 and 2, 2005

State Geological and Natural History Survey of Connecticut  
Department of Environmental Protection  
Guidebook Number 8

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*NEW ENGLAND INTERCOLLEGIATE*

*GEOLOGICAL CONFERENCE*

**97<sup>th</sup> ANNUAL MEETING**

SEPTEMBER 30, OCTOBER 1 AND 2, 2005

**GUIDEBOOK FOR FIELD TRIPS**

**IN CONNECTICUT**

ORGANIZERS

BRIAN J. SKINNER and ANTHONY R. PHILPOTTS

Department of Geology and Geophysics  
Yale University  
New Haven, Connecticut

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## Meetings of the New England Intercollegiate Geological Conference

- 1901 Westfield River Terrace, MA (Davis)  
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1906 Meriden-East Berlin, CT (Gregory)  
1907 Providence, RI (Brown)  
1908 Long Island, NY (Barrel)  
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1910 Hanover, NH (Goldthwait)  
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1938 Rutland, VT (Bain)  
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1940 Hanover, NH (Goldthwait, Denny, Shaub,  
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1941 Northampton, MA (Balk, Jahns, Lochman,  
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1947 Providence, RI (Quinn)  
1948 Burlington, VT (Doll)  
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1952 Williamstown, MA (Perry, Foote, McFaden,  
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1973 Fredericton, NB (Grenier)  
1974 Orono, ME (Osberg)  
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1977 Quebec City, PQ (Beland, LaSalle)  
1978 Calais, ME (Ludman)  
1979 Troy, NY (Friedman)  
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1981 Kingston, RI (Boothroyd, Hermes)  
1982 Storrs, CT (Joesten, Quarrier)  
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1984 Danvers, MA (Hanson)  
1985 New Haven, CT (Tracy)  
1986 Lewiston, ME (Newberg)  
1987 Montpelier, VT (Westerman)  
1988 Keene, NH (Bothner)  
1989 Farmington, ME (Berry)  
1990 La Gaspesie, PQ (Trzecienski)  
1991 Princeton, ME (Ludman)  
1992 Amherst, MA (Robinson, Brady)  
1993 Boston, MA (National GSA: Cheney, Hepburn)  
1994 Millinocket, ME (Hanson)  
1995 Brunswick, ME (Hussey, Johnston)  
1996 Mt. Washington, NH (Van Baalen)  
1997 Killington-Pico, VT (Grover, Mango)  
1998 Kingston, RI (Murray)  
1999 Burlington, VT (Doolan)  
2000 Orono, ME (Yates, Lux, Kelley)  
2001 Fredericton, NB (Fyffe, Lentz, Pickerill, White)  
2002 Lake George, NY (McLelland, Karabinos)  
2003 Amherst/Northampton, MA (Brady, Cheney)  
2004 Boston, MA/Saco Bay, ME (Hanson)  
2005 New Haven, Connecticut (Skinner and Philpotts)

**TO THE TWO JOHNS:**

**JOHN RODGERS (1914-2004)**

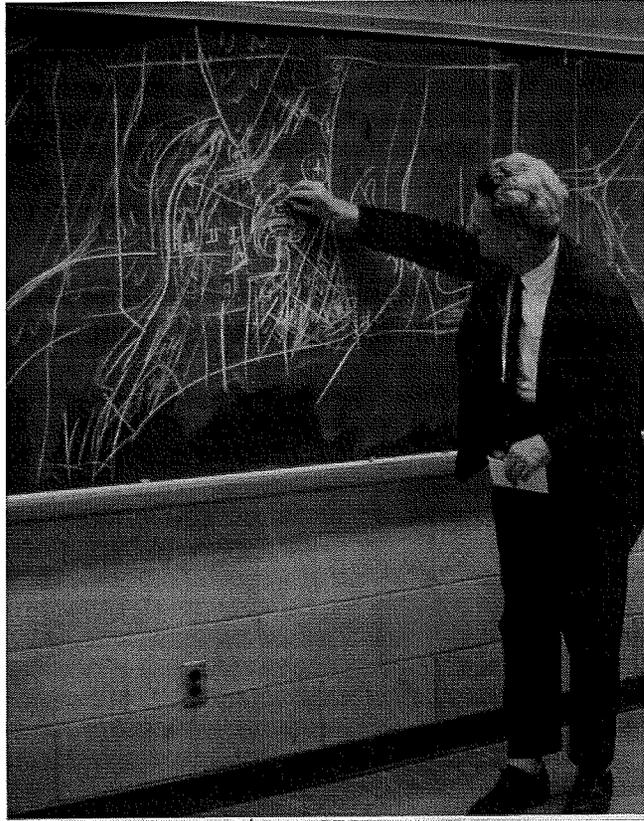
**JOHN H. OSTROM (1928-2005)**

This volume honors the memory of two distinguished geologists, both recently departed, who spent most of their working lives in Connecticut and left major marks on our science.

John Rodgers served on the Yale faculty from 1946 to 1985, a time that spanned the great plate tectonic revolution. Known affectionately as “The Appalachian Mountain Man” for his farsighted studies on the structure and evolution of the Appalachians, John’s work served as a foundation for understanding the evolution of ranges as a consequence of plate motions. Tireless in his pursuit of field evidence, John insisted that “he collected mountains” and his travels took him to the far reaches of the globe. Foreign travel nurtured one of his many talents, language skills. In 1985 John completed a work that he found both challenging and satisfying, the preparation and publication of the Bedrock Geologic Map of Connecticut. That map will be widely used during the field trips planned for the 97th NEIGC.

John Ostrom joined the Yale faculty in 1961 and retired in 1994. He filled many roles on campus. For many years he taught a popular undergraduate course on “The History of Life;” he was curator of Vertebrate Paleontology in Yale’s Peabody Museum, and he was Commissioner of the Connecticut State Geological and Natural History Survey. John is widely known for his work on Hadrosaurs and Ceratopsids (“horned” dinosaurs), and for his discovery of *Deinonychus* (“the terrible claw”), which served as a model for the velociraptor in Michael Crichton’s *Jurassic Park*—a discovery that drew him into the debate on the warm-blooded nature of dinosaurs. As a result of his work on *Archaeopteryx*, John became a leading exponent of the idea that modern birds are descended from dinosaurs.

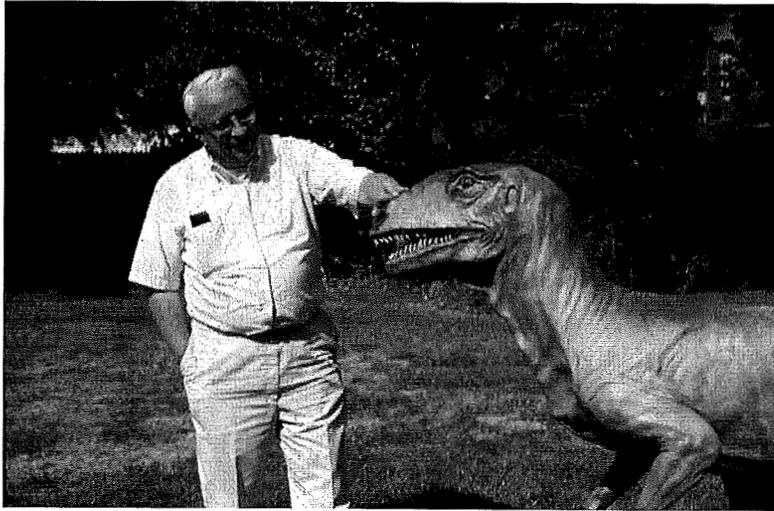
Despite the disparity in their fields of scientific interest, the two Johns served together for many years as Editors of the *American Journal of Science*. Their long careers have left an important imprint on the geological sciences, and we are pleased to honor these contributions with this 97<sup>th</sup> volume of the NEIGC.



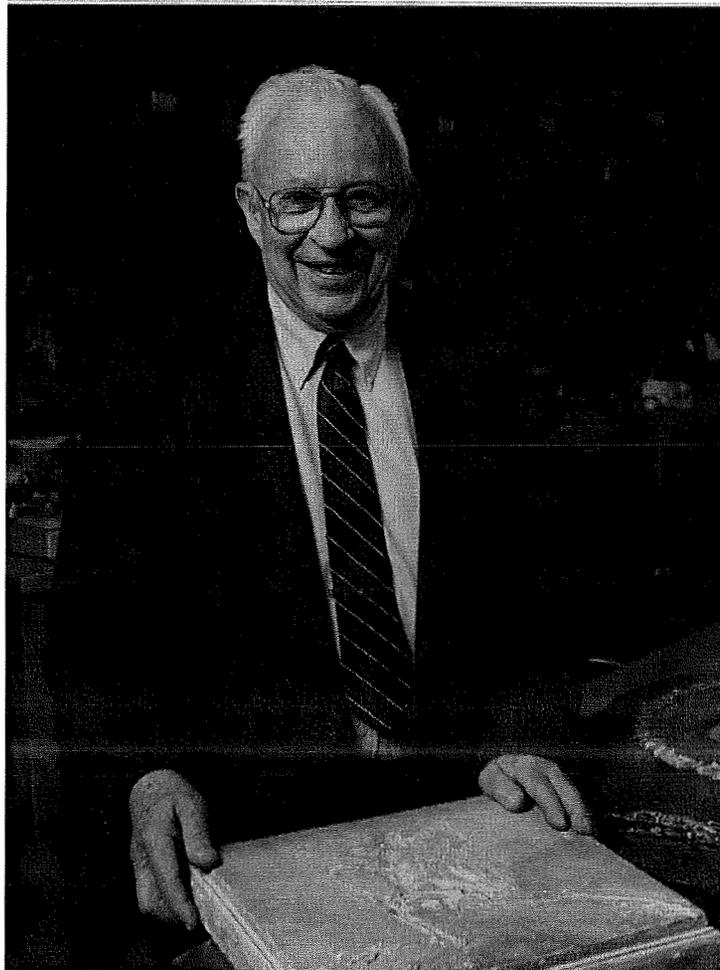
John Rodgers explains the intricacies of Connecticut geology



John Rodgers teaching undergrads how to use a plane table  
Photographs of John Ostrom



John Ostrom and a model of *Deinonychus*



John Ostrom and an engraving of *compsognathus*, a beast in the dinosaur-to-birds chain

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# THE HISTORIC NEW-GATE AND COBALT MINES OF CONNECTICUT

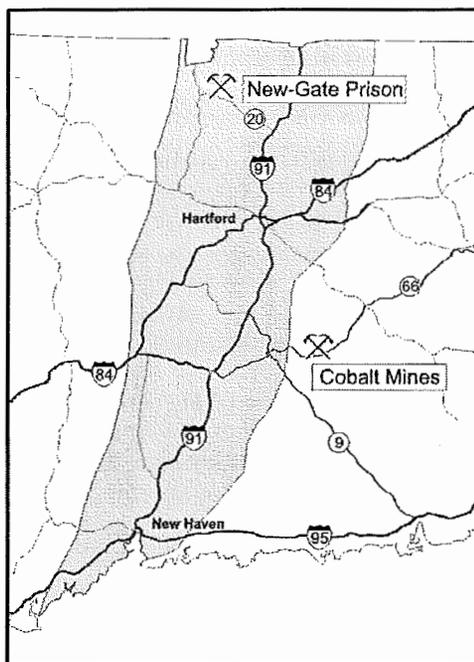
Norman H. Gray, Department of Geology, University of Connecticut, Storrs, CT 06269

## INTRODUCTION

During the 1700 and 1800s Connecticut was the scene of several small entrepreneurial mining ventures. Most of these attempts were short-lived and it is difficult today from the limited tailing piles and collapsed or flooded workings to understand what these early miners were after. The underground workings of two of the more important deposits are still accessible and the nature of the mineralization that attracted the miner's interest can be seen in place (Fig. 1).

The New-gate mines of Simsbury are hosted by grey colored sandstones in the Hartford Basin Mesozoic redbeds. The main ore (grading up to 13% Cu) was a massive chalcocite replacement of the ankerite cemented zones. The mineralization at New-Gate was discovered in the early 1700s, and worked at intervals up to 1900. For a brief period in the late 1700s the underground workings served as Connecticut's State Prison!

The Cobalt mines of East Hampton were first prospected in the 1790s, opened as a cobalt source in the early 1800s and then for nickel in the 1860s. Mineralization occurs in Ordovician metasediments just below the Silurian Clough Quartzite. Local legend has it that the first Governor of Connecticut found gold in the area in the mid 1600s. Later miners make no mention of gold even though stringers of native gold are visible in arsenopyrite veins close to their main workings in the Clough. The early miners, who obviously were very resourceful, traced a single folded thin seam of Fe-Co-Ni arsenide bearing garnet amphibolite along strike for over 1200 ft.



**Figure 1.** Locations of the New-Gate and Cobalt mines in Connecticut. The Mesozoic Hartford Basin is shaded

## NEW-GATE PRISON COPPER MINE

### Introduction

Of the many small redbed type copper occurrences in the Mesozoic Hartford Basin the New-Gate deposit is the largest. Although never profitable, it was actively worked over a period of several decades. The geology in the vicinity of the mine is outlined in detail by Schnabel and Eric (1964), Perrin (1976), and Gray (1982, 1989) on which figure 2 and much of the following is based.

### History

Copper was discovered in the New-Gate prison area in 1705. Two years later the first charter mining company in North America was organized to work the deposits. Although numerous prospect pits, declines, shafts and adits testify to the scale of the effort the venture never proved especially profitable. The mines absorbed much more capital than their output ever provided in return. By 1741, more than \$200,000 had been spent to recover little more than 100 tons of dressed chalcocite ore which at best averaged 13% Cu (Richardson, 1928).

Water was one of the most difficult problems faced by the early miners. Between 1721 and 1730 two neighboring mine owners combined resources and drove a 100 meter tunnel to drain the lower workings of one of the principal deposits. Because it was drained and secure the mine served as the State Prison from 1773 to 1827.

In 1831 New-Gate and number of the other workings in the area were reopened by the Phoenix Mining Company. Forty-two experienced miners were induced to emigrate from Europe to work the deposits. The old workings were pumped, mapped and new level at the lowest point driven 60 meters to the south. Unless purely exploratory, the purpose of these new tunnels is not clear as they were driven through entirely unmineralized strata.

Perhaps that fact alone explains the failure of the Phoenix Company three years later, in 1834. A drainage tunnel designed to intersect a shaft sunk to what must have been extensive underground workings at the North Hill Mines had been driven 65 meters before work was abandoned.

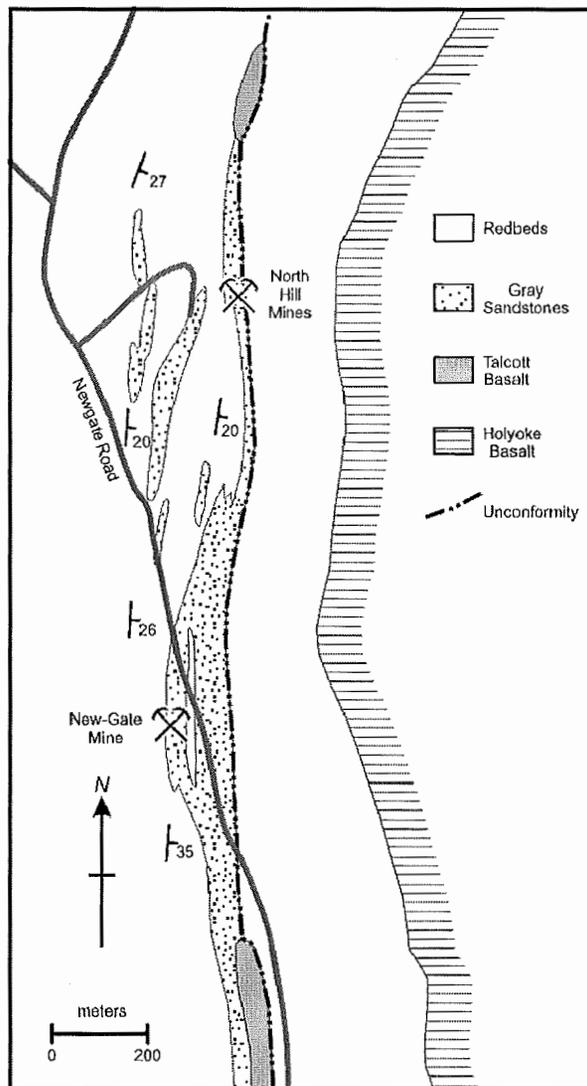
Two other attempts in 1855 and 1907 to reopen the mines also ended in failure. The focus of these ventures seem to have been speculation rather than actual mining but it is possible that the lower workings at New-Gate were dewatered at these times.

### Geology and Mineralization

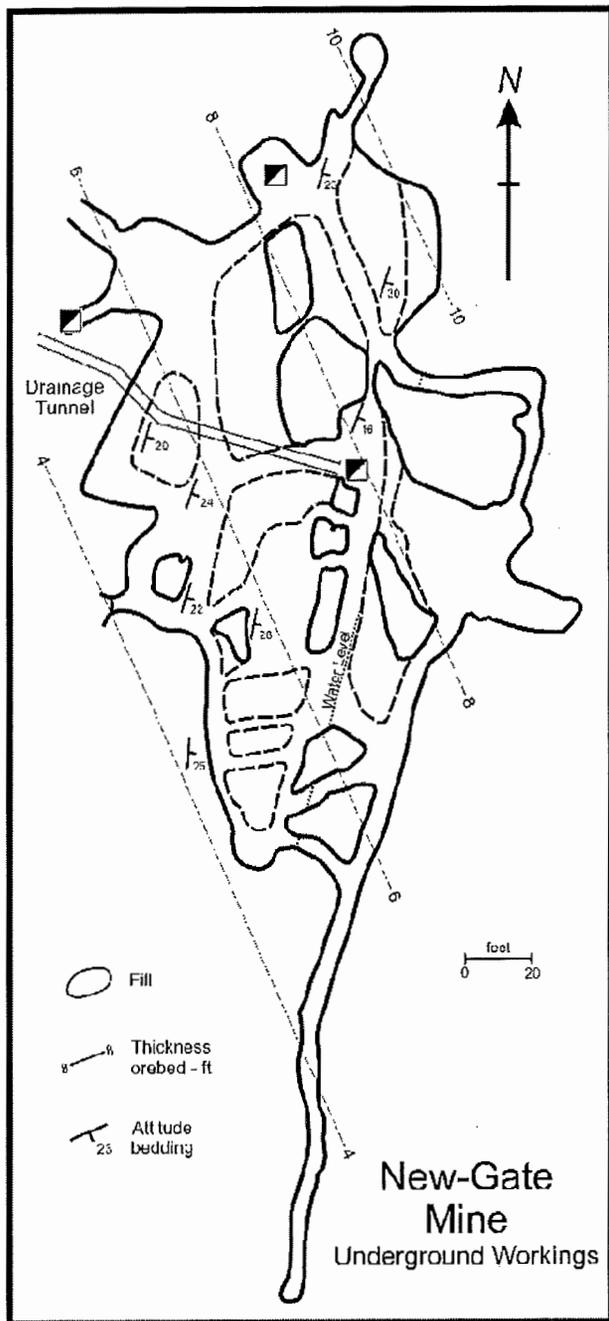
Copper occurrences in the vicinity of the New-Gate mine are associated with a major erosional unconformity which truncates the Talcott basalt (Fig. 2). Disseminated copper mineralization can be traced for over 2 km along strike in a series of small prospect pits just below the unconformity. Bornite and chalcopyrite, in places marginally replaced by chalcocite) fill interstitial pores between detrital grains in the grey siltstones. The mineralization is statabound and is present only in gray to black colored sediments. The interbedded red-colored sandstones and shales are everywhere barren. In the coarser grained grey sandstones the sulfides form millimeter sized nodular concretions. The detrital feldspars in these concretions are replaced by bornite and chalcopyrite while quartz grains are unaffected and even show euhedral diagenetic overgrowths. A small amount of uraninite is associated the nodular copper sulfides. Grey sandstones less than 20m below the unconformity that truncates the Talcott basalt are the most intensely mineralized. Copper averages less than 1% in most of these rocks but in the area around the North Hill mines the average grade of the disseminated mineralization ranges from 2% to as much as 5%

The main copper ore at the New-Gate prison mine is stratigraphically lower and both texturally and mineralogically distinct from the disseminated mineralization. It is also significantly higher grade. Copper averages 2.5 to 10% and silver up to 10 ounces per ton. Chalcocite rather than bornite or chalcopyrite is the principal ore mineral. The highest grade chalcocite occurs in curious mottled zones surrounding centimeter sized poorly cemented porous patches within structureless, bioturbated sandstones. The mottled ore bed is part of a well defined sedimentary fill of a large north-south channel eroded into a thick sequence of red mudstones. Locally the base of the channel is defined by a thin (0-10cm) intraformational breccias containing pebble sized clasts of an iron-rich dolostone. A few thin (cm) gray to black shales are the only clear evidence of stratification within the ore bed. In the area of the highest grade mineralization the ore bed is overlain by festoon bedded medium grained laminated grey sandstones. However in the now flooded southern workings a meter thick black mudstone overlies the stratigraphic equivalent of the ore bed which although unmottled contains the disseminated and millimeter nodular concretions of bornite and chalcopyrite, typical of the regional disseminated Cu-U mineralization.

The ore bed which as befits a channel fill is discordant with the underlying redbeds, thickens to the Northeast, and wedges out to the southwest (Fig. 3). Where less than a couple of meters thick at the edge of the channel fill the ore bed was cemented by Ferroan dolomite prior to compaction and mineralization. The carbonate cemented zones are cut by numerous calcite-chalcocite veins which do not extend into the surrounding non-carbonated cemented



**Figure 2.** Generalized geology of the New-Gate area showing the locations of the New-Gate mine (1001370E, 911260N) and the North Hill workings (1001630E, 913970N). The Higley mine (1000470E, 902640N) is situated in the Talcott Basalt 2 miles south of New-Gate. (Coordinates are SPC83 in feet).

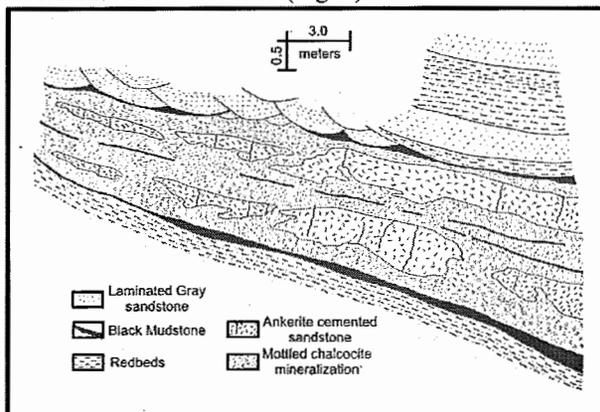


**Figure 3.** Underground workings at the New-Gate Prison mine

a coarse breccia consisting of angular clasts of chloritized basalt and bleached clay-rich hydrothermally altered basalt in a red sandstone matrix. Chalcopyrite-filled fractures crisscross the bleached basalt fragments but nowhere extend into the sandstone matrix. The bleached and mineralized basalt clasts imply extensive hydrothermal activity after the Talcott event during the erosional interval associated with the Talcott unconformity. Hot springs sited along the surface traces of the mapped northwest trending normal faults may have source of the copper bearing waters for both Higley and the disseminated mineralization in the New-Gate area.

The mottled high-grade chalcocite ore at New-Gate was the result of a major change in the chemistry groundwaters late in the diagenetic history of the sediments. Partial dissolution of ferroan dolomite cemented sandstones produced a favorable local environment for the precipitation as chalcocite of copper leached from the surrounding disseminated mineralization. The fact that the syn-sedimentary ferroan dolomite cemented zones may

sandstones. The high grade mottled ore irregularly crosscut and replaces the unmineralized ferroan dolomite cemented zones (Fig. 4).



**Figure 4.** Schematic cross-section of the ore bed at New-Gate

**Origin**

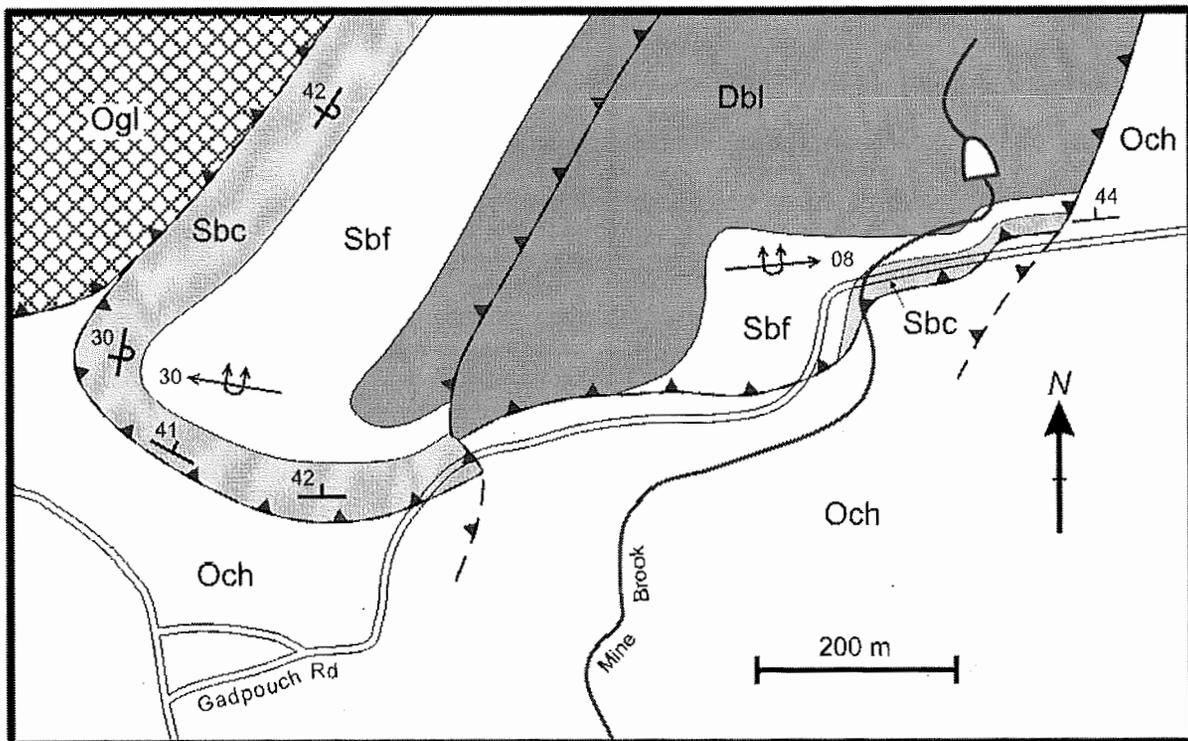
The disseminated copper sulfides either precipitated in pore spaces between detrital grains or formed nodular concretions early in the diagenetic history of the gray sandstones. The close relationship of the disseminated mineralization and the Talcott unconformity suggests saline copper-bearing waters percolated down into the sediments during a prolonged erosional interval. Considering the extent of the mineralized area, the amount of copper precipitated in near surface reducing environments was substantial. The ultimate source of the metals is somewhat of a mystery but one clue may be found at a small, now inaccessible Higley deposit 3 km to the south. In spite of the fact that they both lie at essentially the same stratigraphic level the copper mineralization at the Higley mine is quite unlike that of the New-Gate area. Rather than filling pore spaces in sedimentary rocks, the Higley mineralization occupies fractures and vesicles in altered Talcott Basalt. Vesicles are lined by copper sulfides, calcite and zeolites. The amygdaloidal basalt is extensively replaced by ferroan dolomite, especially where intensely mineralized. Hematite stained fractures and bornite-chalcopyrite veins up to 2 cm wide extend well below the porous vesicular zone. The flow is overlain unconformably by

have shielded some of the adjacent sandstones from the full effects of compaction may also helped preserved some of their original porosity which would have further controlled access by solutions responsible for the chalcocite mottling. The cause of the dramatic change in the water chemistry is not obvious. Perhaps the cooling of the overlying Holyoke generated a thermal regime which flushed new waters through the New-Gate deposit.

## GREAT HILL - COBALT MINES

### Introduction

The geology of the Great Hill-Cobalt mine area is shown in Figure 5 (after Chomiak, 1989). The most obvious feature is the nose of Great Hill or Bolton Syncline, which here because of later deformation, is represented by an antiform gently plunging off to the west. Mineralization is restricted to the north dipping east-west trending limb of this fold which is in fault contact with the underlying Collins Hill schists and gneisses. Cobalt-nickel arsenides are found along a single centimeter thick garnet gneiss bed just below the and gold bearing arsenopyrite-quartz veins just above the Clough-Collins Hill contact. Details of the geology, petrography and ore mineralogy are provided by Chomiak (1989). The following is just a brief synopsis of her work.



**Figure 5.** Generalized geology of the Great Hill – Cobalt Mines area after Chomiak (1989). All rocks are Paleozoic. Ogl- Glastonbury Gneiss, Och – Collins Hill schists and gneisses, Sbc – Clough Quartzite, Sbf – Fitch calc-silicates, Dbl – Littleton schists.

### History

Mining and the Cobalt area of East Hampton have been associated in both fact and legend since John Winthrop the Younger, the first Governor of Connecticut, was reputed to have discovered gold in 1641 at the base of what is now called Great Hill. It is said that he used to resort there to mine gold which he fashioned into gold rings to impress prospective investors. Until the University of Connecticut Summer field school identified native gold at the same site there was no record of anyone having mined or even seen gold since Winthrop's day. Nevertheless, the legend of "the Governor's Ring" (now a nickname for Great Hill) has kept alive by generations of town people.

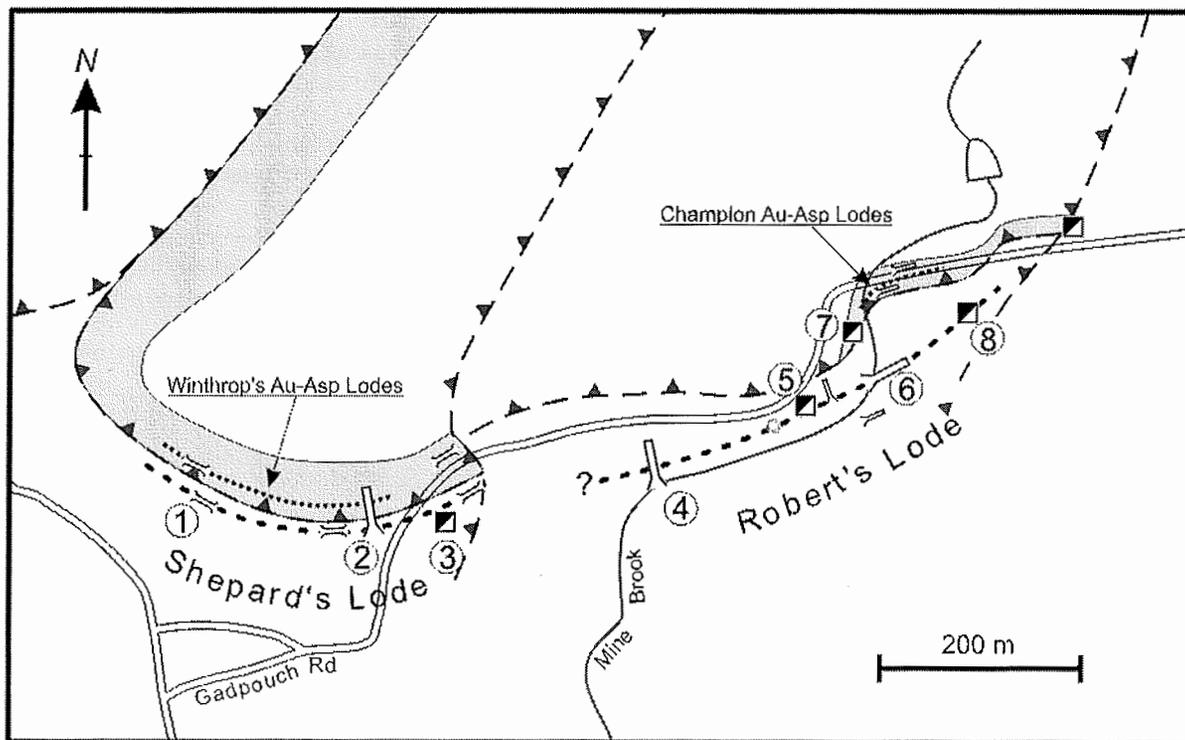
Presumably attracted to the area by reports of gold, a German prospector John Stephauney in 1762 found a small deposit cobalt ore. In 1770 he organized a serious mining effort, opened a major adit and actually shipped some ore. The enterprise was abandoned within a few years and there is no record whether any profit was realized.

The mines then remained dormant until 1818 when Seth Hunt sunk a quarter mile long trench along the main vein and shipped some supposed cobalt ore to England. He abandoned the enterprise after assays indicated that his ore was mostly nickel, which had no economic value at the time.

Soon afterwards, Charles Upton Shepard examined the workings and included a description of the ore mineralogy in his 1837 Report on the Geological Survey of Connecticut. An assay of the dressed ore given by Shepard indicates 28% Co + Ni. Shepard was obviously impressed by this result as he purchased a 99 year lease of Hunts properties. In 1844 Shepard himself opened an inclined adit on the vein and although he abandoned all work within a year he did retain the mineral rights.

In the mid 1800 investor interest shifted to the nickel content of the ore. At the time the US Mint was looking for potential sources of nickel for coinage. In 1850 Brown sunk a 38 foot shaft just east of Shepard's holdings with plans to drive a tunnel along the vein to a new adit in a ravine 700 feet to the east. Funds quickly ran out and the workings were acquired by "The Chatham Cobalt Mining Company" organized by Eugene Francfort, a Middletown Doctor and self-styled mining engineer. Between 1853 and 1859 the Chatham Company sank two major shafts one 120 feet deep, drove at least 100 meters of adits and drifts along the main ore zone and built a stamping mill and assay lab on the site. Although some dressed ore running about 10% nickel and 10% cobalt, was shipped to the US Mint, investor interest lapsed and work ceased by 1859.

A detailed review, and a complete bibliography of the of the mining history of the area can be found in Appendix A of Chomiak's 1989 University of Connecticut's Thesis, "An Integrated Study of the Structure and Mineralization at Great Hill, Cobalt, Connecticut". The locations and historic names of the main workings are shown in Figure 6.



**Figure 6.** Locations of the main Shepard's and Robert's lodes, the Gold-Arsenopyrite veins, and the principal mine workings in the Great Hill - Cobalt Mines area. 1 - Stephauney-Hunt Workings (1052560E, 770400E), 2 - Shepard's Decline (1052860E, 770360N), 3 - Browns Shaft (1053320E, 770460N), 4 - Brown's Ravine Adit (1054020E, 770600N), 5 - Bucks Shaft (1054510E, 771090N), 6 - Stream Adit (1054770E, 771050N), 7 - Engine Shaft (1054760E, 771150N), 8 - Ventilation Shafts (1054900E, 771230N). (Coordinates are SPC83 in feet).

**Cobalt-Nickel Mineralization**

Co-Ni mineralization extends along narrow zone at least 800 meters in length from the base of Great Hill in an easterly direction. The early miners referred to the western portion of the mineralization as “Shepard’s lode, and the eastern segment as “Robert’s lode.

**Shepard’s Lode.** Shepard (1837) described the ore zone originally worked by John Stephauney and Seth Hunt as:

“..[it] forms a thin bed or seam in the mica-slate whose dip and direction conform precisely to the general stratification of the mountain. The stratum embracing the ore [smaltite] appears to run directly across the mountain and has been more or less excavated for nearly the whole extent. Its thickness is with difficulty inferred from the present condition of the mine, but appears to be about one foot. It consists of an aggregate of quartz, garnet and hornblende. The greatest depth to which the ore has been followed is forty-five feet. It is accompanied by copper-nickel [niccolite], blende [sphalerite], galena and traces of yellow copper pyrites [chalcopyrite]”.

Shepard initially identified the Co-Ni bearing arsenide as the cubic di-arsenide, smaltite but after obtaining and studying additional material from his own mine he pronounced it to be a new orthorhombic tri-arsenide for which he proposed the name “Chathamite” after the town in which it was discovered (Chatham – now East Hampton) In the mid 1850s Genth (in Goodrich, 1854) questioned Shepard’s identification and suggested that Chathamite was simply an iron rich variety of the cubic arsenide chloanthite [a misconception that perpetuated up to, and including, the 7<sup>th</sup> edition of Dana’s Manual of Mineralogy]. As it turns out, Shepard’s chathamite is indeed orthorhombic, but today would be classified as a nickel-cobalt rich loellingite.

All of workings on Shepard’s lode are now inaccessible. Loose tailings and scattered mineralized samples near what is probably the site of Shepard’s mine suggest that the lode is essentially a banded amphibolite. Hornblende, manganous garnet, sphene and plagioclase make up the bulk of the rock. 5 to 10% sulphides and arsenides are intergrown with the silicates. The mineral textures are entirely metamorphic. Pyrrhotite and loellingite constitute the bulk of the ore minerals although chalcopyrite, sphalerite, galena, niccolite, and gersdorffite are reasonably common.

**Robert’s Lode.** The Chatham Company followed the eastern mineralized zone in several hundred meters of underground workings, some of which are still accessible although in most of these very little of the lode is either mm thin or has been completely excavated. Tailing, especially in the area of Bucks shaft give some indication of the nature of the mineralization. Francfort (1863) provides the best description of the ore:

“The veinstone is a very fine gneiss, containing a great deal of garnet and black mica, and a sort of quartz which miners call sugary spar. Through this veinstone the smaltine [smaltite] is well disseminated, and it is of so fine a grain that it requires the operation of vaning to show it plainly to the eye”

“We have traced this vein all over the property by pits sunk on the course of the vein, for over 1/4 of a mile, and have driven a long level on the course of it, ... It carries ore well disseminated in every part of it, and in some parts we have met with regular bunches of ore”

Francfort’s mineral collection which is now housed at Wesleyan University contains several ore samples from Cobalt. Quite frankly the specimens are not terribly impressive. We have found similar samples in the rubble around Buck’s Shaft. The most highly mineralized tailings are gneisses containing biotite, manganous garnet, feldspar, quartz and staurolite. Loellingite is the only ore mineral in the samples we’ve examined and is present in small amounts ranging from 1 to 5%. Hornblende is absent. The miners apparently used the lack of hornblende as the principal criteria differentiating the Robert’s from the Shepard’s lode.

The Robert’s lode apparently pinched and swelled along strike, presumably due to folding and boudinage. Where exposed today it is generally less than a few mm thick which raises the question as to how the early miners managed to recognize and trace it over several hundred meters. Presumably they had a good eye for the bright silvery-white loellingite and the characteristic hyacinth-red Mn rich garnets it contains. It is also possible that they were also guided by the easily recognized colors of cobalt and nickel bloom on weathered rock surfaces. The

Francfort mineral collection contains some excellent samples of erythrite from Bucks Shaft although today, after generations of mineral collectors picking over the site, the mineral is difficult to find.

The Chatham Company miners named and excavated a number of other lodes (the Champion and Barrett and Underlay lodes as well as the Brown and flucken veins). None of these contained significant Co or Ni mineralization. All are associated with late quartz veins and the only sulfides generally present are pyrite, pyrrhotite and arsenopyrite.

The characteristics of the main Cobalt lodes suggest a syngenetic seafloor exhalative origin. Both Shepard's and Robert's lodes are conformable, laterally extensive, thin beds characterized by a regionally unique association of Mn rich garnets and Co-Ni arsenides. Shepard's lode differs from Robert's lode only in the presence of hornblende and absence of staurolite. They may be two separate exhalative events or merely different facies of a unique sedimentary event. The rarity of Co-Ni mineralization elsewhere in the Collins Hill Schists may argue for the later interpretation.

### **Arsenopyrite-Gold Mineralization**

The Gold found in the Cobalt area is separate and distinct from the Co-Ni mineralization although both lie stratigraphically within a few tens of meters of each other. Gold occurs in decimeter wide quartz-arsenopyrite veins in the Silurian Clough Quartzite just above its contact with the Ordovician Collins Hill formation. Although the Clough appears to be concordant with the layering in the Robert's and Shepard's lodes the evidence suggests that the contact is tectonic and not an unconformity. The arsenopyrite is these quartz veins occurs as centimeter sized massive concentrations. Pyrrhotite, locally altered to pyrite, is the only other sulfide present in abundance. Native gold, generally as micron sized grains, is found, along with pyrite and chalcopyrite, in a network of thin fractures and veins cutting the arsenopyrite. Although much of the gold is very fine grained and is difficult to see, even with a strong hand lens, grains up to a mm are present and are quite noticeable on bright sunny days.

Although there is no direct genetic link between the cobalt and gold mineralization at Great Hill the quartz-arsenopyrite veins in the Clough may be related to the metamorphic expulsion of arsenic from a metalliferous facies of the Collins Hill formation. The deposition of the gold itself is apparently controlled by the local chemical environment imposed by the presence of the concentrations of arsenopyrite. The actual mobilization and transport of gold may be more directly related to a hydrothermal event accompanying the emplacement of the Permian pegmatites of the Middletown district.

The biggest mystery of the Cobalt area why all the miners, except Winthrop, failed to recognize and exploit the high-grade gold ores on their property. Some of the arsenopyrite-quartz veins, which contain no cobalt or nickel but run up to 6 oz of gold per ton were clearly explored by means of adits, shafts and trenches but there is no written or even verbal record of any gold having been extracted. Were the miners hiding the presence of gold from the owners, and all the inquisitive locals, or were they just totally incompetent?

### **ROAD LOG**

Assemble in the Newgate Prison Museum Parking lot. To reach the site from I-91 take exit 40 and follow Rte. 20W 6.3 miles to a set of traffic lights, turn right onto Newgate Road, the prison is 1.2 miles north Rte 20. The route is also well marked by signs for "Old Newgate Prison". The trip will start at 1:00 pm. If you arrive early the view and picnic tables just outside the walls of the prison are an ideal spot for a brown bag lunch.

**STOP 1. New-Gate Prison mine (70 MINUTES)** About one-third of the underground workings of the mine are easily accessible and moderately well lighted. For those who are willing, and have brought flashlights, we will also visit the southern portion of the mine which is not open to the public. Be prepared to crawl as some of the tunnels are only 3 feet high! No hammers please.

#### **Mileage**

- 0.0 miles. Turn right (south) onto Newgate road.
- 1.1 miles. Traffic lights Rte 20. Continue straight. Newgate Road becomes Holcomb Street.
- 2.0 miles. Park on the left side of road in front of the farmers produce stand.

**STOP 1a. Higley Mine. (10 MINUTES)** There is not much to see here but the small hill to the northeast is the site of the Higley mine which was operated on a small scale in the early 1700's. The first coins minted (illegally) in

America were struck from Higley copper between 1729 and 1739. The metal was purportedly of great purity and the coins were sought after by eighteenth century jewelers as an alloy for their gold. Not surprisingly few survived and Higley coppers are today exceedingly rare.

Return to Rte 20.

- 2.9 miles. At traffic lights turn right (east) onto Rte 20E.
- 5.4 miles. Bear right to stay on Rte 20E, follow signs to I-91.
- 9.0 miles. Bear right onto I-91 South (to Hartford).
- 29.2 miles. Take exit 22S (note this is a LEFT Exit) onto Rte 9 South (to Middletown).
- 34.8 miles. At traffic lights Take exit 16 (note this a right lane exit only) keeping to the right. Follow signs for Rte 66 East.
- 35.0 miles. Turn right at traffic lights onto Rte 66 East.
- 35.7 miles. Turn right at lights to follow Rte 66 East.
- 37.9 miles. Continue straight on Rte 66 East at traffic lights.
- 41.0 miles. Traffic lights at junction of Rte 151. Continue straight on Rte 66 East.
- 41.8 miles. Traffic lights at junction of Rte 16. Bear left to follow Rte 66 East.
- 42.2 miles. Left onto Cone.
- 42.7 miles. At a T-junction first turn left onto Abbey and then almost immediately right onto North Cone.
- 43.0 miles. Turn left onto Gadpouch Road.
- 43.1 miles. Keep left, Gadpouch turns into a dirt road.
- 43.5 miles. Park along edge of road.

**STOP 2 The Cobalt Mine workings (90 MINUTES)** We will work our way down Mine Brook, pausing to look at exposures of the Champions Lode, the Engine Shaft, the stream Adit on Robert's lode, the ruins of the stamp mill, Bucks shaft and finally the Brown's Ravine Adit. If it is a sunny day we might find a grain or two of visible gold in the massive scorodite weathered arsenopyrite of Champions lode. Samples of Robert's lode are best obtained by digging in the tailings around Bucks Shaft. The original underground workings can be entered at the stream Adit (be prepared to crawl) and at the Ravine Adit (be prepared to get wet).

Continue on Gadpouch Road.

- 43.8 miles. Park along the edge of the Road.

**STOP 2a Shepard's Lode (10 MINUTES)** From this spot we can see some of the trenches along Shepard's Lode, Shepard's Decline and the location of Brown's Shaft. All the historic mine workings on Shepard and Winthrop's lodes are on private property. Due to the attention the 1985 rediscovery of gold on their property the local landowners are not particularly fond of prospectors, geologists, and mineral collectors. Their concern was heightened early on when they realized that their deeds stipulated that the mineral rights were reserved to the previous owner. Although the mineral rights were eventually released to them, please respect their wishes and avoid trespassing.

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## SELECTED ARCHEOLOGICAL LITHIC SOURCES IN SOUTHEASTERN NEW ENGLAND WITH LAST STOP AT THE PEQUOT MUSEUM

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### INTRODUCTION

The most important lithic materials used by prehistoric groups in southeastern New England included volcanic felsite<sup>1</sup>, soapstone, chert, argillite, hornfels, and vein quartz. We have characterized the petrology of selected archaeological felsite and soapstone quarry sites that can constrain the sources of materials discovered during archaeological excavations.

Soapstone vessels are useful identifiers of Terminal Archaic occupations in southern New England. Talc-based soapstone's relative softness and its excellent heat retention properties have been well documented making it a highly prized commodity sought by the Eastern Woodland Indians for cooking and boiling during the Terminal Archaic (Klein, 1997; Sassaman, 1999; Turnbaugh et al., 1984; Willoughby, 1935). Use of soapstone vessels peaked between 3400 to 2900 radiocarbon years B.P., and fell into disuse by the end of Terminal Archaic Period concurrent with the adoption of ceramic technology, around 2700 radiocarbon years B.P. (Sassaman, 1999). However, soapstone itself continued to be used by Native Americans in smoking pipe, bead, and pendant manufacture through the Woodland and into the early historic period (3000 to 500 radiocarbon years B.P.).

In addition, a variety of fine-grained, felsitic volcanic rocks that occur primarily around the Boston basin and southward into Rhode Island (Fig. 1) were important sources of lithic raw material for prehistoric groups in the area. For at least 8-9,000 years these felsites were quarried and widely distributed across the region and were a major component of the total lithic resource base. Most of the known prehistoric quarry sites in southeastern New England that served as sources of felsitic rocks are within the Lynn/Mattapan, Blue Hill, and Wamsutta volcanic complexes (Massachusetts Historical Commission, 1980; Ritchie and Hermes, 1992). The Newbury volcanics (located north of the Boston basin; Shride, 1976) and the Spencer Hill volcanics of central Rhode Island (Hermes et al., 1994) also contain felsitic rocks that may represent potential lithic source areas. These complexes contain volcanic rocks displaying a wide range of color, textures, and weathering patterns that present a challenge to lithic sourcing studies.

However, recent geological research has established that volcanic rocks from southeastern New England can be correlated with distinct complexes based on geochemical composition and petrography (Hermes and Murray, 1990; Thompson and Hermes, 1990). As a result, this geologic data base can be used to establish sources of felsitic rocks found as archaeological material.

### GEOLOGICAL BACKGROUND

Felsic volcanic rock suites in southeastern New England range in age from Late Proterozoic to Devonian, and are comprised of two major distinct compositional groups; calcalkaline suites and alkaline suites (Hermes and Murray, 1990).

#### **Calcalkaline Suites**

Calcalkaline volcanics are represented mainly by the Lynn-Mattapan suite that occurs northeast and southwest

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<sup>1</sup> felsite is defined herein as a fine-grained, aphanitic volcanic rock.

of Boston (Fig. 1); recent radiometric data demonstrate that these spatially separated rocks are coeval and of Late Proterozoic age (Kaye and Zartman, 1980; Hepburn et al., 1993; Thompson et al., in press). The Lynn-Mattapan includes a variety of volcanic rock types, including rhyolitic to andesitic tuff, ash-flow tuff, and lava flows; pyroclastic rocks dominate the 1000+ meter-thick sequence. The calcalkaline nature of the Lynn-Mattapan is characterized by its meta-aluminous to peraluminous composition, and by relatively low concentrations of high-field strength trace elements such as Zr, Nb, and Y (Hermes and Murray, 1990; Thompson and Hermes, 1990).

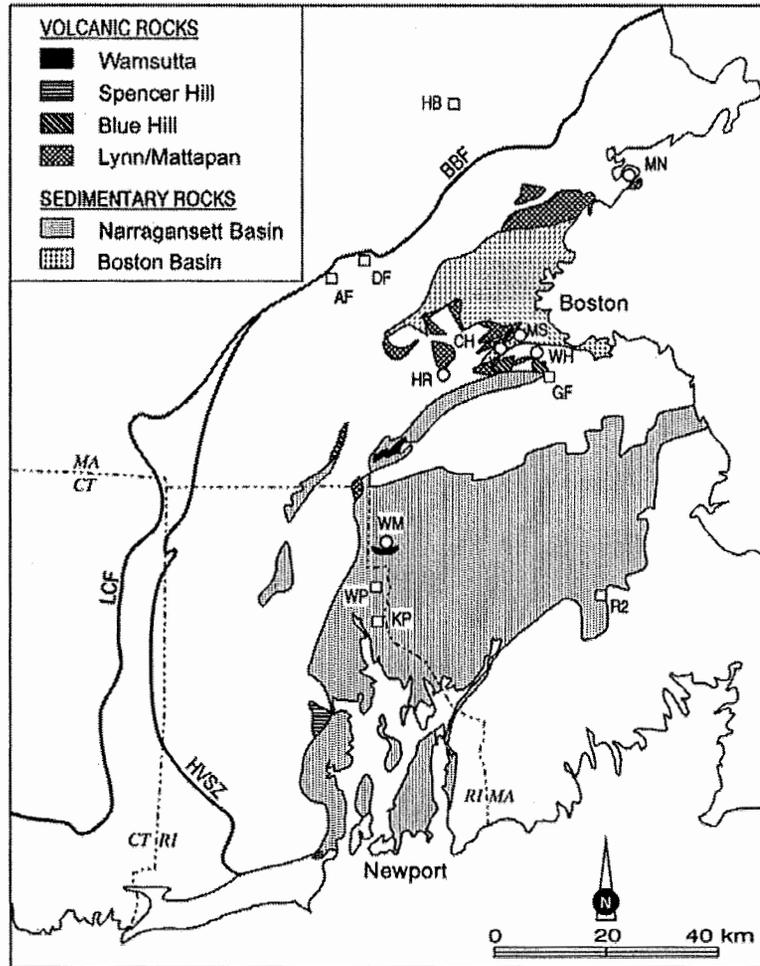


Figure 1: Generalized geologic map of southeastern New England showing distribution of volcanic units as listed in the legend (after Hermes and Murray, 1990; Hermes and Ritchie, 1997). Prehistoric archaeological quarry sites of felsite are shown by open circles: WH=Wampanatuck Hill, HR=Hale Reservation, CH=Clarendon Hill, MS=Mattapan Square, MN=Marblehead Neck, WM=Wamsutta. Archaeological sites for debitage samples discussed in Hermes and Ritchie (1997), but not here, are shown by open squares: DF=Davis Farm, GF=Gill Farm, AF=Atkinson Field, KP=Kettle Point, HB=Heath Brook, WP=Walker Point, R2=Riverside 2.

### Alkalic Suites

In contrast, alkalic suites represented by the Ordovician Blue Hill and Devonian Spencer Hill complexes (Fig. 1) exhibit modest to strong peralkaline character, and are relatively enriched in high-field strength trace elements. The Blue Hill suite is located in the Blue Hill highlands near the southern edge of the Boston basin. The complex contains flows of devitrified rhyolite and abundant subvolcanic granite porphyry; pyroclastic units appear subordinate to other lithologies (Naylor and Sayer, 1976). The Spencer Hill volcanics crop out in central Rhode Island, adjacent to the compositionally similar Scituate Granite pluton just west of the Narragansett basin. The

complex includes fine-grained rhyolite flows, tuff, ash-flow tuff, volcanic agglomerate, and subvolcanic granite porphyry. Although material suitable for tool making is present in the Spencer Hill volcanics, no evidence of prehistoric quarrying in these rocks has been found at this time. The Wamsutta volcanics, located near Attleboro, Massachusetts, are of Devonian age (Thompson and Hermes, 2003) and consist of a bimodal group of at least four basalt and two rhyolite lava flows (Maria, 1990; Maria and Hermes, 2001). The rocks are dominantly lava flows, with only several associated thin layers of tuff and pyroclastic units. The Wamsutta rhyolites generally are alkalic in nature, but less so than the Blue Hill and Spencer Hill suites, and hence are transitional in composition between these rocks and the calcalkaline Lynn-Mattapan suite. Relative enrichment of Fe in the Wamsutta rhyolites causes their characteristic red color.

All of the above volcanic units contain well lithified, very fine-grained varieties of rock that could be conveniently quarried and used by prehistoric people. However, the diversity of colors, textures, and compositions within each group is large, making it extremely difficult to conclusively identify the parent source of archaeological materials on the basis of hand sample identification alone. Hermes and Murray (1990) have demonstrated that petrography and geochemistry, especially some groups of trace elements, are quite useful in distinguishing rock types from these diverse volcanic suites, even where macroscopic features and physical properties are ambiguous.

Quarrying of felsitic rocks from exposed sections of the Lynn/Mattapan and Wamsutta volcanic suites began at least 8500 years ago and continued until the period of first European contact with Native American groups in southeastern New England. Felsites obtained at quarries were chipped into various tool forms such as projectile points and general purpose bifacial tool blades used for cutting and scraping tasks. Artifacts of these felsites diagnostic of the Early Archaic to Late Woodland periods have been found on archaeological sites across central and eastern Massachusetts and Rhode Island.

Both the Lynn/Mattapan and Wamsutta volcanics were important sources of raw material used by groups both in close proximity to the source area and further away from them. A wide variety of visually different felsites from the Lynn/Mattapan were widely used during the Archaic and Woodland periods by prehistoric groups in northeastern, east central and southeastern Massachusetts. Partially completed preforms and finished tools were carried to locations quite far from the lithic source areas and quarries. Artifacts and debitage of felsites from one section of the Lynn/Mattapan complex have been found at distances up to 70 miles (112 km) from their probable source north of Boston. The rhyolite from the Wamsutta volcanics was a primary lithic material used in the Taunton Basin and Narragansett Bay area of southeastern Massachusetts and Rhode Island.

The quarries within sections of the Lynn/Mattapan and Wamsutta volcanic suites attest to the detailed knowledge of the landscape and its resources developed by Native American populations. All the major exposures of fine-grained rocks with felsitic textures were known and used early in the prehistoric period before 8000 years ago, and continued to supply material for many generations. They were important components of prehistoric technology and in some periods use of stone from specific quarry sources was apparently restricted to particular river drainages or group territories.

Prehistoric quarry sites in the Lynn/Mattapan volcanic complex have been studied by natural historians and avocational archaeologists since the late nineteenth century. Some visually distinct felsites from specific sources and quarry sites from different parts of the Lynn/Mattapan have been given names such as "Melrose green," "Saugus jasper" and "Mattapan banded". The felsite within the Wamsutta formation was known to avocational archaeologists as "Attleboro red felsite" and widely recognized in southeastern Massachusetts (Haynes, 1886; Fowler, 1972; Bowman, 1981). Archaeologists have relied on visual, macroscopic features to identify felsites from various sources in southeastern New England. However, the wide range of color, phenocryst type and abundance, and weathering and alteration patterns have lead to frequent errors in assigning archaeological materials to sources.

Prior geological studies have demonstrated that petrography and geochemical composition can be used to provenience volcanic rocks in southeastern New England to igneous complexes (Hermes and Murray, 1990; Thompson and Hermes, 1990). This data can also be used to determine the likely sources of felsites found on archaeological sites. Recent research has applied standard methods used in geological analysis to provide accurate information on the felsites from a number of source areas and quarries in the Lynn/Mattapan and Wamsutta suites.

Samples from several quarry sites in sections of the Lynn/Mattapan north and southwest of Boston and the Wamsutta volcanic near Attleboro have been analyzed (Strauss and Murray, 1988; Strauss and Hermes, 1996; Luedtke, e. al., 1998; Hermes, et al., 2003).

A larger scale pilot study collected petrographic and geochemical data from six prehistoric quarry sources in the Lynn/Mattapan, Blue Hills and Wamsutta volcanic complexes and compared it to a set of debitage samples from archaeological sites in both Massachusetts and Rhode Island. The debitage samples were felsites suspected to be from these major volcanic complexes (Hermes and Ritchie, 1997). All of the felsite locations included in this field trip was part of this pilot study. The objective of these studies was to provide quantifiable data on the fine-grained, felsitic rocks that were the subject of prehistoric quarrying. Petrographic and geochemical features of the rhyolitic rock in these source areas have been of value for determining the probable origin of felsite artifacts found on archaeological sites in Massachusetts and Rhode Island.

This trip will visit examples of outcrops of the Lynn/Mattapan and Wamsutta volcanic complexes quarried during the prehistoric period. Situated in upland terrain with numerous bedrock outcrops, these locations are typical of source areas utilized by Native American in southern New England. At three of these sources, weathered blocks in soil overlying bedrock or talus below outcrops provided the raw material for stone tool production. One quarry is a less common example of extraction directly from an exposed dike of fine-grained rhyolitic material.

The following sections discuss the geological setting or context of quarries and the petrography and geochemistry of the felsite and soapstone present in them. The implications of petrographic and geochemical features for tracing the distribution of artifacts made from these materials found at archaeological sites in southeastern New England are also described.

## GEOLOGICAL CONTEXT

### **Lynn/Mattapan Volcanic Suite**

This suite of calc-alkaline volcanics occurs northeast and southwest of Boston (Fig. 1). It contains a variety of mostly pyroclastic rock types including rhyolite to andesite tuff, ash-flow tuff and lava flows. Features include broken crystals, volcanic clasts, relict pumice and glass shards, phenocrysts of plagioclase, quartz and perthite, often in glomeroporphyritic clusters, accessory titanite and prominent late-stage epidote. Some varieties of rock have distinct flow banding.

Geochemical characteristics of the Lynn/Mattapan include low amounts of high-field strength elements such as Nb, Zr, and Y as well as low Rb; it has relatively high concentrations of Sr and Ba (Table 1, Figs. 2-4). Based on geochemistry, the Lynn/Mattapan south of the Boston basin has been subdivided into two stratigraphic units, the High Rock tuff and Twin Pine tuff. These units display differences in Nb and Zr in comparison to each other and other Lynn volcanics (Thompson and Hermes, 1990). There are lithic source areas and prehistoric quarries in both the High Rock and Twin Pine units, as well as in the Lynn Volcanics which occur northeast of Boston.

### **Blue Hill/Spencer Hill Volcanic Suites**

Rocks of these suites include felsic lava flows, pyroclastic rocks, and porphyritic shallow-seated intrusive rocks. Phenocrysts are dominated by K-feldspar perthite and quartz; plagioclase is subordinate. Common diagnostic accessory minerals include fluorite, riebeckite, and aegerine. In sharp contrast to the Lynn/Mattapan suites, these rocks are enriched in high field strength elements as well as Rb, and they exhibit comparatively low concentrations of Sr and Ba (Table 1, Figs. 2-4)

### **Wamsutta Volcanic Suite**

The Wamsutta suite includes predominantly red, clastic sedimentary rocks, dark basaltic lava flows and pink to dark red or purple rhyolite lava flows, with minor felsic pyroclastic rocks. The rhyolitic rocks contain abundant

	Lynn-Mattapan	Blue Hill	Spencer Hill	Wamsutta
<b>Petrography</b>				
Phenocrysts	plagioclase + k-feldspar + quartz	k-feldspar (perthite) + quartz	k-feldspar (perthite) + quartz	anorthoclase
Key Accessory Minerals	epidote + titanite	aegerine + fluorite	riebeckite + fluorite	riebeckite + fluorite + titanite
Texture	1) pyroclastics > flows 2) fragmental 3) shard texture 4) flow-banding	1) flows > pyroclastics 2) resorbed quartz	1) flows > pyroclastics 2) resorbed quartz	1) flows > pyroclastics 2) flow-banding 3) quartz amygdules 4) quartz bands (lithophysae)
<b>Geochemistry</b>				
Major elements	rhyolite to andesite	rhyolite	rhyolite	rhyolite
Trace elements	high Sr, Ba low Rb, Y, Zr, Nb, Zn	high Rb, Y, Zr, Nb, Zn low Sr, Ba	high Rb, Y, Zr, Nb, Zn low Sr, Ba	intermediate Rb, Y, Zr, Nb, Zn, Sr, Ba

Table 1: Summary of main petrographic and geochemical characteristics that are useful to discriminate volcanic suites in southeastern New England.

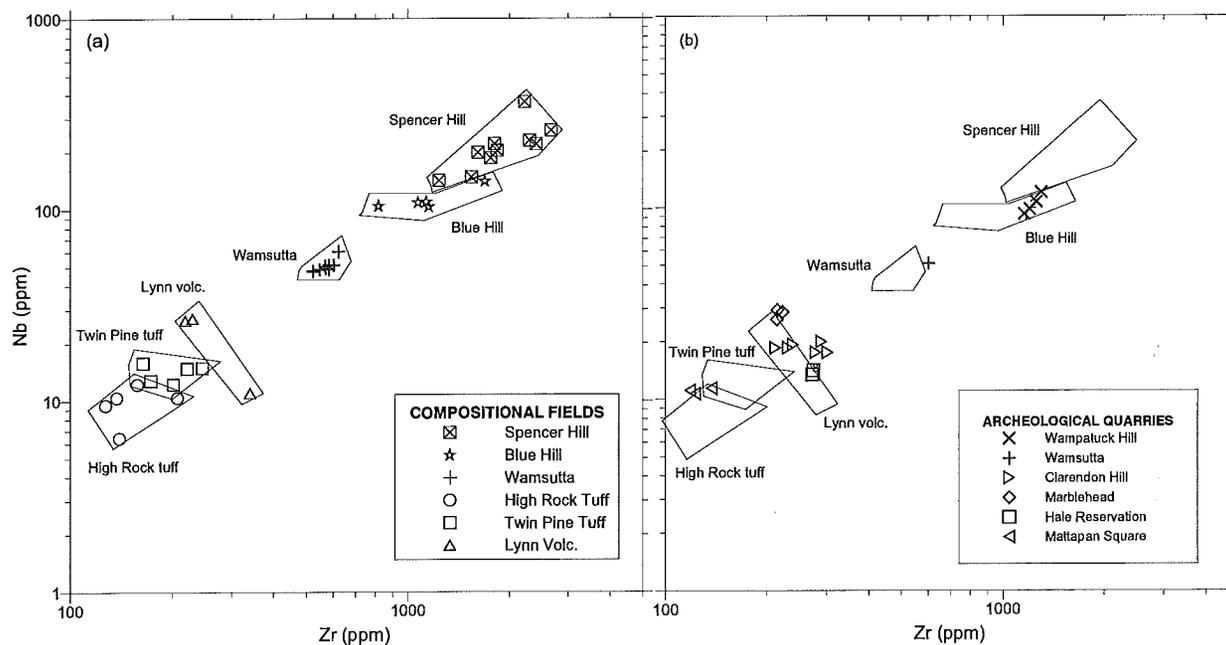


Figure 2: (a) Log-log plot of Zr vs. Nb in volcanic rocks from southeastern New England. The compositional fields are defined by data from Hermes and Murray (1990) and Thompson and Hermes (1990). Note that data for some units is based on a small number of samples; hence, as more data is collected, the size and shapes of the polygons are subject to modification. (b) Log-log plot of Zr vs. Nb in samples from known archaeological quarry sites (fields same as Fig. 2a).

anorthoclase phenocrysts. Quartz and plagioclase phenocrysts are absent. Quartz is abundant as late stage filling of vesicles and as filled areas along sub-parallel planar fractures (possibly volcanic lithophysae); it also occurs in the matrix where it formed from devitrified glass. Accessory minerals include abundant titanite and sparse riebeckite (Table 1).

The Wamsutta rhyolite has a distinct intermediate geochemical composition that allows it to be separated from the other alkalic suites such as Blue Hill and Spencer Hill (Figs 2-4), as well as from the calc-alkaline Lynn/Mattapan suite (Maria, 1990; Strauss and Murray, 1988; Hermes and Ritchie, 1997).

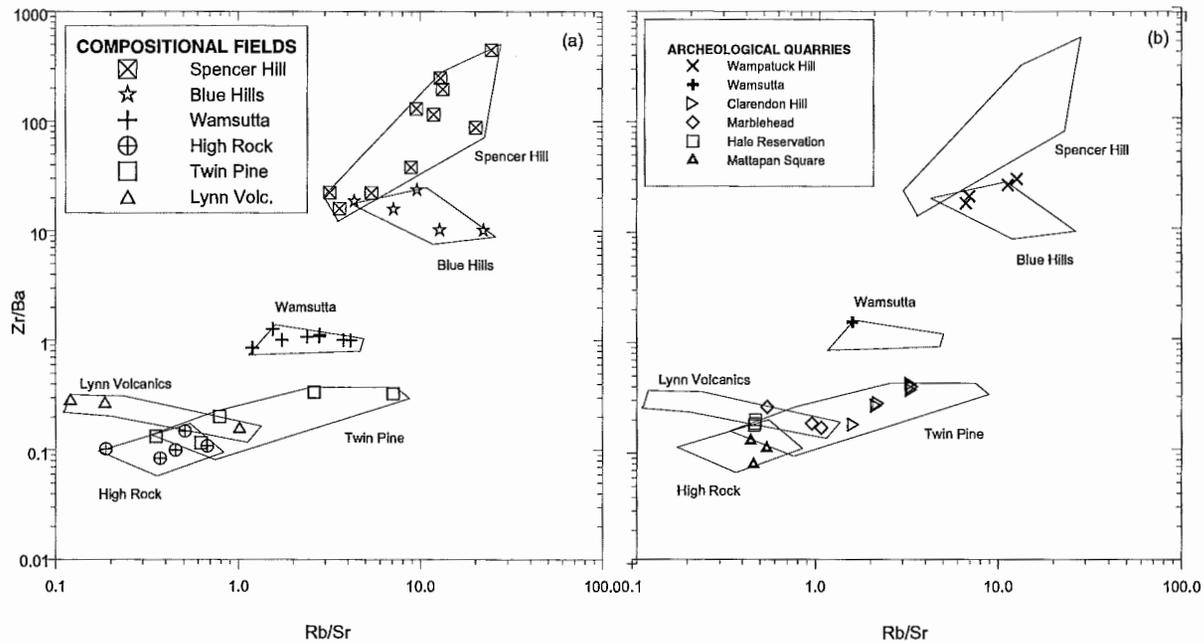


Figure 3: (a) Log-log plot of Rb/Sr vs. Zr/Ba in volcanic rocks from southeastern New England. The compositional fields are defined by data from Hermes and Murray (1990) and Thompson and Hermes (1990). Note that data for some units is based on a small number of samples; hence, as more data is collected, the size and shapes of the polygons are subject to modification. (b) Log-log plot of Rb/Sr vs. Zr/Ba in samples from known archaeological quarry sites (fields same as Fig. 3a).

### Soapstone Quarries

Soapstone is a metamorphic rock comprised primarily of talc with lesser amounts of chlorite, amphibole, magnetite, and other minerals (Rogers et al., 1983; Truncer 2004).

Rare earth or trace elements have proven useful in discriminating Mid-Atlantic soapstone source areas in the US (Allen and Pennell, 1978; Allen et al., 1975; Luckenbach, Allen, and Holland, 1975; Holland et al., 1981; Luckenback Holland, and Allen, 1975; Truncer, 2004; Truncer et al., 1998) and the Southeastern (Allen and Pennell, 1978). Although some major element profiling has been conducted for a few southern New England soapstone source areas (Turnbaugh and Keifer, 1979; Turnbaugh et al., 1984), trace element analyses for these sources are in its infancy. Trace element characterization of southern New England soapstone is being conducted as dissertation project for the Department of Anthropology/University of Connecticut (Joseph Waller) assisted by the Department of Geosciences/University of Rhode Island. A pilot study exploring the feasibility of using Wave-Length Dispersive X-Ray Fluorescence (WDXRF) to identify major and trace elements for some southern New England steatite source areas has produced encouraging results. Soapstone specimens from the two extant Rhode Island soapstone quarries and samples from one Massachusetts quarry suggests that geochemical variability exists within the concentrations of some major and trace elements between these source areas. Further work should assist archaeologists in discerning southern New England Terminal Archaic Period prehistoric trade and exchange networks by correlating the geographic distribution of steatite vessels with their original source areas locales through geochemical means.

Southern New England's steatite or soapstone crops out along a "talca bed" that roughly follows the Appalachian Mountain chain from Georgia to Maine (Turnbaugh et al., 1984). The Office of the Rhode Island State Geologist maps the underlying bedrock in the vicinity of Rhode Island's Ochee Spring soapstone quarry as Late Proterozoic or Older Age epidote and biotite schist, which is included within the Blackstone Group (Hermes et al., 1994). The Blackstone Group is lithologically diverse, and includes greenstone, steatite, and marble units in addition to the more common schist and gneissic rocks. The nearby Oaklawn soapstone quarry is situated atop three intersecting underlying bedrock types that include epidote and biotite schist, Devonian Age diorite/gabbro, and Late Proterozoic Age granite. Each of the quarry sites is situated to the immediate west of the Narragansett Basin Border Fault that separates the Esmond-Dedham West Bay Area Subterranean from that of the East Bay Narragansett Basin. Preliminary geochemical studies indicate the soapstone specimens studied so far are primarily composed of silicon dioxide ( $\text{SiO}_2$ ) and magnesium oxide ( $\text{MgO}$ ). Source area specimens vary widely in their concentrations of minor and trace elements, such as vanadium, chromium, cobalt, nickel, zinc, and barium.

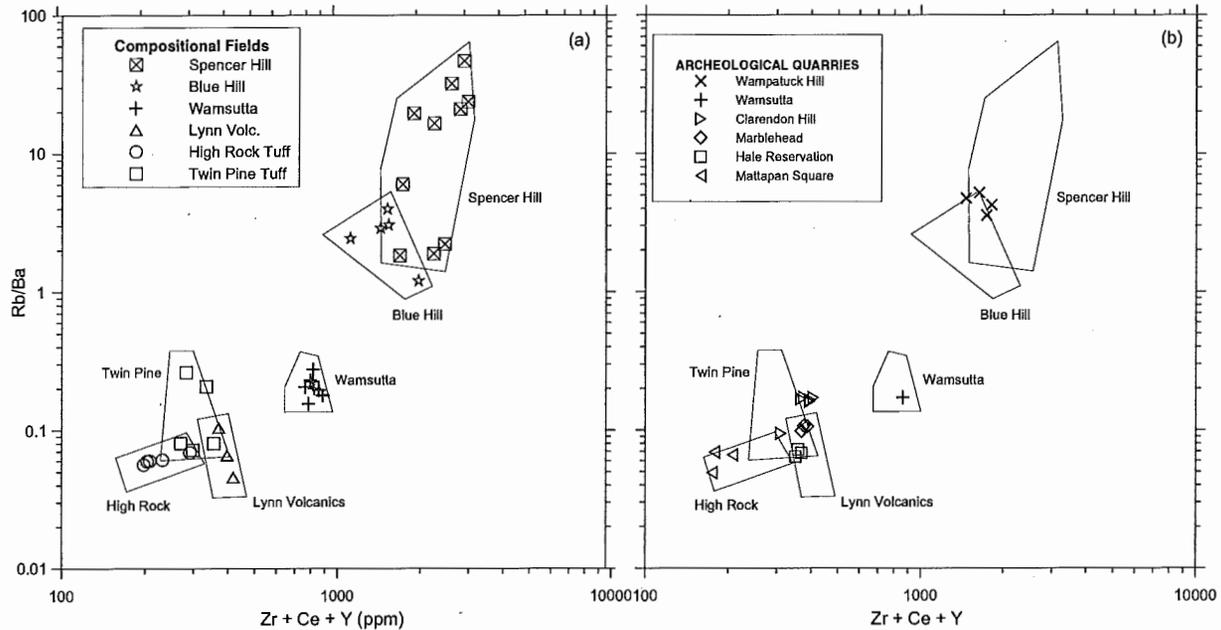


Figure 4: (a) Log-log plot of  $\text{Zr} + \text{Ce} + \text{Y}$  vs.  $\text{Rb}/\text{Ba}$  in volcanic rocks from southeastern New England. The compositional fields are defined by data from Hermes and Murray (1990) and Thompson and Hermes (1990). Note that data for some units is based on a small number of samples; hence, as more data is collected, the size and shapes of the polygons are subject to modification. (b) Log-log plot of  $\text{Zr} + \text{Ce} + \text{Y}$  vs.  $\text{Rb}/\text{Ba}$  in samples from known archaeological quarry sites (fields same as Fig. 4a).

### SOURCE AREAS/ QUARRIES

#### Clarendon Hills

This location is in an area just south of the Boston basin that contains numerous exposures of the Mattapan volcanic. Outcrops of two basic types of visually different felsitic rock occur at this location. They have also been exposed by construction of a shopping center and a modern quarry. Prehistoric quarries have not been reported from here; however any evidence of this type of activity could have been destroyed or obscured by modern development and commercial quarrying. Evidence of prehistoric quarrying of similar material in the general vicinity has been described by avocational archaeologists (Bowman, 1981).

One type of rock is a light grey felsite with conspicuous pyrite on freshly broken surfaces. Weathered surfaces exhibit a bleached tan to white color with rusty iron staining. This rock corresponds to material commonly known as the "Sally Rock felsite" by avocational and professional archaeologists. The second type is a mottled tan-grey to pink felsite with sparse feldspar phenocrysts. Weathered surfaces of this felsite often develop a banded grey-white to light pink color.

Samples of Clarendon Hill material have geochemistry like other Mattapan volcanics and are similar to the Twin Pine tuff unit (Figs. 2-4). The two visually different varieties from this source were found to have slightly different trace element compositions, although they were collected from outcrops in close proximity to each other. This suggests there is some localized heterogeneity of felsitic rock within this source.

Petrographic features of the porphyritic, grey felsite include euhedral-subhedral phenocrysts of perthite with prominent exsolution (Figs. 5-6). Glomeroporphyritic and phenocryst clusters are common. Both quartz and plagioclase phenocrysts are sparse, but there are large, rounded opaque phenocrysts. The matrix consists of massive, fine-grained quartz-feldspar intergrowth with minor opaque minerals, biotite, muscovite and titanite. Some phenocrysts are broken, but no rock clasts were noted. The tan-pink banded felsite from this source is a porphyritic-subporphyritic rock with sparse perthite phenocrysts and some opaque microphenocrysts. Quartz and plagioclase phenocrysts are not present. The groundmass contains abundant small, euhedral-subhedral K-spar crystals and is coarser grained than most other rocks. The feldspar occurs in random orientations, immersed in finer grained quartz and K-spar (Hermes and Ritchie, 1997).



Figure 5: Photomicrograph of sample MTPN-SR from the Clarendon Hill area. Glomeroporphyritic cluster of euhedral-subhedral perthite, with minor alteration to saussaurite. Fine-grained quartz-feldspar rich matrix. Width of field = 3.0 mm, crossed polarizers.



Figure 6: Photomicrograph of sample MTPN-SRBND from the Clarendon Hill area. Angular, broken clast or phenocryst of feldspar in a relatively coarser-grained matrix of intergrown quartz-feldspar. Width of field – 1.2 mm, crossed polarizers.

### Hale/ Noanet Quarry

The Hale or Noanet Quarry is located within the Robert S. Hale Reservation, a private foundation in Dover and Westwood, Massachusetts. The site is now used for environmental and archaeology education programs run by the Hale Reservation and other organizations. This site consists of an exposed section of a dike of fine-grained rhyolitic rock surrounded by country rock. The dike appears to represent a feeder conduit leading from a magma reservoir to the surface where extrusion of a lava flow or pyroclastic material occurred. Rapid cooling of this material at a shallow depth below the surface is indicated by the fine-grained texture of the dike rock. The matrix or country rock for the dike is a medium grained granodiorite composed of plagioclase, quartz and minor potassium feldspar; there are also small veins of epidote.

This quarry was first investigated by an avocational archaeologist in 1970. A more recent study involved limited archaeological sampling and mapping of prehistoric quarry debris in an associated workshop area. Petrographic and x-ray fluorescence analysis of pieces of the dike rock and granodiorite matrix was also conducted to characterize the material in this quarry. This analysis revealed that the rhyolitic dike rock has a trace element geochemistry generally similar to other types of Lynn/Mattapan volcanics (Strauss and Hermes, 1996). The composition is also like that of the Twin Pine tuff unit identified within the Mattapan volcanic suite by Thompson and Hermes, (1990). This unit outcrops nearby in the town of Westwood.

In thin section, the felsite from this quarry site displays a very fine-grained groundmass consisting of feldspar laths in a spherulitic, felty textured mass of feldspar, quartz and opaque minerals. It also contains clusters of twinned plagioclase phenocrysts with accessory epidote (Fig. 7). Unlike other Lynn/Mattapan felsites, no xenoliths, broken crystals or rock fragments typical of pyroclastic texture were present in this dike material (Strauss and Hermes, 1996: 161). Compositionally, these rocks are most similar to the Twin Pine member of the Mattapan volcanics (Figs. 2-4)



Figure 7: Photomicrograph of sample STR-1B from the Hale Reservation. Glomeroporphyritic cluster of plagioclase and accessory epidote embedded in a pilotaxitic matrix of feldspar, quartz, and opaque minerals. Width of field = 3.0 mm; crossed polarizers.

### **Cat Rock**

Cat Rock is another area within the Robert S.Hale Reservation where exposures of rhyolitic rock occur within larger areas of scattered granite outcrops. The primary bedrock unit is the Westwood Granite (Chute, 1966). The topography is typical of uplands bordering the Boston basin consisting of knolls with granite boulders and outcrops bounded by wooded wetlands. At this location felsite outcrops occur on the crest and slopes of a group of prominent knolls.

The felsite at this location consists of a light to medium grey weathering, fine-grained material with a few large feldspar phenocrysts visible on surfaces. On fresh breaks the material is a dark grey to grey-black color. Prehistoric quarrying at this location probably involved excavation of weathered blocks and talus fragments from soil surrounding the outcrops. This observation has not been confirmed by archaeological investigation.

Samples of felsite from Cat Rock were subjected to petrographic and geochemical analysis as part of a pilot study to develop base line data from known lithic source areas within eastern Massachusetts. The Cat Rock felsite has a geochemical composition consistent with the Lynn/Mattapan volcanic suite. Ratios of certain trace elements fall within or near established fields for the Lynn volcanic and Twin Pine tuff unit of the Mattapan volcanic suite. Petrographic features include glomeroporphyritic patches of plagioclase and K spar with Carlsbad twinning (Fig. 8). The plagioclase feldspar is saussauritized. The matrix is a fine-grained intergrowth of feldspar, quartz and lesser amounts of epidote, chlorite and opaque minerals. No rock fragments were noted (Hermes and Ritchie, 1997).

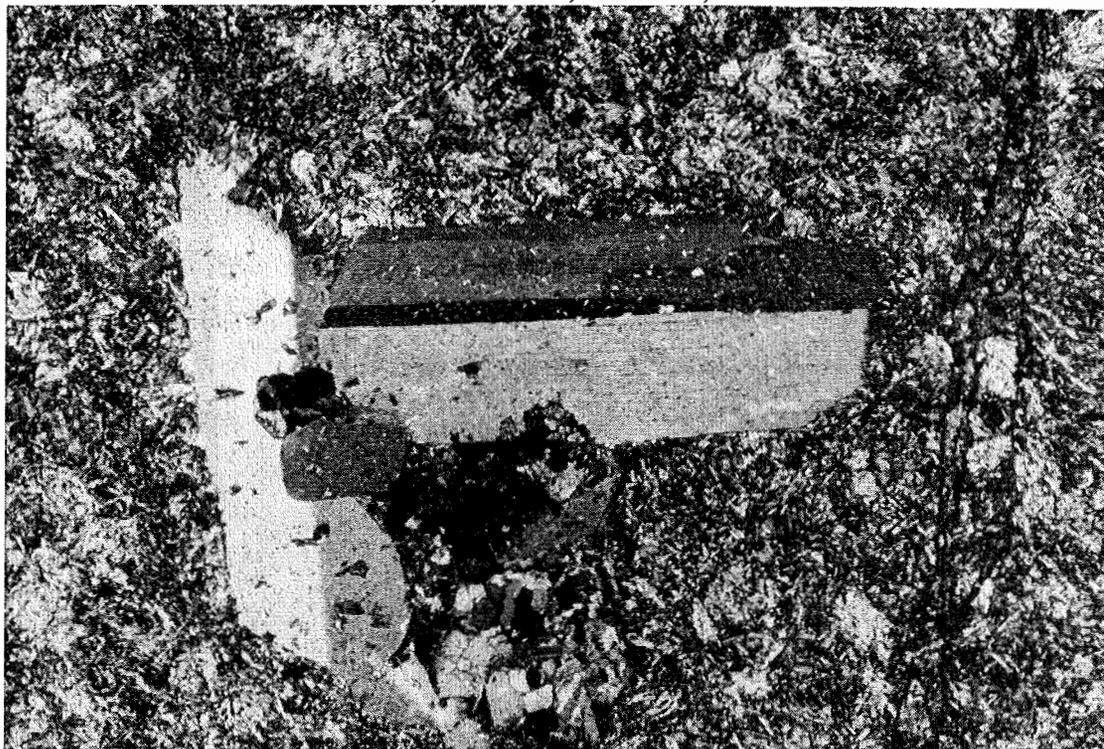


Figure 8: Photomicrograph of sample MAT-W/D from the Hale Reservation. Euhedral-subhedral cluster of plagioclase phenocrysts spatially associated with epidote. Small epidote vein cuts the rock. Quartz-feldspar rich matrix locally exhibits crude spherulitic texture. Width of field – 3.0 mm; crossed polarizers.

#### **Wampatuck Hill Source**

Wampatuck Hill is a source area in the eastern end of Blue Hills Range. Its location within the Blue Hills Reservation, a large tract of state-owned parkland, has afforded this significant source area protection from modern development. Rhyolite outcrops and associated talus areas are dispersed throughout a hillside setting with granite boulders. Native American quarrying methods here appear to have been similar to other source areas with weathered blocks and talus fragments serving as the primary source of raw material for stone tool making.

The rhyolite at this source ranges in color from almost black to medium grey on freshly broken pieces. Pink-tan feldspar phenocrysts and round quartz grains are visible on hand specimens and form key identifying features for this rhyolite. Deeply weathered surfaces are light grey, while prehistoric debitage and stone tools are usually a medium to dark grey color.

Petrographic analysis has revealed that this very hard, fine-grained rhyolite contains abundant euhedral-subhedral perthite phenocrysts, many with Carlsbad twins. Smaller phenocrysts of quartz are abundant, many of which are rounded, embayed, and corroded (Fig. 9). The gray to black matrix is composed of quartz, feldspar, mica, calcite, and opaque minerals. Compositionally, these rocks are similar to the alkalic rocks of the Blue Hill area, and are notably distinct from rocks of the Lynn-Mattapan suite (Figs. 2-4).

#### **Wamsutta Volcanic source**

The Wamsutta volcanic source area is located in an upland area in South Attleboro. Outcrops or exposures of rhyolitic material occur within an arc shaped belt or zone of the Wamsutta volcanics. The extent of this zone and the location of major rhyolite flows within it has been mapped in previous geological research (Woods, 1961) (Maria, 1990). Evidence of prehistoric quarrying in at least one general area was known to avocational archaeologists but

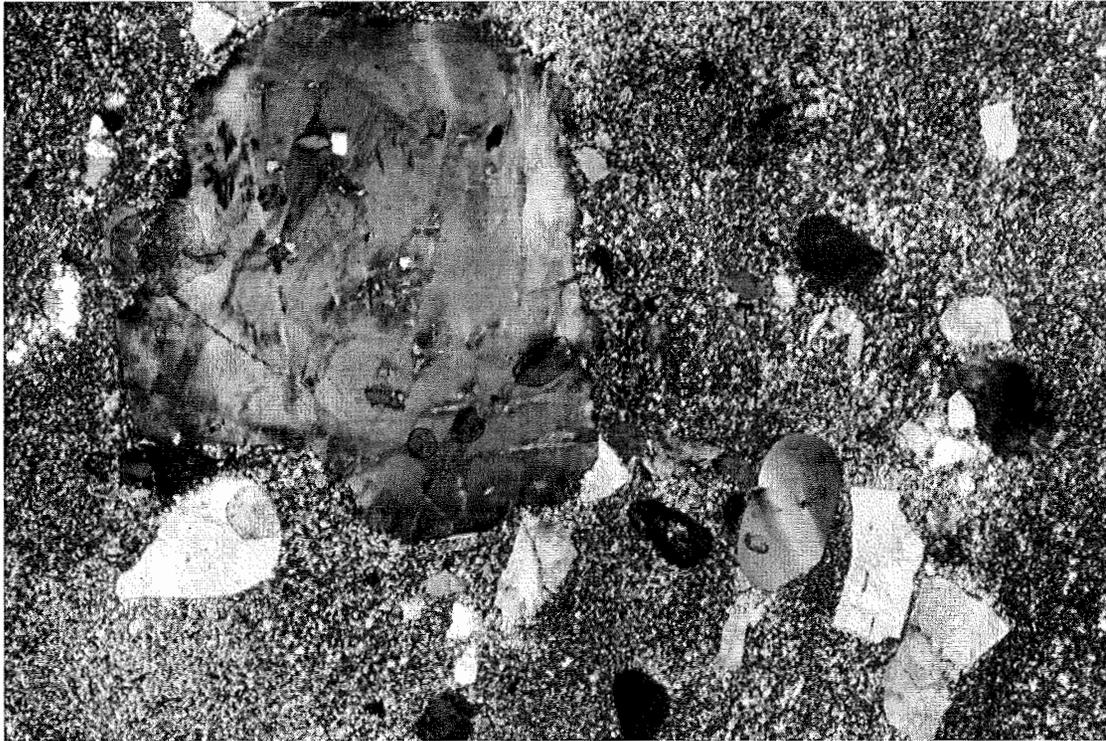


Figure 9: Photomicrograph of sample WH from Wampatuck Hill in the Blue Hills area. Subhedral-anhedral phenocrysts of K-feldspar and quartz phenocrysts (modestly undulatory). Some phenocrysts are angular and broken. Secondary sericite and saussurite in a quartz-feldspar rich matrix. Width of field = 3.0 mm; crossed polarizers.

only briefly described (Fowler, 1972). Other previous archaeological studies identified three outcrop areas with evidence of prehistoric quarrying along this zone. Talus slopes associated with these outcrops were the focus of the quarrying activity. Petrographic and x-ray fluorescence analysis of samples including both outcrop derived material and an artifact (bifacial tool blade) from a nearby archaeological site was also conducted (Strauss and Murray, 1988).

Felsite from this source has a distinct geochemistry (Table 1, Figs. 2-4) that is intermediate to the calc-alkaline Lynn-Mattapan and alkalic Blue Hill/Spencer Hill suites (Hermes and Ritchie, 1997). Thin sections made from outcrop samples with evidence of prehistoric quarrying revealed a groundmass of quartz, potassium feldspar and plagioclase with numerous anorthoclase feldspar phenocrysts and stringers of anhedral quartz (Fig. 10). Hematite plates and other iron oxide provide the red/purple color so characteristic of this felsite. Other accessory minerals are fluorite, riebeckite, chlorite, magnetite and epidote (Strauss and Murray, 1988:46).

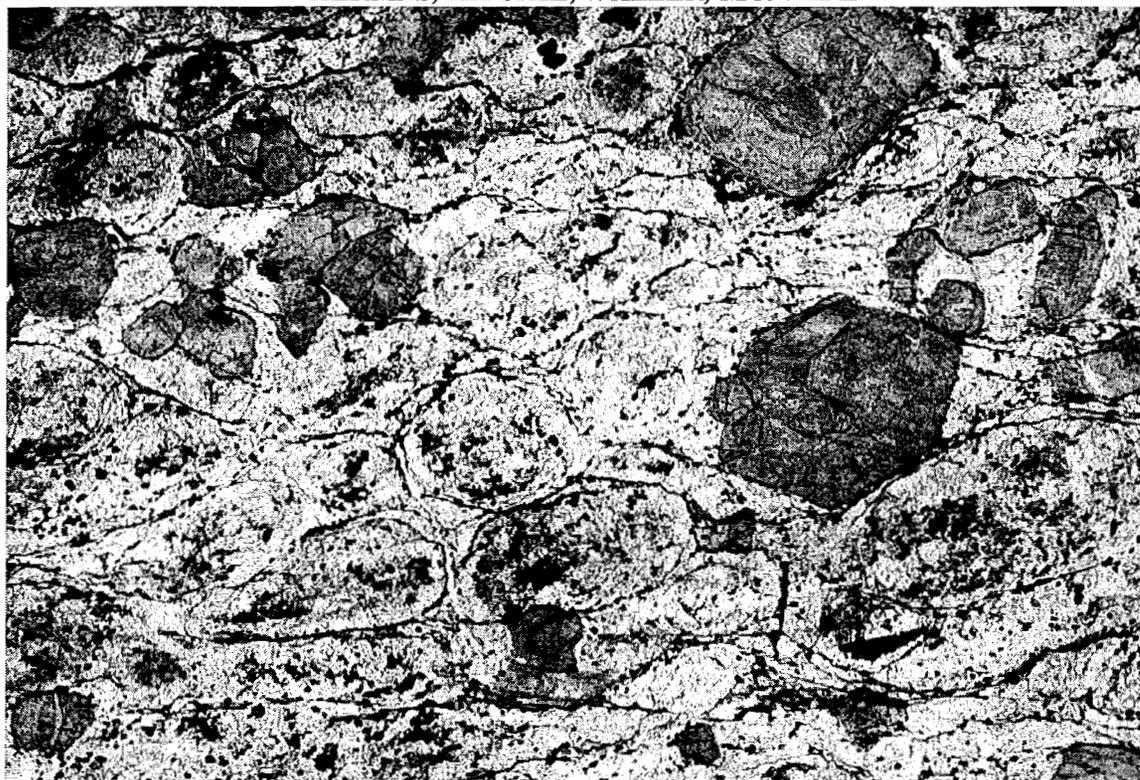


Figure 10: Photomicrograph of one variety of rhyolite from the Wamsutta. Clouded phenocrysts of euhedral-subhedral anorthoclase in a devitrified perlitic textured matrix, now consisting of intergrown quartz, feldspar, and opaque minerals. Width of field = 3.0 mm; plane light.

### SUMMARY

While the characteristics of the Lynn/Mattapan and Wamsutta complexes are fairly well known from the geological perspective the value of these volcanic rock units and the prehistoric quarry sites source areas within them as cultural resources also needs to be emphasized.

Clarendon Hills is typical of lithic source areas within the Boston metropolitan area that have been significantly modified by modern development. This development process can provide fresh exposures of rock, but also leads to destruction of the prehistoric quarry and workshop sites and their surrounding physical context and setting. The Hale/Noanet Quarry and Cat Rock are part of a lithic source area which has fortunately been preserved in a large tract of land in protected open space status. The Wompatuck Hill source area has been well protected by being incorporated in a state reservation or park which was established over a century ago. The Wamsutta volcanic source area is in an area near major highway corridors (Routes 1, 295 and 95) that has recently had intensive commercial development. It is a cultural resource of regional significance that is threatened by future land use.

These lithic source areas represent valuable sources of information about Native American use of specific types of geological resources present in upland sections of the Boston basin and Narragansett basin. The quarry sites can help to reconstruct various aspects of extractive technology, the selection of material for production of chipped stone tools and preliminary reduction, and methods of shaping or preform manufacture done at quarries and associated workshops. Quarry sites also contain information relative to the chronology or sequence of Native American use of particular felsitic volcanics, since quarries were probably not operated with equal intensity during all time periods. Some of the more recent investigations cited above have applied analytical methods used in geology to answer questions about Native American use of felsites from the major volcanic complexes of southeastern New England.

These studies demonstrate the potential of such interdisciplinary research and the need for lithic source areas to be carefully preserved for long term study by both geologists and archaeologists.

#### ACKNOWLEDGEMENTS

Jim Earley, director of the Robert S. Hale Reservation, generously provided access to the Cat Rock and Hale/Noanet Quarry sites.

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## ROAD LOG

### Mileage

- 0.0 Assemble at the Park N Ride lot on Route 138 in Milton, Massachusetts. Leave the parking lot and turn left onto Route 138. Proceed north.
- 4.0 Intersection of Route 138 and Route 28. Continue north on Route 28
- 4.2 Cross the Neponset River on Route 28 to Mattapan Square. Turn left on to Cummins Highway, continue west.
- 5.5 Large cemetery on right side of street. Sign for Stop and Shop and Bradlees, turn left in to parking lot for shopping center. Outcrop is visible to left of store along back edge of parking lot.

**STOP 1** (20 - 30 minutes)

The outcrops of Mattapan volcanic rock here represent the Clarendon Hills source area. The surrounding area has dense residential and commercial development and this stop is similar to some other locations in the Boston metropolitan area where there are small surviving sections of prehistoric lithic source areas and quarries. Sally Rock is the hill behind the shopping center. A modern quarry was formerly located to the rear of the shopping center on the crest of the hill.

The outcrop at the northern end of the parking lot consists of white to yellow weathering, grey felsitic rock with pyrite. Rust stains from oxidation of the pyrite are common. Bowman (1982) described a talus slope in the general area with evidence of prehistoric quarrying. This consisted of deposits of chipping debris, bifacial tool blades or preforms of "Sally Rock felsite" and a full grooved axe made of Braintree Slate, probably used as a quarrying tool. The material at this site apparently had grey/white weathered surfaces with rust stains similar to the rock exposed in the outcrop.

Chipping debris and artifacts of Sally Rock felsite similar to this material are common on prehistoric sites in the southern Boston basin area, particularly along the Neponset and Cochato River drainages. This felsite occurs further south in the Taunton basin, but is rare west of Boston. It appears to have been used primarily in the Middle Archaic period (ca 8000 - 5500 years ago) by people of the Neville complex and to a lesser extent in the Late Archaic period (ca 5500-3000 years ago). Later use was limited to the Middle Woodland period (ca 1600-1000 years ago) Fox Creek complex that made large, lanceolate projectile points. The Woodland period artifacts and chipping debris are not as deeply weathered and still retain a grey-green, light brown or grey color. The lack of weathering may help to distinguish these more recent artifacts from older, Archaic period tools.

At the southern end of the parking lot behind the shopping center were small outcrops and boulders of banded, mottled tan pink felsite. These have recently been covered by soil and shrubbery. Beyond the parking lot, to the south is an open area where topsoil has been removed. Some scattered fragments of felsite are exposed on the surface. An abandoned asphalt access road leads uphill to the former location of the modern quarry pit which has been used as a landfill. The modern quarry pit has been filled recently and used as a dump for concrete demolition rubble and urban soils, now overgrown with small trees and weeds. Approximately 300 feet uphill (east) along the abandoned asphalt road, a large pile of felsitic rock from the modern quarry has survived. The boulder-sized pieces in this pile display the mottled and striped pink, brown and grey color typical of the rock in this specific section of the Mattapan volcanic complex.

Prehistoric use of the mottled pink to purple and tan felsite at this location is likely, but has not been confirmed yet. Artifacts and pieces of chipping debris with pink-grey weathered surfaces found on sites in the Boston area may be from this section of the Mattapan volcanic complex. Compositional characteristics of these rocks are summarized in Figures 2-4.

#### **Mileage (cont.)**

From the shopping center, return by same route described above via Cummins Highway to Neponset River bridge and Route 28.

- 7.8 Turn right and head south on Route 28. Go past intersection of Blue Hill Avenue to point where Route 28 bears left (Brook Road). On Route 28 go past St Mary of the Hills School (on left) and Kelly Field (large playing fields on right).
- 8.9 At intersection of Brook Road, Central Avenue and Reedsdale Road, bear right on Reedsdale Road (Route 28); cross Canton Avenue.
- 9.8 Traffic light at intersection of Reedsdale Road and Randolph Avenue (Route 28). Turn right (south) on Route 28.
- 10.6 Go past entrance to Wollaston Golf Club (on right).

- 11.5 Traffic light at intersection of Route 28 and Chickatawbut Road. Turn left on Chickatawbut Road; you will see large sign for Blue Hills Reservation on the right side of this road as you start to drive through wooded area.
- 12.5 Blue Hills Reservoir is on the right, dam for this body of water is faced with a large rip-rap stone wall. A pump house is located near eastern end of dam.
- 13.0 Bear left on Wampatuck Road at intersection of Chickatawbut Road and Wampatuck Road on south side of Wampatuck Hill.
- 13.3 Gate for fire road is on left. Space for parking here is very limited. Outcrop of rhyolite associated with Blue Hill volcanics is exposed on edge of woods to right of fire road. (Please do not collect or remove any rock from this location; it is within state reservation lands).

**Stop 2 (20-30 minutes)**

Wampatuck Hill. The fine-grained, phenocryst-rich rock here (Fig. 9) is compositionally similar to rocks of the Blue Hill igneous complex, and exhibits significant enrichment of Rb and high field strength elements (Figs. 2-4). This outcrop is typical of the Blue Hills source area. From limited study it is apparent that Native American extraction of raw material relied on easily obtained talus pieces and large fragments found in the soil around outcrops. Actual quarrying or removal of material from outcrops by hammering seems to be rare. The rhyolite here is a dark grey, very fine grained material well suited to stone tool making. Archaeological studies indicate that Blue Hill rhyolite was quarried by Native Americans throughout most of the prehistoric period from the Early Archaic (ca 8500 to 7500 years ago) to Late Woodland periods (ca 1000 to 400 years ago). It was used most extensively in the nearby Neponset, Charles and Taunton river drainage basins to the west and south of the Blue Hills. However, this rhyolite has a very wide distribution across southeastern New England and has been found on archaeological sites in Rhode Island and the Worcester County highlands of central Massachusetts.

**Mileage (cont.)**

Return via Chickatawbut Road to Route 28 (Randolph Avenue).

- 14.8 At Randolph Avenue intersection turn left (south) and continue to exit for Route 93/128 South
- 15.9 Turn right onto ramp for Route 93 South (Route 128). Continue on Route 93/128.
- 18.3 Continue on Route 93/128 past exit for Route 138
- 19.1 Continue past exit for Route 95 South (Providence).
- 21.6 Continue on Route 95/128 past exit for Route 1 (Dedham).
- 23.5 Take exit for Route 109 (South) to Westwood. Continue south on Route 109.
- 24.6 Turn right (west) on Dover Street and continue along this street
- 25.0 Turn right on to Carby Street and follow this street.
- 25.5 Entrance to Robert S. Hale Reservation. After checking in at reservation office, continue up access road.
- 25.8 Drive into parking lot on left side of road.

**STOP 3 (1 1/2 hour)**

From the parking lot walk up the road about 1300 feet to Cat Rock, which is in the woods on the right or north side of the road. This small locus of prehistoric quarrying is located on the crest of a series of knolls with granite outcrops. Felsite dikes in these outcrops were the source of material sought by Native American groups.

Cat Rock is typical of smaller quarry sites in upland terrain where evidence of prehistoric activity is completely hidden by thick forest humus and ground cover vegetation. Extraction of raw material here was probably limited in extent. It may have required excavation of shallow pits into the subsoil to obtain talus blocks or other chunks of material separated from outcrops by frost action or other natural processes. There does not appear to be any evidence of quarrying directly from exposed sections of felsite dikes. Some angular blocks of light grey felsite can usually be observed on the surface. (Please do not collect or remove any of this material). The frequent joints or fracture planes in this felsite were probably a limitation, and may have made it difficult to obtain large pieces suitable for manufacture of bifacial tool blades or preforms. However, those blocks that are free of fracture planes provide a fine-grained material that flakes well.

Return from Cat Rock to the road. Continue walking up the road about 700 feet to sharp left turn. Follow road past turn about 750 feet to Hale/Noanet Quarry located northwest of road.

The Hale or Noanet Quarry is an outcrop that originally contained a dike of fine-grained felsite. Prehistoric quarrying appears to have removed most of the felsite that filled the dike, leaving an opening about 6m in length and 1.2 to 1.4 m in width. Initial testing of the site by an avocational archaeologist yielded a fragment of a Wayland Notched projectile point and bifacial tool blades diagnostic of the Late Archaic period Susquehanna tradition. These artifacts are about 3600 to 3300 yrs old. The quarry may have only been used by people of this tradition which is known for its lithic technology based on good quality rhyolite.

Based on more recent study, it was estimated that the dike at this quarry originally contained about 9.75 cubic m of felsite. Two test pits excavated in a workshop area within 10m radius of the dike revealed dense deposits of debitage and workshop debris up to 1m in depth adjacent to the dike and about 50cm thick further away from it. The density of debitage was about 59,000 pieces per sq meter in the area within 5m of the dike. A few partially completed bifacial tool blades, cores or fragments with evidence of flake removal were also recovered from the workshop. Analysis of the size distribution of debitage showed that small flakes in the 0-1 cm and 1-3 cm size range were common in the workshop debris. This suggests that activity was not restricted to production of large preforms or flake blanks and some reduction of large flakes to bifacial tool blades or projectile points also took place at this quarry (Strauss and Hermes, 1996).

#### **Mileage (cont.)**

Return by same route described above via Carby Street, Dover Street, Route 109 (north) to Route 93 (Route 128) and continue east on Route 93/128.

32.5 Take exit for Route 95 South (Providence).

53.0 Take exit for Route 295, continue west on 295.

55.3 Take Exit for Route 1 (South). Continue on Route 1 past Emerald Square Mall on right. Turn right on Cumberland Avenue.

56.5 Turn right into parking lot for store (Christmas Tree Store). Walk across Cumberland Avenue to parking lot or stores on west side of this street. From rear of parking lot, enter wooded area and walk uphill.

#### **STOP 4 (1 hour)**

This stop is in the western half of the arc shaped zone containing the Wamsutta volcanic complex. The other half of this zone extends east and north of Route 1 toward Walnut Grove Hill, Todds Pond and Manchester Pond Reservoir. While hiking through the woods to the field trip location, some outcrops and or boulders of the red/purple Wamsutta sandstone and other sedimentary rock can be seen. A dike of diabase or basaltic material was also

observed in an earlier visit to this location. Outcrops of dark red to purple felsite are located on the upper slopes and crest of the unnamed hill west of Cumberland Avenue. From cursory examination, felsite outcrops do not appear to have been the object of prehistoric quarrying since talus slope deposits contain a large amount of easily obtained material. The western and southern slopes of the hill contain talus deposits filled with numerous blocks and some platy/tabular fragments with fracture plane surfaces. These talus blocks most likely provided a majority of the raw material for prehistoric tool making. In particular, the tabular or slab shaped pieces bounded by fracture plane surfaces would have made convenient blanks for the manufacture of preforms and bifacial tool blades.

The primary activities carried out by Native American groups visiting this lithic source area were probably procurement or selection of raw material and preliminary shaping of selected pieces into bifacially flaked preforms or tool blades. They may have excavated shallow pits or trenches into talus deposits to find blocks or slabs that were suitable for making chipped stone tools. One aspect of the procurement process was probably the selection of talus blocks without cross-cutting fracture planes or joints that would cause them to break while flaking them into preforms. Pieces of felsite talus were probably tested by removing a few flakes, before expending more effort on them. Any unsuitable material or preforms broken in the early stages of manufacture would have been discarded at the quarry/workshop site.

The Wamsutta volcanic source area is likely to contain loci of quarrying activity and associated workshops created over thousands of years. Diagnostic artifacts from the Early Archaic (ca 9000-7500 years ago) to Late Woodland (ca 1000-450 years ago) period made of this felsite have been found on archaeological sites throughout the Taunton basin and Narragansett Bay area, and to the northwest along the Blackstone River drainage.

**Mileage (cont.):**

56.8 Return to route 1 north; take left onto route 1 at stop light.

58.4 Take exit to I295 south toward Woonsocket, RI.

62.5 Cross Massachusetts-Rhode Island state line.

81.1 Take exit 6A east (Route 6 east) toward Providence

83.7 Take Killingly Street Exit (Route 128), bear right at end of ramp.

84.0 Take right onto Hartford Avenue,

84.2 Pull into the Color Lith parking lot at 805 Hartford Avenue. The National Register quarry site is located immediately in front, behind the chain link fence. Access to the quarry site is afforded by a small gate.

**Stop 5 (30 minutes)**

The Johnston or Ochee Spring soapstone quarry is situated less than a kilometer (km) west of the Woonasquatucket River in eastern Johnston. The Woonasquatucket River meanders east from this point to empty into Providence Harbor, located at the head of Narragansett Bay, approximately 5.5 km east of the quarry site. Flowing waters and large water bodies likely assisted Terminal Archaic peoples in the transport of heavy and bulky vessels to and from processing camps, residential locations, trading locales, and other places.

The Ochee Spring soapstone was quarried by Native Americans for much of the Terminal or Transitional Archaic Period (ca. 3700-2700 B.P.). Individuals would patiently shape the bowls in the ledge outcrop using quartz, quartzite, and basalt picks and stone hammers. After initial rough shaping of the bowl's outline and exterior, blanks were undercut and pried from the ledges. Bowl blanks were then removed to campsites, located not far from quarry sources, where they were pecked, hollowed, and finished. Unlike several of southern New England's other noted soapstone quarries, there has been no evidence to suggest Ochee Spring soapstone was quarried for use in later smoking pipe manufacture.

A2-20

***HERMES, RITCHIE, WALLER, AND McBRIDE***

Currently, the Ochee Spring quarry site exists as a small soapstone ledge limited to an approximate ¼-acre in size. Undoubtedly, the actual quarry area was much larger at one time. The Committee of the Rhode Island Historical Society reports that “about three hundred horse cart loads” of soapstone waste debris and quarrying artifacts were taken from the site area following its discovery in 1878 (Denison et al., 1879). Noted archaeologist F.W. Putnam also reported, upon a visit to the site in 1880, that numerous chisels, picks, and hammer stones were present. Many of these materials ended-up in local or regional museums, historical societies, or private collections. Presently, the soapstone ledge is exposed roughly two to five feet above ground surface and is partially covered by vegetation. Boyd Dixon’s (1987) surface study of the exposed ledge resulted in the documentation of roughly 60 soapstone bowl impressions or bowl removals. The entire range of soapstone bowl manufacture from initial shaping to the removal of the bowl blanks is preserved at this soapstone outcrop. Commercial development, installation and upkeep of local transportation networks, and late nineteenth century antiquarian mining for artifacts has drastically reduced the Ochee Spring quarry from its former size, although many of the bowl impressions and removals remain readily visible. The Ochee Spring soapstone quarry was added to the National Register of Historic Places in 1978 and remains a protected archaeological site.

**Mileage (cont).**

84.6 Retrace path by taking a left onto Hartford Avenue from parking lot. At stop light, take left onto Killingly Street.

84.8 Exit right onto 6 west toward I295 south.

86.4 Take I295 south toward Warwick.

98.2 I295 merges with I95 south; continue on I95 south.

129.2 Take CT-49 exit (CT exit 92) toward Ct-2 and Pawcatuck and North Stonington.

130.0 Merge onto CT-617. Turn right onto CT-2 west, pass through one roundabout.

137.5 Turn left onto Lantern hill Road (CT-214). Continue on CT-214

137.8 Turn right onto Pequot Trail.

139.5 Enter parking lot for museum.

**Stop 6 (1.5-2 hours)**

The group will be provided with an introduction to the Mashantucket Pequot Museum and Research Center exhibits and research facilities, followed by a brief tour of highlighted exhibits including World of Ice, Life in a Cold Climate, and Reconstructing Paleo-Environments. The group also will visit the Early Reservation Period exhibit, followed by a tour of the excavations at the Monhantic Fort Site, a ca. 1675 A.D., which is the basis for the Early Reservation Period exhibit.

To travel onward to New Haven, retrace path to route 2. Take right onto route 2 and continue to intersection with I95. Take I95 south to New Haven.

# THE NEW QUATERNARY GEOLOGIC MAP OF CONNECTICUT AND LONG ISLAND SOUND BASIN PART 1-- Illustrated by a fieldtrip along the coastline

by

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The purpose of this fieldtrip is to demonstrate the glacial and postglacial deposits shown on the new Quaternary Geologic Map of Connecticut and Long Island Sound Basin authored by Janet Radway Stone, John P. Schafer, Elizabeth Haley London, Mary L. DiGiacomo-Cohen, Ralph S. Lewis, Woodrow B. Thompson, and Byron D. Stone. This map has recently been printed and copies will be available at the NEIGC meeting. The discussion below is mostly excerpted from the Quaternary map text and includes some general information about map units in Long Island Sound and along the coast as well as a discussion of chronology of ice retreat and meltwater deposition in the Long Island Sound basin. The first day of this trip will focus on Long Island Sound (LIS) geology illustrated by stops along the coast where on-land units become submerged beneath modern sea level.

## Map Units Beneath Long Island Sound

Glacial and postglacial deposits have long been studied on land in Connecticut. Similar deposits are also present beneath modern marine sediments in Long Island Sound; in fact, it is here that the thickest and most extensive Pleistocene deposits are found. Map units beneath Long Island Sound differ from those on land in two ways. First, offshore geologic units are mapped largely from analysis of seismic-reflection profile data instead of from direct observation. Secondly, due to the three-dimensional aspect of seismic-reflection data, map units in many places in the Long Island Sound basin are superposed and show vertical as well as areal distribution. Nearly all of the offshore geologic units of the Long Island Sound basin occur beneath a generally ubiquitous blanket of Holocene marine mud a few to 15 m (33 ft) in thickness which is not shown on the map (Lewis and DiGiacomo-Cohen, 2000).

The distribution of late Quaternary geologic units beneath Long Island Sound was mapped from more than 4,000 line-km of high-resolution, seismic-reflection profiles (see Figure 1 for locations) supplemented by vibracore data. These data were collected beginning in 1982 as part of an ongoing marine geologic mapping program conducted by the Connecticut Geological and Natural History Survey (CGNHS) in cooperation with the U.S. Geological Survey. Spacing between seismic lines that were collected perpendicular to the coast is about 1.6 km (1 mi) (locally 0.8 km (0.5 mi)), and about 4.4 km (2.7 mi) between tie lines paralleling the coast. Vibracores (obtained through CGNHS cooperative programs with the U.S. Minerals Management Service) and submersible and remotely operated vehicle dives (made possible by cooperative programs with National Oceanic and Atmospheric Administration's National Undersea Research Center at the University of Connecticut) provided verification of near-surface geologic interpretations made from seismic-reflection profiles.

Varved lake clays and marine muds in the Long Island Sound basin have distinctive seismic signatures and were easily differentiated with minimal sampling (Lewis and Stone, 1991; Lewis and others, 1993; Lewis and DiGiacomo-Cohen, 2000). Other geologic units (including moraine deposits, proximal and distal glaciolacustrine fan deposits, glaciodeltaic deposits, channel-fill deposits, and marine delta deposits, see Figures 2, 3, and 4) were mapped by analyzing internal-reflection characteristics of seismic units in a synergistic basinwide context. The offshore mapping was made possible by collaborative interpretation of seismic lines by glacial geologists familiar with the distribution and internal structure of terrestrial Quaternary deposits and by marine geologists familiar with the offshore seismic record. Systematic seismic-survey coverage of the entire basin (see Figure 1) was also an important factor in the offshore mapping effort because knowledge of the areal distribution of seismic units is necessary to their interpretation as specific depositional facies and correlation with terrestrial geologic units.

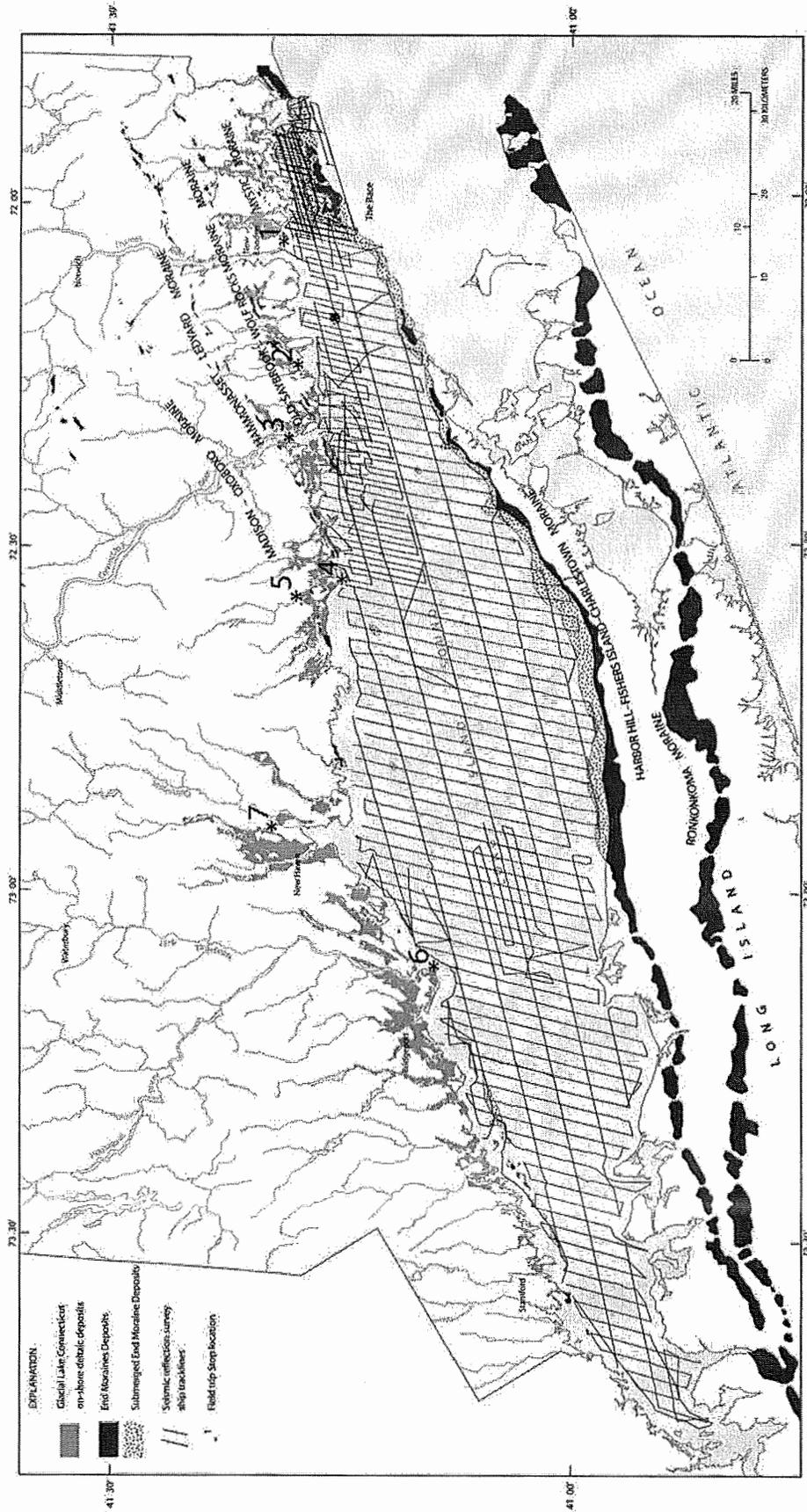


Figure 1. Field trip stop locations (1-7), location of on-shore deltas of Glacial Lake Connecticut, recessional moraines, and seismic reflection data coverage in LIS.



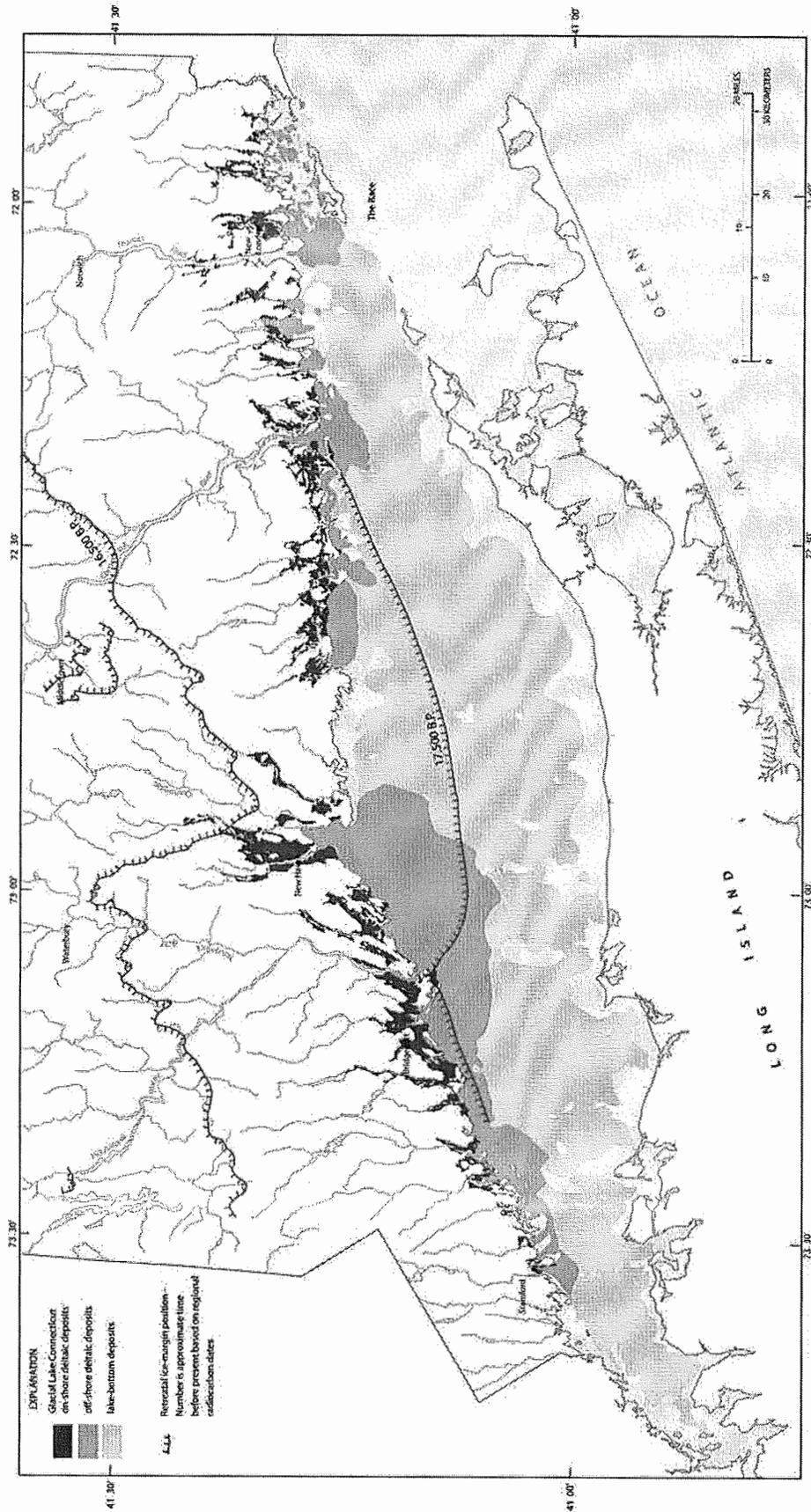


Figure 3. Deltaic and lake-bottom deposits of Glacial Lake Connecticut.

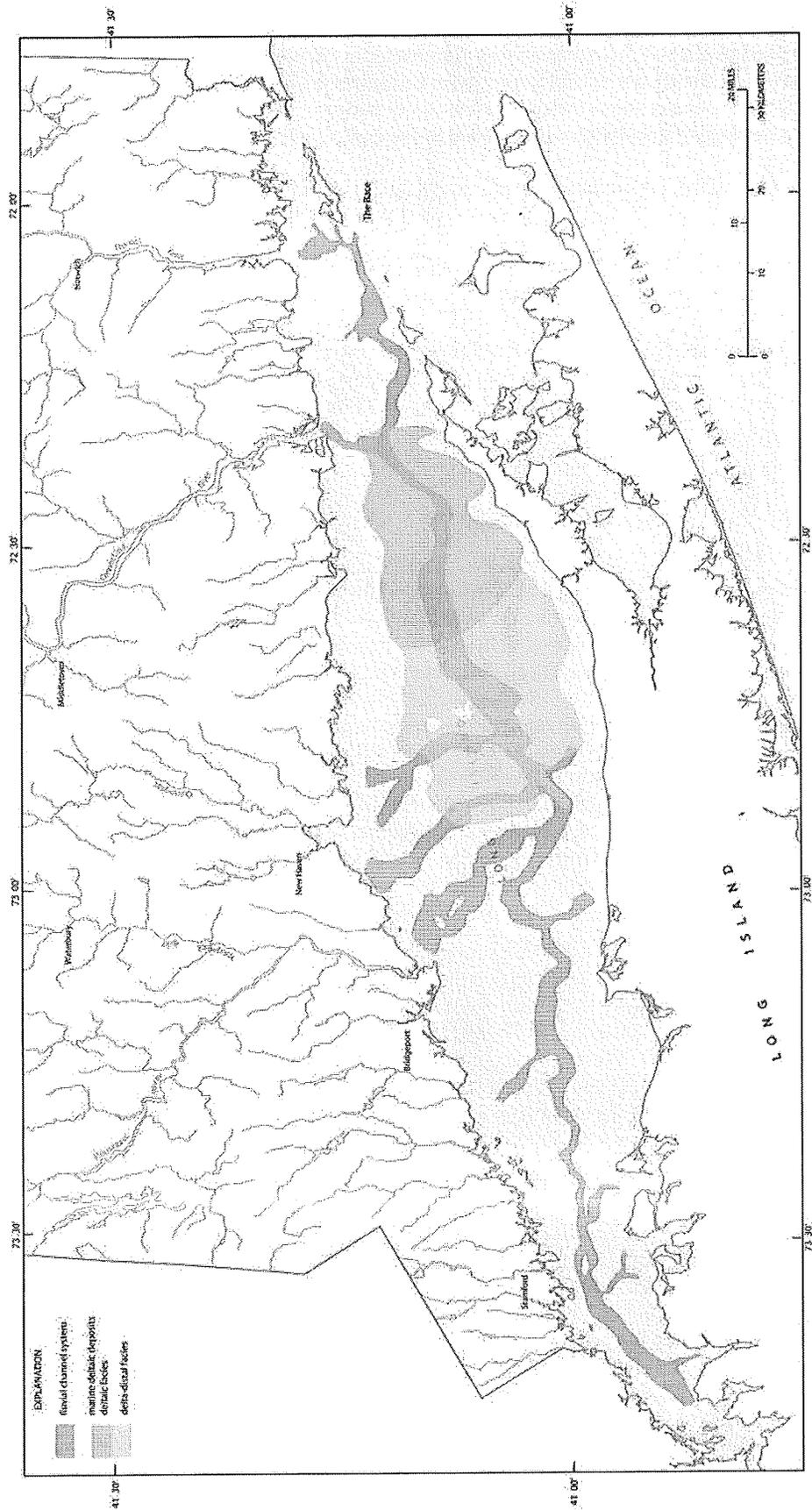


Figure 4. Fluvial channel system through which Glacial Lake Connecticut drained to the east, and the post-glacial marine delta.

### Long Island Sound Basin--Glacial Lake Connecticut

Deglaciation of the Long Island Sound basin was dominated by the presence of glacial Lake Connecticut (Figure 3), which was impounded in the Long Island Sound basin behind the Harbor Hill-Fishers Island-Charlestown moraine. Formation of the lake began at about 19 ka when the ice front began to recede from the terminal moraine position (Stone and Borns, 1986). Meltwater was impounded in the expanding long, narrow basin between the moraine to the south and the retreating ice margin to the north. Initially, when lake levels were highest, the impounded water was probably coextensive with a lake in Block Island Sound, and the whole system spilled across a notch in the terminal moraine at the head of Block Channel (Lewis and Stone, 1991). As the Block Channel spillway was eroded, lake levels lowered and, eventually, the surface of the moraine emerged in the relatively low area between Fishers Island and Orient Point. The result was the formation of a separate body of water in the Long Island Sound basin defined by Stone and others (1985) as glacial Lake Connecticut. Glacial Lake Connecticut water levels were controlled by a spillway across the lowest point on the moraine at the place today called The Race. From a relative initial spillway altitude of about -10 m (-33 ft), the lake gradually lowered by erosion at The Race to a final spillway altitude of about -60 m (-197 ft); the rate of this erosion was controlled by the rate of lowering of the lake in Block Island Sound to the south.

Systematic northward retreat of the ice margin through the Long Island Sound basin is recorded by sequential ice-marginal lacustrine-fan deposits (Figure 2) built on the lake bottom by meltwater that issued from tunnels at the grounding line of the ice. The proximal parts of the largest of these lacustrine-fan deposits are linear, ice-margin-parallel features deposited in deeper water areas along the trends of submerged extensions of the recessional moraines of coastal Connecticut.

Deltas deposited in glacial Lake Connecticut (lcms, lcp, lcjl, lcn, lcwoo, lchm, lcew, lcnh, lcnh, lcdm, lcl, lcss, lcsnw) exist along the Connecticut shoreline near the mouths of most rivers entering Long Island Sound. Delta topset-foreset contacts are paleo-water-level indicators of the lake; they occur at altitudes of 0 to 10 m (0 to 33 ft) above present sea level in the coastal deltas and indicate that, at its highest levels (when the ice margin stood at positions near the present Connecticut shoreline), glacial Lake Connecticut occupied all of the Long Island Sound basin and extended into the present river estuaries. When their present altitudes are adjusted to the regionally established glacio-isostatic rebound slope of 0.9 m/km (4.74 ft/mi) to the N. 21° W. (Koteff and others, 1988), coastal deltas record slowly lowering lake levels during their sequential construction. Because of its east-northeast trend, the ice margin first retreated onto higher ground out of the lake basin in the southeast corner of Connecticut (east of the mouth of the Thames River). Meltwater flowed from onshore ice-margin positions down bedrock valleys and built deltas into the lake near the present coastline; these deltas record the highest lake levels, which were 0 to about 2 m (0 to about 7 ft) below present sea level, which in turn projects to The Race spillway (TRS) at about -9 to -10 m (-30 to -33 ft) altitude.

By about 17.6 ka, the ice sheet had established a recessional position just offshore of the present western Connecticut coast, along the Norwalk Islands moraine and its submerged eastward extension. This timeframe is indicated by correlation of the Norwalk Islands moraine with morainal positions in New York and New Jersey (Stone and Borns, 1986). Eastward along this trend, the ice margin stood at the head of an emergent delta at Lordship (Stop 6) that marks an inter-lobate angle. Lobation of the ice front southeastward from Lordship, across the deep offshore extension of the Hartford basin, is marked by an extensive lacustrine-fan deposit (Lewis and others, 1988); this fan deposit appears to correlate with a line of less extensive fans and submerged moraine segments that trend northeast and pass onshore at Old Saybrook. Deltas at the mouth of the Connecticut River (lcwoo) (Stop 3) built in front of this ice-margin position indicate that lake levels had lowered only a few meters from the initial spillway altitude of -10 m (-33 ft) by this time.

Further ice retreat progressively deglaciated shoreline areas west of Old Saybrook. A recessional position, marked by the Hammonasset-Ledyard moraine line (hlm) that passes offshore at Hammonasset Point in Madison (Stop 4), appears to correlate with a line of lacustrine fans south of New Haven that mimics the lobate shape of the Lordship lacustrine-fan deposit and lies about 5 km (3 mi) to the north of it. West of Milford, the ice front also retreated out of the lake. Deltas were built into the lake directly in front of the Hammonasset moraine (hlm) and the slightly younger Madison moraine (mom) to the northwest. At this time, west of Milford, meltwater streams flowed southward down bedrock valleys and built deltas into the lake at altitudes that indicated that the lake had lowered about 5 to 7 m (16 to 23 ft).

As deglaciation progressed, the topography of the Central Lowland produced a lobe of ice extending southward from New Haven that lingered the longest in the lake. As this ice lobe retreated out of the lake, a complex of deltas

was built northeast of Milford in West Haven, New Haven (Stop 7), and East Haven (lcmd, lcnh, lcnh); the lake had lowered about 10 m (33 ft) by this time. These deltas are extensive both on land and offshore. In the deep central part of the lake basin, much of the varved lake-clay section, which is commonly 100 m (328 ft) in thickness, settled out in the lake concurrent with deposition of the New Haven area deltas; their stratigraphic equivalence is clearly indicated by continuous internal reflectors on seismic-reflection profiles. Onshore evidence indicates that this delta building took place during the time of ice retreat from the Lordship position to about 16 km (10 mi) north of New Haven (see 16.5 ka ice position on Figure 2). Current estimates of deglacial chronology suggest that this period was perhaps on the order of 1,000 years.

A much thinner section of varved lake clay overlies the section that is stratigraphically equivalent to the New Haven deltas and provides evidence that the lake continued to exist as the ice margin retreated northward; however, the coarse-grained sediment supply to the lake was largely cut off when the ice margin left the shorter drainage basins feeding the lake and when new glacial lakes trapped much of the sediment in larger drainage basins to the north. During this time, the level of glacial Lake Connecticut continued to lower due to erosion at the spillway. As shoreward portions of the lakebed were subaerially exposed, streams locally entrenched older, higher level delta deposits and redeposited coarser material farther out into the lake basin; this material is seen in places at the very top of the varved lake-clay section. The gradually shrinking glacial lake may have lasted another 1,000 years; this seems a reasonable estimate for the duration of varve deposition in the upper lake section. If so, by about 15.5 ka, the lake was completely drained and the lakebed was subaerially exposed.

### **Long Island Sound Basin-- Postglacial deposits**

In the Long Island Sound basin, significant postglacial events include the drainage of glacial Lake Connecticut and subsequent sea-level rise. The remnant glacial lake was probably completely drained by 15.5 ka and a fluvial channel system (Figure 4) was being carved on the lake floor by meteoric streams flowing to the south in coastal Connecticut, and to the north on the north shore of Long Island; these tributary channels joined a major east-west-trending trunk channel which also received distal meltwater drainage from the Hudson River valley to the west (Stanford and Harper, 1991). The channel system exited the basin through the lake-spillway notch in the end moraine at The Race and provided a path through the moraine for the subsequent transgression of the sea from the south (Lewis and Stone, 1991). Minor fluvial sediments were deposited in the bottoms of the channels during the time that they were occupied by streams. The channels are filled predominantly with estuarine sediment (ch) deposited as the early postglacial sea flooded these low-lying areas of the drained lake basin when eustatic sea level began to rise significantly between 16 and 15 ka (Fairbanks, 1989; Bard and others, 1990) and before glacio-isostatic rebound began.

A major wave-cut marine unconformity was cut across the top of the estuarine channel fill and over higher lake deposits as sea level rose. The marine unconformity is present in seismic sections up to altitudes of about -25 m (-82 ft), indicating that sea level probably rose to this height in central Long Island Sound before crustal rebound began.

A figure on sheet 2 of the Quaternary map shows a conceptual relative sea-level curve for central Long Island Sound. The curve was derived by combining the eustatic sea-level curve from Barbados (Fairbanks, 1989; Bard and others, 1990) with a curve representing the timing and total depth of glacio-isostatic depression in central Long Island Sound. The uplift curve is based on several assumptions (rebound began shortly after 16,500 c.y.a.; maximum depression in LIS was about 100 m) that are indicated from regional evidence, some of which is presented in this report and in Koteff and Larsen (1989) and Stone and Ashley (1995). The presence of the extensive marine delta that records a -40-m (-131-ft) to -50 m (-164 ft) relative sea level in central Long Island Sound (Lewis and Stone, 1991; Stone and Lewis, 1991) provides good evidence for the conceptualized relative sea-level curve. The large volume of deltaic sediment required a significant length of time for construction. The topset-foreset contact of the marine delta indicates that sea level was relatively (at -40 to -45 m (-131 to -147 ft)) stable during the deposition of the delta.

The only possible source of the great volume of sediment contained within the marine delta was the drained lakebed of glacial Lake Hitchcock in the Connecticut Valley to the north. This sediment supply became available only when the "Stable Phase" of glacial Lake Hitchcock ended at about 13.5 ka; as previously discussed, glacio-isostatic uplift had to occur in order for glacial Lake Hitchcock to drain. Regional evidence from northern New England (Barnhardt and others, 1995; Koteff, Robinson, and others, 1993; Koteff, Thompson, and others, 1995) also indicates that isostatic rebound began around this time. Thus, the early rapid rate of uplift was balanced with the equally rapid rate of eustatic sea-level rise resulting in a sea-level stand in Long Island Sound at about -40 m (-131 ft) for several

thousand years (between 13.0 and 9.5 ka). During that time, the marine delta was built and the Connecticut River terrace and flood-plain surfaces were incised. Recently obtained <sup>14</sup>C dates (9,370±100 Beta-52257, 8,530±80 Beta-52256) on basal organic material beneath the lowest terrace surfaces along the Connecticut River in Massachusetts indicate that most of the postlake incision into the lakebed had been accomplished by ~9.0 ka (Stone and Ashley, 1992). The volume of eroded lakebed sediment, as calculated from the area and depth of incised terraces, is 12 billion m<sup>3</sup>; this material now composes the marine delta, the calculated volume of which is 11.5 billion m<sup>3</sup>.

As eustatic rise overtook the rate of isostatic rebound, relative sea level in central Long Island Sound rose continuously; the transgression submerged the marine delta and a blanket of marine mud accumulated over the entire basin. As marine waters deepened, intense tidal-scour conditions developed in eastern Long Island Sound, resulting in the local reworking of marine-delta sediments and the development of a very large sand-wave field in that part of Long Island Sound (Fenster and others, 1990). A record of 4 to 5 m (13 to 16 ft) of sea-level rise during the last 4,000 to 5,000 years is preserved in coastal salt-marsh deposits (Bloom and Stuiver, 1963; van de Plassche and others, 1989; Patton and Horne, 1991; van de Plassche, 1991).

## ROAD LOG

Our trip begins at the "office" - the Marine Sciences building on the University of CT campus at Avery Point in Groton, which offers views of bedrock outcrops, recessional moraines, and ice-marginal deltas built into Glacial Lake Connecticut. Exposures are rare and fleeting in this part of southern New England, so fieldtrip stops along the Connecticut shoreline will mostly illustrate the morphology of recessional moraines and ice-marginal deltas as they extend offshore and locally are aligned with subaqueous lacustrine fans deposited at the grounding line of the ice-margin as it retreated sequentially northward in glacial Lake Connecticut. Seismic records and photographs of cores from LIS will help illustrate the character and distribution of extensive varved lake clay that overlies morainal and lacustrine fan deposits and interfingers with glaciodeltaic deposits along the Connecticut coastline. We will discuss these features as well as the postglacial chronology at additional stops along the coast: Rocky Neck State Park in Niantic, the Connecticut River in Old Lyme, Meigs Point at Hammonasset State Park in Madison, Lordship Point in Stratford, and East Rock in New Haven.

### **STOP 1. MARINE SCIENCES AND TECHNOLOGY BUILDING, AVERY POINT CAMPUS, UNIVERSITY OF CONNECTICUT, Groton, CT, (Uncasville Quadrangle)**

The Avery Point campus occupies a peninsula on the east side of the Thames River where it enters Long Island Sound (Figure 1). Avalonian terrane bedrock (New London Gneiss) underlies the peninsula and is locally exposed through a thin mantle of Late Wisconsinan till. To the east, the Hope Valley Alaskite Gneiss forms the seaward end of a similar south-trending promontory and, in the intervening valley; deltaic deposits associated with the Poquonock River (lcp) underlie the tidal marshes (sm) and supply the sediments that sustain the beach. The topset-foreset contact in this large glacial delta is reported at 1-2 m (2 to 5 ft.) altitude [-11 to -12m (36-39 ft.) projected to The Race Spillway (TRS)]. Deltas associated with the Thames and Poquonock drainages extend seaward (lcd) to the vicinity of Fishers Island where they have been truncated by tidal scour. Variations on the theme of south-trending, till-mantled, bedrock promontories bracketing marshes and pocket beaches that developed on meltwater deposits, are repeated all along the Connecticut Coast. We will use this third floor vantage point to view the expanse formerly occupied by glacial Lake Connecticut and its spillway at The Race. If we have a clear day, four morainal positions of the late-Wisconsinan ice sheet (Sirkin, 1982 would argue for Gardiners Island as a fifth) will be visible. In the far southeasterly distance, we will see 1) the Late Wisconsinan terminal moraine where it forms the south fluke of Long Island at Montauk Point; 2) the Harbor Hill-Roanoke Point-Fishers Island-Charlestown recessional moraine line, along the north shore of Long Island and Fishers Island; and, slightly closer, 3) the Clumps recessional moraine where it forms a line of small islands (the Dumplings) north of Fishers Island. The small islands composed of bouldery ablation till seen to the southeast immediately offshore from the campus are Pine Island and Bushy Point, which form a segment of the Mystic moraine (Goldsmith, 1962; 1982). This is the earliest morainal position that occurs on land in southeastern Connecticut. The Mystic moraine continues westward from Pine Island as submerged segments beneath Long Island Sound, mapped using seismic reflection techniques. We will discuss the Quaternary geology beneath the Sound as shown on the recently published Quaternary Geologic Map of Connecticut and Long Island Sound Basin (U.S.G.S. Scientific Investigations Map 2784) and display some of the seismic evidence that

supports the mapping.

- 0.0 Leave parking lot D via the campus exit road
- 0.1 Turn left at end of exit road and backtrack to I-95 by following I-95 signs
- 2.0 Proceed up Eastern Point Road, past the Pfizer complex. Stay straight; as Eastern Point Road becomes Benham Road (do not follow Rte.349 to the left). Turn right onto Rainville Ave. at the next light and proceed one long block
- 2.2 At next light (Brandegge Ave.) turn left and follow I-95 SOUTH signs
- 3.8 Take left exit for I-95 SOUTH, New Haven and proceed to Exit 72 for Rocky Neck State Park (about 12.2 miles)
- 16.0 Take exit 72 (sharp curve to the right) and follow the Rocky Neck Connector to the intersection with Rte. 156
- 16.9 At light turn left onto Rte 156 and proceed to park entrance (on right)
- 17.2 Turn right into Rocky Neck State Park and proceed past ticket booths. At rotary go  $\frac{3}{4}$  around and follow road to the beach
- 18.4 Park near buildings in the far left corner of the parking areas

## **STOP 2. ROCKY NECK STATE PARK, East Lyme, CT, (Niantic Quadrangle)**

The overall coastal setting here is a very nice example of what we just saw on a larger scale at Avery Point. Rocky Neck, to the west, and Giants Neck, to the east, are south-trending bedrock [Potter Hill Granite Gneiss of the Avalonian (Rodgers, 1985) or possibly Gander? (Wintch and others, 2005) Terrane] promontories that are thinly mantled by till. They bracket and confine the fluviodeltaic deposits that choke the Bride Brook (lcn) valley. Just to the east in Niantic, the topset-foreset contact in deltas built into glacial Lake Connecticut lies at an altitude of about 1.2 m (4ft.) or -12 m (-39 ft.) projected to TRS. The north-south grain of the bedrock-controlled landscape results from the west to east accretion of terranes that occurred as the Iapetus Ocean closed (Coleman, 2005), and the subsequent rifting associated with the opening of the Atlantic Ocean. Exploitation of preferentially aligned rock units, faults and fractures by streams, and at least two glaciers, has produced a predominant pattern of south-draining bedrock valleys along the Connecticut coast. A double segment of the Old Saybrook-Wolf Rocks (owm) recessional moraine traverses the Park from southwest to northeast (Goldsmith, 1964). To the northeast, the double aligned segments of this moraine belt can be traced into Rhode Island. To the southwest, the moraine crosses the mouth of the Connecticut River and extends offshore at Cornfield Point in Old Saybrook. We have correlated the Old Saybrook-Wolf Rocks moraine with a large lacustrine fan offshore of Stratford Point (Stop 6) and the with the Captain Islands-Norwalk Islands (cnm) moraine farther to the west (17,500 B.P. position, Figure 2). Immediately offshore of the beach, the seismic data show a nicely developed delta (lcd) that built into glacial Lake Connecticut as meltwater issued from the ice position (owm) just to the north. A cluster of lacustrine fans, and submerged moraine segments that lie about half way across the Sound may be correlative with the Clumps (cm) moraine of Fishers Island Sound (Stop 1). The Roanoke Point-Fishers Island-Charlestown (hcm) recessional moraine is represented on the horizon by Orient Point and Plum Island.

- 19.6 Backtrack out of the Park and turn left onto Rte.156
- 19.9 Turn right onto Rock Neck Connector and follow I-95 SOUTH signs (left lane)
- 20.8 Take entrance ramp for I-95 SOUTH, New Haven and proceed to Exit 70 Old Lyme (about 4 miles)
- 24.8 Take Exit 70 for Old Lyme and proceed to traffic light at end of ramp (intersection with Lyme Street)
- 24.9 Go straight through intersection with Lyme Street, and follow I-95 SOUTH sign
- 25.5 Make a left turn at the end of the road (intersection with Rte. 156) and proceed southward under I-95 to Ferry Road
- 26.0 Turn right onto Ferry Road (sign for DEP Marine Headquarters) and proceed just past DEP Marine Headquarters Building to a sharp left turn
- 26.6 Take a sharp left turn at the end of the building and an immediate right turn into parking lot adjacent to railroad tracks

**STOP 3. DEP MARINE HEADQUARTERS, Old Lyme, CT, (Old Lyme Quadrangle)**

Core data relating to the construction of the I-95 Bridge, which spans the Connecticut River just north of this site, have been used to produce a cross-section of the underlying, 76 meter (250 ft.) deep, bedrock valley (Figure 5). We will discuss this cross-section as it relates to the geologic history of the Lower River and adjacent Long Island Sound. A remnant delta surface, at an elevation of about 25ft. (7.5m), is visible on the west side of the river. This is part of the ice-marginal delta (lcwoo) that built into glacial Lake Connecticut from the Hammonasset moraine (hlm, Stop 4 also) position that is mapped just north of the bridge. The topset-foreset contact in this delta is reported at altitudes of -13 to -15 m (-43 to -49 ft.) projected to TRS. Tidal marsh deposits (sm) that have built up as the sea encroached up the river mouth (Patton and Horne, 1991) are exposed along the walkway under the railroad bridge. The view to the south includes the mouth of the Connecticut River with Orient Point, Long Island (Roanoke Point-Fishers Island-Charlestown (hcm) recessional moraine) on the horizon. Cross-section C-C' on the "Quaternary Map" nicely illustrates the features that lie under Long Island Sound in this area. These include inferred remnants of Cretaceous strata, lake bottom deposits of glacial Lake Connecticut (lclb), channel-fill deposits (ch) associated with the draining of the Lake, the marine transgressive (ravinement) surface, the marine delta (md) that was deposited as glacial Lake Hitchcock drained down the Connecticut River valley, the erosional remnants that form the foundation for parts of Long Sand Shoal, marine mud and reworked marine deposits, evidence (sand waves up to 17 meters high) for extensive modern erosion and transport of earlier deposits (Fenster and others, 1990). Along the west side of the river, the lighthouse at Lynde Point marks the position of the Old Saybrook-Wolf Rocks (owm) moraine (seen at Stop 2 also) and the on-land portion (lcwoo) of its associated ice-marginal delta (lcd). The thalweg of the buried bedrock valley of the Connecticut River lies under these deposits at Fenwick (just to the west of the lighthouse). The seismic data indicate that this valley is slightly over 100 meters (328 ft.) below sea level just offshore. A later ice position forms the north shore of Saybrook Point. By the time the ice margin had retreated to its position just north of the I-95 bridge (Hammonasset moraine, hlm), the waters of glacial Lake Connecticut were a bit more extensive than the estuary we see here today.

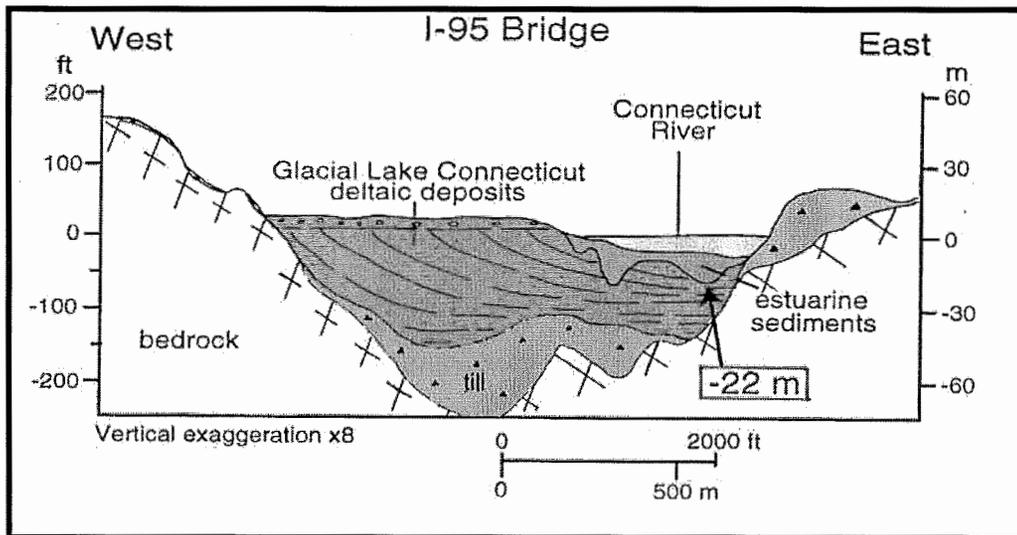


Figure 5. Geologic cross section at I-95 bridge showing the deep bedrock valley of the Connecticut River (west of the present river) filled with ice-marginal deltaic deposits, which were, in turn, incised and filled with estuarine deposits.

- 27.2 Backtrack out of the parking lot and up Ferry Road to Rte 156. Turn left onto Rte.156
- 27.6 Proceed through the traffic light and under I-95, and then make a left turn onto the entrance ramp for I-95 SOUTH
- 40.2 Stay on I-95 South for about 12.6 miles and take Exit 62 for Hammonasset State Park
- 40.3 Turn left at the end of the ramp and proceed down the Hammonasset Connector
- 41.7 At the light marking the intersection with Rte.1 proceed straight into the park entrance. Once past the ticket

- booths follow the signs to Meigs Point
- 42.7 At the rotary keep left  $\frac{3}{4}$  of the way around and exit to the right, following the Meigs Point signs
- 43.9 Proceed on the park road past the Meigs Point Nature Center and around a sweeping curve to the right (Hammonasset Point). At the end of the curve, park in the lot that is just before the entrance to the car-top boat launching area

**STOP 4. HAMMONASSET POINT, HAMMONASSET STATE PARK, Madison, CT, (Clinton Quadrangle)**

A succession of ice-marginal, fluviodeltaic deposits bury most of the bedrock surface between the Hammonasset-Ledyard (hlm) and Madison-Oxoboxo (mom) moraine positions. These deposits (lchm) are associated with the Hammonasset River and Menunketsuck River, and were emplaced as the ice front melted back to its Madison-Oxoboxo position. Delta surfaces that were graded to glacial Lake Connecticut have surface elevations of about -14m (-46 ft.) projected to TRS. East and north of the road to Hammonasset Point, extensive tidal marshes (sm) have developed on the surface of these deposits. The Hammonasset Point segment of the Hammonasset-Ledyard (hlm) moraine is a northeast-southwest trending belt of paired, parallel, linear till bodies that are partially buried by the surrounding meltwater deposits. Wave action along the seaward flank of the moraine belt has created till exposures and a beach composed of a bouldery lag deposit. From the observation deck and other vantage points on the moraine we can see its northeast trend across Clinton Harbor. The moraine continues as a belt of double linear features for about 10km (6miles) northeastward to Westbrook, CT. From there its trace is more discontinuous as it crosses the Connecticut River in Old Saybrook (Stop 3) and continues northeastward through Ledyard, CT and toward Rhode Island (possibly the Queens River moraine). A southwestward-trending seaward extension of the Hammonasset-Ledyard moraine (osm) has also been detected offshore from Hammonasset Point. To the southwest, a series of submerged and possibly related moraine segments (osm) and lacustrine fans (lcf) trend toward Charles Island (hlm) in Milford, CT. As seen from the observation deck, the submerged moraine segment is marked by a buoy and trends southwestward toward Falkner Island (tt). Looking to the north, we can see the extent of the tidal marsh (sm) deposits, and a small parallel ridge of till that is the northerly component of the double moraine belt. To the south, the Horton Point (Long Island) segment of the Roanoke Point-Fishers Island-Charlestown (hcm) recessional moraine lies on the horizon. Offshore seismic data reveal the internal structure of the marine delta (md, Stop 3 also) overlying the marine transgressive (ravinement) surface.

- 47.5 Backtrack out of the Park and proceed up the Hammonasset Connector to the interchange with I-95. DO NOT TAKE I-95 but proceed straight over the highway where the connector becomes Duck Hole Road.
- 49.2 Follow Duck Hole Road to Dairy Hill Road and make a right turn onto Dairy Hill Rd.
- 49.4 Turn right into pit area

**STOP 5. GRAVEL PIT IN HAMMONASSET RIVER DEPOSITS (hr) OFF DAIRY HILL RD., Madison, CT, (Clinton Quadrangle)**

This pit exposes bedrock (Middletown Formation) and overlying delta foreset and topset beds of the Hammonasset River (hr) deposits. These deposits were ponded behind the head of the Hammonasset River-Menunketsuck River (lchm) deposits as the ice front retreated up the Hammonasset River valley. In this area delta surfaces reach approximately 18m (60 ft) and are located just north of an ice-margin position (beds have the potential to be slightly collapsed). This exposure is representative of the SP Depositional System (Related Series of Sediment-Dammed Ponds) which is colored brown on the Quaternary Map.

- 51.3 Backtrack to I-95 and make a left turn onto the entrance ramp for I-95 SOUTH
- 86.6 Go about 35.3 miles on I-95 South to Exit 32. At the end of the ramp turn left
- 86.7 Make an immediate left and go under I-95 to a traffic circle. Follow the Lordship signs
- 87.0 Go to the left at the traffic circle, onto West Broad St., and follow the signs to Lordship/Short Beach
- 87.2 Turn right on Main Street, Rte 113, and follow the signs. (Note at the intersection with Rte. 130 stay straight on Rte 113 and at approximately mile 88.5 stay straight on Rte.113 at the light)

- 89.3 As you go around this sharp curve to the right you are passing over the ice-contact slope at the head of the Lordship deposits (lcl)
- 89.9 Make a left turn onto Prospect Street at the light. There is a church across from this turn
- 90.7 Proceed down Prospect to a left turn into the Lordship Coastal Grasslands Management Area (Just before the lighthouse)

#### **STOP 6. STRATFORD POINT, Stratford, CT, (Milford Quadrangle)**

The Lordship deposits (lcl) are the terrestrial component of a large ice-marginal delta that extends seaward (lcd) from the Lordship section of Stratford, CT. The topset-foreset contact in this delta is reported (Hokans, 1952) at an altitude of -12m (-39ft.) projected to TRS. Ice contact slopes, like the one we drove across on the way in, can be traced along the northwestern and northeastern sides of the delta. These slopes mark the interlobate angle between the western Connecticut ice margin and a stand of the Connecticut Valley ice lobe. The western Connecticut ice margin stand can be traced westward to the Captain Islands-Norwalk Islands (cnm) moraine (17,500 B.P. position, Figure 2), and evidence of the Connecticut Valley lobe stand is inferred to lie offshore in the form of the large, submerged, ice-marginal lacustrine fan deposit (lcf) that extends southeastward from Lordship and has been correlated with the Old Saybrook-Wolf Rocks moraine (own, Stops 2 and 3 also). This large lacustrine fan can be traced along a gentle arc, 17.3 km (10.75 miles) seaward (southeastward) from Stratford Point. Its main body averages slightly less than 3 km (1.4 miles) in width and is commonly over 65 m (213 ft.) thick. Further to the south, the offshore seismic data reveal the sequential northward retreat of the ice margin in the form a shingled series of lacustrine fans that are progressively younger to the north. The southward transition from submerged delta foreset facies (lcd) into distal lake bottom deposits of glacial Lake Connecticut, and the channeling associated with the draining of the Lake, are also nicely represented on the seismic records. The Roanoke Point segment (in the vicinity of Riverhead, Long Island) of the Harbor Hill-Roanoke Point-Fishers Island-Charlestown recessional moraine can be seen on the horizon to the southeast. Port Jefferson, Long Island lies to the south.

- 94.2 Backtrack over the same route toward I-95 NORTH (right onto Rte. 113 from Prospect and follow the Rte 113 to a left turn onto West Broad St.)
- 94.4 Proceed .2 of a mile on West Broad to the entrance ramp for I-95 NORTH. Turn right onto ramp at traffic circle
- 111.2 Drive about 16.8 miles on I-95 North to Exit 48 for I-91 NORTH (left lane)
- 112.0 Proceed north on I-91 to Exit 3, Trumbull Street (right exit). Follow the exit ramp to the first traffic light
- 112.7 At the first light (Orange Street) turn right and follow Orange Street to the end.
- 113.9 At the end of Orange Street turn left and follow the summit signs for East Rock summit
- 114.9 Follow the winding road for about one mile to a right turn for the summit
- 116.1 Continue on access road to a fork in the road, with a large clearing ahead. Make a sharp right turn, keeping the clearing and communication tower to your left
- 116.4 Park to the right of the monument

#### **STOP 7. EAST ROCK PARK, New Haven, CT, (New Haven Quadrangle)**

At the monument at East Rock we are atop an outcrop of the West Rock Dolerite (diabase), at an altitude of about 110 m (360 ft.), overlooking the city of New Haven, CT. From this vantage point (on a clear day) the central part of the Long Island Sound basin, which was formerly occupied by glacial Lake Connecticut, and the north shore of Long Island, in the vicinity of Port Jefferson and Wading River, are clearly visible to the south. To the southwest, smooth crests of south to southwest-trending drumlins at altitudes as high as 107 m (350 ft.), which lie on the irregular crystalline bedrock surface west of the Mesozoic lowland, can be seen. The broad New Haven delta plain, at about 14m (45 ft.) altitude in downtown New Haven, extends from west of Yale Bowl (seen to the west) to Fair Haven (seen to the south-southeast). This delta was graded to glacial Lake Connecticut, and at Yale Bowl and Fair Haven the altitude of the topset-foreset contacts are reported to be at 7 to 9 m (22 to 30 ft) which projects to -18 to -20 m (-59 to -66 ft.) at TRS. The western scarp of the Mill River meltwater terrace is incised into the New Haven delta plain and is just behind the row of three churches on the green in downtown New Haven. To the east and

northeast, the extensive salt and freshwater marshes of the Quinnipiac valley are dotted by ponds that fill abandoned brickyard pits in the Quinnipiac clay. The Fair Haven part of the New Haven delta overlies a deep bedrock valley. The Quinnipiac River flows between the eastern valley wall and this part of the New Haven delta; the original extent of the delta at Fair Haven dammed glacial meltwater in the Quinnipiac basin to the north, thus impounding glacial Lake Quinnipiac. Striations and grooves preserved in the glacially polished diabase on top of East Rock record southwesterly ice flow. Thirty-six feet northwest of the stone block wall that is west of the monument, grooves 1-2 cm wide and 20 cm long trend S19°W. Thirty feet further northwest, grooves 1-2 cm wide and 30 cm long trend S32°W. The size of the submerged lacustrine fans (lcf, Stop 6) and deltas (lcd) that lie offshore of New Haven, and the reddish tint that pervades lake bottom deposits throughout the Long Island Sound basin, provide evidence that the Mesozoic basin was a major source of the huge volume of sediment that was delivered to glacial Lake Connecticut during the retreat of the Wisconsinan glacier. Coastal Connecticut was deglaciated from east to west between Stonington and New Haven. When projected to the spillway at the Race, recorded topset-foreset contacts in deltas built into glacial Lake Connecticut show a progressive drop in elevation westward [about -12 m (39 ft.) near Stop #1 at Avery Point to about -20 m (66 ft.) in the youngest delta at Yale Bowl]. We take this as good evidence that glacial Lake Connecticut drained away slowly (through the channel system (ch) shown Figure 4) as the spillway at the Race was eroded. This scenario is consistent with the fact that there was almost no gradient in the stream systems of southern New England and the adjacent exposed continental shelf until post-glacial rebound initiated southward tilting of the area. This occurred after glacial Lake Connecticut had drained.

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# JURASSIC CYCLOSTRATIGRAPHY AND PALEONTOLOGY OF THE HARTFORD BASIN

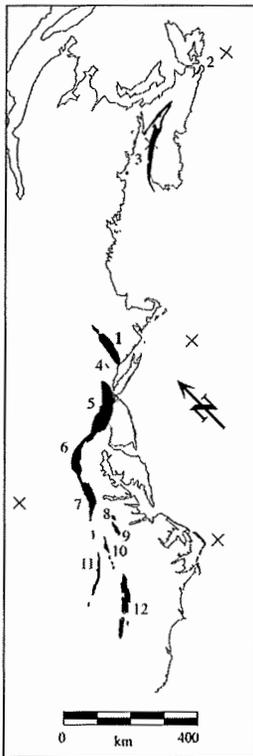
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## INTRODUCTION

Jurassic age lacustrine strata of the Hartford rift basin (Figs. 1, 2) are profoundly cyclical. Interbedded with the three laterally extensive basalt flow sequences of the Central Atlantic magmatic province (CAMP; Marzoli, 1999), the Shuttle Meadow and East Berlin sedimentary formations show a dominance of the familiar ~20 ky climatic precession cycle as well as the ~100, ~400, and longer eccentricity cycles that are strongly influenced in frequency by the chaotic behavior of the Solar System. Less well-studied, the lower 2 km of strata of the 4 km-thick Portland Formation immediately above the basalts are also cyclical.

In this guidebook, we describe and interpret the cyclicity and biofacies through most of the major Jurassic intervals in the south-central Hartford basin. We use the permeating Milankovitch cyclostratigraphy as the basis for an astronomically calibrated time scale for the Hettangian and possibly early Sinemurian ages, placing into environmental and temporal context the rich paleontological and biostratigraphically useful assemblages in the aftermath of the Triassic-Jurassic mass extinction and the great CAMP flood basalt eruptions.



## ENVIRONMENTAL, GEOLOGICAL, AND BIOLOGICAL CONTEXT

With its nearly symmetrical meridional supercontinent, Pangea, the Triassic and Early Jurassic represent an extreme end member of Earth's geography and climate. With no evidence of polar ice (Frakes, 1979), this "hot house" world is marked by coal deposition in polar and equatorial regions and plausibly extremely high  $p\text{CO}_2$ , as inferred from soil carbonate data (Wang et al., 1998; Ekart et al., 1999; Tanner et al., 2001) stomatal density trends (McElwain et al., 1999; Retallack, 2001), and geochemical models (Beerling and Berner, 2002). Despite vast climate differences from the present, a humid equatorial zone of modern dimensions (Kent and Olsen, 2000; Kent and Tauxe, 2005) existed (Fig. 3). Within the transition zone between this humid region and the arid, central subtropics to the north, a series of continental rifts from Nova Scotia to North Carolina filled with lacustrine and fluvial strata of the Newark Supergroup during the crustal extension that ultimately led to the fragmentation of Pangea. The Hartford basin is one of the largest segments of these extensive central and north Atlantic margin rifts (Fig. 1).

In central Pangea, including eastern North America, continental synrift sedimentation started in the middle Permian (Olsen et al., 2002d) and continued through the Early Jurassic (Olsen, 1997). The synrift sequence is comprised of four tectonostratigraphic sequences (Olsen, 1997; Fig. 4). Tectonostratigraphic sequences (TS) are conceptually similar to marine sequence stratigraphic units in that they are largely unconformity-bound, genetically-related packages, but are controlled largely by tectonic events. Tectonostratigraphic sequence I (TS I) is median Permian in age and while known from the Fundy basin of maritime Canada and various Moroccan basins,

**Figure 1.** The Hartford basin within the Newark Supergroup. 1, Hartford and Deerfield basins; 2, Chedabucto or Orpheus basin; 3, Fundy basin; 4, Pomperaug basin; 5, Newark basin; 6, Gettysburg basin and mostly buried; 7, Culpeper basin; 8, Taylorsville basin; 9, Richmond basin; 10, Farmville and associated basins; 11, Dan River basin; 12, Deep River basin (modified from Olsen, 1997).

could also exist in the subsurface in other basins. Tectonostratigraphic sequence II (TS II) is of Middle (Anisian-Ladinian) Triassic to early Late Triassic (Early to early Late Carnian) age and although present in most Newark Supergroup basins, is not known to exist in the Hartford basin. Tectonostratigraphic sequence III (TS III), of early Late Triassic (Late Carnian through early Late Rhaetian) age, is the most widespread sequence, dominating nearly all Newark Supergroup basins, and is represented in the Hartford basin by the lower +90% of the New Haven Formation.

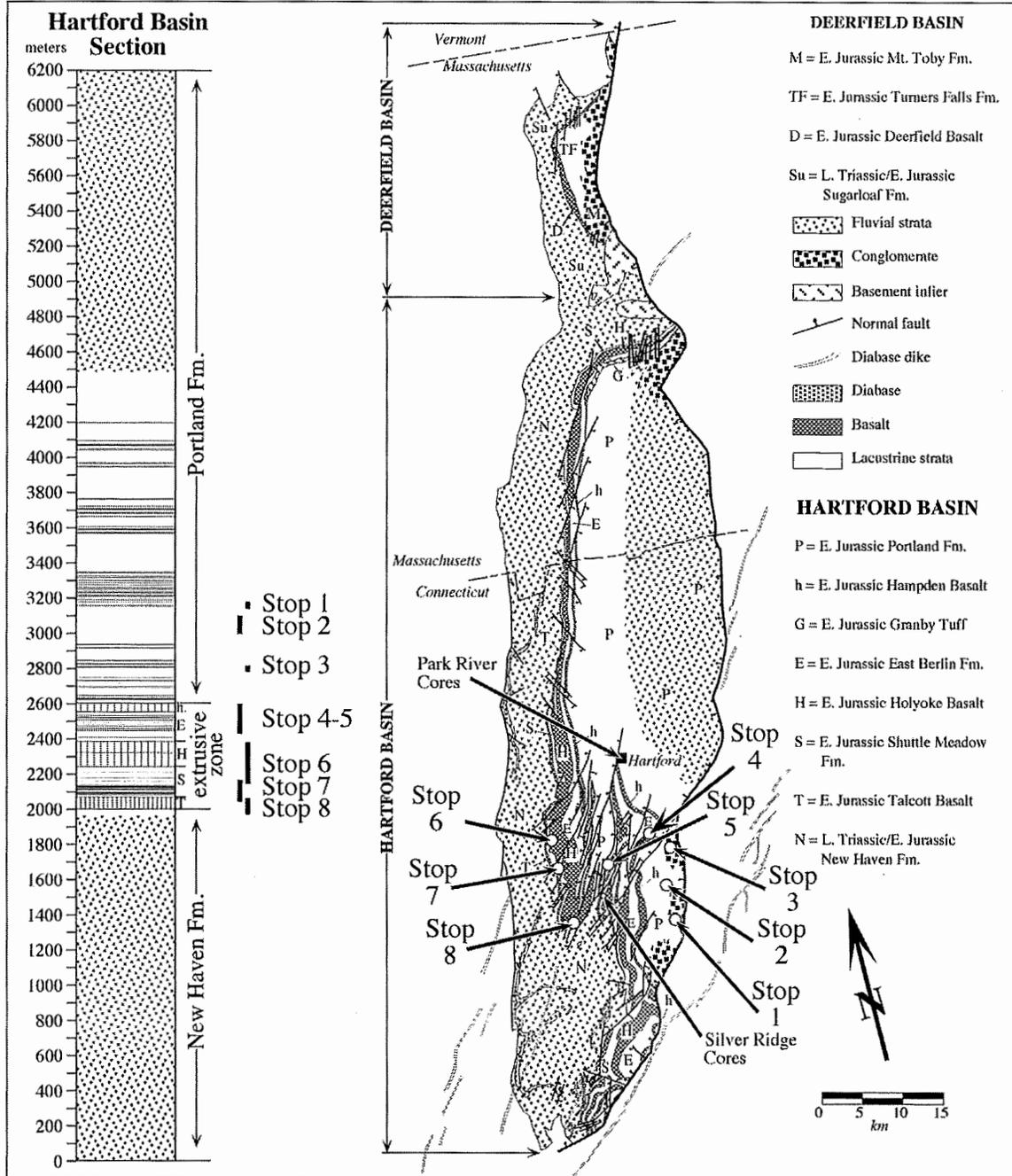
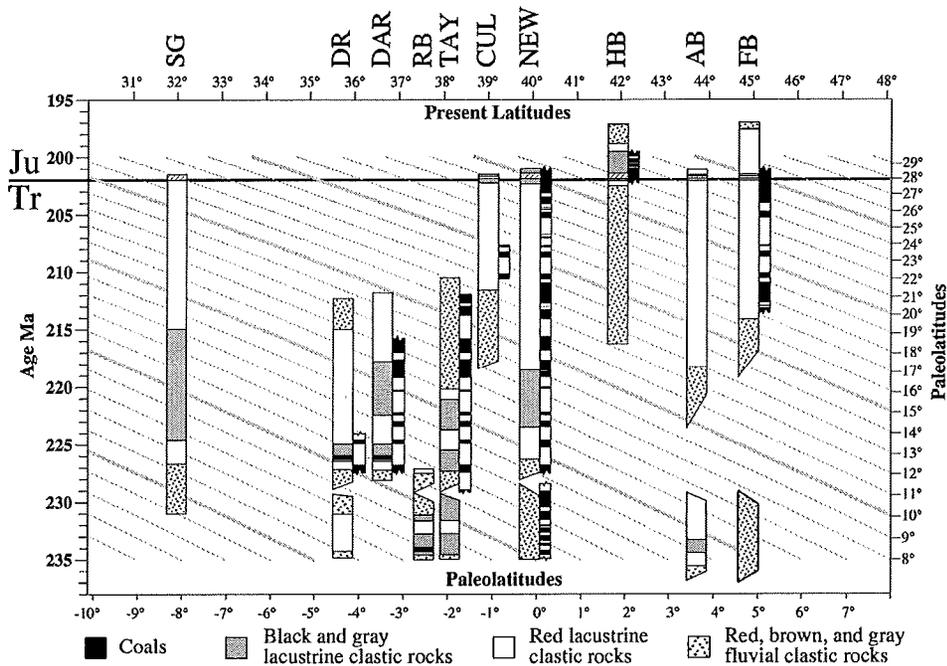
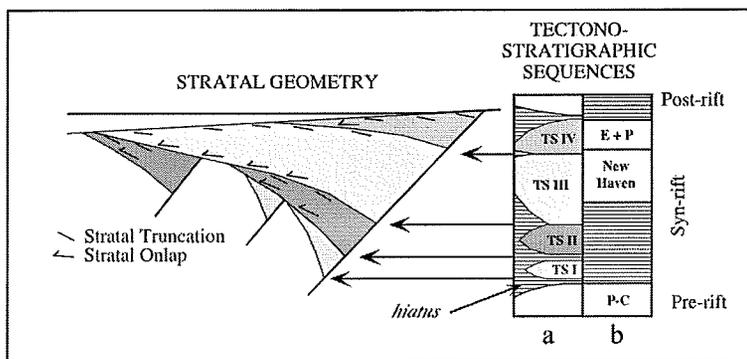


Figure 2. Bedrock geological map of the Hartford basin and basin section showing fieldtrip stops. Modified from Olsen et al., (2003a).



**Figure 3.** Time-geography nomogram showing the relationship between the main climate sensitive lithologies, age, geography, and latitude. Sections are correlated by magnetostratigraphy as indicated (black, normal; white, reverse). Slightly curved diagonal lines are lines of equal paleolatitude. Modified from Olsen and Kent (2000), Kent and Olsen (2000), and Kent and Tauxe (2005). Basins are: SG, South Georgia Rift; DR, Deep River Basin; DAR, Dan River basin; RB, Richmond basin; TAY, Taylorsville basin; CUL, Culpeper basin; NEW, Newark basin; HB, Hartford basin; AB, Argana basin; FB, Fundy basin.

Tectonostratigraphic sequence IV (TS IV) is of latest Triassic (Late Rhaetian) to Early Jurassic (Hettangian and Sinemurian) age. It contains the Triassic-Jurassic boundary, extrusive tholeiitic basalts of the CAMP, and in some basins extensive post-CAMP sedimentary strata. TS IV is very well represented in the Hartford basin where more Jurassic strata are preserved than elsewhere in eastern North America. The uppermost few meters of the New Haven Formation make up the lowest portions of TS IV, followed by the three basaltic lava formations of the basin, the Talcott, Holyoke, and Hampden basalts, intercalated with and overlain by the profoundly cyclical, lacustrine Shuttle Meadow, East Berlin, and lower Portland formations. The upper Portland Formation, still within TS IV, returns to fluvial deposits.



**Figure 4.** Tectonostratigraphic divisions of the Fundy basin. Adapted from Olsen (1997).

The Triassic and Early Jurassic was a time of evolutionary advent for terrestrial communities. Dinosaurs and mammals along with other major groups of extant vertebrates evolved during the Triassic. The crurotarsians occupied certain key ecological positions, including that of the top predators. Although modern groups had evolved, this was a world largely populated by unfamiliar taxa and with latitudinally distributed communities of different composition. Hence, the terrestrial Late Triassic was not merely an intermediate between the Early to Middle Triassic and Jurassic, it was a fully mature world unto itself, unsuspectingly diverse and unique. In contrast, the Early Jurassic was the dawn of the modern era of terrestrial vertebrate communities, exemplified in TS IV of the Hartford basin. All of the top predatory and herbivorous crurotarsians are gone. The strong biotic provinciality of the

Late Triassic has been replaced by an essentially global distribution of many genera and species. The Jurassic biological record of the Hartford basin features ecological dominance by dinosaurs with crocodylomorphs as the only other remaining archosaurs. Most of the continental tetrapod groups (e.g., lizards, turtles and lissamphibians) that survived the boundary are extant; mammals and the mammal-like tritheodonts and tritylodonts also survived; pterosaurs survived the Triassic-Jurassic boundary, but perished at the Cretaceous-Tertiary boundary.

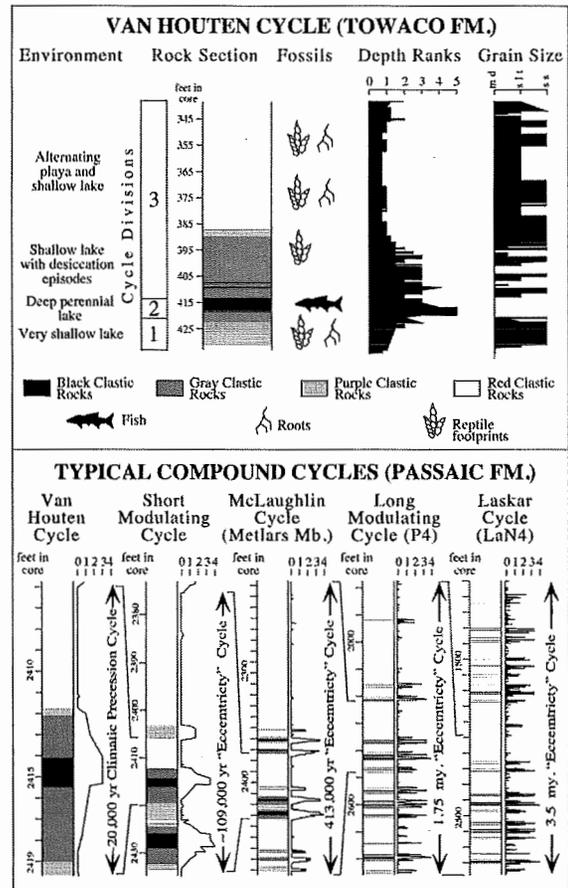
In both the Triassic and Early Jurassic, seed plants (e.g., conifers and cycadophytes) and various ferns and fern allies including many extant families were abundant. While certainly not plentiful, angiosperms (flowering plants) may have evolved by the Late Triassic (Cornet, 1989a,b; Wolfe et al., 1989). Through the Late Triassic, an extinct conifer group, the Cheirolepidiaceae or cheiroleps became relatively abundant. After the Triassic-Jurassic mass extinction, the cheiroleps became the most conspicuous element of tropical terrestrial communities for 80 million years until angiosperm proliferation. The extraordinary preponderance of the cheirolepidiaceous conifers and the rise dominance of the dinosaurs after the boundary are two of the main features of the biological record in the Hartford basin.

The subject of this field guide is the cyclicity seen in the sedimentary record of the Hartford basin played out against the background of Pangean rifting, tropical climates, and post-Triassic-Jurassic boundary biotic recovery. Given this context, not all of our observations fit neatly into paradigms based on our modern world. Some will certainly reflect grand repetitive patterns driven by incessant climate cyclicity, while others will be completely contingent on the unique peculiarities of the times. This dialectic between the recurring and the historical is a theme of the discussions at all of the field stops.

**CYCLICITY AND PALEOECOLOGICAL SUCCESSIONS**

Lacustrine rocks of the Newark Supergroup record repetitive and permeating sequences called Van Houten cycles (Olsen, 1986), after their discoverer, Van Houten (1962, 1964, 1969, 1980) (Fig. 5). The Van Houten cycle is comprised of three lithologically distinct divisions that represent lacustrine transgression (division 1), high stand (division 2), and regression followed by lowstand deposits (division 3) controlled by precipitation and evaporation changes attributed to forcing of the 20ky climatic precession cycle. Van Houten cycles are modulated in vertical succession by a hierarchy of four orders of cycles (Fig. 5). These are ascribed to modulation of precession by "eccentricity" cycles, which average approximately 100 ky, 405 ky (named the McLaughlin cycle after a University of Michigan astronomer who mapped 405 ky cycles over much of the Newark basin; Olsen and Kent, 1996), and 1.75 and 3.5 m.y. cycles (Olsen, 2001; Olsen and Rainforth, 2002b). The 405 ky McLaughlin cycle in the Newark basin serves as a basis for an astronomically-calibrated geomagnetic polarity time scale for the Late Triassic (Olsen and Kent, 1996, 1999), and earliest Jurassic (with data added from the Hartford basin), which is pinned in absolute time by radiometric dates from CAMP igneous rocks. Employing the 405 ky cycle for an interval hundreds of millions of years ago is justified because this eccentricity cycle is caused by the gravitational interaction of Jupiter and Venus, a cycle which should be stable on the scale of billions of years (Laskar et al., 2004).

The lacustrine cyclicity pervades the lower three quarters of the Jurassic age section in the Hartford basin



**Figure 5.** Van Houten and compound cycles. Modified from Olsen and Kent (1999).

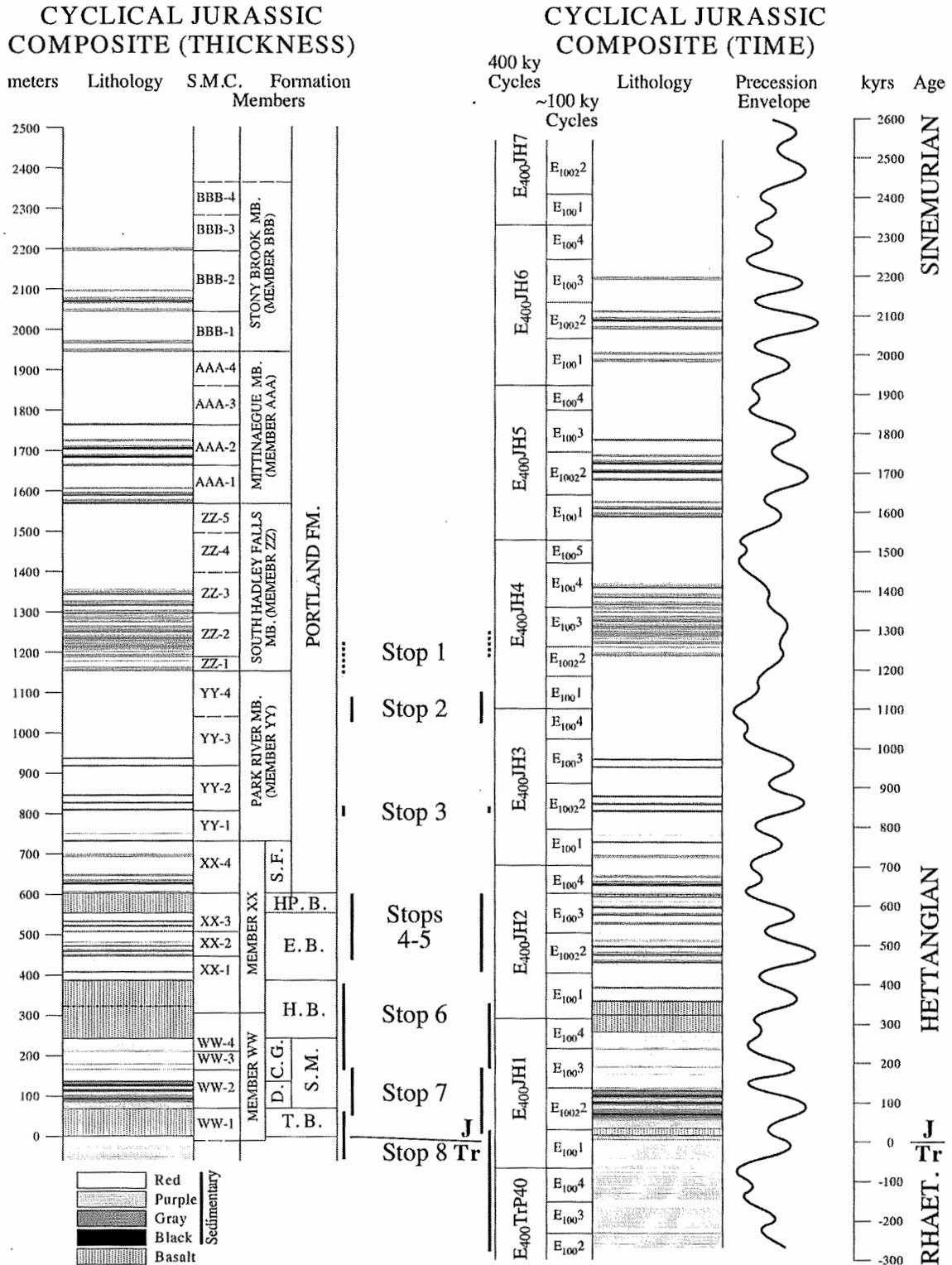
(Fig. 6). While understood in broad outline for decades (Hubert et al., 1976), the detailed pattern of this cyclicity has been worked out only recently as a result of detailed fieldwork and study of industry and Army Corps of Engineers cores (Kent and Olsen, 1999a; Olsen et al., 2002c, 2002d, 2003a; Pienkowski and Steinen, 1995). Van Houten cycles in the Hartford basin range from 10 to 30 m in thickness, depending on stratigraphic and geographic position. Within single formations in specific areas of the basin, Van Houten cycle thickness tends to vary only ~25%, but there are consistent differences between formations. The modal thickness of Van Houten cycles in the central Hartford basin is ~15 m in the Durham member of the Shuttle Meadow Formation; ~11 m in the East Berlin Formation; and ~20 m in the Portland Formation.

Six McLaughlin cycles are represented in TS IV in the Hartford basin. The lower two are segmented by the CAMP basalt flow sequences (Figs. 6, 7). The lowest begins in the uppermost New Haven Formation, the lithologic expression of which is termed the Farmington member, interrupted at its top by the Talcott Formation. This oldest McLaughlin cycle continues in the overlying Shuttle Meadow Formation where the lower mostly gray portion is termed the Durham member and the upper mostly red portion the Cooks Gap member. The Holyoke Basalt nearly divides the lowest McLaughlin cycle from the next, most of which comprises the East Berlin Formation. The top of this second McLaughlin cycle is truncated by the Hampden Basalt but resumes in the lower Portland Formation. The uppermost portion of this second McLaughlin cycle we term the Smiths Ferry member. Four more McLaughlin cycles are present in the rest of the lower Portland Formation; we have designated them in ascending order: the Park River member, the South Hadley Falls member, the Mittineague member, and the Stony Brook member. The relatively unknown remaining ~2 km of the Portland Formation appears to be completely fluvial.

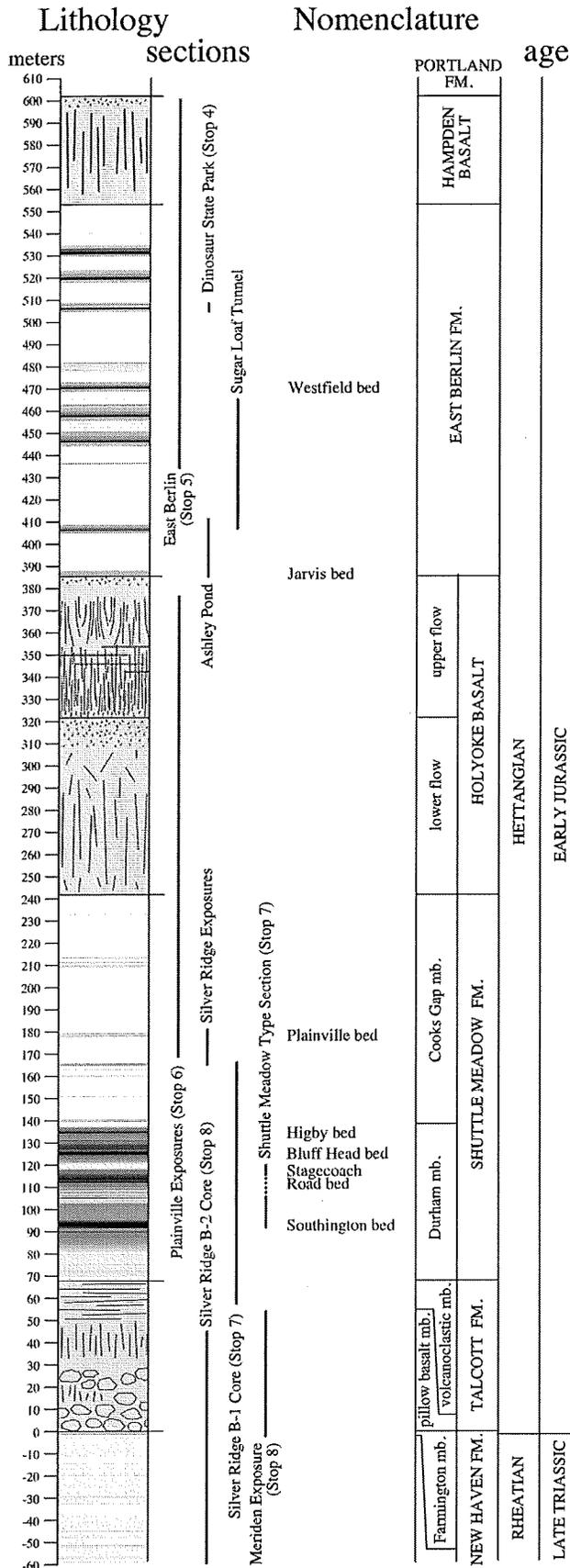
The style of the cyclicity shifts through the Jurassic succession of the Hartford basin, in part in response to the longest period cycles (1.75 and 3.5 m.y.), and in part due to tectonic, climatic, and biotic changes related to the rifting history, the CAMP episode, and Triassic-Jurassic boundary phenomena. The major trends include: 1) the unusual prevalence of gray beds with a muted cyclicity in the middle of the profoundly cyclical lower Portland Formation (Whiteside et al., 2005a); 2) the appearance of lacustrine strata in the basin during a 2.5 m.y. interval in the Early Jurassic; and 3) the prevalence of calcareous units in the Shuttle Meadow Formation in association with the fern *Clathropteris*.

Influence of the longer period eccentricity cycles can be seen in the expression of the distinctness of the cyclicity itself. The ~20 ky and 100 ky cyclicity is very obvious during the wetter phases of the 405 ky (McLaughlin) cycles. The oldest 405 ky cycle beginning in the Farmington member of the New Haven Formation is not itself cyclical. Based on correlation between the overlying Shuttle Meadow Formation with the Feltsville Formation of the Newark basin, the Farmington member should represent part of the first 100 ky cycle of the first Jurassic 405 ky cycle, the bulk of which is in the overlying Shuttle Meadow Formation. The next 405 ky cycle is comprised of the East Berlin Formation and overlying basal Portland Formation (Smiths Ferry member). These first two 405 ky cycles and the next, represented by the Park River member, have a very obvious cyclicity in which most 20 ky cycles have a black and gray portion and a distinct and thicker red portion. A large proportion of these cycles have a division 2 that is partially microlaminated, which we interpret as the deepest water facies. Thus, although these cycles have the best representation of the deepest water facies, they have a very large proportion of red beds suggesting that they formed during times of maximum precessional variance.

In dramatic contrast, the 405 ky cycle represented by the South Hadley Falls member has relatively muted Van Houten cycles—for the wetter (grayer) intervals, red bed development is poor. The development of the 100 ky cyclicity is very poor as well. Seemingly paradoxically, although the development of gray beds is the greatest of any of the 405 ky cycles in the Hartford basin, suggesting wet conditions, there is also greater development of evaporates than elsewhere and virtually no microlaminated intervals, suggesting drier conditions. This apparent paradox can be resolved if deposition occurred during a time in which precessional and 100 ky eccentricity forcing were at a minimum. In that case, both wet and dry extremes would be mollified. But, as a consequence, the lakes may never have deepened enough to reach the outlet so that solutes concentrated during the previous precessional cycle would not have been flushed from the system, allowing unusual concentrations to build up through succeeding cycles. Just such conditions occurred during the precessional minima tracking the g3-g4 eccentricity cycle that during the Triassic had a period of 1.75 m.y. If we look at the 1.75 million year cycle in light of independent correlation of the Newark and Hartford basins, and project forward from the Passaic Formation, the first Jurassic precessional minimum should in fact occur in the South Hadley Falls member.



**Figure 6.** Jurassic cyclostratigraphy expressed as thickness and lithological units on left and in time and astronomical cycles on right, showing stratigraphic and temporal position of stops. Abbreviations are: T.B., Talcott Formation; D., Durham member; C.G., Cooks Gap member; S.M., Shuttle Meadow Formation; H.B., Holyoke Basalt; E.B., East Berlin Formation; HP.B., Hampden Basalt; S.F., Smiths Ferry member;

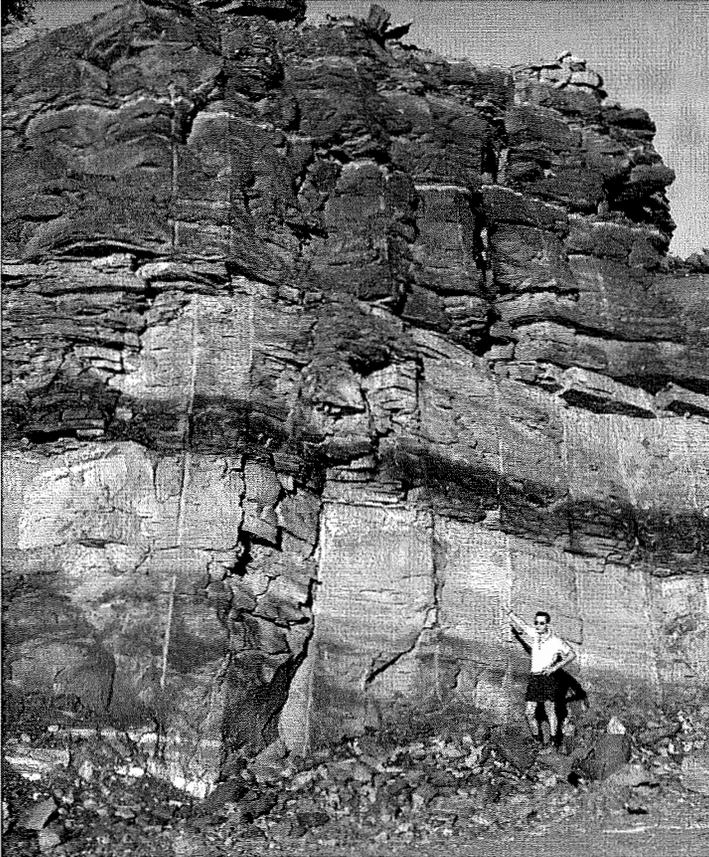


The largest-scale pattern of the appearance and disappearance of lacustrine strata is clearly of tectonic origin. Accommodation space growth had to first increase and then decrease in order to allow the transition from the fluvial New Haven Formation to the lacustrine deposits of the Shuttle Meadow Formation with an associated increase in sediment accumulation rate, in spite of the additional basin fill represented by the CAMP basalts. Similarly, accommodation space growth had to decrease in the middle Portland Formation to produce the transition back to purely fluvial deposits. As expected in a half graben these transitions are accompanied by changes in paleocurrent direction from largely axial and from the northeast during deposition of the New Haven Formation to largely from the west (hanging wall) during deposition of the Shuttle Meadow through the lower Portland Formation and back from the northeast during deposition of the upper Portland Formation (McInerny, 2003; Hubert et al., 1992), reflecting a pulse of accelerated fault-controlled asymmetrical basin subsidence as explained by the models of Schlische and Olsen (1990) and Contreras et al. (1997). LeTourneau (2002) termed this predictable sequence of facies caused by a pulse of extension a Schlische cycle.

Characterized by deeper water units rich in carbonates (see Stops 5a and 5b), the cycles of the Shuttle Meadow Formation are amongst the most distinctive in the basin. The carbonates span a range of facies from black to light gray microlaminated deep-water lacustrine carbonates and calcareous mudstones (Cornet et al., 1975) of the Durham member to shallow-water and pedogenically-modified massive and nodular micrites (Gierlowski-Kordesch and Huber, 1995) of the Plainville limestone bed of the Cooks Gap member. The only other significant carbonate unit in the Hartford basin section occurs at the base of the East Berlin Formation (e.g., Starquist, 1943). Gray and tan sandstones and siltstones of the Shuttle Meadow Formation also tend to have macrofossil remains of the large pinnate-leafed fern *Clathropteris meniscoides* (Dipteridaceae) sometimes in growth position. The laminated to microlaminated units tend to have fronds of *Clathropteris* as well. Within the Hartford basin, such occurrences are restricted to this one formation, although fragments of the fern are preserved throughout the Jurassic sequence.

Figure 7. Stratigraphy of the extrusive zone in the Hartford basin and sources of stratigraphic data.

The prevalence of carbonates and the abundance of *Clathropteris* characterize the Shuttle Meadow Formation, remarkably similar to homotaxial units in the Culpeper, Newark, Pomperaug, Fundy, and in part Moroccan basins (no *Clathropteris* yet). That these distinctive sequences occur in the same homotaxial position relative to the basalts suggests that they represent deposition during the same time interval within their respective basins. The pattern probably represents a facies syndrome associated with the underlying Triassic-Jurassic boundary, perhaps related to the super-greenhouse conditions postulated for this time by McElwain et al. (1999) (Whiteside et al., 2005b).



**Figure 8.** Single Van Houten cycle containing the Westfield bed at Stop 5. This is the uppermost of the second black-shale baring triplet from the top (i.e. short modulating cycle AAA-2 of Fig. 6). PML for scale.

depositional mudstone mélangé consisting of dark gray to black mudstone with microlaminated to laminated clasts derived from the underlying (and occasionally overlying) units. While interpreted by some authors as beds of rip up clasts, Olsen et al. (1989a) interpreted these sequences as a partly penecontemporaneously sheared dewatering phenomenon occurring within the beds after deposition of some thickness of overlying strata. Hence, they cannot be easily interpreted in terms of depositional environments (see Stops 5 and 7a). The upwardly shoaling rest of division 2 of the cycles tends to consist of laminated mudstones and minor sandstone (Stop 5) to sandstones and conglomerates (Stops 2 and 3), and frequently contain abundant conifer remains and reptile footprints (Stops 4, 5, 7, 7a). The succeeding division 3 is usually comprised of fine to coarse red beds, with abundant mud cracks and root structures, and less common reptile footprints (Stops 4, 5, 7).

Laterally, most of the cycles behave in a similar manner; they are thinnest over much of the central Hartford basin, thickening rapidly toward the eastern basin edge, as the cycles in general coarsen. The divisions of the cycles do not increase in proportion, however. Division 2 of the cycles increases at a much faster rate than divisions 1 or 3. This thickening is accomplished by an increase in the individual laminae and increase in the total fraction of the cycle taken up by the beds, and an increase in clastic content. The increase in clastic content is partially because of

At the opposite scale, Van Houten cycles of the Hartford basin demonstrate a predictable pattern of biofacies and depositional environments, fully developed within only a few of the cycles. These cycles have distinct, informally named microlaminated beds that have produced the bulk of fossil fish in the basin. These cycles contain in ascending order the Southington, Stagecoach Road, Bluff Head and Higby beds (Shuttle Meadow Formation: Stops 7 and 7a), the Westfield bed (East Berlin Formation; Stop 5), and the Middlefield (Stop 3) and Chicopee beds of the Portland Formation. The transgressive division 1 contains mudcracked clastics often with abundant reptile, especially dinosaur tracks deposited on an infrequently exposed shallow lake floor (Stop 5). The transition to division 2 is relatively abrupt with only a few centimeters of fossil-poor non-mudcracked mudstone or claystone. The usually calcareous microlaminated division 2 contains a stereotyped biofacies with articulated fishes, coprolites, occasionally the conchostracan *Cornia*, allochthonous vascular plant material (sometimes spectacularly preserved), and often allochthonous charophyte fragments. These laminites have individual laminae of great lateral extent (Olsen, 1988) and were deposited in perennial lakes exceeding tens of meters in depth (Olsen, 1990) (Stop 5). In fine grained settings, the microlaminated portion of division 2 is often overlain by a post-

the presence of numerous small graded beds we interpret as turbidites (Stop 7). A similar pattern characterizes all of the black and dark shale beds of the Hartford basin and was outlined by LeTourneau and McDonald (1985).

Fish preservation within laminites of division 2 also changes along this thickening gradient. At several localities in the central Hartford basin, articulated fish are preserved as organic films with virtually no relief and sometimes as only very faint pyritic halos (i.e. dephosphatization) (Stop 5). McDonald and LeTourneau (1989) described the trend from these regions towards the eastern basin edge where preservation of bony tissue vastly improves and the fish have progressive higher relief. The better preservation seems associated with higher clastic content, but the mechanism is very poorly understood. A very similar pattern of dephosphatization occurs in the fish-bearing laminites of the Devonian Caithness Flagstone of Scotland (Westoll, pers. comm., 1982).

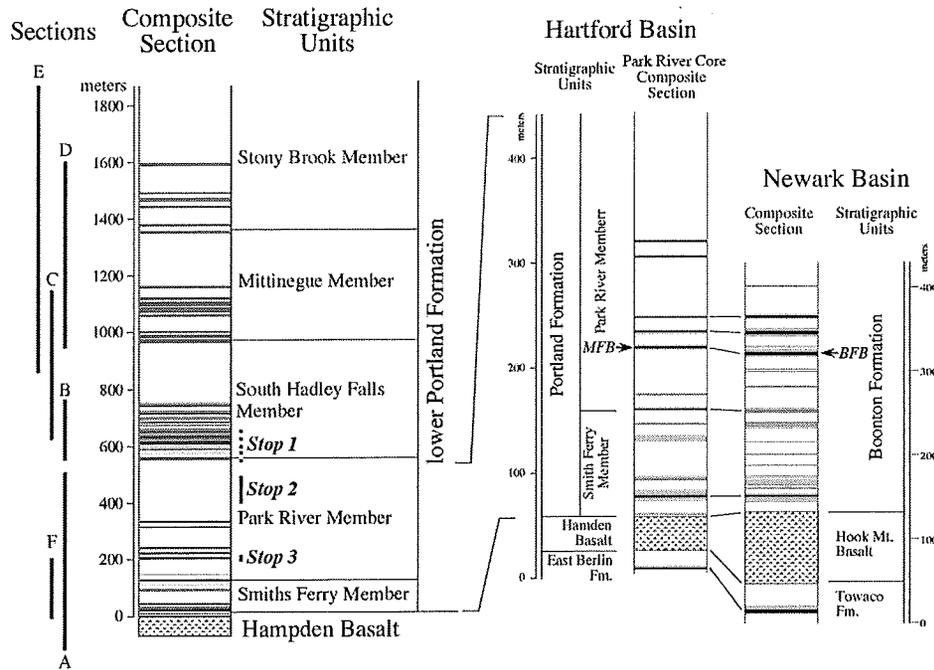


Figure 9. Stratigraphy of the Portland Formation, informal members of the Portland formation, and correlation to the Newark basin. Colors as in Fig. 6.

LITHOSTRATIGRAPHY AND BIOTA OF THE HARTFORD BASIN

New Haven Formation

The New Haven Arkose was named by Krynine (1950) for the thick succession of red, brown, tan, and minor gray siltstone, sandstone and conglomerate that represent the lower portion of the Hartford basin stratigraphic column. These strata were previously called "Western Sandstone" by Percival (1842) and Davis (1898), and in part, "Sugarloaf Arkose" by Emerson (1898; 1917). Krynine attempted a generalized stratigraphy of the formation based on a small number of short (2-20 m thick), scattered outcrops in the vicinities of New Haven and Meriden, CT, that extended through the formation's entire thickness (Krynine, 1950, fig. 10, table 2). Krynine designated 5 type sections, each of which represented what he interpreted to be a distinctive sedimentary facies assemblage, with much of his facies descriptions being based on petrologic data (e.g., heavy mineral assemblages) rather than stratigraphic and/or sedimentological criteria.

Gray (1987), among others, have recognized that the New Haven Formation contains diverse associations of lithofacies and have used the lithologic descriptor "Formation" instead of "Arkose". Several workers since Krynine (1950) have described the stratigraphy and sedimentology of several long stratigraphic sections of New Haven Formation rocks or have attempted generalized characterization of the formation (e.g., Wessal, 1969; Hubert et al., 1978; Lorenz, 1987; McInerny, 1993; McInerny and Hubert; 2002) though no one has attempted a comprehensive stratigraphic synthesis of the formation. The New Haven Formation varies in thickness from ~1500 m in the New Haven, Connecticut region to as much as 2000 m at, and west of Meriden, Connecticut.

Unlike its lacustrine time equivalents in the Lockatong and Passaic formations of the Newark basin, virtually all of the New Haven Formation fluvial strata lack black mudstones. The basal New Haven Formation locally has

beds of gray sandstone that at one locality (Forestville, CT: Krynine, 1950; Cornet, 1977) produced a palynoflora closely comparable to that of the lower Passaic Formation. Hence the basal New Haven Formation is conventionally assigned a basal Norian age (Cornet, 1977).

The rest of the lower New Haven Formation consists of cyclical fluvial strata that have been interpreted as meandering river sequences (McInerney, 1993; McInerney and Hubert, 2003; Horne et al., 1993). These have common and locally well-developed pedogenic soil carbonates. Pure pedogenic micritic calcite from one such carbonate provided a U-Pb date of  $211.9 \pm 2.1$  Ma (Wang et al., 1998), a Norian age on most time scales, including the Newark GPTS (Gradstein et al., 1995; Kent and Olsen, 1999). The same exposure has produced a partial skull of the crocodylomorph *Erpetosuchus*, known otherwise from the Lossiemouth Sandstone of Scotland, conventionally assigned a Carnian age (Olsen et al., 2000b). Previously described reptilian skeletal material from the New Haven Formation in the southern Hartford basin comprises the holotype of the stagonolepidid *Stegomus arcuatus* Marsh, 1896. Lucas et al. (1997) considered *Stegomus* a subjective junior synonym of *Aetosaurus*, an index fossil for continental strata that again suggests an early to middle Norian age and referred the lower to middle New Haven Formation to the Neshanician Land Vertebrate Faunachron (Lucas and Huber, 1993, 2003).

Virtually unstudied, the middle New Haven Formation in the central Hartford basin consists of mostly red massive sandstone with much less well-developed pedogenic carbonates (Krynine, 1950). Apart from abundant *Scoyenia* burrows and root casts, the only fossil recovered is a scapula of an indeterminate phytosaur ("*Belodon validus*" Marsh, 1893), indicating only a Late Triassic age. Much more varied lithologies categorize the upper part of TS III and the upper New Haven Formation (Hubert, et al., 1978), including meandering and braided river deposits and minor eolian sandstones (Smoot, 1991). Vertebrates from these strata include an indeterminate sphenodontian (Sues and Baird, 1993) and the procolophonid *Hypsognathus fenneri* (Sues et al., 2000). The presence of *Hypsognathus* indicates correlation to the upper Passaic Formation of the Newark basin and thus a Cliftonian Land Vertebrate Faunachron age (middle to late Norian and Rhaetian age).

We informally coin the name "Farmington member" for the distinctive uppermost 3-20 m of strata of the New Haven Formation (Stop 8, 8a). These strata consist of laterally-continuous, red to gray, well-bedded to laminated sandstone, siltstone and minor shale facies that are interbedded and/or intertongue in places with arkosic conglomerate, and sometimes volcanoclastic strata. The Farmington member is named for exposures of uppermost New Haven Formation strata that outcrop on the north side of Route 4 in Farmington, Connecticut. Davis (1898) and Gray (1982b) described facets of the geology of this locality. The Farmington member has produced a florule of *Brachyphyllum* conifers at multiple localities and in its uppermost few centimeters a palynoflorule of typical Jurassic aspect at one locality (see discussion at Stop 8a). The member plausibly contains the Triassic-Jurassic boundary, unless cut out by a TS III – TS IV unconformity.

### **Talcott Formation**

The Talcott Formation was originally named the Talcott Diabase by Emerson (1917), the lower lava flow of the Meriden Formation by Krynine (1950) and raised in rank to formation by Sanders (1968). The Talcott Formation is a high-titanium quartz normative tholeiitic basalt (HTQ basalt; terminology of Puffer, 1992) that attains a maximum thickness of ~130 m in central Connecticut, and is composed of a variety of textures that display complex relationships. In central and southernmost Connecticut, the lower ~75 m of the Talcott Formation includes ubiquitous pillowed horizons that grade laterally and vertically into columnar-jointed massive basalt. The upper part of the formation is made up of volcanoclastic breccia and thin interbeds of conglomeratic, fluvial sandstone and conglomerate. Locally, the uppermost beds of the New Haven Formation interfinger with the Talcott Formation. This can be seen where forsets of pillow basalts overlie the New Haven Formation and tongues of red beds extend between them such as at Stop 8 in Meriden, CT.

Between the southernmost and central basin areas, the Talcott Formation consists of similar lithologies that can laterally interfinger along strike and throughout the entire thickness of the formation and reflects the proximity to sites of fissure eruptions where Talcott feeder dikes intersected the surface (cf. Philpotts and Martello, 1986). In this region, Sanders (1970) proposed a sweeping lithostratigraphic redefinition of the Talcott Formation to include four basalt and three sedimentary units that possessed a cumulative thickness of 300+ m. Sander's (1970) proposed scheme relied implicitly upon the assumption that the relevant units were tabular, parallel, conformable lithosomes,

but this is not the case. Instead, the outcrop consists of growth structures developed with the feeder system of the Talcott Formation (Philpotts and Martello, 1986; Olsen, et al., 2004).

The nature of the upper contact of the Talcott Formation is variable, and is gradational over a short stratigraphic distances often consisting of volcanoclastics where the largely gray lower Shuttle Meadow Formation (Durham member) is present (see Stop 7). In other areas of the basin in the vicinity of Meriden and also north of Cooks Gap, the Durham member is not present and the contact is a disconformity, and basal Shuttle Meadow strata locally consists of up to 0.5 m of well rounded/well-sorted conglomerate composed exclusively of clasts of Talcott Formation.

### Shuttle Meadow Formation

The Shuttle Meadow Formation was named by Lehmann (1959) to replace the anterior shales of Percival (1842) and Davis (1898) and the lower sedimentary division of the Meriden Formation of Krynine (1950); it is the oldest of the Hartford basin sedimentary formations that is entirely of Jurassic age and the oldest unit expressing well-developed lacustrine cyclicity (Figs. 6, 7). The type section is near the Shuttle Meadow Reservoir and was described by Krynine (1950). The formation attains a maximum estimated thickness of ~250 m near the border fault, and is herein defined to consist of two member-rank units: a lower sequence of gray, black and red strata, up to ~100 m thick that we informally designate the Durham member after outcrops comprising the famous fossil fish locality of the same name in Durham, CT (Newberry, 1888; Davis and Loper, 1891; McDonald, 1975); and an upper mostly red sequence we informally call the Cooks Gap member, named after the exposures at Cooks Gap, Plainville, CT.

Having produced fossil fish for over a century, the Durham member is exposed at many locations in the Hartford basin, but no section revealed the complete succession of beds until the recovery of the Silver Ridge B-1 core (see Stop 7). We recognize three well-developed Van Houten cycles in this member, each with a distinctive and laterally recognizable division 2 to which we give informal names.

The contact of the Durham and Cook's Gap members is defined by the transition of gray- green shale and siltstone of the Higby cycle into red, finely laminated similar lithologies which pass into red, ripple cross laminated siltstone. Where the Durham member is absent, redbeds of the Cooks Gap member rest directly on basalt of the Talcott Formation. The Cooks Gap member is overlain everywhere by the Holyoke Basalt. We hypothesize that the pinching out of the Durham member between the Talcott Formation and the Cooks Gap member is accomplished by progressive onlap of the Durham member against the Talcott Formation, rather than a dramatic thinning of its component parts or a lateral change in facies of the Durham member to that of the Cooks Gap member. The striking changes in thickness laterally of the Shuttle Meadow Formation along strike along the western outcrop belt of the formation is thus largely due to a onlap of progressively younger strata onto the Talcott Formation.

The sequence of Van Houten cycles seen in the Durham member as well as the transition into red beds as seen in the upper parts of the Silver Ridge B1 core (Fig. 7; Stop 7) is typical of a short modulating cycle (Fig. 5). We regard the top of this short modulating cycle (WW2; Fig. 6) to be at the base of the weakly variegated beds at the top of the Silver Ridge Core. The largest continuous exposure of the Cooks Gap member is at Cook Gap, Plainville, CT (Stop 6), where virtually the entire member is exposed. As described by Gierlowski-Kordesch and Huber (1995) most of the section consists of red and red-brown ripple-cross-laminated siltstone and sandstone and interbedded red, mostly massive, mudstones. The red units have abundant desiccation cracks and dinosaur tracks, as well as some locally abundant burrows and root casts. There are a series of gray, purplish, and tan beds that mark out successive, very weakly developed Van Houten cycles, and one set of limestone beds near the base of the member that we designate informally, the "Plainville limestone bed" (see Stop 6). This bed is laterally continuous and has been identified at all sections where the suitable interval is exposed. It falls stratigraphically at the appropriate place for the peak of the next short modulating cycle (WW3; Fig. 6) after the cycle containing the Durham member.

Lateral facies changes in the Cooks Gap member are poorly known due to limited outcrop as well as the lateral variability of beds as seen in the Cooks Gap section. At East Haven the Cooks Gap member is much coarser. Pebbly sandstone and conglomerate beds are abundant, reflecting the proximity of the eastern border fault 1.2 km to the southeast. Here the Plainville limestone bed is well developed, containing macerated plant debris and isolated fish scales and bones.

Overall, in the Shuttle Meadow Formation, microfossil assemblages are present in most gray claystones and siltstones, which is not the case in the other formations. In all cases the palynoflorules are dominated by the cheirolepidiaceus pollen genus *Corollina* (Cornet, 1977). Floral macrofossils are often present in the same units. Assemblages bearing *Clathropteris* and *Equisetites* are common throughout the Shuttle Meadow Formation. The Bluff Head bed has produced a relatively diverse macroflora of ferns, cycadeoides, ginkophytes, and cheirolepidiaceus conifers (Newberry, 1888) (See Stop 7).

Invertebrates are represented by burrows and walking and crawling traces of arthropods, along with locally by common clams, ostracodes, the conchostracan *Cornia*, and the beetle larva *Mormolucoides* (McDonald, 1992; Huber, et al., 2003). Articulated fossil fish, often beautifully preserved and very abundant, occur in the five named beds with the formation and pertain to four taxa: the ptycholepid palaeonisciform *Ptycholepis marshi*, the redfieldiid palaeonisciform *Redfieldius gracilis*, the holostean neopterygian *Semionotus* sp., and the coelacanth *Diplurus longicadatus* (Newberry, 1888; Cornet et al., 1975; Schaeffer et al., 1975; Schaeffer and McDonald, 1978; Olsen and McCune, 1991). Additionally, two isolated possible theropod teeth were recovered from the lower Shuttle Meadow Formation (McDonald, 1992).

While the Jurassic strata of the Hartford basin comprise the type area of the famous Connecticut Valley footprint assemblage (e.g., Hitchcock, 1836, 1848, 1858, 1865; Lull, 1904, 1915, 1953; Olsen et al., 1998; Olsen and Rainforth, 2002a; Rainforth, 2005), footprint assemblages from the Shuttle Meadow Formation are amongst the most poorly known in the Jurassic of eastern North America. However, the assemblage is clearly of Connecticut Valley aspect. The theropod dinosaur tracks *Eubrontes giganteus*, and a variety of smaller forms traditionally called *Anchisauripus* and *Grallator* are present along with much less common *Anomoepus* (ornithischian dinosaur: Lull, 1953; Olsen and Rainforth, 2002a), *Batrachopus* (crocodylomorph), and a recently discovered example of the synapsid track *Ameghinichnus* (found by G. McHone). The only significant Connecticut Valley genus missing is the prosauropod dinosaur track *Otozoum* (Rainforth, 2003).

### **Holyoke Basalt**

The Holyoke Basalt is the most widespread of the CAMP flow sequences in the Connecticut Valley, always being present at the appropriate stratigraphic level (Stop 6). It was named the Holyoke diabase bed by Emerson (1891) for what previously had been called the main trap sheet by Percival (1842) and Davis (1898), and renamed the Holyoke Basalt by Lehmann (1959). The Holyoke Basalt is an HFQ-type (high-iron quartz normative) basalt, essentially identical in chemistry to the lower three flows of the Preakness Basalt of the Newark basin (Puffer et al., 1981; Puffer, 1992). It reaches a maximum thickness in excess of 200 m (Philpotts and McHone, 2004). Two flows are present in central Connecticut, but only the upper extends throughout the basin (Davis, 1898; Gray, 1987). The lower flow tends to be very massive with a thick and reddened vesicular top. The second has a colonnade with a characteristic splintery jointing, and a curvilinear upper entablature (Philpotts and McHone, 2004). Prevot and McWilliams (1989) identified an anomalous paleomagnetic direction (an excursion) in the lower part of the Holyoke Basalt at several localities in northern Connecticut and Massachusetts where the lower flow is absent. The excursion is recorded in the second flow of the Preakness Basalt of the Newark basin (P2 and P3 units of Tollo and Gotfried, 1992), characterized by the same kind of splintery fracture, and again in the upper flow of the Deerfield Basalt. The presence of this excursion in these different basins strongly suggests that these flows were erupted simultaneously, within hundreds of years. It also implies that the lower flow of the Holyoke Basalt is represented in the Newark basin by the thin discontinuous unit at the base of the formation (unit P1 of Tollo and Gotfried, 1992), or the lowest, pillowed flow of the Deerfield Basin.

### **East Berlin Formation**

Lehmann (1959) named the sedimentary unit between the Holyoke and Hampden basalts, the East Berlin Formation, replacing the informal name middle shale of Percival (1842), posterior shales Davis (1898), and upper sedimentary division of the Meriden Formation of Krynine (1950). It was the first Hartford basin unit in which cyclical strata were described (Krynine, 1950; Klein, 1968; Hubert and Reed, 1978, Hubert et al., 1976, 1978; Demicco and Gierlowski-Kordes, 1986; Olsen et al., 1989a). The formation is ~200 m thick in its type area, thickening toward the east (Figs. 6, 7).

The Van Houten cycles and short modulating cycles of the East Berlin Formation are the most obvious in the basin, due partly to the excellent exposures at which they are displayed (e.g., Stop 5), and partly to the thick red bed-dominated division 3 that contrasts obviously with the gray divisions 1 and 2 and the black deepest water portion of division 2 (Fig. 8). The two or three Van Houten cycles of each short modulating cycle divisions 1 and 2 that are gray purple or red, make the longer frequency cycles obvious as well.

Three short modulating cycles comprise the East Berlin Formation (Figs. 6, 7). The lowest has Van Houten cycles in which each has a relatively poorly developed division 2, lacking a microlaminated bed, while the upper two have a two Van Houten cycles each with a microlaminated division 2. The thickest microlaminated bed is within division 2 of the third Van Houten cycle of the middle short modulating cycle (Westfield bed). Hence, the middle of the East Berlin Formation marks the wettest phase of a McLaughlin cycle. Fourier analysis of depth ranks of the section at Stop 5 reveals a clear hierarchy of periodicities in thicknesses of 12.0 m and 68.3 m. Assuming that the 12.0-m-thick cycles are the 20 ky precession cycles, the 68.3-m-thick cycles have periodicities in time of 113 ky, consistent with a Milankovitch interpretation (Olsen et al., 1989a).

The lateral continuity of Van Houten cycles in the East Berlin Formation was demonstrated by Hubert et al. (1976, 1978) who showed the strong stratigraphic relationship between the mapped distribution of the Hampden Basalt and the underlying Van Houten cycles over much of the Hartford basin. Olsen et al. (1989a) elaborated upon the lateral correlation of the cycles and showed that correlation of the microlaminae and turbidites in polished slabs from the Westfield, providing compelling independent evidence of the lateral continuity of the cycles (Olsen, 1988). Changes towards the border faults are much like in the other cyclical units (e.g., Shuttle Meadow Formation), as is the fish dephosphatization pattern, best exemplified in the Westfield bed (McDonald and LeTourneau, 1989).

Despite the abundance of gray beds in the East Berlin Formation, palyniferous units are rare. This is not a function of the thermal maturity of the strata, because even those that are of very low thermal maturity ( $R_o = 0.5$ ; Pratt et al., 1988) have rare preservation of pollen and spores. Nonetheless some palynofloules have been recovered and all are typically Early Jurassic in aspect being very strongly dominated by *Corollina meyeriana* (Cornet, 1977; Cornet and Olsen, 1985). While macrofloral remains are common in the gray beds, they are almost totally dominated by shoots of *Brachyphyllum* (see Stop 5) with scraps of *Otozamites* and even more rare *Clathropteris*, making a dramatic contrast with the underlying Shuttle Meadow Formation.

The fish assemblage is similar to the underlying Shuttle Meadow Formation. *Semionotus*, *Redfieldius*, very rare *Ptycholepis* and *Diplurus* are present (McDonald, 1975, 1992, pers. comm.), as well as the abundant *Diplurus* coprolites. In addition, six or more, slightly recurved, non-serrated conical reptile teeth have been found in the Westfield bed that cannot yet be attributed to any taxon with confidence (H. D. Sues, pers. comm., 1995).

As usual, invertebrates are represented by burrows and walking and crawling traces of arthropods, along with locally by common examples of the conchostracan *Cornia*, and rare insect body fossils, specifically the beetle larva, *Mormolucoides* (Lull, 1953; McDonald, 1992; Huber, et al., 2003).

Although the East Berlin Formation is justifiably famous for the spectacular display of large theropod dinosaur tracks (*Eubrontes giganteus*) at Dinosaur State Park (Stop 4), the overall assemblage is actually relatively poorly known. Like the underlying Shuttle Meadow Formation, the usual Connecticut Valley taxa (*Eubrontes*, *Anchisauripus*, *Grallator*, *Anomoepus*, and *Batrachopus*) are present, with the exception of *Otozoum*.

### Hampden Basalt

Emerson (1898) named the Hampden Diabase for the posterior sheet of Percival (1842) and Davis (1898) that was later termed the upper lava flow of the Meriden Formation by Krynine (1950) and finally the Hampden Basalt by Lehmann (1959). The Hampden Basalt is a high-TiO, high-Fe, quartz-normative tholeiite (HFTQ) identical in composition to the Hook Mountain Basalt, which appears to be its exact temporal correlate. According to Chapman (1965), there may be as many as eight flow units based on the presence of thin vesicle zones, although Gray (1982a) suggested that the maximum is two flows in Massachusetts (Colton and Hartshorn, 1966), which agrees with our observations and one flow in central Connecticut (Stop 5). The flows are typically massive but can be very vesicular at the base and top. Tilted pipestem vesicles are common at the lower contact and indicate a northeasterly flow direction (Gray, 1982a). The Hampden Basalt is the thinnest of the extrusives in the Hartford basin, reaching a

maximum thickness of 30 m, and thinning and disappearing near the Massachusetts border where it is largely replaced by the equivalent Granby Tuff (Emerson, 1898; Robinson and Luttrell, 1985), reappearing again on the back slope of the Mount Tom and Holyoke ranges at the northern end of the Hartford basin.

### **Portland Formation**

Krynine (1950) coined the name Portland Arkose to apply to all sedimentary rocks above the Hampden Basalt and its equivalent, the Granby Tuff. Previous names for these strata include the eastern sandstone of Silliman (1826), Percival (1842) and Davis (1898), and Longmeadow Sandstone (in part) and Chicopee Shale (in part) of Emerson (1898, 1917). Krynine's (1950) stratotype was designated at the Portland brownstone quarries along the east bank of the Connecticut River in Portland, Connecticut (see Stop 2). Leo et al., (1977) recognized that the Portland contains a diverse range of lithofacies, and modified the unit name to Portland Formation. The Portland Formation ranges in thickness from ~450 m in southern Connecticut, to ~1 km in the vicinity of Middletown, to as much as 4 km thick in the central portion of the Hartford basin north of Hartford and in south-central Massachusetts (Fig. 2).

The Portland Formation is composed of a lower half consisting of cyclical lacustrine and marginal fluvio-lacustrine red, gray and black clastic rocks closely comparable to the East Berlin Formation, and an upper half made up almost entirely of fluvial red mudstone sandstone and conglomerate and minor red eolian strata (total maximum thickness ~5 km). Olsen et al. (2002c, 2003a) divide the lower Portland Formation into members in a parallel manner to the Passaic Formation of the Newark basin. They recognize four full McLaughlin cycles in the lower Portland, and one continuing from the underlying East Berlin Formation. These mappable units are proposed as informal members as follows (from the bottom up): "Smiths Ferry", "Park River", "South Hadley Falls", "Mittineague", and "Stony Brook" members (Figs. 6, 9). These units are critical to establishing the cyclostratigraphy and time scale for the Hettangian and Sinemurian sites and hence reconstructing the sedimentologic and structural history from the Triassic-Jurassic.

We recognize five members in the lower, 2 km-thick lacustrine-dominated portion of the Portland Formation. The upper 2 km-thick interval of the formation, dominated by fluvial and alluvial-fan facies is undivided. As the cyclicity of the lower Portland strongly resembles the lithologic expression of cyclicity displayed by the Passaic Formation in the Newark basin, we have modified McLaughlin's (1933) and Olsen et al.'s (1996) approach by adopting a lithostratigraphic nomenclatural convention for the Portland Formation from the Passaic Formation in which lithologically identified McLaughlin cycles are given formal member names (e.g., McLaughlin, 1933; Olsen et al., 1996) (Fig. 6). In this convention, we divide the Portland Formation into a series of lithologically identified members, each consisting of a lower cyclical portion with significant amounts of gray and black strata, and an upper portion that is predominately red. While these members do broadly correspond to McLaughlin cycles, they are defined by purely lithological criteria, and each member has, in fact, has its own lithologic and boundary characteristics that do not necessarily correspond to the natural peaks or troughs of cycles derived from Fourier analysis.

During the Park River diversionary tunnel project, the Army Corps of Engineers cored the lowermost ~500 m of the Portland Formation the near Hartford (Pienkowski and Steinen, 1995). These cores, a number of much less extensive cores and borings as well as outcrops along Stony Brook River, the Connecticut River, the Westfield River, and the Chicopee River in south central Massachusetts (Fig. 9) provide the most significant stratigraphic information on the Portland Formation. Combined with other outcrops and exposures, these sections allow us to compile a composite stratigraphy for the lower half of the Portland Formation, the informally designated members of which are described below.

**Smiths Ferry member:** We define the base of the Portland Formation as the stratigraphically lowest sedimentary bed that lies above the Hampden Basalt or Granby Tuff. The 110 m of the lower Portland Formation as seen in the Park River cores comprise the Smiths Ferry member, named for extensive outcrops at, and in the vicinity of, the Dinosaur Footprint Reservation at Smiths Ferry, Massachusetts. The Park River cores penetrated the entire thickness of the Smiths Ferry member including its contact with the Hampden Basalt, and provide an excellent reference section that allows correlation of other lengthy exposures in different areas of the basin.

Depending on basin location, the Smiths Ferry member ranges between 100-150 m-thick. In the Park River cores, the member begins with thin, red basal strata followed by a series of three Van Houten cycles each with gray to dark gray division 2. The basal red strata and triplet of gray shales up to the top of the third shale unit have a cumulative thickness of 40 m, and are in turn overlain by 30 m of entirely red siltstone and sandstone which contain two red Van Houten cycles. The upper Smiths Ferry member includes two additional Van Houten cycles, each with a gray or purple division 2, and intercalated red siltstone and sandstone with a cumulative thickness of 30 m. The top of the Smiths Ferry member is defined as the base of the prominent black shale marking the overlying Park River member in the Park River cores.

The lower Smiths Ferry member constitutes the uppermost, short modulating cycle of the ~400 ky McLaughlin cycle that began in and comprises the East Berlin Formation. The last two Van Houten cycles within the upper 30 m of the member (as seen in the Park River cores) actually belong to the basal short modulating (~100 ky) cycle partially contained within the overlying Park River member, but the non-red parts of those two cycles are too indistinct to be practical for field recognition as an upper boundary for the Smiths Ferry member.

At its type section, the Smiths Ferry member consists of well-exposed, basal red siltstone and sandy siltstone that include the main dinosaur track-bearing surface of the Holyoke Footprint preserve described by Ostrom (1972). It is also the locality for the type specimen of *Eubrontes giganteus*, the first dinosaur footprint to be described and figured (Hitchcock, 1836; Buckland, 1836; Olsen et al., 1998). Above the aerially-extensive footprint horizons, only the next two Van Houten cycles of the member are in part exposed. The gray strata of these cycles contain abundant oscillatory ripples, dinosaur footprints, and oriented plant debris. Laterally-equivalent exposures include extensive outcrops in the Middletown and adjacent quadrangles mapped and described in part by Lehman (1959) and Gilchrist (1979). The Smiths Ferry member is well-represented within the confines of Wadsworth Falls State Park and nearby in the bed and banks of the Connichaug River in Middletown. Most of the gray and dark gray lakebeds visible in the Park Cores can be readily identified in the field.

**Park River member:** We informally name cyclical lacustrine strata between the Smiths Ferry member and the overlying South Hadley Falls member the Park River member after the type section contained by the Park River cores (Fig. 9). In the cores, the type Park River member is represented by 280 m of section, and begins at the base of the first prominent black mudstone above the mostly red beds of the upper Smiths Ferry member as discussed above. Its top is defined as the base of the first well-developed black shale of the succeeding South Hadley Falls member, which is not represented in the cores. The lower half of the Park River member is profoundly cyclical, with a pattern remarkably similar in lithology and overall properties to the East Berlin Formation, but different than overlying members.

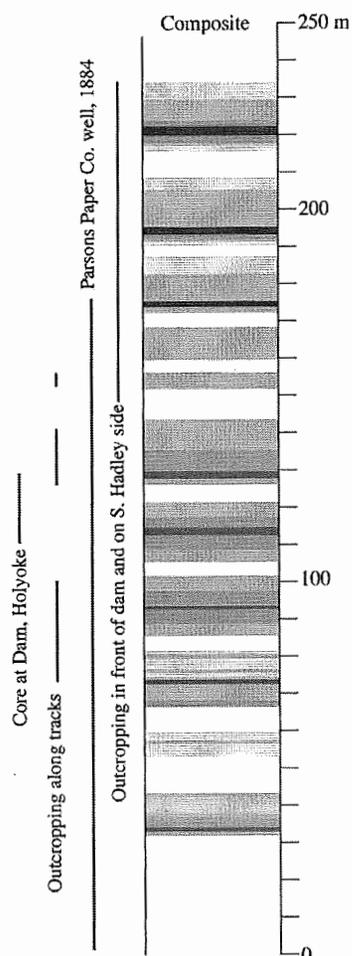
The upper half of the Park River member is represented by extensive outcrops in the Middletown area first documented in detail by McDonald (1975) and LeTourneau (1985a). The most distinctive portion of the Park River member is the second short modulating cycle, the lowest Van Houten cycle of which has a microlaminated division 2, termed the Middlefield bed after outcrops along Laurel Brook in Middlefield, Connecticut. Known since at least the 1820's (Redfield, 1836; Davis and Loper, 1891; McDonald, 1975, 1992), this unit has produced the holotype of *Redfieldius gracilis* Hay 1899 (Schaeffer and McDonald, 1978) along with several morphotypes of neopterygian *Semionotus* (possibly "species flocks" similar to those of McCune (1986) and Olsen and McCune (1991). Other exposures of this bed and the enclosing Van Houten cycle include outcrops in Bachelor Brook, South Hadley, Massachusetts (Hubert et al., 1992), Long Brook, Cromwell, Connecticut, the foundation excavations (now covered) for the One State House Square building, Hartford, Connecticut, and the unit uncovered in a gravel pit at Glastonbury, Connecticut described by LeTourneau and McDonald (1997) (Stop 3). These locations have collectively produced a wealth of fossils including fish and coprolites and most importantly conchostracans, plants and dinosaur footprints.

The third short modulating cycle of the Park River member contains a doublet of Van Houten cycles each with black shale-bearing division 2 (Figs. 6, 9). These lack a microlaminated portion and as a rule are not well exposed. One outcrop that is probably of this short modulating cycle is exposed along Long Hill Brook, Middletown, Connecticut and one of the thin black mudstones of division 2 of the upper Van Houten cycle of this doublet does contain associated if not completely articulated fossil fish (McDonald, 1975).

The sandstone facies of the Park River member in the northern and southern portions of the Hartford basin are distinctive in preferentially being used for building stone, compared to other members of the cyclical portion of the formation. These include quarries in South Hadley, Massachusetts that produced the type of *Otozoum moodii*, *Grallator parallelum* (Hitchcock, 1847; Olsen et al., 1998; Rainforth, 2003), and numerous other footprint taxa of dubious validity. Red beds in the fourth and uppermost short modulating cycle comprise the extensively exposed sandstones in the famous Portland Brownstone Quarries (Hubert et al., 1982, 1992, McInerney, 1993; Guinness, 2003) that have produced very abundant and well-preserved reptile footprints (i.e., Hitchcock, 1865) as well as one example of dinosaur bone casts (Colbert and Baird, 1958) (Stop 2).

Lateral changes in the Park River member are comparable to those seen in the East Berlin Formation, again with a disproportionate increase in the thickness of division 2 of Van Houten cycles – best seen in the cycle bearing the Middlefield fish bed (Stop 3). Extensive conglomerates are present in this member in southern Connecticut and have been described by LeTourneau and McDonald (1985, 1997), Smoot et al. (1985) and Smoot (1991).

**South Hadley Falls member:** We informally designate the largely gray strata and succeeding red beds between the underlying Park River member and the overlying Mittineague member the South Hadley member after its type section along the Connecticut River at South Hadley Falls and adjacent Holyoke, Massachusetts (Olsen et al., 2003a).



**Figure 10.** Composite section in vicinity of the Holyoke dam. Parsons Paper Co. well data from Emerson (1898). Rock color as in Fig. 5.

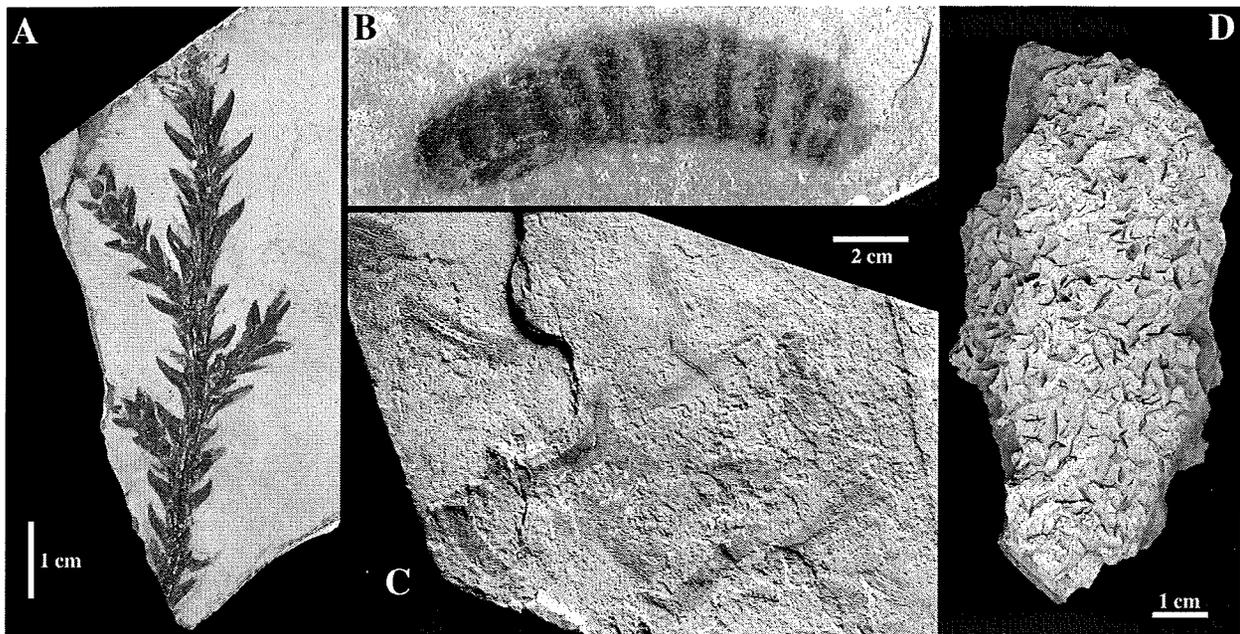
The South Hadley Falls member distinctively has a very large proportion of gray strata, a relatively muted expression of the short modulating cycles, and a high frequency of evaporite pseudomorphs in the lower half of the member (Fig. 10, 11). The base of the South Hadley Falls member is defined by the base of the first well-developed gray Van Houten cycle above the red bed sequence that comprises the upper Park River member. The top is defined by base of the gray strata of the overlying Mittineague member. The type section consists of exposures that comprise four transects along the Conrail railroad cut in Holyoke, Massachusetts and adjacent exposures in the bed and banks of the Connecticut River (and canals) in Holyoke and South Hadley Falls, Massachusetts. These outcrops are augmented by a short core (DH-1, Holyoke Power and Light) and records of a water well drilled for the Parsons Paper Company (documented by Emerson, 1898).

As seen at the type section, most Van Houten cycles have an unusually thick and gray division 2 composed of laminated mudstone but generally lack a microlaminated portion (see Stop 3). Division 1 tends to be relatively thin and silty as does division 3, with relatively uncommon desiccation cracks and footprints. Most of division 2 tends to be dominated by thin-bedded mudstones with macroscopic pyrite and delicate pinch and swell lamination. Evaporite pseudomorphs, probably after halite and gypsum, or glauberite and halite, are abundant in many layers, especially in the transition between division 2 and 3, sometimes constituting more than 50% of a bed by volume (Fig. 11). Some of these evaporite pseudomorphs have been described by Parnell (1983) and were commented on much earlier by Emerson (1898; 1917). Because of the relatively thick and gray division 2 in many Van Houten cycles, the short modulating cycles are relatively muted although still discernable compared to other parts of the Hartford Jurassic section.

Other areas in Connecticut that expose the lower South Hadley Falls member include those at Prout Brook (Stop 3) and Petzold's Marina where LeTourneau (1985a) and LeTourneau and McDonald (1988) first documented the apparent thickening of individual Van Houten cycle black shale beds across the basin approaching the border fault system. In Massachusetts, the member is exposed along streams in northern West Springfield, and in northeastern South Hadley (M. McMenamin, pers. comm.) and Granby, Massachusetts, where the entire member coarsens.

The red beds that comprise the upper South Hadley member are very poorly exposed. The only good outcrops are along Stony Brook at West Suffield (Chard Pond), Connecticut and at Granby, Massachusetts. Almost nothing is known about lateral changes in the red upper portions of this member because there are virtually no outcrops.

The thick-fine grained division 2 of the gray cycles have abundant conifer shoots and isolated cone scales to complete cones, equisetalian stems, and large, pieces of wood. Fragmentary fish are also present, and at least one bedding surface contains a mass mortality accumulation of insect larvae (Huber, et al, 2003). Division 3 of the mostly gray Van Houten cycles tend also to be thin- to flaggy-bedded and often have shallow and poorly preserved reptile footprints, especially brontozoids. The red beds at Granby have produced a large number of tridactyl dinosaurian footprints, mostly brontozoids, from a small but productive dinosaur footprint quarry (William Gringas Granby Dinosaur Museum).



**Figure 11.** A, large-leafed conifer, south side of canal, Holyoke dam (N.G. McDonald collection); B, unidentified larva, collected by B. K. Emerson in 1901 near Holyoke, Massachusetts; C, Unidentified larvae, downstream from the Holyoke dam, Portland Formation, Holyoke, Massachusetts; D, evaporite pseudomorphs after a ?sulfate, railroad cut near dam, Holyoke. B and C from Huber et al., 2003.

**Mittineague member:** We informally name the cyclical strata between the South Hadley Falls member and the Stony Brook member the Mittineague member after the village adjacent to the type section along the banks of the Westfield River (Figs. 6, 9). Strata of the Mittineague member to some extent resemble the Park River member and the East Berlin Formation, except that the Mittineague member tends to be more calcareous in both its lower and upper parts. This member, like the overlying member of the Portland Formation, is known almost entirely from outcrops, and although there are no long continuous sections a composite section can be readily constructed because of the distinctiveness of individual beds, in particular, the Chicopee bed.

Van Houten cycles of the lower half of the Mittineague member differ from those in the underlying South Hadley Falls member in having a much smaller proportion of gray strata, very much like the Park River member. The basal short modulating cycle of this member has a triplet of Van Houten cycles each with a gray and black division 2 lacking microlaminations. The base of the black shale in the lowest cycle of the triplet defines the base of the member. Red beds in this short modulating cycle have abundant septarian nodules that characterize much of the rest of the red beds of the lower half of the Mittineague member. Thin red claystone beds with conchostracans and ostracodes are also present. The second short modulating cycle of the member has a basal Van Houten cycle with a

thick black, but not microlaminated division 2. The second Van Houten cycle has a very distinctive division 2 that is very calcareous and microlaminated informally called the Chicopee bed (Olsen et al., 1989b).

Overall, the Chicopee bed lithologically resembles portions of the Westfield fish bed of the East Berlin Formation, but it contains a unique fossil fish assemblage dominated by the probable pholidophoridiform holostean "*Acentrophorus*" *chicopensis* (Newberry, 1888). It is easily recognizable at the three major sections of the Mittineague member and microlaminae are traceable for at least 12 km, assuring correlation of the sections.

Divisions 1 and lower parts of division 3 of Van Houten cycles in the lower part of the Mittineague member tend to have abundant beds of climbing ripple cross laminated siltstone and fine sandstone usually with abundant dinosaur footprints of the "leptodactylus" form (penetrating deeply into the bed). Desiccation cracks are usually present as well as septarian and other carbonate nodules.

The upper strata of the Mittineague member are characterized by meter-scale (sub-Van Houten cycle-scale) alternations of red mudstones and thin but laterally continuous gray claystones and associated carbonate layers. These palyniferous (Cornet, 1977) gray layers usually have abundant conifer fragments, conchostracans and ostracodes.

**Stony Brook member:** The unit we informally designate the Stony Brook member is the uppermost recognized cyclical member of the Portland Formation falling between the underlying Mittineague member and the overlying coarse clastic rocks of the undivided upper Portland Formation. It is named for the outcrops in the eponymous creek in Suffield, Connecticut near its mouth on the Connecticut River. Its base is defined as the base of the lowest black shale of the basal short modulating cycle of the member above the largely red beds of the upper Mittineague member.

Unlike the underlying members, Van Houten cycles of the Stony Brook member often have well-developed inclined beds (delta forests) even in the central portion of the basin. McDonald and LeTourneau (1988) described the geometry of one of the cycles. None of the cycles have microlaminated strata. As a whole, the member is less well-cemented than lower units and locally was used in brick making. The last active pit, the Kelsey-Ferguson quarry, closed in about 1992 and is now reclaimed.

Fossils tend to be abundant and vary in the laterally variable division 2 of the Van Houten cycles. Carbonate pellet lags at the base of inclined strata include bivalve escape structures, unionid bivalves, fish scales and bones, and oncolites (McDonald and LeTourneau, 1988). Red and gray claystones locally have abundant and well-preserved plant remains, fragmentary fish, conchostracans, ostracodes, and insects (Cornet and McDonald, 1995; Huber et al, 2003). The macrofloral assemblages are dominated by cheirolepidiaceus conifers, notably *Brachyphyllum* and *Pagiophyllum* and their reproductive structures (Cornet, 1977; Olsen et al., 1989a; Olsen et al., 2003a). Ripple crosslaminated siltstones and fine sandstones associated with divisions 1 and 3 of Van Houten cycles tend to have abundant leptodactylus dinosaur footprints. Footprints are also present as impressions, usually collectable as natural casts (sole marks).

The upper red strata of the Stony Brook member are substantially coarser than the lower members in the same basinal position with abundant interbedded bioturbated sandstone and thin beds of massive red mudstone. The best exposure of this facies is the CT Route 190 cut at Thompsonville, Connecticut. No Van Houten cycles are readily visible in this outcrop and this facies seems transitional into the overlying, yet coarser clastic rocks of the upper undivided Portland Formation.

**Upper Portland Formation (undivided):** The upper ~2 km of the Portland Formation is dominated, or perhaps composed exclusively of red fluvial strata (Hubert et al., 1982; McInerney, 1993). Very little is known about these strata otherwise. Several localities in the upper Portland have produced fragmentary to nearly complete skeletons of the prosauropod genera *Anchisaurus* and *Ammosaurus* and the crocodylomorph genus *Stegomosuchus* (Lull, 1953).

#### ACKNOWLEDGEMENTS

We would like to thank the landowners of the field stops for permission to visit their property. This manuscript benefited from discussions with Emma Rainforth on ichnotaxonomy, Bruce Cornet on cheirolepidiaceus conifers



Prout Brook flows east through a gorge from the dam near north end of Crystal Lake, Middletown, CT. Exposed in the beds of the brook and in the walls of the gorge is a 39 m thick section of mostly gray sandstone, pebbly sandstone, and dark gray mudstone overlain red mostly coarse clastics (Fig. 12). These outcrops were briefly mentioned by Rice and Foye (1927) and were described by LeTourneau (1985).

According to LeTourneau (1985a), the Prout Brook section lies about 530 m above the Middlefield bed outcrops at Laurel Brook (its type section) lying within the lower Park River member. The intervening strata are mostly red, coarse clastics with an intervening set of gray intervals that best outcrop along Long Hill Brook that almost certainly correlate with the two well-developed Van Houten cycles above the triplet that contains the Middlefield bed. Assuming no thickening, this would place the Prout Brook section within the South Hadley Falls member, with which it is lithologically consistent. Assuming 25% thickening towards the border fault, the Prout Brook section would fall somewhere in the mostly gray portion of the South Hadley Falls member.

The lower part of the Prout Brook section consists predominantly of planer to trough cross-bedded grayish brown conglomerate overlain by fining up cross-bedded gray sandstone (Fig. 13). This is followed by a dark gray to black laminated mudstone and siltstone sequence with abundant microfaults (*sensu* Olsen et al., 1989; Ackermann et al., 2003) and conifer fragments, that coarsens up through a series of beds alternating sandstone and siltstone beds, with ripple cross-lamination upwards. The upper half of the section is red with abundant planer and trough cross-bedded sandstone and some ripple cross-laminated siltstone, with the uppermost exposed section consisting of cross-bedded red conglomerate.

The finer-grained gray part of the section is characterized by both an unusually small amount of evidence for desiccation and no microlaminated unit. Unfortunately the outcrops do not permit us to determine what part of the lower South Hadley Falls member this section represents. However, the fact that the adjacent underlying section exposed along the shores of Crystal Lake consists entirely of red strata, plausibly of the upper Park River member, suggests that the Prout Brook section represents the lowest short modulating cycle within the lower South Hadley Falls member. A similar outcrop is present at Petzold's Marina in northern Portland, along the Connecticut River and this section presumably correlates with the Prout Brook member. These are amongst the only known outcrops of the South Hadley Falls member in Southern Connecticut.

Return to vehicle.

- 25.9 Start out going EAST on PROUT HILL RD toward SUNNYSLOPE DR.
- 26.0 Turn LEFT onto MILLBROOK RD.
- 26.8 MILLBROOK RD becomes E MAIN ST.
- 27.5 Stay STRAIGHT to go onto MAIN ST EXT.
- 28.1 MAIN ST EXT becomes CRESCENT ST.
- 28.2 CRESCENT ST becomes MAIN ST.
- 28.9 Turn SLIGHT RIGHT onto CT-66 / CT-17 / ARRIGONI BRIDGE. Continue to follow CT-66 / CT-17.
- 29.7 Turn LEFT onto SILVER ST.
- 29.9 Turn RIGHT onto BROWNSTONE AVE.
- 30.0 Turn right into viewing area for Portland Quarries, Portland, CT

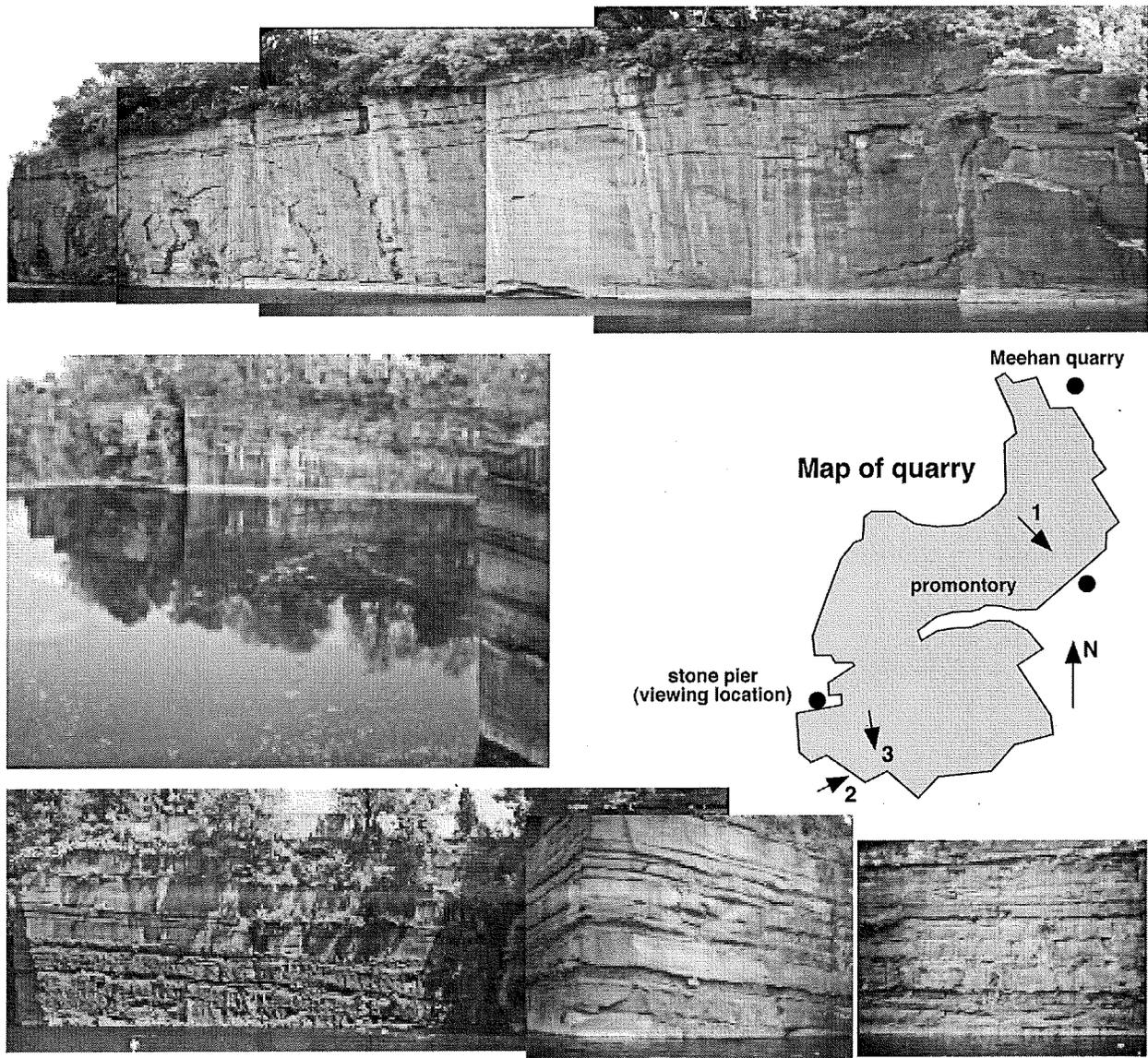
**STOP 2. BROWNSTONE QUARRY, PORTLAND FORMATION (30 MINUTES)** Central Middletown Quadrangle, (approx.) 41°34.50' N, 72°38.50' W; Tectonostratigraphic sequence TS IV; upper Park River member, Portland Formation; Hettangian age, 200 Ma. Main points are: extensive exposures of red and brown fluvial and eolian of the dry phase of a 405 ky cycle; strata quarried for building stone for over two centuries; abundant and very-well preserved reptile footprints and other fossils.

**Location, Stratigraphy, Sedimentology (contributed by Peter M. LeTourneau)**

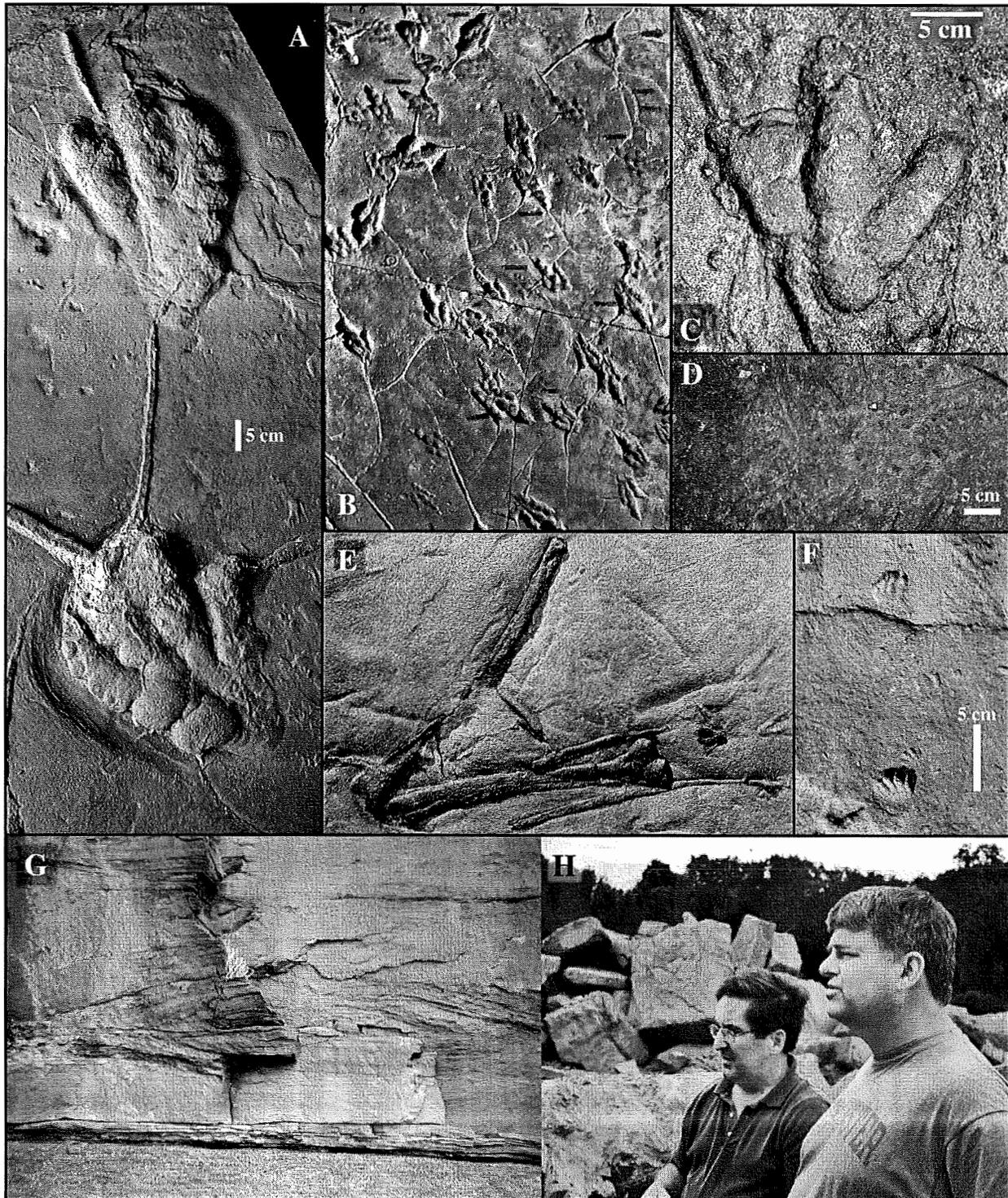
Located less than 3 kilometers from the eastern border fault of the Hartford basin, the Portland quarries occupy a position on the western edge of the Crow Hill Fan complex that has its depocenter near the high school on the top of Crow Hill. The eastern edge of the fan complex is observed along Route 17 in Portland, near the golf course, where boulder conglomerate is exposed in road cuts and natural outcrops. The southern margin of the fan complex is observed along the north side of Route 66 in the vicinity of the miniature golf course where thinning and fining

wedges of the fan conglomerate lithosomes are observed. The Connecticut River laps against the northern flank of the fan complex in the vicinity of Petzold's Marina where a long stratigraphic section of interbedded alluvial fan and dark gray to black, relatively deep water lacustrine deposits of the South Hadley Falls member, reminiscent of the Hales Brook Fan-Delta (Stop 3), but correlative to the lower South Hadley Falls member (Stop 1) are exposed.

Brownstone is the trade name for the Portland Formation arkosic sandstone ("Portland arkose"), which is made of quartz and feldspar sand with calcite and hematite cement. Re-examination of sedimentary features reveal that eolian deposits are a significant component of this sequence. These rocks contain sedimentary features attributable to sand sheets, low angle dunes, and linear "coppice" dunes. The eolian beds were apparently preferred for building stone because of their grain size and texture. The eolian beds alternate with fluvial beds in intervals about 15 m thick indicating possible cyclic climatic control on deposition, tracing out subtle Van Houten cycles.



**Figure 13.** Portland brownstone quarries. Top, view of southeast wall, Middlesex pit (see map for direction of view); center left, view of northeast wall, Brainerd pit (see map for direction of view); center right, map of quarry; bottom; View of southwest wall, Brainerd pit (see map for direction of view),



**Figure 14.** Portland quarries: A, *Otozoum*, trackway on exhibit at Dinosaur State Park (Stop 4); B, slab of numerous *Anchisauripus* that was a sidewalk stone in Middletown, CT (Amherst College, Pratt Museum, AC 9/14), probably from the Portland quarries; C, The largest known eastern North American *Anomoepus* on display at Dinosaur State Park (Olsen and Rainforth, 2003a); D, plant impressions, float, Meehan Quarry; E, natural cast in sandstone of the impression of a partial hind limb and pelvis of a theropod dinosaur (from Colbert and Baird, 1958); F, *Batrachopus* cf. *deweyii*, on block at quarry site; G, coppice dune in quarry wall; H, quarryman Michael Meehan (left) discusses the finer points of brownstone with paleontologist Nick McDonald.

The Portland brownstone quarries presently consist of two large, water-filled pits. The main (northern) pit, where we will focus, consists of the Middlesex quarry on the north side of the promontory and the Brainerd quarry on the south side. South of Silver Street, which bisects the quarries, are the Shaler and Hall workings. It is surprising that, given their obvious appeal to geologists, relatively little work has been done on the sedimentology and structure of the rocks, perhaps due to difficulties in approaching the towering quarry walls in the water-filled pits. Nevertheless, several studies including Krynine (1950), Lehmann (1959), Gilchrist (1979) and Hubert, Gilchrist and Reed (1982) provided partial descriptions of the Portland Formation rocks exposed in the quarries. The type section for the Portland Formation (Arkose), atypical though it is for the formation as a whole, was described by Krynine (1950). A detailed and captivating account of the history of the brownstone quarries may be found in Guinness (2003). LeTourneau (1985a, 1985b) discusses the rocks of the vicinity in context of the paleogeographic distribution of alluvial fan complexes located along the rift margin in central Connecticut.

The main pit ranges in depth from about 25 to 150 ft and is separated by a promontory that emerges from the eastern side of the quarry. South and upsection of the promontory the quarry walls consist of alternating layers of sandstone and sandy siltstone deposited mainly in a fluvial environment. LeTourneau (2002) and LeTourneau and Huber (2005) describe thin eolian beds intercalated with the fluvial rocks in the southern part of the main pit.

A view to the north and east shows the dramatically different character of the downsection rocks in the Middlesex pit, particularly the massive sandstone forming the large wall on the eastern side. Notably, the intercalated fine-grained rocks are nearly absent in this part of the quarry. A close view of the large east wall of the Middlesex pit reveals an abundance of inverse-graded low-angle inclined planar stratification indicative of migrating wind ripples (pin-stripe lamination) (Fig. 13). In addition several enigmatic large-scale convex-up dune forms may be observed. LeTourneau (2002) ascribed these unique sedimentary structures as "coppice dunes" formed around clumps of plants. Evidence for the coppice dune origin of these features includes, complex internal stratification with root traces, inverse-graded wind ripple lamination. The eolian beds were apparently preferred for building stone because of their grain size and texture. A modern model for the Portland brownstone eolian deposits is the Stovepipe Wells dune field in Death Valley, California. There a relatively thin sheet of dune sand overlaps alluvial fan and fluvial deposits on the edge of the extensional basin. Small coppice dunes anchored by plants are found in the interdune areas.

North of the Middlesex pit, Michael Meehan (Fig. 14) has brought the quarries full circle from early development, to abandonment, to renewed extraction of the historic brownstone. The Meehan quarry uses modern, non-explosive methods of extraction, and rather than the horse- and steam-powered equipment of the past, uses electric and diesel power to cut, shape, and transport brownstone. Standing on the promontory at the Meehan quarry we can peer into the geologic and historic past in the old quarries and see the modern re-emergence of layers that have not basked in the sun for over 200 million years.

Now the quarries enter yet another phase as well. In the 1990's the Town of Portland purchased the brownstone quarries and in 2000 the quarries were designated a National Historic Landmark. Currently, plans are underway to develop the quarries as a center for deep-water SCUBA training, rock-climbing, education, and outdoor recreation. Where, once, steam whistles signaling the start of the day for immigrant stone workers climbing steep ladders into the shadowed pits, shouts of "on-belay!" may soon reverberate from sun-drenched climbers clinging to thin holds on the sheer rock faces. From small crocodylomorphs scampering across sand flats in the Early Jurassic, to wind-blown dunes, to the peak of brownstone production, to the now placid waters of the abandoned pits, the geologic and historic story of the Portland brownstone quarries is truly remarkable.

Eolian sedimentation in the upper Park River member of the Portland Formation at Portland was promoted by both favorable paleolatitudinal position, deposition within the dry climatic interval of a 405 ky cycle, and proximity to fan-related sand. These deposits formed at about 22° paleolatitude, on the transition between the relatively humid tropics and the arid subtropics. The high-resolution correlations with arid to semi-arid intervals in the nearby Newark basin and the reinterpretation of paleomagnetic data of Kent and Tauxe (2005) support the hypothesis that the eolian sandstones are indicators of regional paleoclimate conditions, rather than just local depositional environments (Fig. 3).

### Paleontology

Quarrying of the brownstone in the Portland quarries revealed many fine reptile footprints, many of which are among the best preserved of their kind (Lull, 1953; Guinness, 2003; Olsen et al., 1998; Rainforth, 2003). Footprint taxa of biological significance from the quarries include the stratigraphically lowest occurrence of the sauropodomorph track *Otozoum* (Rainforth, 2003), the small brontozooids (*Anchisauripus* and *Grallator*), representing theropods, and the crocodylomorph ichnite *Batrachopus* (Fig. 14). *Otozoum* is relatively abundant in these strata possibly correlated with that, the large theropod track, *Eubrontes giganeteus* so abundant elsewhere, is conspicuous by its absence. A similar pattern is seen at least two other occurrences: first at the Moody Homestead locality (Olsen et al., 1998) also in the Park River member that produced the type of *Otozoum* (Rainforth, 2003); and at the McKay Head and Blue Sac localities in the McCoy Brook Formation of the Fundy basin (Olsen et al., 2005). This suggests ecological segregation of the track makers of the largest herbivore, represented by *Otozoum* and the largest carnivore *Eubrontes*, during the Early Jurassic.

The vast majority of footprints were collected as natural casts. The tracks themselves were made in thin mud layers, but lithified, the mud layers proved incompetent, crumbling away during the quarrying process. The sedimentological events that these mud layers represent is unclear. They may represent flooding events from adjacent streams, or the transgression of playas. The overlying sandstone beds that preserved the tracks as natural casts could crevasse splays or sheet floods. Despite the abundance of these tracks in numerous institutions their sedimentology remains essential unexplored.

In addition to footprints, the Portland quarries also has produced a fragmentary skeleton of a theropod dinosaur represented by a natural cast in sandstone of parts of the hind limb(s) of a small theropod (Colbert and Baird, 1958) (Fig. 14). Evidently, bones of a small theropod settled into a mud surface, were washed away leaving a natural mould, and like a footprint, were buried by sand washed in at a later time. This specimen was long thought to be lost, but it is archived at the Museum of Science in Boston (formerly, Boston Natural History Museum) (Rainforth pers. comm., 2003). Plants are represented by natural casts of tree limbs and trunks (Newberry, 1888; Guinness, 2003) and natural casts of root traces (Fig. 14).

Return to vehicle.

- 30.0 Start out going SOUTHWEST on BROWNSTONE AVE toward SILVER ST.
- 30.1 Turn LEFT onto SILVER ST.
- 30.3 Turn LEFT onto CT-66 / CT-17 / MAIN ST. Continue to follow MAIN ST.
- 32.7 MAIN ST becomes CT-17A / MEADOW RD.
- 33.3 Turn LEFT onto CT-17 / GLASTONBURY TURNPIKE. Continue to follow CT-17.
- 36.0 Turn LEFT onto OLD MAIDS LN.
- 36.7 Turn left into dirt road leading to sand pit, northern Portland, CT, walk to end.

**STOP 3. FAN DELTA OF THE HALES BROOK FAN, PORTLAND CONNECTICUT AND MIDDLEFIELD FISH BED OF THE PARK RIVER MEMBER (60 MINUTES).** Central Middletown Quadrangle, (approx.) 41°38.00' N, 72°37.25' W; Tectonostratigraphic sequence TS IV; lower Park River member, Portland Formation; Hettangian age, 200 Ma. Main points are: very coarse alluvial fan conglomerate and microlaminated, fossiliferous black shale of the wet phase of a 405 ky cycle; proximity to border fault; Middlefield fish bed.

### Stratigraphy and Sedimentology (contributed by Peter M. LeTourneau)

In 1991 an extraordinarily illustrative exposure of very coarse alluvial fan conglomerate and microlaminated, fossiliferous black shale was uncovered in a sand and gravel quarry in northern Portland, Connecticut. This exposure is an unsurpassed example of fan-delta deposition at the faulted margin of the Hartford Basin. The purpose of this stop is to illustrate this remarkable fan delta sequence within the context of lower Jurassic depositional environments of the Portland Formation in central Connecticut.

Stratigraphically, the section lies about 100 m above the projection of the Hampden Basalt. The exposures therefore lie within the expected position of the lower Park River member. Lithologically, only the Middlefield bed

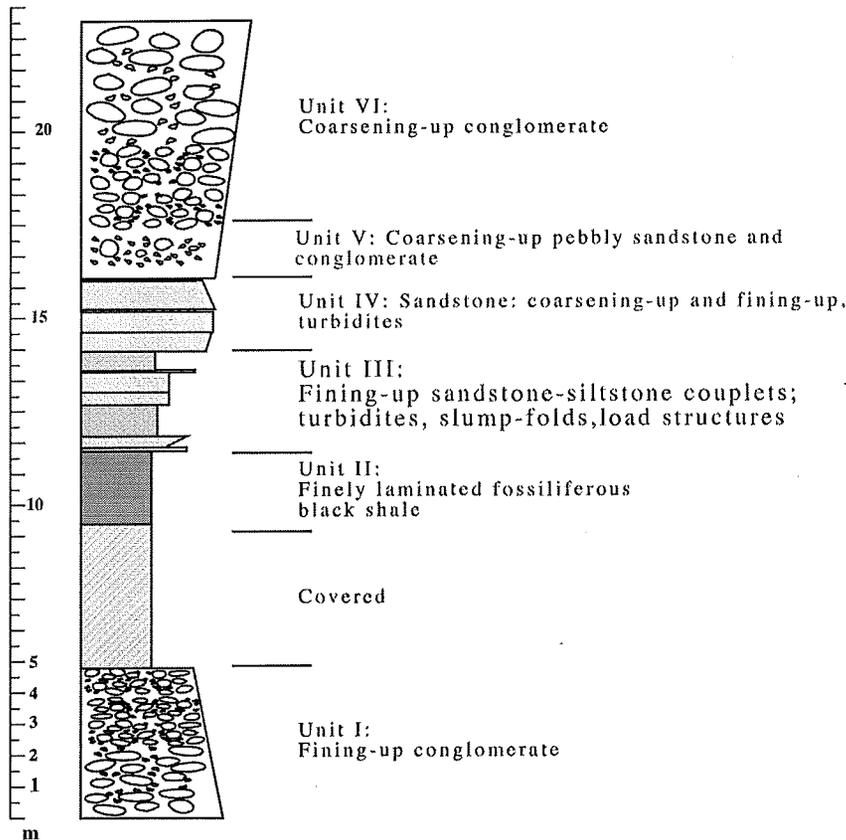


Figure 15. (Right) Measured section at Stop 3.

within the Park River member resembles the exposed fish-bearing unit and therefore we conclude that the exposed unit represents that bed and the overlying portions of the same cycle.

In the context of previous studies of the alluvial fan deposits of the Portland Formation, this new occurrence is considered the type fan delta facies association in the Hartford Basin for the following reasons: 1) accessibility and lack of vegetative or soil cover; 2) great range of grain size, from clay to boulders over 2 m in length in close stratigraphic proximity; 3) evidence of rapid sub-aqueous fan deposition including turbidite beds and extraordinary slump folds; 4) abundant fossils including fish,

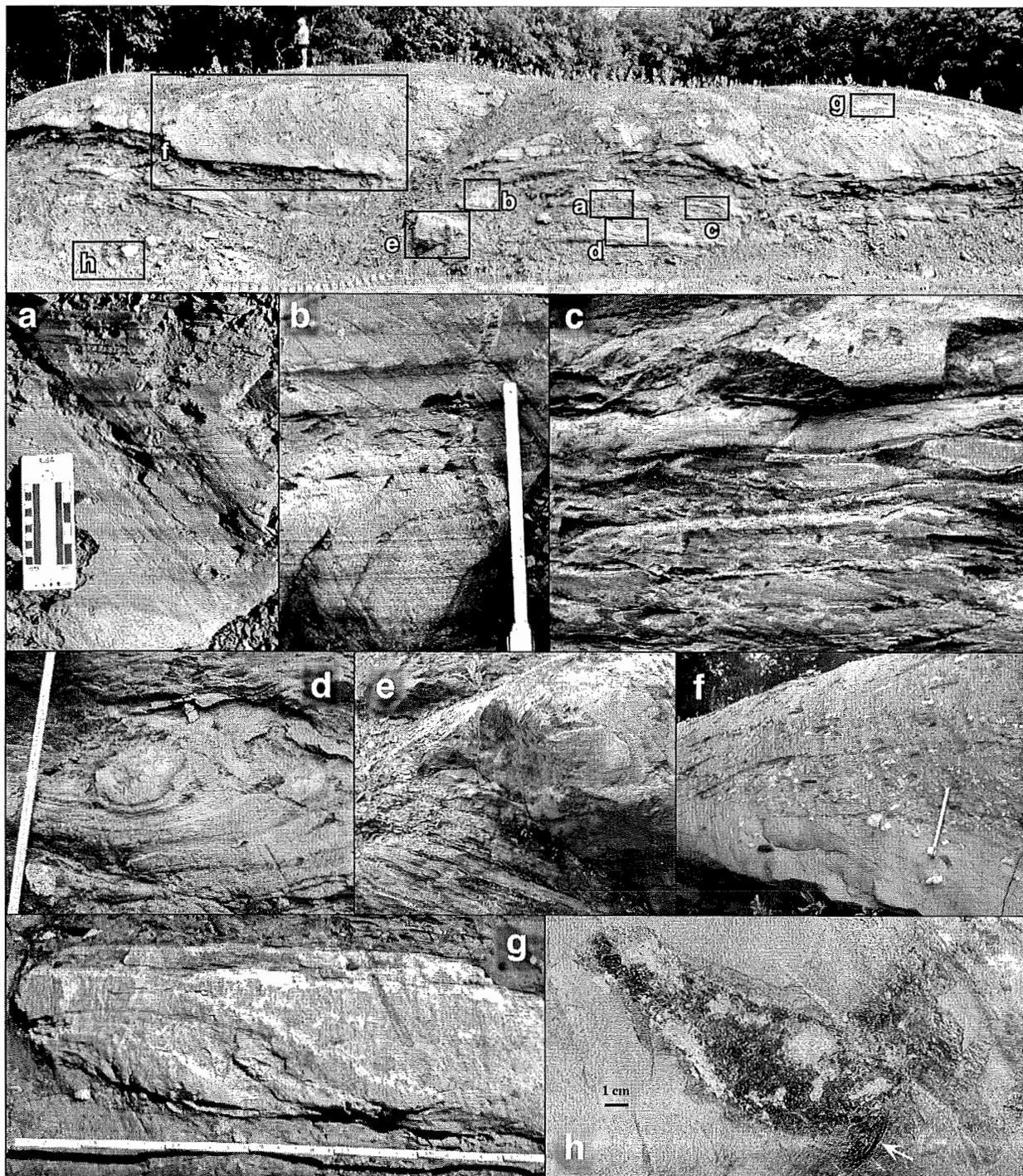
conchostracans, and plants; 5) dramatic coarsening-up sequence; 6) one of the few well-exposed fan-lake sequences found within 1 km of the eastern fault margin of the basin; and 7) extraordinary three-dimensional exposure showing bedding plane, along-strike and along-dip views of beds and sedimentary structures.

The exposure has great utility for educational purposes for demonstrating diverse geological principles including the following:

- *Sedimentology*: finely laminated lake shale, shallow water near shore sandstone, sub-aqueous mass flow deposition (turbidites), slump folds and soft-sediment deformation, sub-aqueous and sub-aerial conglomerate; fan progradation, Walther's Law;
- *Structure*: post-depositional normal faults, effects of bed strength on fault plane attitude, bedding plane shear
- *Paleontology*: modes of fossil fish preservation, paleoecology, lacustrine environments
- *Glacial Geology*: glacial striations, effect of bedrock on ice flow; ice contact deposits.

The exposure was uncovered during routine excavation of glacial outwash for sand and gravel. Previous geological reconnaissance revealed the presence of small scattered outcrops of boulder conglomerate at elevations above 200 ft (msl) on the small isolated hill located east of the sand and gravel quarry, but no dark shale beds were previously found in the area (LeTourneau, 1985a).

The exposure is about 150 - 200 ft. long and consists of a north-facing vertical section of dark shale, sandstone and conglomerate and a broad, south-facing dip slope of boulder conglomerate. Bedding strikes N 55° E and dips about 25° SE (110-25). The eastern and western ends of the outcrop are terminated by normal faults striking roughly north-south. The entire outcrop shows evidence of glacial scour with deep, sub-parallel grooves and scratches that envelope bedrock surfaces. The glacial striations also wrap around the eastern and western ends of the exposure, providing dramatic evidence that bedrock influenced the local flow path of the base of the overlying ice sheet.



**Figure 16.** Fan delta sequence at Stop 4: Top, composite of view of units II – VI and key to following; a, turbidite layers; b, Fining-up turbidite thin beds and laminae with pseudo-flame structures caused by loading; C, Contorted turbidite layers caused by loading and possible shear; d, Large turbidite load slump (Note fining-up layers within); e, large turbidite load slump with highly contorted internal structure; f, coarsening-up conglomerate of unit V; g, gneiss boulder from unit VI; h, articulated *Redfieldius* from Middlefield bed (unit II), McDonald collection.

The exposed strata coarsen dramatically upward and exhibit an extremely wide range of grain sizes and bedding style, from thin-bedded finely laminated dark shale to crudely stratified boulder conglomerate within the relatively thin, but continuous vertical section. The depositional environments represented include deep-water lacustrine, shallow or littoral lacustrine, sub-aqueous fan-delta, and sub-aerial alluvial fan. A measured section (Fig. 15) shows the stratigraphic succession from very fine grained lacustrine beds through exceedingly coarse boulder conglomerate beds.

**Hales Brook Fan Delta:** This outcrop is located within eastern fault margin of the basin in an area of very coarse conglomerate previously termed the Hales Brook Fan (LeTourneau, 1985b). The fan delta sequence is divided for reference into 6 units (Fig. 15), from base to top. Unit I at the base of the section is a poorly exposed pebble and cobble conglomerate more than 4 m thick. The exposed portion of conglomerate is undoubtedly part of a larger conglomerate sequence as suggested by bedrock "float" and the east-west oriented resistant ridge that contains Unit I. A covered interval of approximately 4.5 m overlies Unit I and forms the base of Unit II; 2.2 m of finely laminated, fossiliferous black shale. Unit III is about 4 m thick and consists of three fining-up sub-units of interbedded sandstone and siltstone. Unit IV is a 0.5 m bed of coarse to pebbly sandstone with a sharp basal contact and gradational upper contact. Unit V forms the top of the exposed section and consists of very coarse boulder conglomerate. The sedimentary features and paleoenvironmental interpretation of these units are discussed below.

**Unit I:** The base of the outcrop consists of poorly-sorted, poorly-stratified conglomerate with few thin silty sandstone partings or thin beds. Although the exposure of this bed is poor, a fining-up trend can be observed in the sized of the major conglomerate clasts. Clast composition is polymict and reflects the varied low- to high-grade metamorphic rocks and plutonic igneous rocks. Clasts are sub-rounded to sub-angular; no preferred orientation could be observed. Unit I is interpreted as an alluvial fan deposit based on comparison with other well-known ancient examples including the Hartford Basin (e.g., Steel, 1976; Nilsen, 1982; Letourneau, 1985a) and analysis of modern depositional environments (e.g., Bull, 1972; Nilsen, 1982).

**Unit II:** The base of Unit II is a 2 m covered interval. The upper 0.75 m consists of microlaminated to finely laminated, thin bedded black to dark gray, organic-rich, fossiliferous shale (Fig. 16). Bedding contacts are sharp and planar and laminations may be traced through the exposed portion of Unit II. The upper portion of Unit II contains a few fining-up siltstone interbeds or lenses. Laminations consist of clastic and carbonate couplets (*varves*, sensu Olsen, 1986). Evidence of subaerial exposure such as mud cracks or root burrows is not present nor are invertebrate burrows observed. Fish fossils are whole and articulated, but are in some cases disrupted by normal faults and bedding plane faults of small displacement that penetrate portions of the outcrop. These features correspond to those described by Olsen (1986, 1990) for deep water lacustrine strata within the Newark Supergroup. The sedimentary structures and preservation of whole articulated fish are indicative of deposition below wave base and the thermal or chemical stratification of the lake water column. Therefore, Unit II is interpreted as a perennial, stratified lake deposit based on comparison to examples of modern and ancient rift lakes (e.g., Olsen, 1990; Dean, 1981; Trewin, 1986).

Fossils from Unit II include the holostean *Semionotus*, conchostracans, and plants. The excellent preservation of bone in the wholly articulated fish is typical of fossil fish found within a lateral distance of 1 to 2 km of the eastern basin margin. McDonald and LeTourneau (1989) attribute geographic trends in fossil fish preservation to relatively high rates of sedimentation adjacent to the basin margin. The relatively rapid burial of fish carcasses prevents microbially mediated dephosphatization of bones (Nriagu, 1983) that is occurs in lake beds located in central and western areas of the Hartford Basin. In addition, the size of Unit II conforms the bed thickness predicted by the previously determined relationship of the thickness of division 2 and distance from the faulted basin margin for the Hartford Basin (Fig. 17).

**Unit III:** Unit III consists of 1.3 m of interbedded gray mudstone, thin normal-graded sandstone to siltstone beds, and light gray to brown medium to coarse sandstone with subordinate granule lenses and small pebble layers and lenses. The gray mudstone is thin-bedded with planar horizontal lamination, and minor pinch and swell lamination. The mudstone beds include rusty-weathering ferroan dolomite-rich (e.g., Hubert et al., 1978) beds and nodules. The thin normal-graded sandstone-siltstone beds occur in repetitive bed sets throughout Unit III. Thick slump folds and soft sediment deformation are noteworthy in this Unit.

A large fold is observed in the central portion of the exposure. The recumbent fold is about 1 m thick and 1.5 m wide with a rounded lower boundary and a shallow, convex upward upper surface. Laminations of fine and coarse sand, often as normal-graded couplets, within the slump are crenulated, recumbent, and of different attitude than the laminations on the outermost portion of the slump fold, indicating that the internal deformation was somewhat independent of the shear forces that shaped the outer boundaries of the slump fold. About 3 m to the west of the largest slump fold is another smaller slump about 0.5 m thick by 1.2 m wide. This fold is also has complex internal deformation and recumbent, multiply folded laminations. The slumps appear to be part of a formerly continuous coarse sandstone wedge that tapers toward the west, away from the basin margin.

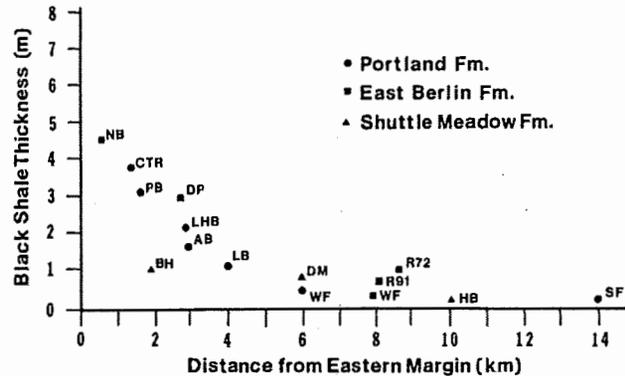
Beds overlying the folded horizon are thin bedded and laminated gray mudstone and normal-graded sandstone-siltstone couplets. In the upper few centimeters of the mudstone beds a complexly deformed 2-4 cm medium sand bed forms a "train" of recumbent and overturned folds. We interpret that the folded sand layer is a result of bed-shearing forces during the mass flow emplacement of the overlying thick sand wedge that forms the base of Unit IV.

Unit III is the product of mass-flow, turbidite-dominated sedimentation in a pro-delta environment. The normal-graded couplets and folds are interbedded with gray mudstone as opposed to finely laminated fossiliferous black shale, indicating the progradation of the fan-delta over organic-rich lake bottom sediment.

**Unit IV:** Unit IV is similar to the underlying Unit III but overall coarser-grained and less dominated by the large folds that characterize Unit III. The base of Unit IV is formed by a massive coarse to granule sandstone bed with a wedge-like shape that progressively thins from 40 cm in the eastern portion of the outcrop to 20 cm in the western portion. The lower surface of the sand bed is generally sharp and planar, although some soft sediment load structures are observed. The upper surface is hummocky to irregular and is "onlapped" by normal-graded sandstone beds in the western portion of the outcrop. Internally, the sand bed is massive to laminated; cross stratification is not observed. The normal-graded sandstone beds that overlie the massive sandstone wedge are about 20 to 30 cm thick and consist, internally, of thin, 10 to 3 cm, normal-graded couplets. The apparent "onlap" of these beds on the massive sandstone bed is caused by the westward thickening of the normal-graded beds.

**Unit V:** Unit V is an abrupt grain-size transition from the underlying turbidite sandstone. Very coarse sand and fine gravel predominate the lower part of the unit and pebbles and cobbles appear in higher abundance toward the top of the unit. A few scattered pebbles and cobbles "float" within the matrix throughout the unit. Typical fluvial sedimentary structures are missing from Unit V, including crossbedding and lag gravels, nor are typical deltaic features, such as, climbing ripple cross-laminae, load casts, or sorted layers and lenses, present. Furthermore, features of typical sub-aerial alluvial fans, including fining-up layers and lenses with silt drapes, or debris-flow lenses are not in evidence. Therefore the depositional environment of the Unit remains somewhat enigmatic, although a sub-lacustrine origin seems likely.

**Unit VI.** Unit VI is a spectacular, coarsening-up cobble and boulder conglomerate. The conglomerate may be observed in three dimensions, including breathtaking bedding plane views on the south side of the outcrop. The largest clasts, ranging up to 2 meters in length, are observed in the highest stratigraphic levels in the outcrop. The polymict conglomerate includes both low- and high-grade metamorphic rocks derived from the Paleozoic eastern highlands; phyllite, gneiss, and quartzite are all common, and one basalt clast was observed. Stratification within Unit VI is obscure. In many places grain size segregations suggestive of bedding are observed, but, except in a few locations, crossbedding is faint or absent. As in Unit V evidence of sub-aerial alluvial fan deposition, including, but not limited to, well-defined channels, silt drapes, or debris flow lobes, is absent. This unit was likely deposited at



**Figure 17.** Relationship between "black shale" thickness (division 2) and distance from the boarder fault (from LeTourneau, 1985a).

and below the lake margin where rapid sedimentation below water resulted in poorly sorted and poorly segregated, amalgamated conglomerate with faint bedding.

**Paleoenvironments - Alluvial fans:** Well-developed alluvial fan deposits are found along the eastern fault margin of the Hartford Basin from North Branford to South Glastonbury, Connecticut, specifically at this stop. In particular, the alluvial fan deposits of the lower Portland Formation in Durham, Middletown, and Portland have been extensively studied (Gilchrist 1979, LeTourneau, 1985a, 1985b). These deposits consist of coarse sandstone and conglomerate in coarsening- and fining-up sequences with intercalated, fossiliferous, finely laminated lacustrine dark shale. The alluvial fan deposits form discrete prisms of coarse-grained rocks within finer-grained fluvial sandstone and siltstone, and dark shale lacustrine mudstone deposits of the basin floor. The alluvial fan deposits can be separated into two general types of facies associations. The relatively large (1 to 4 km radius) basin margin alluvial fans, typically show evidence of fluvial activity, such as cross stratification, channels, and thin fining-up beds and features of wave reworking and shoreline modification of fan sediment is often observed. Large boulders may be found, but these are generally a small percentage of the total conglomerate clast population and they occur in discrete boulder layers, including debris flows, or as isolated clasts. In contrast, the smaller radius fans (0.5 - 1 km), perhaps more related to steep talus fans, are: characterized by extremely poor stratification and internal fabric organization; are exceedingly coarse-grained, and are thick bedded with little cross-stratification or channelization evident. Large boulders comprise a significant percentage of the conglomerate clasts often within a fine sand and silt matrix.

The contrast in the two fan types indicate differences in the dominant depositional processes (e.g., stream-flow, debris-flow, colluvial) which are linked to source area drainage basin size, catchment and fan slopes, and stage of fan development (Blair and McPherson, 1994, fig 20, pg. 474). For reference, the alluvial fan deposits of the Portland Formation in central Connecticut have been identified by their association with modern geographic features, e.g., Round Hill Fan, Reservoir Brook Fan (LeTourneau, 1985b). The alluvial fan - lacustrine sequence described here is located within the Hales Brook Fan complex.

**Paleoenvironments - Perennial Lakes:** The intercalated dark shale beds are typically very finely laminated, lack evidence of bioturbation or sub-aerial exposure, contain fine carbonate (calcite and dolomite) laminae, often contain exquisitely preserved fish fossils, and are laterally continuous over broad areas of the Hartford Basin. The dark shale beds are the result of sedimentation in deep-water, stratified, perennial lakes (Olsen, 1984) Stratification of the water column promotes anoxic bottom water conditions which in turn allows the preservation of abundant organic matter and excludes benthic epifauna and infauna resulting in finely laminated, organic-rich black shale with fully articulated fish fossils. As demonstrated in the Newark basin and the East Berlin Formation in the Hartford Basin (Stop 5), the periodic occurrence of perennial lakes within the vertical stratigraphic section of these basins is a result of astronomical forcing (Milankovitch cycles) of regional climate in the late Triassic and early Jurassic (Olsen, 1986, 1990; Olsen et al. 1989a). Therefore, the presence of fossiliferous dark shale beds within conglomerate beds as seen at this stop represent the climatically controlled transgression, high stand, and regression of perennial lakes over coarse basin margin alluvial fan deposits.

**Fan Deltas.** Fan delta deposits are formed where alluvial fans intersect marine or perennial lake waters (Wescott and Ethridge, 1990). Distinct from wholly sub-aerial alluvial fans, fan deltas consist of two portions: 1) the sub-aerial alluvial fan deposited above mean water level and: 2) the sub-aqueous portion deposits below standing water (Nemec and Steel, 1988; Blair and McPherson, 1994). Although alluvial fans are deposited by water-laden mass flows, ephemeral to perennial stream flow, and sheetfloods, in this context "sub-aerial" means all deposition above mean water level of a perennial lake. The sub-aqueous portions of fan deltas show evidence of coarse-grained deposition into profundal waters dominated by fine grained sedimentation, and typically contain turbidite beds and soft sediment slumps and folds generated on unstable sub-aqueous slopes (Postma, 1984). Fan delta sequences typically coarsen-upward as a result of the progradation of coarse-grained littoral, shoreline, and alluvial fan sediment. These features are well represented in the Hales Brook Fan fan delta sequence. A progradational sequence may result from sediment deposition on the alluvial fan during relatively static lake levels or from the "forced" regression of lake margin deposits during receding lake levels. We interpret the Hales Brook Fan delta as the progradation of an alluvial fan *into* deep water during a lake high stand because of the intimate association of exceedingly coarse grained deposits with anoxic, profundal lake water and the absence of features indicative of sub-aerial exposure. Other alluvial fan deposits in the lower Portland Formation show evidence of progradation following climate-driven lake regression.

### Paleoecology and Evolution

Characteristic of the fish assemblage of the Middlefield bed is the large amount of variation seen in individuals of the holostean genus *Semionotus*. Most of this variation is apparently attributed to morphological differences between different species. The presence of many endemic species of the same genus (in closely related genera) is termed a species flock. This *Semionotus* species flock are akin to their counterparts in the Westfield bed of the East Berlin Formation, and the Bluff Head bed of the Shuttle Meadow Formation (Stops 5 and 7) and are analogous to species flocks of cichlid fishes of East African great lakes (Olsen, 1980; McCune et al., 1984; Olsen and McCune, 1991; McCune, 1996). In this case, all the species seem to belong to one limited clade (the *S. elegans* species flock of Olsen and McCune (1991), which is also the case in the precisely correlative Van Houten cycle in the Boonton Formation of the Newark basin. *Semionotus* species flocks are limited to Newark Supergroup strata above the Triassic-Jurassic boundary, although fish assemblages are abundant in older Newarkian strata, and suggesting that perhaps the evolution of the species flocks was fostered by the extremely low post-Triassic-Jurassic boundary fish diversity in the watersheds of the rift basins.

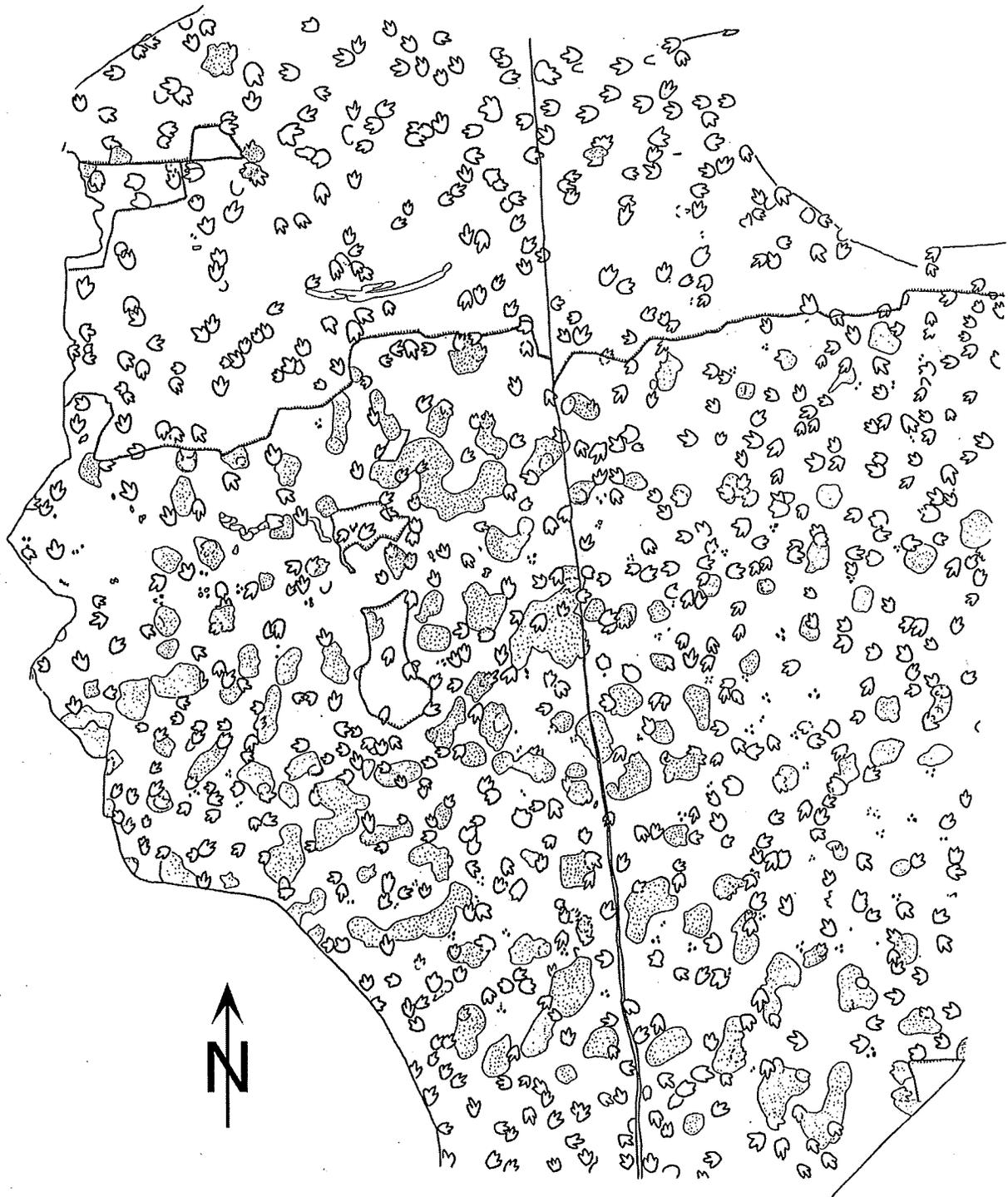
The only other fish genus found in the Middlefield bed here or elsewhere is the palaeonisciform *Redfieldius*. However, the presence of abundant phosphatic coprolites strongly suggests the presence of the coelacanth *Diplurus longicaudatus* that has been found with such coprolites within its body (Gilfillian and Olsen, 2000). The conchostracan *Cornia* is also abundant as are various plant fragments, including large stems.

Return to vehicle.

- 36.7 Start out going EAST on OLD MAIDS LN toward CT-17 / MAIN ST.
- 37.4 Take CT-17 N.
- 42.9 CT-17 N becomes CT-2 W.
- 43.8 Merge onto CT-3 S via EXIT 5D toward WETHERSFIELD.
- 46.8 Merge onto I-91 S via the exit on the LEFT toward NEW HAVEN.
- 50.9 Take the WEST STREET exit- EXIT 23- toward CT-3 / ROCKY HILL.
- 51.1 Turn LEFT onto WEST ST.
- 52.0 Turn right into parking lot for at Dinosaur State Park, 400 West St Rocky Hill, CT

**STOP 4. DINOSAUR STATE PARK, EAST BERLIN FORMATION (LUNCH - 60 MINUTES)** SE Hartford South Quadrangle, (approx.) 41°39.03' N, 72°36.48' W; Tectonostratigraphic sequence TS IV; East Berlin Fm.; Hettangian age, 201.5 Ma. Main points are: Abundant large theropod dinosaur tracks (*Eubrontes*) in regressive portion of upper gray and black Van Houten cycle of East Berlin; extreme rarity of herbivores; post boundary ecological release; no evidence for herding in theropod dinosaurs.

This site was discovered in 1966 during excavation for the foundation of a state building. Exposures here have revealed nearly 2000 reptile tracks, most of which have been buried for preservation and future exhibition. The present geodesic building at the park houses approximately 500 tracks (Fig. 18). The tracks are found in the gray arkoses, siltstones, and mudstones of the East Berlin Formation. The stratigraphic position of the main track-bearing horizon has been a bit of conundrum. Byrne (1972) correlated the track-bearing unit with the uppermost Van Houten cycle that has black mudstone in the East Berlin Formation, placing it about 20 m (17.4 m) below the contact with the Hampden Basalt, a correlation followed by Olsen et al. (2003a). However, close examination of these sections through Byrnes' descriptions of the cores shows that the thickness of red beds below the track level is too great to be accommodated by Byrne's proposed correlation. In fact, the only interval with such a long red bed sequence in the upper East Berlin is the third black shale-bearing cycle from the top. In outcrop along Rt, 9 (Stop 5) this cycle also has a second dark shale unit in the upper part of division 2, a feature matched at the track site. Given the extremely slow lateral change in thickness of Van Houten and short modulating cycles observed laterally, the latter is the most parsimonious hypothesis, and that would place the track horizon at about 38 m below the Hampden Basalt.



**Figure 18.** Dinosaur footprints on display *in situ* at Dinosaur State Park (from Farlow and Galton, 2003). Most footprints are 30-40 cm long. Stipple illustrates where overlying rock did not separate cleanly from the upper track-bearing layer, cross-hatched lines indicate boundary between the upper and the lower track-bearing layers (cross-hatches directed toward the lower layer). Groups of circles in a triangular pattern indicate footprints made by swimming (?) dinosaurs.

Thus, the track-bearing surface in the enclosure is part of the regressive portion of a Van Houten cycle in part of the wettest phase of a short modulating cycle. Ripple marks, raindrop impressions, mudcracks, and the footprints indicate shallow-water conditions and some subareal exposure. The track-bearing beds grade upward into dark gray mudstone followed by red sandstone and mudstone, suggesting a return to slightly deeper water conditions prior to conversion of the lake into a playa.

The typical Connecticut Valley ichnogenera *Eubrontes*, *Anchisauripus*, *Grallator*, and *Batrachopus* have been identified at this locality (Ostrom and Quarrier, 1968). All but *Batrachopus* were made by small to large theropod (carnivorous) dinosaurs. *Batrachopus* was made by a small, early, fully terrestrial protosuchian crocodylian. *Eubrontes giganteus* tracks are the most common and are the only clear tracks visible in situ within the geodesic dome. Because of the popularity of this site, *Eubrontes* is now the Connecticut state fossil. Based on what appear to be claw drags without any pad impressions, Coombs (1980) suggested some of the track makers were swimming, but Farlow and Galton (2003) argue that equivalent tracks were made by the *Eubrontes* trackmaker walking on a hard substrate. This is yet another example of the low diversity, theropod dominated assemblages typical of strata only a few hundred thousand years younger than the boundary.

*Eubrontes giganteus* has the appropriate size and pedal morphology to be made by a dinosaur the size of the ceratosaurian theropod *Dilophosaurus*. Olsen et al. (2002a) show that *Eubrontes giganteus* appears within 10 ky after the Triassic-Jurassic boundary and that it represents the first evidence of truly large theropod dinosaurs. The largest Triassic theropods *Gojirosaurus* and *Liliensternus* were less than 80% the size of the larger *Dilophosaurus*, despite statements to the contrary by Lucas (2002). This size difference compares well to the disparity between the largest Newarkian Triassic *Anchisauripus* (25 cm) and *Eubrontes giganteus* (35-40 cm) Olsen et al., 2002a). This difference in length scales to roughly more than a doubling of mass. As described by Olsen et al. (2002a, 2003a), the appearance of these larger theropods could be due to either an abrupt evolutionary event or an immigration event. Although we cannot currently distinguish between these two possibilities, we favor the former in which the size increase is an evolutionary response to "ecological release" in which extinction of the Triassic top predators (rauisuchians and the phytosaurs), allowed a very rapid increase in size in the absence of competitors.

- 52.0 Start out going WEST on WEST ST toward GILBERT AVE.
- 52.9 Merge onto I-91 S via the ramp on the LEFT toward NEW HAVEN.
- 54.5 Merge onto CT-9 N via EXIT 22N toward NEW BRITAIN.
- 57.5 Take EXIT 22 toward US-5 S / NEW HAVEN / CT-15 S.
- 57.6 Turn RIGHT onto FRONTAGE RD.
- 57.7 Turn RIGHT onto CT-372 / WORTHINGTON RDG.
- 57.8 Go straight onto ramp for CT-9 S
- 57.9 Park on dirt on right, off road.

**STOP 5. CT RT 9 ROAD CUTS IN EAST BERLIN AND HAMPDEN BASALT, BERLIN, CT. (60 MINUTES).** Northeast corner Meriden Quadrangle (41°37.37'N, 72°44.33'W); Tectonostratigraphic sequence TS IV; East Berlin Fm. and Hampden Basalt; Hettangian, ~201.5 Ma. Main points are: cyclical middle and upper East Berlin Formation; remarkably close match to Towaco Formation; deep-water lacustrine carbonate with dephosphatized fish; superb pattern of Van Houten and short modulating cycles; "Dead horses" dewatering? structures; very thin basaltic crystal tuff; HFTQ-type quartz basalt of the Hampden Basalt.

### Stratigraphy and Cycles

The intersection of US 15 and CT 72 reveals over 120 meters of the upper two-thirds of the East Berlin Formation and almost the entire thickness of the Hampden Basalt (Figs. 19, 20). The exposure consists of cyclical red, gray, and black lacustrine units and subordinate fluvial strata that document dramatic changes in the depths of the lakes over relatively short periods of time. Krynine (1950) first recognized the lacustrine cycles here; later work was carried out by Klein (1968), Hubert and Reed (1978), Hubert et al. (1976, 1978), Demicco and Gierlowski-Kordes (1986), Suchecki et al. (1988), and Kruger et al. (1990). The Van Houten cycles are virtually identical in form, but not in all fabrics, to the Van Houten cycles described from the Towaco Formation (Olsen, 1980, Olsen et al., 1996b).

Hubert et al. (1976) demonstrated the lateral continuity of the Van Houten cycles. The three cycles in the upper part of this section correlate with the three cycles exposed a few hundred meters to the north on CT 72 and the three cycles exposed at the I-91/CT-9 cloverleaf, ~2 km to the east (Hubert et al., 1978). Visible at this stop and on CT 72, a 35m thick section of red and minor gray and purple clastics separates the upper cycles from three underlying cycles in the middle East Berlin Formation. The uppermost of these cycles contains a black, microlaminated carbonate called the Westfield bed. The most distinctive bed in the East Berlin Formation, it preserves a characteristic assemblage of fishes that include *Semionotus*, *Redfieldius* and surprisingly common large coelacanths (*Diplurus* cf. *D. longicaudatus*). Correlation of the microlaminae and turbidites in polished slabs from the middle of the fish-bearing carbonate is further evidence for lateral continuity of these units (Olsen, 1988).

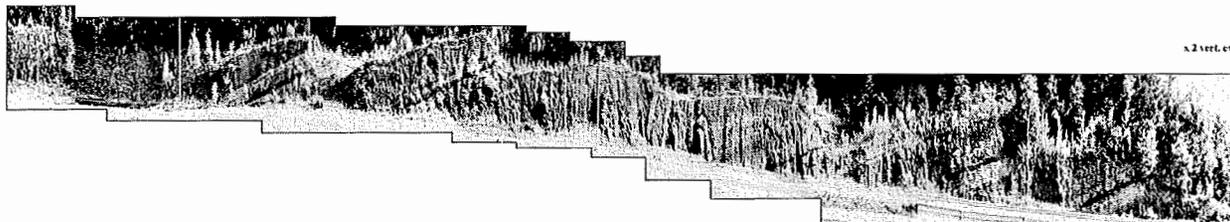


Figure 19. Photo mosaic of exposures on south side of southbound entrance ramp for CT Rt. 9, Stop 5.

Fourier analysis of proxies of water depth of this section reveals a clear hierarchy of periodicities in thicknesses of 12.0 m and 68.3 m. Assuming that the 12.0-m-thick Van Houten cycles are the 21,000-year precession cycles, the 68.3-m-thick short modulating cycles have time periodicities of 119,000 years. There is a clear hierarchy of cycles similar to that seen in the Lockatong and Passaic Formations. This sequence represents the wet part of a McLaughlin cycle within the wet part of long modulating cycle. Evidently, the lakes that produced these cycles in the Newark and Hartford basins rose and fell synchronically, controlled by regional climatic change, without necessarily being contiguous bodies of water (e.g., Fig. 9).

During the deposition of TS III of the underlying New Haven Arkose, sediment and water supply were mostly from the footwall (east) side of the basin. During Early Jurassic time, in TS IV the drainages reversed with the western hinge side of the basin becoming the source for the bulk of the sediment fill. This can be seen in the cross bedding at this outcrop (LeTourneau and McDonald, 1988). The reversal of drainage probably reflects increased asymmetry of the basin and correspondingly high rates of footwall uplift during an Early Jurassic pulse of increased extension associated with the igneous episode and TS IV. Extensive lacustrine deposition occurred in the Hartford basin only during this period of increased extension. Lacustrine deposition is favored by high extension rates, whereas low extension rates favor fluvial deposition.

**“Dead Horses”:** The black portions of division 2 of these Van Houten cycles show some of the deformation features common to Newark Supergroup lacustrine strata. Small-scale thrust fault complexes associated with bedding-parallel shear zones are commonly developed in the microlaminated black shales of division 2 as seen here. At this outcrop and in the East Berlin and Towaco formations, black mudstone units above and sometimes within the microlaminated intervals contain what termed by Olsen et al., (1989a) “dead horses”, small (usually <20 cm long) quadrangular pods of rock composed of laminae and layers oblique to the bedding surface. Similar structures occur in the Eocene Green River Formation and in a variety of other laminated shale sequences (J. Stanley, pers. comm., 1988; M. Machlus, pers. comm., 2002). We interpret the “dead horses” as compacted and dissociated blocks, after the term “horse” for a fault-bounded sliver of rock and their “dead” or dismembered condition. Most commonly, these “dead horses” float in a matrix of massive black mudstone; however, there is a progression from jumbled and partly associated masses of these blocks with no matrix, to isolated blocks in layers composed mostly of matrix. The associated massive mudstone suggests that partial liquefaction accompanied the deformation that produced the dead horses. Beds composed of these black massive mudstones and “dead horses” have usually been interpreted as depositional units, but we interpret them as a form of low-pressure structural mélange.

Depending on depth of burial and pore-fluid pressure, we suggest that bedding plane-parallel shear lead to three intergradational types of deformation: 1) shallow burial bedding plane shear with the development of plastic folds,

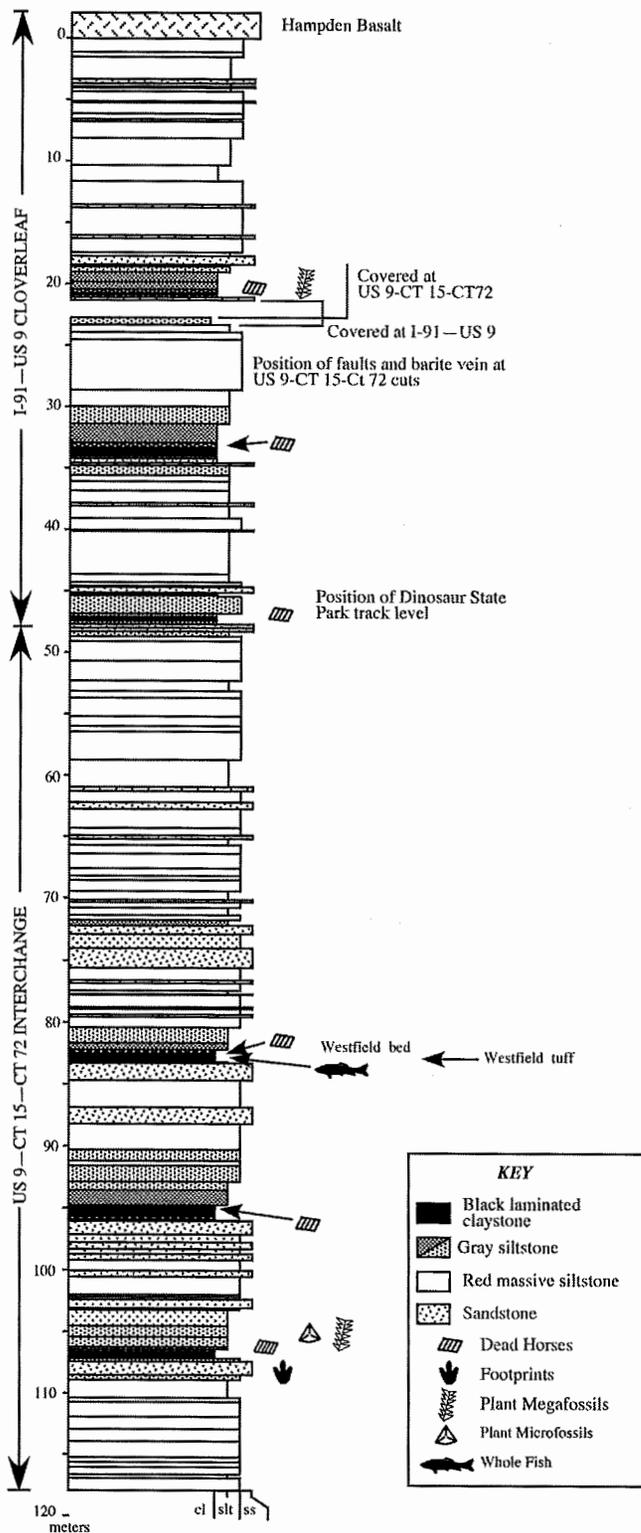


Figure 20. Section exposed at Stop 5. Modified from Olsen et al. (1989).

duplexes, considerable liquefaction, and the production of "dead horses;" 2) intermediate depth bedding-plane shear with abundant slickenside formation, some brecciation, some mineralization of voids, and small amounts of liquefaction; 3) completely brittle shear with extensive slickensiding and decalcification, with the production of platy, polished horses. All of these structures involve at least small-scale thrust faults and concomitant fault-bend folds. However, the sense of shear of the brittle bedding plane faults is up to the west, consistent with late, probably related to tectonic inversion-related compression (e.g., Withjack et al., 1998). In contrast, the dead horses and associated folding seem to indicate down to the east shear, consistent with syn-rifting tilting. All of these structures can easily be confused with "slump" structures, which require a free upper surface, and are usually interpreted as indicators of paleoslope. Although the ultimate tectonic origin of these bedding-parallel shear-related structures is not fully worked out, the folds cannot be assumed to be related to depositional paleoslope.

**Westfield Tuff (contributed by Anthony Philpotts):** A very thin (3 mm) bed tends to erode proud of the surrounding basal portion of the Westfield bed (Fig. 20). This unit resembles thin airfall tuffs in microlaminated units in the Eocene Green River Formation (P.E.O., pers. obs.). A photomicrograph of this layer (Fig. 21) shows that the only detectable particles in the layer are all euhedral plagioclase laths with a high aspect ratio. They show no signs of rounding. They must have originally been enclosed in glass that is now converted to a red clay or to what appears to be chalcedony. There are small circular areas that contain a mottled gray birefringent material that looks very much like chert. In addition, large areas of interstitial carbonate all go into extinction at the same time (poikioblastic?). The laths in Fig. 21 are 0.1 mm long that is typical of the rest in the section. As soon as you move into the material immediately above or below the layer, angular fragments of quartz and muscovite are found, but none is present in the layer itself.

These observations are entirely consistent with the small laths of plagioclase being part of a basaltic crystal tuff. The question is where did it come from? The camptonite dikes on Higby Mountain (Charney and Philpotts, 2004) are vesicular and probably exploded onto the surface, but the plagioclase in those rocks was the last mineral to crystallize and does not form crystals like the ones in the tuff. They can't be the source. The tuff would seem to

represent an explosive eruption of the CAMP that is separate from those that produced the stereotypical sequence of known CAMP flows.

**Hampden Basalt:** The Hampden Basalt is a high titanium, high-iron, quartz-normative tholeiite (HFTQ) identical in composition and age to the Hook Mountain Basalt of the Newark basin (Puffer et al., 1981). The basalt is typically massive but is vesicular at its base. Tilted pipestem vesicles are common at the lower contact and indicate a northeasterly flow direction (Gray, 1982a). The visible thermal effects of the flows are minimal, restricted to the upper meter of the East Berlin. The Hampden Basalt is the thinnest of the extrusives in the Hartford basin, reaching a maximum thickness of 30 m.

### Paleontology

The gray mudstones and gray-black shale beds have generated most of the fossils at this site. The gray units are somewhat palynologically productive and have also yielded carbonized leaf and twig fragments of the conifers *Brachyphyllum* and *Pagiophyllum* and the cycadophyte *Otozamites* (Cornet, 1977a). Other fossils include the conchostracan *Cornia* sp., coprolites, articulated but dephosphatized *Semionotus* and *Redfieldius* (in the Westfield bed), dinosaur tracks, and burrows and invertebrate trails in various gray and red lithologies (McDonald, 1982, McDonald and LeTourneau, 1988a).



Figure 21. Photomicrograph of Westfield tuff. Laths are about 0.1 mm long.

Return to vehicle.

- 57.9 Continue Merge onto CT-9 S toward MIDDLETOWN.
  - 60.6 Merge onto I-91 N via EXIT 20N toward HARTFORD / SPRINGFIELD.
  - 62.7 Take the WEST STREET exit- EXIT 23- toward CT-3 / ROCKY HILL.
  - 63.2 Turn left onto West Street
  - 63.3 Merge onto I-91 S via the ramp on the LEFT toward NEW HAVEN.
  - 64.9 Merge onto CT-9 N via EXIT 22N toward NEW BRITAIN.
  - 75.9 Merge onto I-84 W / US-6 W via EXIT 32 on the LEFT toward WATERBURY.
- Begin drive by of locations at Cooks Gap.

**STOP 6. COOKS GAP, UPPER SHUTTLE MEADOW FORMATION** (Drive by-10 MINUTES). North-central New Britain Quadrangle (very approx. 41°40.50'N, 72°49.25'W). Tectonostratigraphic sequence TS IV; Holyoke basalt, Shuttle Meadow Fm.; Hettangian age, 200 Ma. Main points are: two lava flows; paleomagnetic excursion; Cooks Gap member of Shuttle Meadow Formation; Plainville bed, red beds; footprints; conchostracans; dry phase of 405 ky cycle with muted cyclicity.

The two flows of the Holyoke Basalt are obvious in the road cuts along I-84 and quarries at Cooks Gap, one of the best places to see them in clear superposition. The poorly known lower flow is massive with a thick and purple to red upper vesicular zone. The second flow displays a colonnade with the splintery fracture characteristic of the upper flow. Just north of Cooks Gap, Irving and Banks (1961) reports paleomagnetic directions from outcrops of the lower flow at CT Rt. 10 similar to the Talcott and Hampden. Prevot and McWilliams (1989) sampled the upper flow in Northern Connecticut and Massachusetts, where the lower flow does not extend (Gray, 1982a) and found that it had anomalous directions compared to the other Hartford basin flows. The Cooks Gap section would be perhaps the best place to further examine the paleomagnetic and geochemical properties of the two flows in the Holyoke, where they can be seen in superposition. These anomalous directions imply a magnetic excursion of very short interval (a few thousand years at most) that provides a time marker within the basalts (Prevot and McWilliams, 1989).

Continuous exposures of the upper 55 m of the Shuttle Meadow Formation and its contact with the overlying Holyoke Basalt are well exposed along an old quarry face at the north side of Cook's Gap in Plainville. This section was shown on Simpson's (1966) Bedrock Map of the New Britain Quadrangle, and was first described by Hubert et al. (1978), and since has been a popular field trip stop (e.g., Gierlowski-Kordesch and Huber, 1995). At this location near the western edge of the Shuttle Meadow outcrop belt, the formation is relatively thin when compared to downdip areas such as Silver Ridge, or farther southeast and adjacent to the eastern border fault where this same stratigraphic interval is as much as ~200 m thick.

The base of the section begins at the northwest corner of the small parking lot behind the basket shop, and consists of 10 m dominated by decimeter-scale fining upward packages of ripple cross laminated siltstone and thin mud drapes. At the northeast corner of the parking lot, a prominent, ~ 1 m thick bench of buff, massive limestone and bracketing green and gray shale and siltstone are exposed on the small cliff face. The limestone has been described by Hubert et al. (1978), Gierlowski-Kordesch and Huber (1995) and DeWet et al. (2002) to consist of micritic carbonate, locally with a brecciated or cracked texture with tubules of varying diameter, gradational into immediately underlying and overlying units. The latter are brecciated as well, but more clastic-rich with more obvious rhizoliths and some thin bedding. The limestone contains ostracodes and fish fragments and possible poorly preserved charophyte debris. We refer informally to these beds as the Plainville limestone, and recognize the unit to be traceable over at least the southern half of the Hartford basin. Other outcrops of the Plainville limestone are found at Silver Ridge on Lamentation Mountain, along I-91 below Higby Mountain, along an unnamed stream at the northern tip of Totoket Mountain at Durham, and on the north side of U.S. Route 1 in East Haven, at the southern end of Lake Saltsonstall. The unit contains isolated to patches of scales of semionotid fossil fishes at several localities and the Higby Mountain section produced the skull of a large coelacanth (cf. *Diplurus* sp.), along with clams and the fern *Clathropteris*. We interpret the Plainville limestone as a division 2 of Van Houten cycle 2, the best developed of those in the second short modulating cycles within the Shuttle Meadow Formation (Figs. 6, 7).

The remaining ~40 m of section is largely composed of ripple cross laminated and massive siltstone with clay interbeds of various thickness and four thicker and more finely bedded intervals we interpret as the division 2 of weakly developed Van Houten cycles. Locally, the upper two intervals are gray, the lowest having been described by Hubert et al. (1982). The uppermost beds in the Cooks Gap section consist of slightly metamorphosed siltstone to pebbly sandstone.

These predominately red beds of the upper Cooks Gap member contain abundant reptile footprints and invertebrate trails. Large and small brontozoids, including *Eubrontes giganteus*, are the most common dinosaurian forms and *Batrachopus* is the only other reptilian track form present. Fragmentary plant remains are present in the gray beds and conchostracans are present in the lower two thicker claystone intervals.

- 82.3 Take unnumbered exit on right toward CT-372 / NEW BRITAIN AVE / PLAINVILLE.
- 82.5 Turn right onto CT-372 / NEW BRITAIN AVE
- 83.2 Turn right onto CROOKED ST.
- 83.3 Turn RIGHT onto WHITE OAK AVE.
- 83.4 Turn LEFT onto LEDGE RD.
- 84.7 LEDGE RD becomes SHUTTLE MEADOW RD.
- 85.8 SHUTTLE MEADOW RD becomes LONG BOTTOM ST.
- 86.5 Turn RIGHT onto ANDREWS ST.
- 86.6 Park next to overgrown small quarry on left, Southington, CT

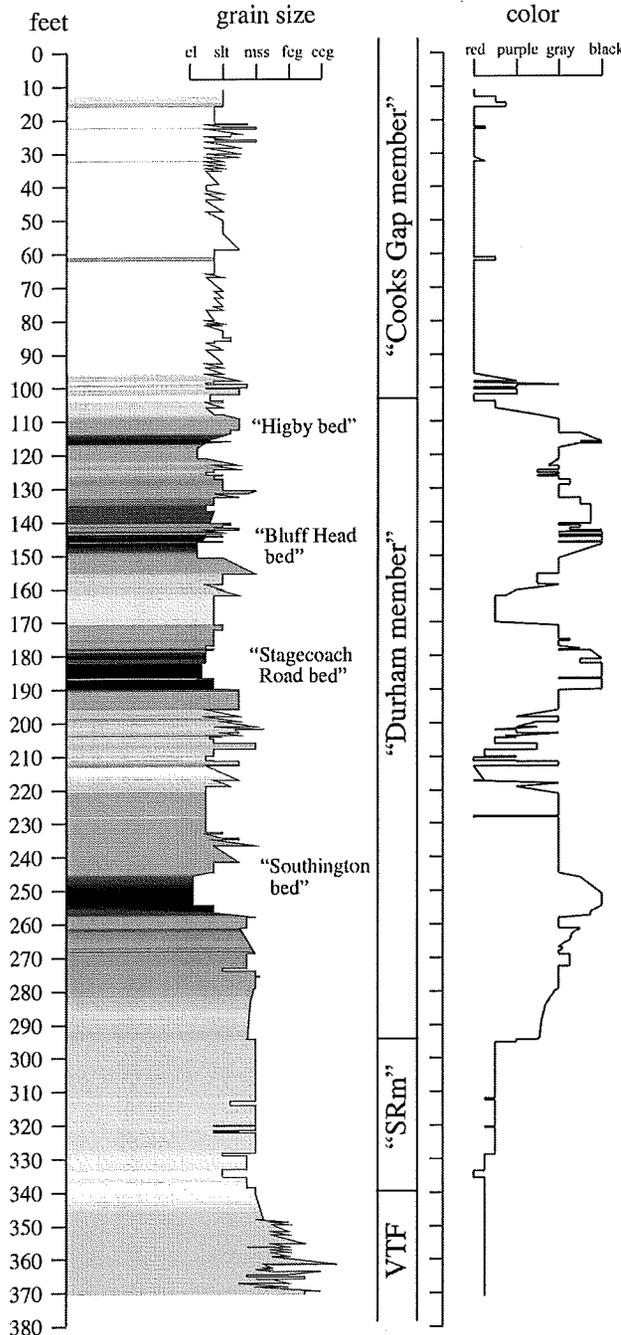
#### STOP 7. SILVER RIDGE CORE B-1 AND TYPE SHUTTLE MEADOW FORMATION (30 MINUTES).

Stop location in north central New Britain Quadrangle (41°37.694' N, 072°50.013' W); core location in northeastern Meriden Quadrangle (41°35.102' N, 072°45.388' W). Talcott Formation, and Shuttle Meadow Fm.; Hettangian age, 200 Ma. Main points are: type section of Shuttle Meadow Formation correlated to complete stratigraphy derived from core; core with continuous section of uppermost volcanoclastic Talcott Formation upward into cyclical Durham Member; cyclicity of Durham member with fish-bearing carbonates; lateral persistence of cycles; fauna and flora of Durham member.

This small, long-abandoned quarry comprises the section described by Krynine (1950) near the Shuttle Meadow Reservoir that was designated the type sections of the formation of the same name. The stop consists of two parts: 1)

examination of the Silver Ridge B-1 core that displays a complete section of the Durham member of the Shuttle Meadow Formation; and 2) examination of the type section that proves to be the lowest Van Houten cycle in the formation, containing the Southington bed.

The Silver Ridge B-1 core, representative portions of which will be on displayed at this stop, penetrates nearly the entire lower half of the Shuttle Meadow Formation as well as most of the volcanoclastic equivalent of the Talcott



Formation (Fig. 22). This core revealed for the first time the stratigraphy of the lower, mostly gray and black, highly fossiliferous portion of the Shuttle Meadow Formation that we call the Durham member.

The lowest Van Houten cycle has a distinctive black and dark gray division 2 bearing a limestone with rather thick and distinctive laminae with a pelleted or flocculent appearance (254.6-251.0 ft.). This bed, which we call the "Southington bed", crops out in a variety of places, most importantly at small, long-abandoned quarries to the south in Southington, Connecticut and this stop that is the section described by Krynine (1950, p. 61) near the Shuttle Meadow Reservoir. Rather poorly preserved (partly dephosphatized), but articulated fish are present in this core (e.g., at 252.7 ft, Fig. 23) and adjacent temporary exposures. We believe that the "Southington bed" is the lateral equivalent of the Coe Quarry limestone near Northford, Connecticut described by Krynine (1950), Mooney (1979), Steinen et al. (1987) and DeWet et al. (2002).

The second Van Houten cycle lacks a well-developed limestone bed but does have a division 2 with significant black shale (186.7-175.0 ft, Fig. 23). This shale is distinctive in that there are many silty interbeds with some of them bearing reptile footprints (as seen in adjacent exposures, to the core). We correlate this interval with the lower fish-bearing black shale (the "crinkle bed") that crops out at the Durham fish locality and name it the "Stagecoach Road bed" after the road immediately east of the stream on which the latter locality is located.

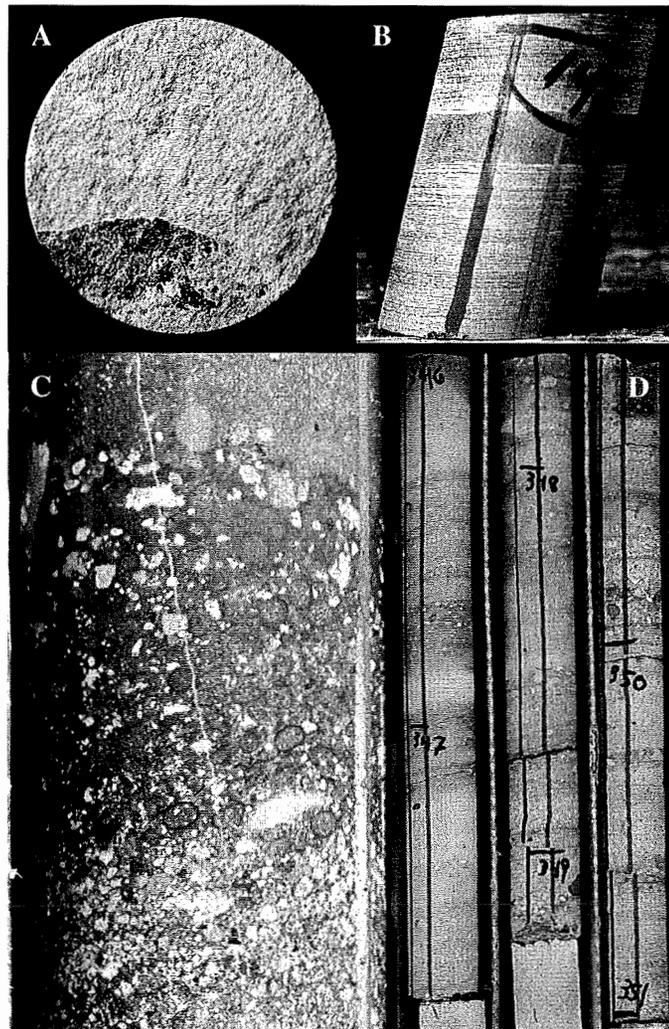
The third Van Houten cycle has two distinctive dark gray to black intervals (Fig. 22). The lower bed (146.5 - 132.8 ft) is highly calcareous and finely microlaminated and contains abundant fish and coprolites (Figs. 23, 24). This bed is evidently the main fish-bearing unit that crops out at the famous Durham and Bluff Head fish localities (McDonald, 1992), and we thus call it the "Bluff Head bed". Some of the best-preserved fish in the Newark Supergroup (Fig. 24) have been recovered from this unit, including the subholostean *Semionotus*, redfieldiid palaeonisciforms *Redfieldius* and *Pycholepis* and the coelocanth *Diplurus longicaudatus* (McDonald, 1992.) The Bluff Head bed also provides the best example of a *Semionotus* species flock in the Hartford basin.

Figure 22. Silver Ridge Core B-1 (Stop 7).

The upper of the two dark gray to black beds (118 – 113.6 ft; Fig. 22) is neither clearly microlaminated nor as calcareous. It apparently correlates with a black shale that crops out along Highland Brook on the East face of Higby Mountain, and we term this bed the “Higby bed”. At the latter locality, the “Higby bed” produces articulated fish and conchostracans.

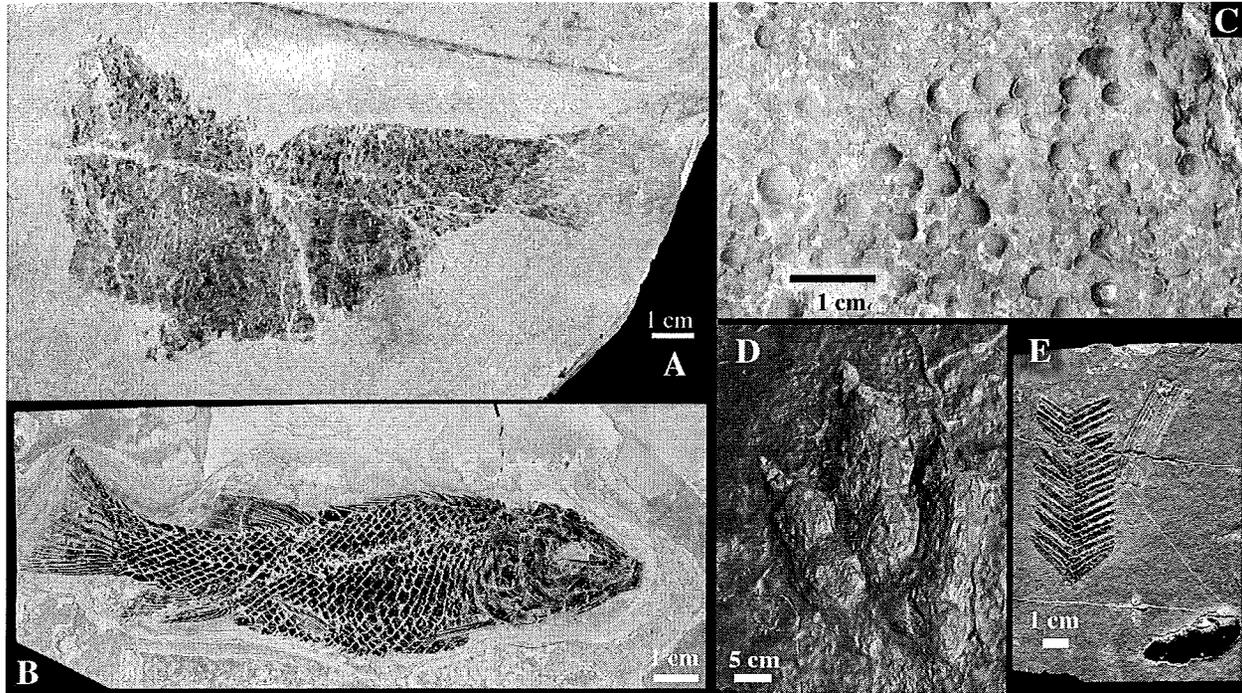
The Silver Ridge B-1 core unveiled for the first time the surprisingly complex stratigraphic relationships of the numerous black shale-and carbonate-bearing localities in the lower Shuttle Meadow Formation. In addition, the core showed that the cyclostratigraphy of the Durham member is profoundly similar to the homotaxial portion of the Feltville Formation of the Newark basin, known from multiple cores (Olsen et al., 1996a) (Fig. 25), suggesting that three Van Houten cycles exist in the lower Feltville Formation, rather than the two reported in Olsen et al. (1996b; 2003b.) The total duration of the CAMP basalt flows is thus extended ~20ky, totaling ~610ky (Whiteside et al., 2005).

The carbonate-rich nature of the “Durham member” is distinctive compared to most Jurassic age strata in the Newark Supergroup. There is a strong cyclostratigraphic similarity between the sections above these initial flows throughout eastern North America and Morocco (Olsen et al., 2003b; Whiteside et al. (2005). A similar stratigraphy of two HTQ flow sequences with two sedimentary cycles is also present in eastern Morocco; however, there the entire interbed is limestone-dominated. The apparent relationship between the initial HTQ CAMP flows, the underlying Triassic-Jurassic boundary and the overlying carbonate rich sequence suggest to us two possible interpretations (Olsen et al., 2003b; Whiteside et al., 2005): 1) The limestone deposition is a response to a super-greenhouse effect at Triassic-Jurassic boundary with effects tapering off over some hundreds of thousands of years; 2) the limestones are weathering products of vast drainage areas newly floored by relatively Ca-rich basalt, and the ferns are a consequence of unusually heterogeneous depositional environments caused by the accelerated tilting and subsidence associated with the eruption of the basalts (and the initiation of TS IV. These scenarios are not mutually incompatible, but they do predict completely different far field effects that can be looked for in other regions far from the CAMP.



**Figure 23.** Silver Ridge B-1 core: A, Fish (*Redfieldius*) at 146 ft; B, lateral view of microlaminated calcareous mudstone with fish (“Bluff Head bed”) at 146 ft – note small turbidite; C, spherules from 349 ft. (examples at arrows); D, typical Talcott volcanoclastics (footage on core). From Olsen et al. (2003a).

There are 13.5 m (339.0 to 294.6 ft) of largely red beds between the largely gray “Durham member” and underlying volcanoclastics (Fig. 22, 23). The volcanoclastics begin as a few clasts of basalt in sandstone at 339 ft but rapidly becomes downward a conglomerate of basalt clasts and basaltic sandstone to the base of the core at 371 ft. At 362.5 ft and 367 ft there are well-developed layers of reddish spherule layers (Figs. 23, 24) that might represent basaltic lapilli. These are well-displayed at the adjacent outcrops (Olsen et al., 2003a).



**Figure 23.** A, *Semionotus* from the “Southington limestone bed” in exposure, Silver Ridge; B, *Semionotus* from the “Bluff Head bed”, Bluff Head, North Guilford, CT N.G. McDonald Collection; C, Spherules from Talcott volcanoclastics, Silver Ridge, (Stop 6) (N.G. McDonald collection); D, *Eubrontes giganteus*, Highland Brook, between Higby and Bluff Head beds, Huber collection; E, *Otozamites* and coprolite, Bluff Head bed, Durham locality, Durham, CT.

At this field stop, as presently exposed, Krynine’s (1950) type section consists of about 5 m of laminated and thin-bedded dark to medium gray clay- and siltstone with plant fragments (mostly conifers). This is followed upward by 0.6 m of thin-bedded gray mudstone that grades upward into flaggy-bedded red siltstone which is overlain by 3.6 m of red mudstone to the top of the exposure. Small exposures along Andrews Street to the northwest and in the woods reveal an additional 4 m of gray grading upward into red siltstone beginning about 9.6 m above the top of the type section exposures. The character of the gray laminated units is consistent with the gray beds above the Southington member and the gray-into-red siltstone higher up is consistent with the cycle bearing the Stagecoach Road bed as seen in the Silver Ridge B-1 core and adjacent outcrops. To test these hypotheses, we excavated at the base of the type section and uncovered a black limestone with a pelleted appearance characteristic of the Southington bed. Although Krynine (1950) mentioned a fish-bearing limestone at this section, it had not been identified. It seems clear that the type section does indeed represent the cycle that bears the Southington bed.

The relationship between the accelerated tilting in the lower part of TS IV can be seen in a comparison of the stratigraphy in the lower Shuttle Meadow Formation across the various fault blocks in the Meriden area. At Higby Mountain, gray siltstones and sandstones of the “Durham member” rest directly on the Talcott Formation. In the next fault block east with Lamentation Mountain, there are 10 m of intervening red beds, suggesting additional accommodation due to syndepositional subsidence along the intervening fault. On the next major fault block to the west with Cathole Mountain and Stop 8, the Durham member is missing entirely and the entire formation is much thinner. Based on outcrop width, this thinning continues further to the west. To the northwest the Shuttle Meadow thins, then thickens to this field stop and then thins again towards Plainville and is only 14 m thick at Tarrifville, CT

with the Durham member again missing. In contrast there appears to be much less variability of the Talcott Formation along this same interval. The simplest interpretation of these observations is that considerable structurally controlled depositional relief developed between the emplacement of the Talcott Formation and the deposition of the upper Shuttle Meadow Formation, an interval of time probably no more that a couple of hundred thousand years.

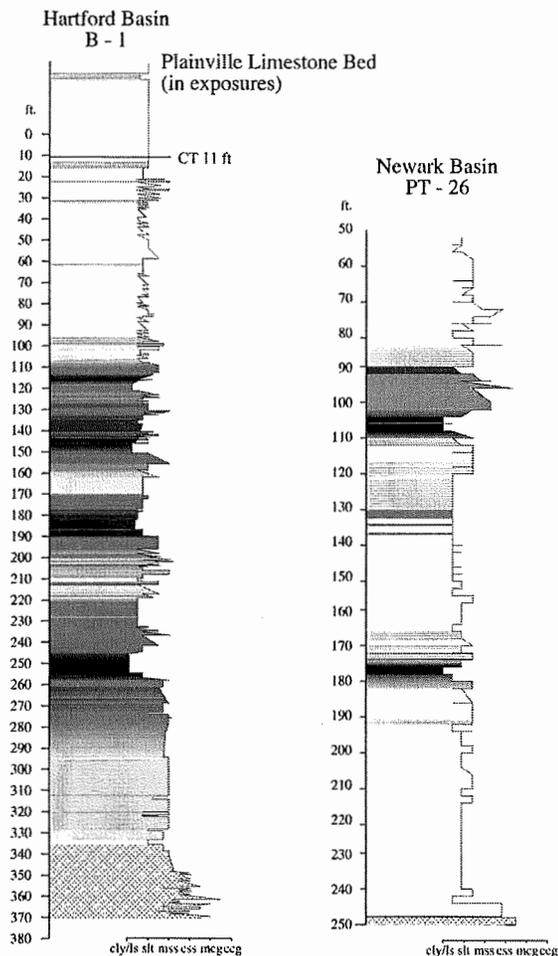
Return to vehicle.

- 86.6 Start out going SOUTH on ANDREWS ST toward SMITH ST.  
 87.8 Turn SLIGHT LEFT onto CAREY ST.  
 88.2 CAREY ST becomes RESERVOIR RD.  
 89.6 Turn RIGHT onto CHAMBERLAIN HWY / CT-71A. Continue to follow CHAMBERLAIN HWY.  
 94.0 Turn right into parking lot for TARGET (474 Chamberlain Hwy Meriden, CT)  
 94.1 Drive to southwest corner of parking lot.

**STOP 8. TALCOTT PILLOW BASALT SEQUENCE ON NEW HAVEN FORMATION. (45 MINUTES).** Central Meriden Quadrangle, (approx.) 41°33.12'N, 072°48.91'W and Silver Ridge Core B-2 (41°35.053'N, 072°45.647'W). Tectonostratigraphic sequence TS IV; New Haven and Talcott formations; Hettangian age, 200 Ma. Main points are: northeast prograding pillow lava forsets and flow lobes in the; onlapping red sandstones; graded beds altered pyroclastics; gray beds in the upper New Haven in the Silver Ridge B-2 core; and transition downward into typical New Haven Formation.

This spectacular exposure reveals nearly the entire thickness of the Talcott Formation as well as the uppermost few meters of the New Haven Formation. Most of the lower half of the Talcott Formation here consists of forsets and tongues of pillow lava, while most of the upper part consists of non-pillowed flow lobes and vesicular basalt. It is unclear how many eruptive events are represented.

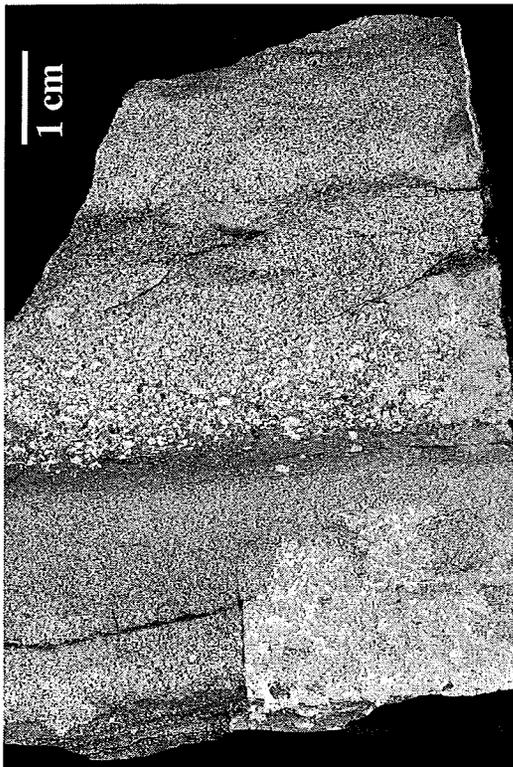
The base of the Talcott Formation at this exposure displays an extremely informative relationship to the underlying Farmington member of the New Haven Formation (Fig. 26). Northeast tapering wedges of pillowed basalt onlap each other in the lower 10 m of the flow complex. These are easily traced by following the red sandstone and siltstone beds that extend upward from the underlying New Haven Formation into the basalt. In several places these beds are internally stratified and contain numerous large to small clasts of basalt and highly altered basalt. Tracing the red beds downward, they merge with the underlying New Haven Formation. Beneath the pillow forsets and associated red beds there is well-bedded red sandstone and mudstone of the New Haven Formation. In its upper few decimeters, there are beautiful graded beds comprised of basaltic gravel and sand (Fig. 27). The basalt and what presumably was basaltic glass is generally altered to a yellow or tan material, superficially resembling a carbonate. All stages of alteration from unquestionable nearly unaltered basaltic material to the tan to yellow clasts seem to be visible. Based solely on macroscopic examination, the lowest beds of the New Haven Formation at this outcrop seem to lack basaltic material.



**Figure 24.** Comparison of the Silver Ridge B-1 core and core PT-26 from the lower Feltville Formation of the Newark basin. Note the homotaxiality of the black units.



**Figure 25.** Exposures of Talcott Formation and underlying Farmington member of the New Haven Formation at Stop 8: left, pillow forests and onlapping red siltstone (A) and underlying New Haven Formation with tuff layers (B); right, pillowed basalt in forests with overlying massive flow lobe.



**Figure 26.** Graded volcanoclastic (tuff) beds in uppermost New Haven Formation at Stop 8 at the Target.

Higher up in the Talcott Formation, amongst the wedges of basalt pillows are larger lobes of massive basalt without pillows. Yet higher in the Talcott Formation, some of these are over 6 m thick and scores of meters long. These are almost certainly flow lobes that were the sources of the pillowed wedges. The uppermost Talcott here is highly vesicular, not pillowed and appears to have been deposited subareally.

Although locally deformed by the overlying pillow wedges, the series of graded volcanoclastic beds can be traced across the admittedly limited exposure of the New Haven Formation. Very similar graded volcanoclastic beds have been observed in the uppermost New Haven (by PH) at Hubbard Park at approximately 41°33'22"N and 072°49'33"W. These volcanoclastics would appear to be relatively proximal altered air-fall(?) pyroclastics from a CAMP eruption that began some time before the flows that we see here. This interpretation is different than that given in Olsen et al. (2003a) in which these graded beds were assumed to be related to the basalt pillow wedges.

We interpret the observations on the pillow forsets to indicate that the basalt was flowing to the east and north into a large lake as a series of lava streams. At the advancing front of these lava streams the cooling crust constantly ruptured, sending basalt pillows tumbling down in front, making cones and wedges of pillows. These wedges shed hyaloclastite into muds onlapping the pillow wedges. Hyaloclastite is a hydrated tuff-like rock composed of angular, flat fragments 1 mm to a few cm across formed by granulation of the lava front due to quenching when lava flows into, or beneath water. It would be good to know how much of the red matrix is composed of locally derived, very fine grained hyaloclastite and how much is sediment derived from the highlands or reworked older sedimentary strata.

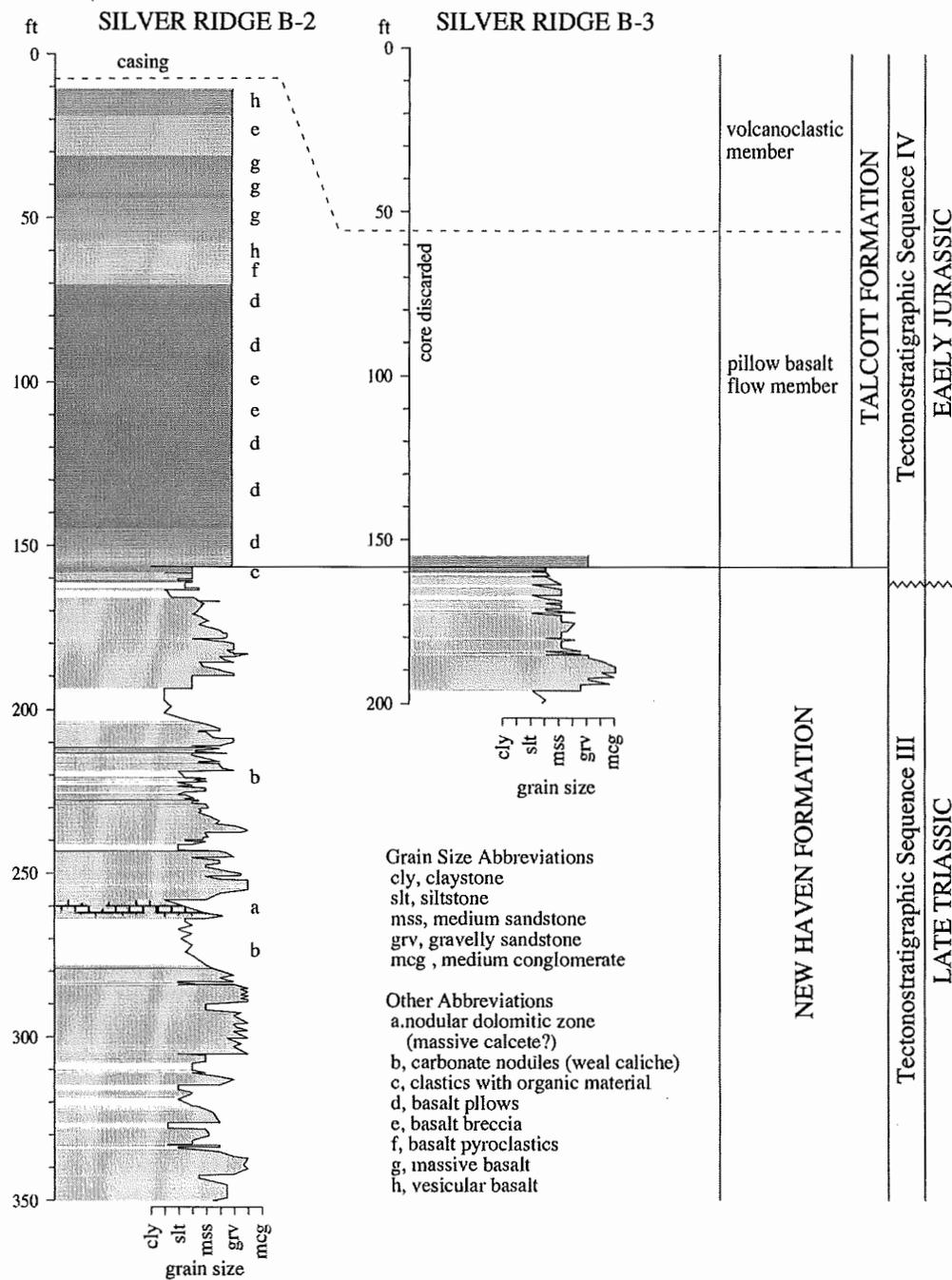


Figure 27. Cores B-2 and B-3 from Silver Ridge on display at Stop 8. From Olsen et al. (2003a).

Assuming that the various pillow wedges and flow lobes represent one major eruptive event constrains the accumulation rate and water depth of the uppermost New Haven Formation. The couple or so meters of basalt-bearing sedimentary strata obviously took no more time to accumulate than the flow complex took to advance over the site, which would seem to be on a time scale of years to at most hundreds of years. The lake into which the lava poured had to be at least the depth of the high points of the individual wedges of pillowed basalt (i.e., 5 m or so), but probably not the depth of the entire Talcott Basalt because lava displaces water. We have found no desiccation cracks in the basalt-bearing red beds, suggesting the lake did not locally dry up during the advance of the Talcott Basalt. It is worth noting, however, that no sedimentological criteria indicates that these red beds were deposited under a significant body of water, or even that they are lacustrine strata, yet they were and are.

Core B-2, portions of which will be on display at this stop, samples 44.5 m (146 ft) of the lower Talcott Formation and the 58.8 m (193 ft) of the upper New Haven Formation and hence the Triassic–Jurassic and TS III – TS IV boundaries (Fig. 28). In core B-2 the lower 56.5 m (below 164.5 ft) is typical of the New Haven Formation and characteristic of the alternating “Red stone” and Lamentation facies of Krynine (1950), itself typical of the Meriden area. The “Redstone” facies consists of red micaceous feldspathic sandstone and interbedded bioturbated mudstones that can be seen in the exposures along near by I-691, while the Lamentation facies consists of often conglomeritic coarse gray or purplish white arkose. To the east, the Lamentation facies predominates; to the west the “Redstone” facies is dominant (Krynine, 1950). Both facies tend to be heavily bioturbated with roots and the burrow *Scoyenia*, and consequently there is very little preservation of fine sedimentary structures (e.g., ripples).

Above 164.5 ft, the New Haven Formation facies change abruptly to finely interbedded, red and tan fine sandstone that grades upward into gray fine sandstone and minor mudstone with abundant plant fragments, including conifer shoots. This facies is much more similar to the overlying Shuttle Meadow Formation in the preservation of small-scale sedimentary structure, the reduction of bioturbation, and the preservation of organic matter. We have not yet extracted pollen from these gray beds, however, just below the Talcott Formation, a similar sequence of conifer-bearing gray beds produced a palynoflora of earliest Jurassic aspect, strongly dominated by *Corollina* (Robbins, quoted in Heilman, 1987; pers. obs.), and therefore we believe the gray beds to be earliest Jurassic in age. The abrupt facies change at 164.5 ft and 166.9 ft in cores B-2 and B-3, respectively, is best interpreted at the TS III – TS IV boundary. The same facies transition could also represent the Triassic–Jurassic boundary.

The contact with the overlying Talcott Formation at 156 ft in core B-2 is sharp, but there is some deformation of the uppermost New Haven Formation. From 156 ft to 70 ft, the Talcott is pillowed and brecciated, looking very much like the exposures at this stop. Radial pipe vesicles are evident as are the chilled margins of pillows. From 70 ft to 32 ft there is massive locally vesicular basalt. This is overlain by breccia unit to 19 ft, which itself is overlain by vesicular basalt to the top of the core (11 ft). It is impossible to tell whether the contacts at 19 ft and 70 ft are contacts between different major flows or contacts between lobes of one eruption, although we favor the latter interpretation based on the geometry visible at this stop.

There is a dike trending towards the Berlin area in Meriden (mapped as present near Hanover Pond). This could be a local feeder for the pillowed basalts, pyroclastics, and volcanoclastics seen at this stop and the Silver Ridge cores and outcrops. Such a feeder, graben system, and associated Talcott flows, has been described in East Haven, CT by Philpotts and Martello (1986) and Olsen et al. (2003a). Assigning the Meriden dike to the feeder system, in a less speculative fashion, however, requires much additional work. It is clear however that there is much to be learned about the dynamics of the Talcott eruption in southern Connecticut.

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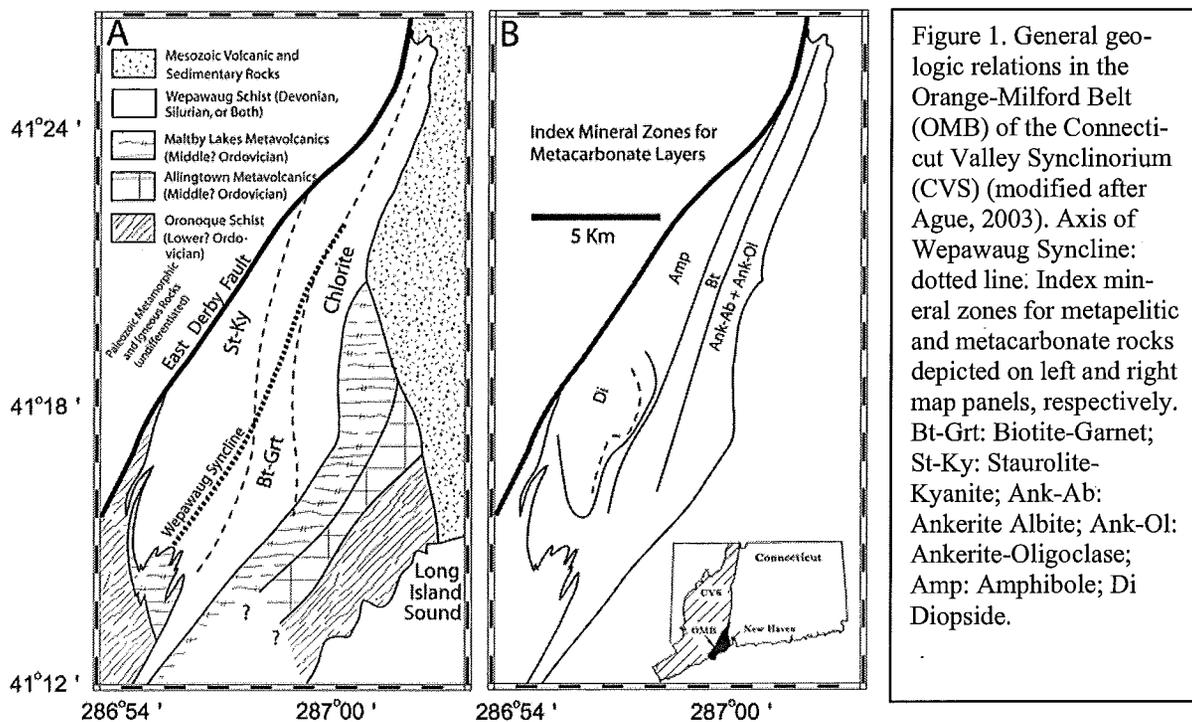


# Deep Crustal Metamorphism of South-Central Connecticut

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## INTRODUCTION

The Wepawaug Schist of the Orange-Milford belt is one of the classic field areas of southern New England (Fig. 1). Here, Barrovian-style metamorphism affected a dominantly metasedimentary sequence. A notable feature of the Schist is that metamorphic grade increases systematically from greenschist facies in the east to amphibolite facies in the west. To the author's knowledge, the eastern part of the Orange-Milford belt is the only place in Connecticut where prograde greenschist facies metamorphism of the Acadian orogeny is exposed. The Schist is dominated by metaclastic rocks (metapelites and lesser metapsammites), but also includes metacarbonate rocks and felsic igneous bodies. Numerous petrologic, isotopic, and tectonic studies have focused on the Wepawaug Schist, owing to the well-defined increase in metamorphic grade across the area and the rich variety of mineral assemblages and exposed structures (e.g., Burger et al., 1968; Dieterich, 1968; Hewitt, 1973; Bahr, 1976; Silverman, 1976; Tracy et al., 1983; Breault, 1992; Palin, 1992; Ague, 1994a, b; Ague, 1995; Lanzirrotti and Hanson, 1996; van Haren et al., 1996; Karsh, 1998; Ague and Rye, 1999; Ague, 2000, 2002, 2003; Wilbur, 2003; Wilbur and Ague, 2005). These studies have been aided immensely by the outstanding geologic maps of the Mt. Carmel, Ansonia, and Milford Quadrangles by Fritts (1963, 1965a, 1965b) and, more recently, by John Rodgers' (1985) classic geologic map of Connecticut. The Wepawaug Schist has proven to be an excellent location for the study of metamorphic fluid flow and its impact on, for example, the genesis of staurolite and kyanite, and the prograde reactions that release CO<sub>2</sub> from metacarbonate rocks.



## OVERVIEW OF REGIONAL STRUCTURE

The Wepawaug Syncline (Fritts, 1965a) is the predominant major structure in the area (Fig. 1). It plunges to the north-northeast and was defined by Fritts because bedding in the eastern part of the area dips steeply to the west and bedding in the west dips steeply to the east. Given the large-scale syncline interpretation, it follows that the Maltby Lakes metavolcanics, Allingtown metavolcanics, and Oronoque Schist underlie the Wepawaug Schist (Fig. 1). Dieterich (1968) mapped minor fold asymmetries ("s-folds" and "z-folds") and concluded that Fritts' interpreta-

tion of a synclinal structure was correct. Fold geometries are generally tight to isoclinal. Several generations of structures are visible in outcrop. The main deformational foliation or cleavage that characterizes the rocks is sub-parallel to relic bedding and the axial planes of most folds, but can be seen to cut bedding in fold hinges. This cleavage is often deformed by a later crenulation cleavage which, in some localities, is further deformed by kink-like folds. Several faults are inferred to cut the interior of the Wepawaug Schist. The largest of these is the Mixville fault which Fritts (1965a) suggested records a vertical displacement of at least ~100 m. The existence of this fault remains controversial owing to lack of diagnostic outcrops. The East Derby fault (Rodgers, 1985) lies along the western margin (amphibolite facies portion) of the Wepawaug Schist (Fig. 1). The adjacent Oronoque Schist in this area is a heavily retrograded metapelitic rock that may have actually been derived from the Wepawaug. Garnets are strongly chloritized and the rocks contain what appear to be staurolite and/or kyanite crystals pseudomorphically replaced by sericite and chlorite. The fluids required for retrogression may have been introduced during exhumation of the sequence along the adjacent fault zone, but much more work is needed to document geologic relations in this area.

### PROGRADE METAMORPHISM

Metamorphic grade increases from the greenschist facies to the amphibolite facies from east to west across the Wepawaug Schist and adjacent rock units. Metamorphic temperatures ( $T$ ) recorded by a variety of geothermometers indicate that  $T$  increases from 400-450 deg. C in the eastern, Chlorite zone part of the Wepawaug Schist to in excess of 600 deg. C in the amphibolite facies (Fig. 2A; Ague, 2002). The  $T$  increase occurs over a remarkably short distance of ~4 km, yielding a steep metamorphic field  $T$  gradient of ~58 deg. C per km. Results from quantitative thermobarometry show that the metamorphism took place at deep crustal levels. Recorded pressures increase from ~7 kbar (~25 km depth) in the east to ~10 kbar (36 km depth) in the west (Fig. 2B). The ~10 kbar pressures are among the highest recorded in the Acadian orogen.

The age of the metamorphism and the depositional age of the Wepawaug Schist are under active investigation. Fritts (1962) correlated the Wepawaug with the Waits River and Northfield Formations in Vermont, and assigned a Siluro-Devonian depositional age to the Wepawaug. These correlations are plausible but remain to be fully tested. Recently, Lancaster et al. (2005) used the Sm/Nd garnet-whole rock method to determine ages of garnet growth in the Wepawaug Schist. Results for a Biotite-Garnet zone sample and a Staurolite-Kyanite zone sample are indistinguishable at 379.2 +/- 6.7 Ma and 379.9 +/- 6.9 Ma, respectively. These results firmly establish an Acadian age for the metamorphism, consistent with geochronological studies done on other nearby metapelitic rocks in southwestern Connecticut, including the Trap Falls Formation which outcrops a few km to the west of the Wepawaug (Sevigny and Hanson, 1993). It is interesting to compare the results of Lancaster et al. (2005) to the earlier U/Pb dating of monazite done by Lanzirrotti and Hanson (1996). In one sample, they found monazites that yielded close to concordant ages; their best age estimate was 388 +/- 2 Ma. In another sample, they found monazites which defined a chord with an upper intercept age of 411 +/- 18 Ma. The 388 Ma age was interpreted to be the age of retrograde greenschist facies chloritization, but the Sm/Nd garnet results of Lancaster et al. (2005) preclude this interpretation. The age could, however, represent a stage of monazite growth during part of the prograde history. One, admittedly speculative, possibility for the ~410 Ma result is that it represents the age of detrital grains (see Lanzirrotti and Hanson, 1996); in this case the age would establish a maximum for the time of deposition of the Wepawaug Schist. Clearly, much work remains to resolve the depositional and prograde metamorphic

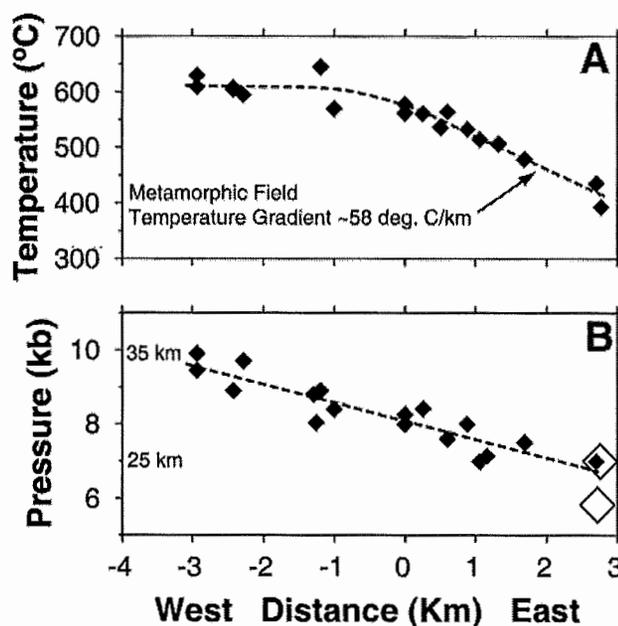


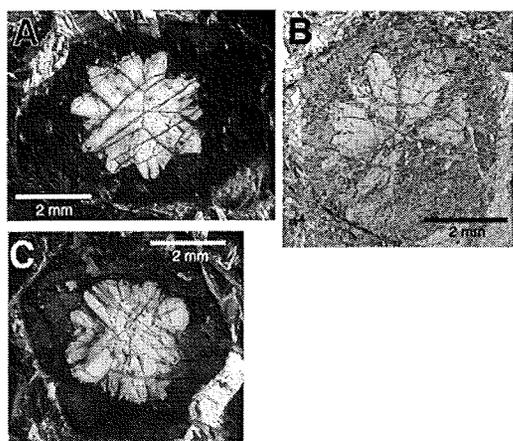
Figure 2. Estimates of metamorphic temperatures and pressures across the Wepawaug Schist. Modified after Ague (2002).

age relations in the area. The  $^{40}\text{Ar}/^{39}\text{Ar}$  work of Moecher et al. (1997) showed that muscovite from a sample of Wepawaug Schist cooled through its closure  $T$  at 294 +/- 2 Ma.

### Metamorphism of metapelitic rocks

The progression of prograde mineral assemblages in metapelitic rocks across the Wepawaug Schist has much in common with the classical Barrovian zones defined in the late 19<sup>th</sup> and early 20<sup>th</sup> centuries by George Barrow in the Scottish Highlands. The lowest grade Chlorite zone assemblages typically contain muscovite, chlorite, quartz, and albitic plagioclase; calcite and/or ankerite are present in some samples, and most contain accessory rutile and/or ilmenite, organic matter, pyrite, monazite, zircon, apatite, and tourmaline. With increasing metamorphic grade biotite and garnet were produced. Separate biotite and garnet zones are not mapped in Fig. 1 because these phases appear together in nearly all samples as a result of a dehydration reaction that destroyed chlorite and muscovite and produced garnet and biotite (van Haren et al., 1996).

Garnets preserve a striking texture comprising relatively inclusion-free, anhedral to star-shaped cores that transition abruptly to more euhedral rims (Fig. 3; see also Fig. 9 in Ague, 1994b). Wilbur (2003) and Wilbur and Ague (2005, and in review) interpret these textures to be the result of an initial growth phase that took place far from chemical equilibrium and produced the characteristic core morphologies, followed by rim growth that was considerably closer to chemical equilibrium. Monte Carlo simulations of crystal growth confirm that anhedral, "star-shaped", and even dendritic growth can be produced at large supersaturations (Fig. 3; Wilbur and Ague, in review). Sluggish garnet nucleation could have caused the equilibrium  $T$  conditions for garnet-producing dehydration reactions to be overstepped during heating. Once growth began, however, it was probably rapid and released considerable water. This fluid, in turn, could have weakened the rocks, produced hydrofracturing, and even seismic failure in tectonically-active regions (Ague et al., 1998; Wilbur and Ague, in review). As garnet growth continued, supersaturation was relieved promoting a transition to more euhedral rim growth (Wilbur, 2003; Wilbur and Ague, 2005, and in review).



### Model of Crystal Growth

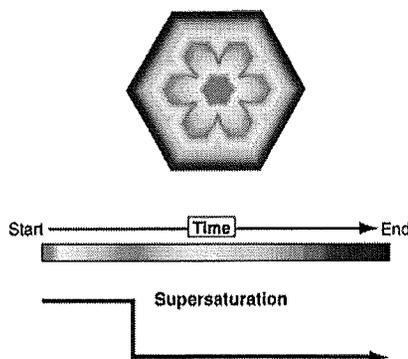


Figure 3. *Left panels:* Thin section views of garnet textures. *Right panel:* Monte Carlo computer simulations of crystal growth. Growth begins under highly super-saturated conditions producing core region; supersaturation then decreases to produce more euhedral rim. Modified from Wilbur & Ague (in review).

With increasing metamorphic intensity, staurolite and kyanite were produced. These minerals are often concentrated around quartz-rich veins, as a result of reactions that destroyed plagioclase and/or micas, liberated alkali and alkaline earth metals including Na, K, and Ba, and produced aluminous bulk compositions amenable to the growth of staurolite and/or kyanite (Ague, 1994b). The veins are interpreted to be mineralized fractures, and increase in abundance from the greenschist to the amphibolite facies (Fig. 4). In the greenschist facies, most of the vein quartz appears to have been derived from adjacent wallrocks. However, in the amphibolite facies, a substantial component of externally-derived silica is present in many of the veins, leading to the interpretation that they were conduits for aqueous metamorphic fluids that transported mass and heat through the Schist (Ague, 1994b; van Haren et al., 1996). These fluids are inferred to have leached alkali and alkaline earth metals from cm to dm-wide altered zones or "selvages" adjacent to the veins, producing open-system staurolite and kyanite in the process. Some staurolite and kyanite in the Wepawaug undoubtedly formed by the "closed-system" processes commonly discussed in textbooks, but the reactions adjacent to veins often resulted in copious growth and produced large crystals (kyanite

crystals can reach 5-6 cm in length). The average time-integrated fluid flux needed to transport the heat and mass through the amphibolite facies was large: about 60,000 m<sup>3</sup> fluid per m<sup>2</sup> rock area (Ague, 1994b).

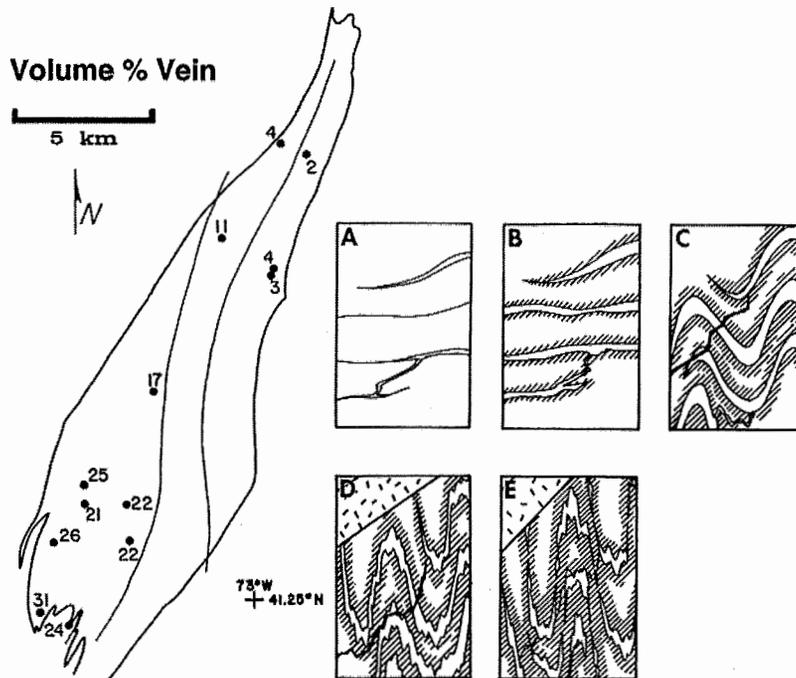


Figure 4. *Left figure:* Volume % vein in outcrop. Note increase in vein percentage with increasing metamorphic grade. *Right panels:* Cartoon depiction of syn-metamorphic vein initiation and ultimate deformation during regional orogenesis. Alteration zones or “selvages” that developed around the veins denoted by diagonal rule. A syn-metamorphic intrusion is depicted in parts D and E. Modified from Ague (1994b).

### Metamorphism of metacarbonate rocks

Metacarbonate rocks also record systematic changes in mineral assemblages with increasing metamorphic grade, as recognized by Fritts (1965a) and Hewitt (1973) (Fig. 1). The metacarbonate mineral zones are very similar to those recognized elsewhere in New England, such as in the Waits River Formation of Vermont (Ferry, 1992). Low-grade examples in the Ankerite-Albite zone consist mostly of calcite, ankerite, quartz, muscovite, and albitic plagioclase, and most samples include accessory rutile and/or ilmenite, organic matter, pyrite, monazite, zircon, apatite, and tourmaline. With increasing metamorphic intensity, biotite, amphibole, and diopside were produced (Fig. 1). Considerable open system mass transfer of volatiles and rock-forming and trace elements occurred at lithologic contacts and in alteration selvages around quartz veins, as recognized first by Hewitt (1973) and Tracy et al. (1983). Influx of water-rich fluids from surrounding schists or intrusions drove prograde reactions that liberated carbon dioxide (Hewitt, 1973; Ague and Rye, 1999). Infiltration into the metacarbonate layers occurred via various combinations of diffusion, mechanical dispersion, and fluid flow (Fig. 5; Ague and Rye, 1999; Ague, 2000; 2002; 2003). In the Amphibole and Diopside zones, the most devolatilized rocks are present at the contacts with metapelitic rocks or veins, and the less reacted rocks are found in bed interiors (Figs. 5, 6). Hewitt (1973) postulated that steep gradients in fluid composition were present across contacts between metacarbonate layers and metapelitic rocks or quartz veins; subsequent work has shown that gradients in water and carbon dioxide activities could have been very subtle but still able to drive significant decarbonation (Ague, 2000; 2002). Very water-rich fluids were required to produce diopside ( $X_{\text{CO}_2} \sim 0.03-0.13$ , Ague, 2002); Tracy et al. (1983) and Palin (1992) showed that quartz veins associated with diopside production contained isotopically-light oxygen. This result may indicate derivation or equilibration of at least some of the diopside-producing fluids from syn-metamorphic magmas or perhaps the underlying Maltby Lakes Metavolcanics (Palin, 1992; van Haren et al., 1996; Ague, 2002).

Calc-silicate rocks poor in carbon-minerals but rich in zoisite (or clinozoisite) hornblende and, appropriate metamorphic grade, diopside, are commonly present at lithologic contacts and in vein selvages (Fig. 6). The calc-silicate metasomatism involved substantial addition of and Al into the metacarbonate rocks, surprising result given commonly-assumed immobility of Al in metamorphic fluids (Ague, 2003). The loss of CO<sub>2</sub> from the calc-silicates was extreme—about 70 to over 95 percent—and represents a source of CO<sub>2</sub> that is commonly overlooked when estimating CO<sub>2</sub> contributions to the atmosphere from mountain belts (Ague, 2003).

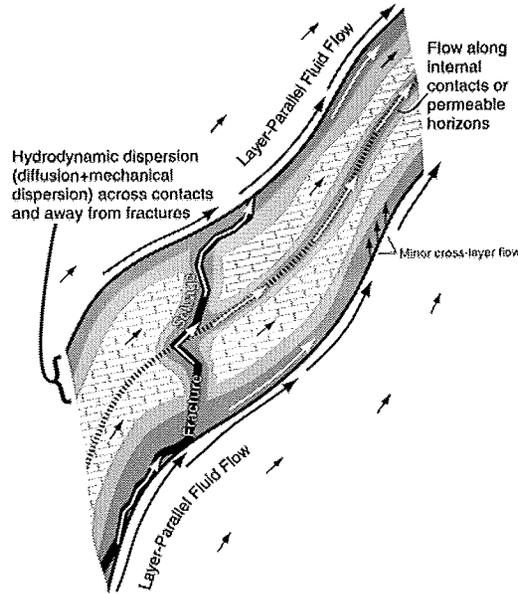


Figure 5. Schematic representation of possible fluid flow and mass transfer pathways in metacarbonate rocks. Increased reaction progress and carbon dioxide loss along lithologic contacts and vein margins denoted by darker shading. From Ague (2003).

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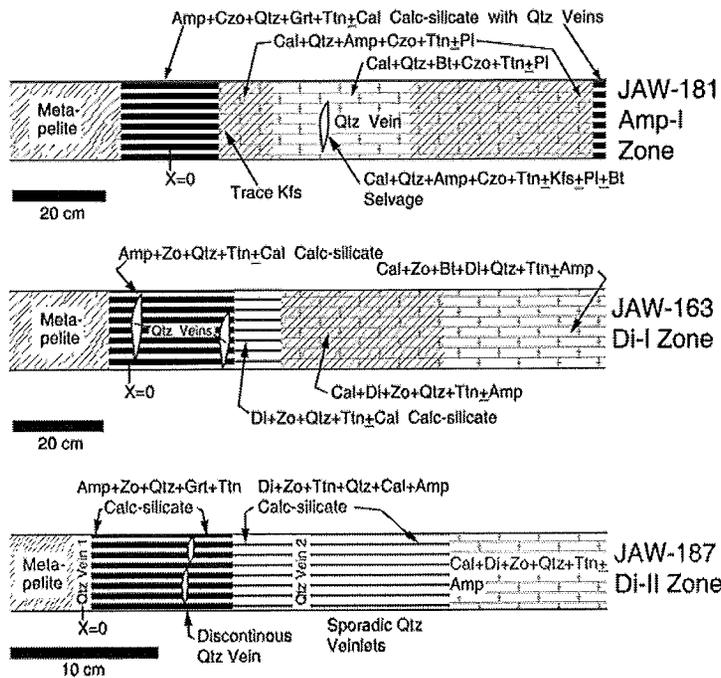


Figure 6. Lithologic profiles across metacarbonate rocks at contacts with metapelitic rocks and quartz veins (from Ague, 2003). Amp-I denotes regional Amphibole-I zone in which amphibole is present only on layer margins. Di-I and Di-II denote Diopside-I and II zones in which Diopside is present on layer margins and throughout layers, respectively. The Di-II zone layer interiors are the most devolatilized metacarbonate layer interiors in the Wepawaug Schist. Mineral abbreviations after Kretz (1983).

**Igneous rocks**

The Wepawaug Schist contains scattered pods, dikes, and lenses of leucocratic igneous rocks that are variously mapped as “Woodbridge Granite” and Devonian pegmatite (Fritts, 1963, 1965a). In the greenschist facies, the rocks are mostly tonalites composed predominantly of oligoclase and quartz with lesser muscovite, K-feldspar, and biotite; more granitic compositions are also present. Bahr (1976) concluded that the Woodbridge Granite comprised

volcanic rocks or shallowly-emplaced intrusions; field relations in the greenschist facies are reasonably well-preserved and are clearly consistent with such origins. At higher metamorphic grades, deformation is more intense and field relations are often complex. Xenoliths of pelitic schist can be found that contain truncated quartz veins, and xenocrysts of metapelitic garnet ripped from adjacent wallrocks are not uncommon (Ague, 1994b). These relations indicate that intrusion did not predate the metamorphism. Furthermore, many of the amphibolite facies igneous rocks have been deformed and some contain staurolite and kyanite of apparently metamorphic origin, strongly suggesting that at least some of the intrusions underwent peak metamorphism (Ague, 1994b). Thus, at least some of the intrusive activity in the amphibolite facies was broadly coeval with the metamorphism. This interpretation is not inconsistent with unpublished U-Pb-Th ion probe data for zircons obtained by C.M. Breeding at UCLA, which suggest an igneous age in the range of ~385 Ma. However, considerably more work is needed to determine the absolute timing and origins of igneous activity. For example, it is still unclear whether or not rocks mapped as Woodbridge Granite are in fact the same unit in both the greenschist and amphibolite facies.

### IMPLICATIONS FOR BARROVIAN METAMORPHISM

Barrovian metamorphism is commonly thought to result from the thermal and baric relaxation of tectonically-overthickened crust. This relaxation produces isograd reactions that are widely spaced in time (10's of millions of years). However, Baxter et al. (2002) showed that peak metamorphic heating in the classical Barrovian zones of Scotland was essentially simultaneous in the garnet through sillimanite zones, strongly suggesting advective heat input via syn-metamorphic magmas and associated fluid flow. Interestingly, this interpretation is consistent with the original views of Barrow (1893). It is possible that advective heating played a role in the Wepawaug Schist as well (Lancaster et al., 2005). For example, the Biotite-Garnet and Staurolite-Kyanite zone metamorphic ages are statistically identical, the results of Ague (1994b) indicate advective heat transfer by fluids, and field relations indicate that some of the felsic intrusions were penecontemporaneous with the metamorphism. The steep metamorphic field temperature gradient may be indicative of a relatively rapid influx of heat centered on the amphibolite facies portion of the schist.

### ROAD LOG

The field trip will focus predominantly on the progressive metamorphism of the metapelitic rocks of the Wepawaug Schist, but will also include opportunities to examine metacarbonate rocks, leucocratic igneous rocks, and the underlying Maltby Lakes Metavolcanics (Fig. 7). The exposures that we can visit with a large group are somewhat limited for a variety of reasons. Most of the problems involve access to private property, temporary flooding of Regional Water Authority property, safety issues on roads and railroad grades, and the destruction of outcrops by new residential and commercial construction. For example, we will be unable to visit the classic Staurolite-Kyanite zone exposures of metapelitic, metacarbonate, and igneous rocks along the active railroad grade that runs along the east side of the Housatonic river. Nonetheless, we will be able to see many key components of the prograde sequence in some of the most beautifully serene settings in southwestern Connecticut.

Trip begins at 9:00 AM at the Lime Kiln on Rt. 69 (Stop 1). Note that because the trip is on a Friday, traffic will be heavy, particularly on Rts. 15, 34, and 69. Exercise extreme caution when driving and crossing roads. Lunch is planned for Stop 5. Stores and restaurants are not close by so please bring your own food and water for lunch.

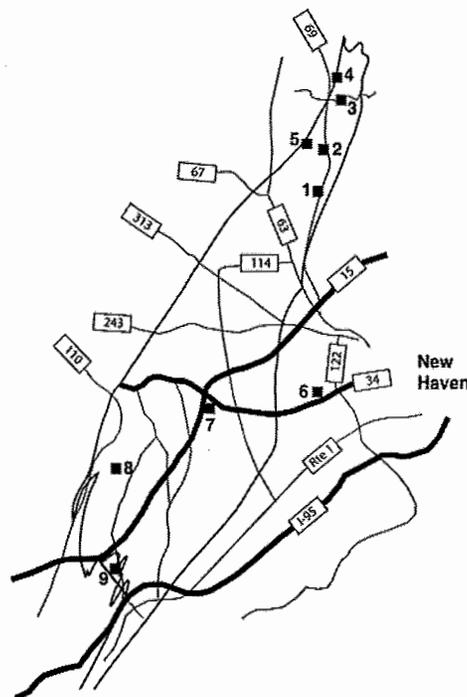


Figure 7. Roads and field trip stops.

Take exit 59 off Rt. 15 and head north on Rt. 69.

2.7 miles to Dillon Rd. (on left when heading north). Turn left and then pull around on southbound Rt. 69 shoulder. The road is busy so please pull off as much as possible.

**STOP 1.** Historic Lime Kiln (private Regional Water Authority property; access by permission). This stop is in the Chlorite zone of the Wepawaug Schist. Metapelitic and metacarbonate layers are well-exposed as a result of quarrying; this is a particularly good place to view the low-grade metacarbonate rocks containing calcite, ankerite, quartz, muscovite, and albitic plagioclase. Metacarbonate layers range from centimeters to over a meter in thickness.

Cement production by the Durgy Cement and Umber Co. took place in the 19<sup>th</sup> century after the Civil War. Note the abundant muscovite and considerable quantities of pyrite in the metacarbonate rocks. As a result of these mineralogical characteristics, the cement products made at the kiln were of relatively low quality, and the operation soon folded (historical information provided by Barbara Narendra, Yale Peabody Museum).

Turn back onto northbound Rt. 69.

0.65 miles north on Rt. 69 pull off onto shoulder; a large roadcut will be present on the left. *Exercise extreme caution when crossing the road as it is very busy.*

**STOP 2.** Chlorite zone metaclastic rocks. This is perhaps the most commonly-visited exposure of the Wepawaug Schist. Metaclastic rocks are composed mostly of muscovite, chlorite, quartz, and albitic plagioclase; pyrite is relatively common and a few layers contain accessory calcite and/or ankerite. The main deformational foliation is axial planar and, like the relic bedding, dips steeply; the axes of associated tight folds run sub-parallel to the long axis of the roadcut. These folds are crenulated in some areas, and a later stage of kink-folding is also apparent. Quartz-rich veins are obvious but comprise a relatively small volume of the exposure – around 2-3 percent. They commonly contain inclusions of wallrock ripped from the vein margins during fracturing. Note that muscovite and chlorite can be somewhat coarser grained in the veins and vein margins; this feature probably indicates the presence of fluid in the fractures during at least some of the metamorphism. At the southern end of the exposure thin layers of metacarbonate rock and Woodbridge Granite can be found. At some exposures, particularly at higher grades, the felsic igneous rocks appear intrusive, but the rock at this locality may represent bedded, tuffaceous material (Burger et al., 1968; Bahr, 1976).

Continue northward on Rt. 69.

1.95 miles to Hatfield Hill Rd. Turn right and proceed 0.5 miles to small parking area on right beneath power lines. Walk back along road to outcrops by dam.

**STOP 3.** Biotite-Garnet zone, south Lake Bethany dam (private Regional Water Authority property; access by permission). This exposure is near the low-grade boundary of the Biotite-Garnet zone. Here, mm scale porphyroblasts of biotite and garnet can be found in metapelitic schist rich in organic matter. The porphyroblasts are variably retrograded to chlorite-bearing mineral assemblages, but fresh examples are present. Relic bedding, deformational fabrics, and quartz veins are well-exposed in and around the dam spillway.

Retrace path back to Rt. 69 (also called the Litchfield Turnpike), and turn right (northbound).

0.95 miles, turn right onto Hoadley road. Travel 0.2 miles to stop sign, turn left and go over small bridge, then immediately turn right.

0.4 miles to gated entrance to Lake Bethany area on right. Enter dirt road and park so as not to block road. Walk eastward through trees across dry (hopefully!) reservoir lake bed to outcrops.

**STOP 4.** Upper Biotite-Garnet zone including metacarbonate layers (private Regional Water Authority property; access by permission). These exposures may or may not be accessible depending on the water level in the reservoir.

Walking westward across the lake bed one encounters glacially-polished outcrops dominated by garnet-biotite schist. Note that the garnets are significantly larger here (some up to ~0.5 mm), even though we are only a short distance up-grade from Stop 3 (this stop is visible at the southern end of the reservoir). The considerable increase in grain size over a relatively short distance reflects the steep metamorphic field temperature gradient of ~58 deg. C per km. Walking southward along the reservoir margin one encounters outcrops of the Woodbridge Granite. Is it an intrusive rock at this locality? Is it pre-, syn-, or post-metamorphic?

Further to the west, glacially-polished outcrops of garnet-biotite schist containing intercalated metacarbonate layers are present. **PLEASE DO NOT DESTROY THE METACARBONATE LAYERS BY HAMMERING.** The metacarbonate layers are mineralogically zoned such that the least-reacted, biotite-bearing assemblages are present in the layer interiors; more-reacted hornblende-bearing layers occur near contacts with metapelitic rocks and in vein selvages; and strongly metasomatic, veined calc-silicate rocks rich in zoisite, hornblende, and garnet are found directly at lithologic contacts and vein contacts (Fig. 8; Ague, 2003). These relations are the result of transport of reactive fluids across contacts and fractures which drove CO<sub>2</sub> production.

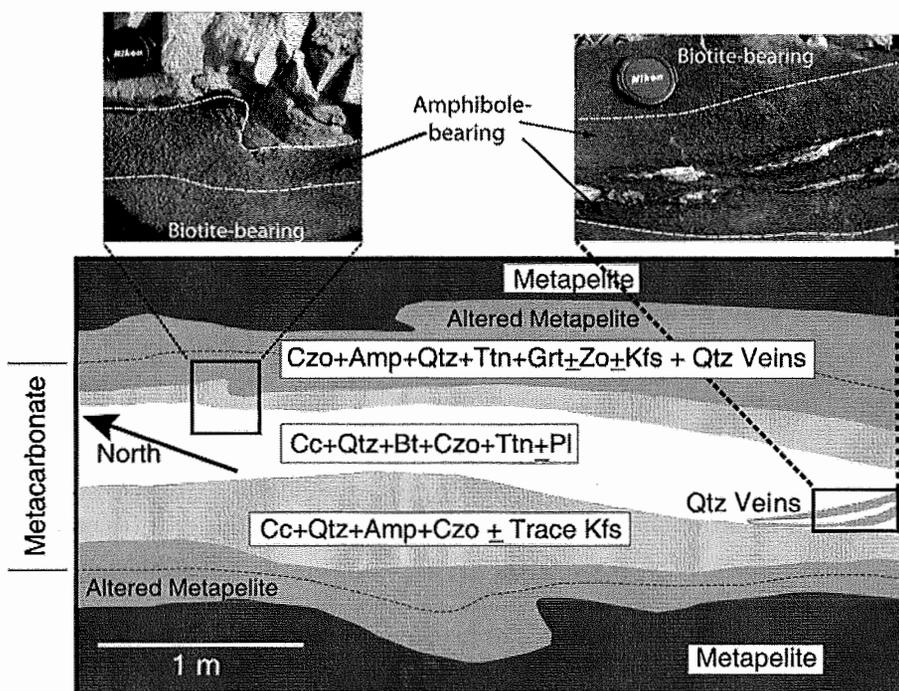


Figure 8. Map of geologic relations among metacarbonate and metapelitic rocks at Stop 4. Mapping and photographs by J.J. Ague.

Retrace path back to Rt. 69.

Turn left onto southbound Rt. 69 and travel to Morris Rd. (on right, just to the north of Stop 2).

0.6 miles on Morris Rd. to stop sign at intersection with Sperry Rd. Turn left, proceed a few tens of feet, then turn right into gated entrance to Lake Chamberlain area.

**STOP 5.** Upper Biotite-Garnet zone (private Regional Water Authority property; access by permission). Lunch stop. Relatively coarse-grained garnet-biotite schist, with lesser amounts of intercalated Woodbridge Granite; at least one layer of hornblende-rich calc-silicate rock can also be found. Vein abundance in the Biotite-Garnet zone exceeds that in the Chlorite zone, and in general increases as metamorphic grade increases (Fig. 4; Ague, 1994b). The veins are deformed, but this fact does not indicate that veining pre-dated metamorphism. The deformation of fractures is to be expected as a natural outcome of the overall deformation of the rock mass (Fig. 4). Fracturing of garnets (Ague, 1994b) is a clear indication that veining did not pre-date the metamorphism.

Exit Lake Chamberlain area and retrace path back to Rt. 69.

Turn left (southbound) onto Rt. 69 and head back about 3.5 miles to intersection with Rt. 15. Take entrance ramp for southbound (New York) Rt. 15.

Travel about 4 miles on Rt. 15 and get off at exit 57. Turn right (eastbound) at intersection with Rt. 34.

2.9 miles, turn left at big yellow 35 miles per hour sign, then turn left onto westbound Rt. 34. Get over in right lane, travel a few hundred yards, and enter gated Maltby Lakes recreation area on right.

**STOP 6.** Type locality of the Maltby Lakes Metavolcanics (private Regional Water Authority property; access by permission). The type locality of the Maltby Lakes Metavolcanics, the rock unit that directly underlies the Wepawaug Schist. The rocks are in the Chlorite zone of Barrovian metamorphism, and are mostly green basic to intermediate metavolcanic rocks with some intercalated phyllites. The metavolcanics are dominated by chlorite, actinolite, epidote, and albite, but also include smaller amounts of quartz and carbonate minerals. Veins containing quartz and carbonate minerals, often with epidote, are common. Given that the Maltby Lakes metavolcanics underly the Wepawaug Schist, was dehydration of the metavolcanics during metamorphism a source for fluids that flowed upward through the Wepawaug?

Elsewhere in the area scattered occurrences of serpentinized, carbonated, ultramafic rocks can be found. Higher-grade equivalents a few miles to the southwest were quarried by Baldwin's Verde-Antique Marble Company. These rocks, originally discovered by Solomon Baldwin on a class outing led by Silliman in 1811, were used for ornamental purposes and in gravestones. The rock was quite attractive and was used in a number of buildings, including mantelpieces in the East Room of the White House and the Smithsonian Institution. The quarrying operation was relatively short-lived, at least in part because the sulfides (including Ni-sulfides) contained in the rocks broke down rapidly when exposed to the weather, dissolved the carbonate minerals, and thus severely degraded the ornamental stone. The ultramafic lithologies were also extracted at the Maltby Lakes locality, but the rocks were difficult to extract and polish. The quarry, which operated at a loss, was soon closed (historical information provided by Barbara Narendra, Yale Peabody Museum).

Return to intersection with Rt. 34 and turn right.

2.6 miles, turn left onto Mapledale Rd. (note that there are other "Maple..." roads around, like Mapleview"; the one you want is Mapledale). Immediately pull over to the right along shoulder and park, but take care not to block the road.

**STOP 7.** Type locality of the Wepawaug Schist. Here metapelitic rocks of the Biotite-Garnet zone crop out along the banks of the Wepawaug river. The exposure also contains biotite-bearing metacarbonate layers and generally conformable layers of Woodbridge Granite. The degree of accessibility will depend on the water level in the river.

Turn around and head back to Rt. 34. Pass over to opposite side of road and turn left onto Rt. 34, heading west.

2.2 miles, turn left onto Derby-Milford Rd.

1.35 miles, turn right onto Wheeler's Farms Rd.

1.4 miles, turn right onto Herbert Lane (a different road than Herbert Street!).

0.15 miles to Wolcott Ln., turn right.

0.15 miles to Windy Hill, turn left.

0.1 miles to Wagon Trail, turn right.

0.25 miles to stop sign. Continue through stop sign to Broadview Rd, and then turn right.

0.15 miles to Aspen Ln., turn right and proceed 0.2 miles to intersection of Aspen Ln. and Cedar Grove.

**STOP 8.** Granitic intrusive rocks and quartz veins with comments on diopside zone metamorphism in the area (private property, access by permission only, no hammers). This exposure lies within the regional Staurolite-Kyanite zone (amphibolite facies). The igneous mass contains plagioclase, quartz, muscovite, quite variable amounts of K-feldspar, and small quantities of garnet and biotite (biotite altered to chlorite in some areas). Considerable quartz veining is present; vein widths reach at least 1.5 meters. Is this igneous body part of the Woodbridge Granite visited at the lower-grade exposures, or is it a different rock unit altogether? The magma intruded metapelitic schists that are visible in the surrounding lawns (do not visit/hammer these exposures). Note that the rocks dip to the east-southeast, as opposed to the lower-grade rocks on the eastern limb of the Wepawaug syncline which dip mostly to the west-northwest.

An example of diopside zone metacarbonate rock illustrating metasomatic zoning taken from the nearby railroad track will be displayed and discussed (see Ague, 2003). Palin (1992) and van Haren et al. (1996) found evidence for isotopically-light oxygen in some metamorphic quartz veins, and these fluids drove at least some of the diopside-producing reactions in the metacarbonate rocks. Were these fluids derived from intrusive bodies like the one at this stop?

Retrace path out of subdivision (exit along Aspen Ln., turn left onto Broadview, left onto Wagon Trail, left onto Windy Hill, right onto Wolcott, and left onto Herbert Lane).

Proceed to Wheeler's Farms Rd. and turn right (southbound).

2.3 miles, turn right onto E. Rutland Road (past the Rt. 15 underpass).

0.15 miles, turn left into small residential court and park (street is "Chevelle Place" but you can't see the sign easily from the direction we will be headed). **PLEASE TAKE CARE NOT TO DISTURB OR WALK ON THE NEIGHBOR'S LAWNS AND LANDSCAPING.**

Walk across street through bush and trees to outcrops under power lines.

**STOP 9.** Staurolite-Kyanite zone Wepawaug Schist and quartz veins (access to outcrops via private property by permission only; hammering float is OK but outcrop needs to be preserved). Quartz veins are abundant and have alteration selvages rich in kyanite and, to a lesser degree, staurolite. These aluminous minerals are, in contrast, rare or absent in the matrix schist away from the veins. Silica was lost from the selvages, and deposited in the adjacent veins. Metamorphic garnets record this process, as quartz inclusions are abundant in garnet cores and become progressively smaller and less abundant toward rims (Fig. 9). On average, about 70 volume percent of the silica in amphibolite facies veins was derived from the selvages, and about 30 percent was precipitated from externally-derived fluids (Ague, 1994b). These fluids were able to leach alkali and alkaline earth metals from the selvages, leaving behind an aluminous residue in which abundant kyanite (+/- staurolite) could crystallize (Fig. 10). Where did these large volumes of fluid come from, and did they transport heat as well as mass?

If the exposure is not too overgrown with vegetation, some isolated exposures of amphibolite of the Maltby Lakes metavolcanics may also be seen in the area.



Figure 9. Photomicrograph of Staurolite-Kyanite zone metapelite rock from quartz vein selvage. Note that quartz inclusions in garnet (arrow) become progressively less abundant from core to rim, reflecting loss of silica to quartz vein during garnet growth. Selvage matrix is now nearly devoid of quartz. Other phases in photo are mostly biotite and kyanite. Plane-polarized light, field of view 4.4 mm wide. From Ague (1994b).

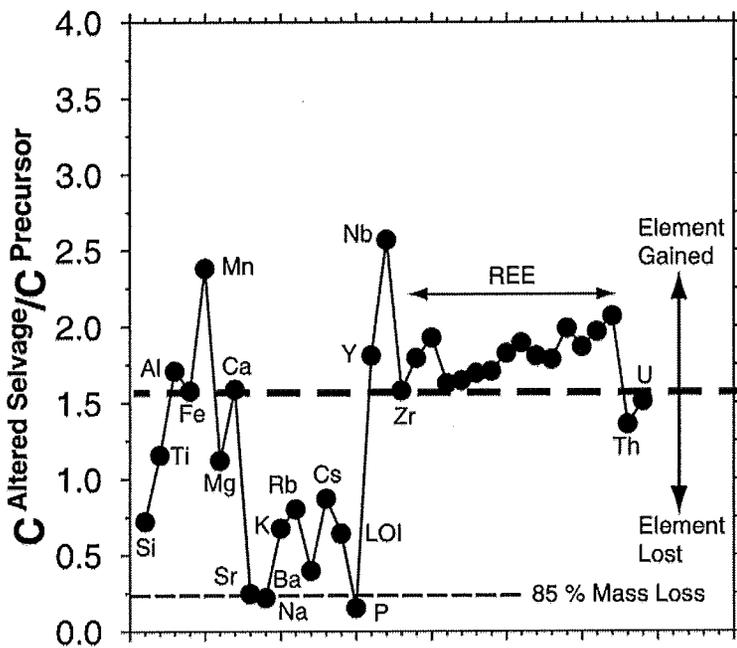


Figure 10. Concentration ratio diagram for altered selvage relative to less altered precursors far-removed from veins in same outcrop. Heavy dashed line represents geochemical reference frame for elements that underwent little or no gains/losses; the corresponding concentration ratio of ~1.55 reflects about -35 % mass loss from selvage (dominated by silica transfer to adjacent quartz vein). Ti was also deposited in the vein in rutile. Note pronounced losses of many alkali and alkaline earth metals. The loss of metals like Na and K produced a highly aluminous selvage rock in which abundant kyanite and staurolite could crystallize. Increases in Mn and heavy REE reflect growth of garnet in selvage. J.J. Ague, preliminary unpublished data. Concentration ratio diagrams discussed in detail by Ague (2003).

Return to vehicles, and retrace path back along East Rutland Rd. to Wheeler's Farms Rd. Turn left (northbound); road intersects Rt. 15. End of trip.

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## **ARCHAEOLOGY OF MINERAL AND WATERPOWER RESOURCES IN NORTHWEST CONNECTICUT**

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### **INTRODUCTION**

Starting in 1734, the people of the Salisbury district in northwest Connecticut exploited the region's iron ore and waterpower resources with fuel from sustained-yield forestry to build a nationally-important iron industry that endured to 1923. They supplied the Continental Army with cannon. In the early years of the republic they made the metal parts for machinery and the supplies of raw materials used by Eli Whitney and other New England entrepreneurs in launching manufacturing with interchangeable parts. By 1825 the Salisbury district iron makers has established themselves as the sole-source supplier of high-quality gun-iron to the national armories at Springfield and Harpers Ferry. Completion of the Housatonic Railroad in 1842 enabled a new generation of entrepreneurs to move to a new line of products. By 1855 Horatio Ames's works at Falls Village was one of the nation's two largest makers of locomotive tires. After 1865, and continuing until 1923, the region's ironmasters concentrated on casting railroad car wheels, which they sold to lines as far west as the Union Pacific.

Ironmaking is energy-intensive. The ironmasters in northwest Connecticut used water power to drive their machinery, and wood fuel usually converted to charcoal for their furnaces and forges. Rivers such as the Blackberry were of a size that enabled them to obtain the mechanical power they needed with a capital investment within their means. Ironworks proprietors gradually acquired tracts of woodland ranging to 10,000 acres or more for their fuel supplies. They introduced the coppice system of forest management to assure a continuous yield of wood for coaling. The region's wood, water, and ore resources remained adequate to sustain iron making until metallurgical advances in the 1920s eliminated the demand for charcoal-smelted iron.

At the dawn of the 20<sup>th</sup> century the development of long-distance, high-voltage electric transmission by Eli Terry of Hartford and others opened a new use for the northwest's water resources. Falls along the Housatonic River were excellent sites for hydroelectric power plants. These could generate far more power than was needed locally. Transmission lines exported power to Waterbury and other cities that lacked hydropower resources of their own. The system of interconnected power plants along the Housatonic from Falls Village through Derby includes the nation's first pump-storage facility (at Rocky River in New Milford), and remains a valuable resource for peaking power on the Connecticut grid.

Even while mining, smelting, and forest harvesting created wealth in the northwest in the mid-19<sup>th</sup> century, residents began a gradual transition from dependence on mining and smelting to residential and recreational uses of their land. In 1846 they changed the name of Furnace Village to Lakeville, and Furnace Pond to Lake Wonoskopomuc. The newly-opened Housatonic Railroad brought summer visitors such as Henry Ward Beecher to fish the streams, hike woodlands and, for variety, visit the ironworks. The gradual transition to new uses enabled northwest Connecticut to avoid the economic and social decay that faces many regions at one time dependent on extractive industries. Today woodland tracts assembled for ironworks are forest preserves; the furnaces and forges are today's historical and archeological sites.

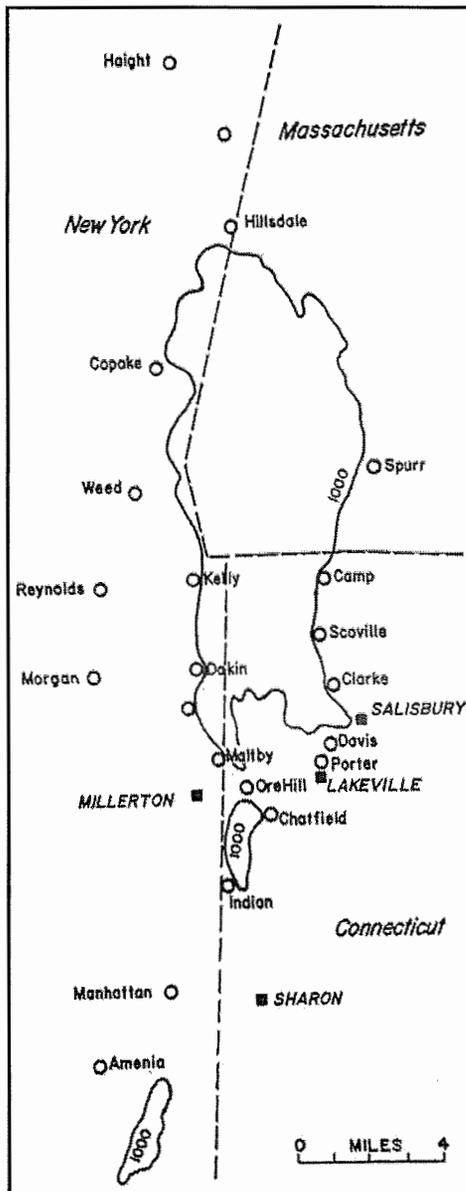
### **GEOLOGICAL SETTING**

Northwest Connecticut is made up of two terranes. The first, the Proto-North-American Terrane, was built on a core of the oldest rock in North America. Nearly a billion years ago, continental collision attached additional rock to this core, and pushed it upward to form the Grenville Mountains. The eroded remnants of the Grenvilles, generally consisting of granite-gneiss, are the oldest rock in Connecticut. Next, about 200 million years later, motion of tectonic plates broke the old continent apart, leaving the Grenville Mountains on the coast of the newly-formed Iapetus Ocean. Sediment eroded from the Grenville Mountains accumulated in the shallow margins of the Iapetus Ocean, which also held in its shallows a great bank of coral reefs. Today the limestone formed from the

coral appears as the band of Stockbridge dolomitic marble that runs diagonally across northwestern Connecticut. The remnants of the Grenville Mountains, the rock formed from the sediment that eroded off the mountains, and the limestone together formed the continent called Laurentia, and today make up the proto-North-American Terrane in Connecticut.

About 500 million years ago a collision between Laurentia and a large offshore island pushed deep-water sediments and volcanic rock from the floor of the Iapetos Ocean onto the edge of Laurentia. The rocks formed from this newly-accreted material now make up the region's second terrane, the Iapetos. A line generally trending southwest to northeast (Cameron's Line) separates the proto-North-American Terrane (on the northwest side) from the Iapetos Terrane (on the southeast side). Cameron's Line is not a perfect boundary because the collision with Laurentia pushed pieces of Iapetos Terrane on top of the older proto-North-American Terrane. After this last continental collision, hundreds of millions of years of erosion removed much of the rock of both terranes.

All the rock that is left from these events was once deeply buried. It owes its toughness to the heat and pressure it encountered deep in the earth. The alignment of the rock units of the two terranes caused by the processes of accretion, together with their differing resistances to weathering, determines the bedrock topography of the region. The easily-weathered marble forms the lowland areas. The schist, with intermediate erosion resistance, forms high ground with moderately rugged topography, as on Mount Riga and along the route from Litchfield north through Goshen. The granites-gneiss of the Grenville formation is the most resistant to weathering, and forms the region's most rugged topography.



**Figure 1. Most of the iron mines in the Salisbury District (open circles) are near the contacts between the Stockbridge marbles, below the 1000-foot contours, and the Walloomsac schist and other erosion resistant rocks that form the higher elevations.**

Deposits of iron ores, primarily goethite formed by the weathering of limestone, are located along the contacts of the Stockbridge marbles of the lowlands and the Walloomsac schist of the adjacent mountains (Figure 1). The largest of these deposits formed Ore Hill in Lakeville, acquired in 1734 by a company of proprietors that included Jared Eliot, the distinguished colonial minister, doctor, metallurgist, and agricultural reformer. Open pit mining continued into the 20<sup>th</sup> century. Ore Hill is today a lake with its mine dumps reforested.

The field trip route follows the Blackberry and Housatonic rivers as they traverse the Stockbridge marble and pierce the Grenville formation in water gaps.

### Industrial Archaeology of the Blackberry River

The Blackberry River (Figure 2) offered settlers in Canaan and Norfolk numerous water privileges of the ideal size for starting industries. Richard Seymour's 1738 forge was the first of a cluster of ironworks that formed North Canaan's principal industry for a hundred and fifty years, and continued to 1923. By 1795 Samuel Forbes with partner John Adam, Jr. had enlarged Seymour's old forge on the Blackberry into works supplying metal and machinery to customers throughout southern New England. Through the early decades of the republic, Forbes & Adam supplied New England's infant manufacturing industries with heavy forgings and machine parts from its works on the Blackberry. Members of the Adam family remain in their houses on the Blackberry, now as proprietors of a vineyard located at the site the last operating blast furnace.

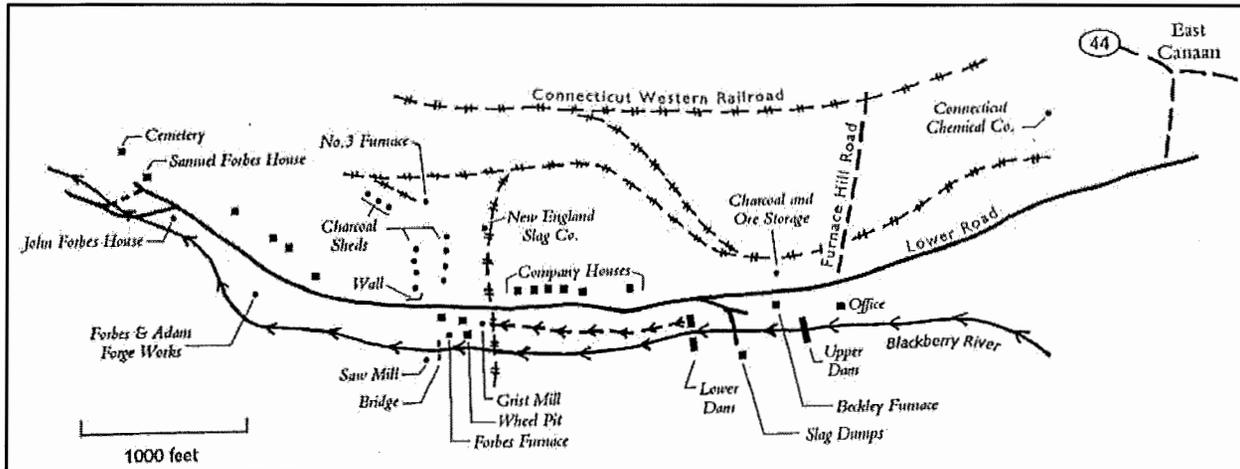


Figure 2. Industrial and Archaeological Sites along Lower Road in East Canaan.

### ROAD LOG

0.0 miles.

**Stop 1. Beckley Furnace Site.** (1 HOUR) The trip starts at 9 AM at the Beckley Furnace State Industrial Monument. The site is 0.5 miles from the intersection of Lower Road with Route 44; the East Canaan church is a landmark near the intersection. Park cars at the adjacent education center, at the furnace, or along the road.

John Adam Beckley built this blast furnace in 1847 at the site of the Forbes & Adam rolling and slitting mill, then obsolete and occupying a valuable water privilege at the first of three dams on the Blackberry River that powered industrial works. The terrace north of Lower Road provided the large, level area needed for the furnace bank (the storage area for ore and charcoal). A bridge crossed over the road to the charging house on the top of the furnace stack. In its original configuration a hot blast stove was placed at the top of the stack. Blast air was carried by a pipe within the furnaces structure to a bustle pipe that surrounded the hearth at the bottom of the stack, and injected onto the furnace by three or more tuyeres. The casting shed extended outward from the west side of the stack.

The Upper Dam, a short walk upstream from the furnace, held water to drive the blast engine. A two-cylinder blast pump, driven by an overshot wheel, had cylinders 72 inches in diameter made of wooden staves with cast iron heads. The cylinders were held together by long rods. The piston stroke was 6 feet. A weighted piston in a third cylinder regulated the pressure. This form of blower was cheap, but not very efficient.

The furnace was rebuilt in 1856, and again in 1880, when the hot blast stoves were moved to ground level on the east side of the stack. Probably at this time the overshot wheel was replaced by the turbine now in place at the dam, and an auxiliary steam engine was installed for use during periods of low river flow. The furnace plant burnt

in 1896, was rebuilt, and was back in blast in 1898. The new casting and engine houses were rebuilt with brick and metal roofs. In its final configuration after the 1880 rebuild the furnace shaft as 40 feet high. The stack above the crucible was lined with fire brick; the crucible was made of schistose quartzite from the Dalton formation. The favored source was on Mine Mountain is Sharon adjoining Mount Easter (Bartram's quarry). Blast air was supplied to five tuyeres at 650°F and a pressure of 1 psi. The furnace made about 15 tons of iron per day, using 100 bushels of charcoal to the ton of iron.

The Beckley furnace was closed when a salamander blocked the hearth in 1919. Salvagers carried away the brick casting house and other materials. In 1946, at the behest of Charles R. Harte, concerned citizens raised money to buy the site for the state. The great weight of a stone furnace stack consolidates the soil under it. Non-uniform consolidation can throw the stonework out of shape, and may lead to collapse of the stack. Distortion of the front arch of the Beckley furnace stack visible in pictures taken after its abandonment in 1919 indicates outward movement of the south wall of the furnace, probably due to slippage of the stone retaining wall along the river bank. Since further movement of this wall could put the furnace stack at risk, a committee of volunteers with state support undertook restoration work to stabilize the stack in 1998. They have since added a museum in the former furnace company office, and interpretive signs. The reconstructed furnace hearth, based on partial archaeological evidence, may be compared with an original hearth still in place at the Kent furnace.

Today the stack, dam with its in-place turbine and tail race, and several retaining walls remain as standing structures from the extensive furnace plant. The ashlar wall that once supported the north end of the charging bridge stands on the north side of the road above the stack. A pile of salamanders lines the north bank of the river just below the bridge, making, with the one in front of the furnace, a record of the number of times the furnace hearth was rebuilt. The huge slag piles reached by the bridge across the river indicate the overall amount of iron the furnace made. The pond behind the dam no longer has significant storage capacity, being filled with silt. We lack a study of the sediment load carried by the Blackberry. However, the silt accumulation in the pond may be a record of erosion of the surrounding hillsides that resulted from the clear cutting technique used in the coppice system of harvesting wood for coaling.

Before leaving this stop be sure that you have signed the list for admittance to the Falls Village hydro station.

(10 AM) Return to cars and proceed west on Lower Road.

0.2 miles. A row of houses built by the Barnum-Richardson Company for the staff of the Canaan furnaces stands on the north side of the road. Julia Michaels, daughter of a worker at the furnaces who moved into the fifth house from the west in 1898, remembered that the houses had no electricity, plumbing, or running water. A pump in the kitchen drew water from a cistern filled by a spring on the hillside. Monthly rent was four dollars, then two and a half day's wages.

Immediately west of the company houses the Canaan No. 3 furnace, built in 1872 and in blast until 1923, stood with its charcoal sheds on the north side of Lower Road in what is now a vineyard. Here spur tracks for delivery of ore and charcoal from the Connecticut Western Railroad ran directly to the furnace bank. The New England Slag Company began recycling here in the early twentieth century. It shipped out car loads of crushed slag for use as aggregate, roofing material, and road surfacing. After excavating most of the No. 3 furnace's slag banks, the company build a narrow gauge railway over the Blackberry River to get at the Forbes furnace slag on the south side of the river. Concrete abutments of the bridge stand upstream of the gristmill site.

Salvagers have carried off everything of value from the No. 3 furnace plant, including the stone blocks of the stack itself. The Adam family has converted the field in which the furnace stood into a vineyard. The foundations of the slag company's crushing plant now stand in the midst of the vines. A mass of slag that once lined the furnace crucible marks the eastern entrance of the vineyard. Two holes in it show where tuyeres entered the furnace. Nearby are two salamanders, masses of solid iron and slag that accumulated at the bottom of the blast furnace hearth over a period of months, and had to be removed when the hearth was rebuilt.

0.6 miles. Bear left across the Forbes Bridge. The earliest ironworks, Samuel Forbes's forge, was located just above Forbes Bridge. Samuel Forbes's house (now on the National Register of Historic Places) with its adjacent

cemetery is ahead on the right. Forbes built a center-chimney salt box in 1754, and lived here until his death at age ninety-eight in 1828. Later owners rebuilt the house in its present two-chimney design. Houses owned by John Adam and his children are to the east.

1.1 miles. The large dolomite quarry, opened by Alexander Maxwell, is on the left. In 1874 Maxwell used the then newly-constructed Connecticut Western Railway to ship dolomite marble blocks to Hartford for construction of the new state capitol. The rock here is so nearly free of cracks that no ground water flows into the pit; the quarrymen need only pump out rainwater. The piles of white quarry debris adjacent to the pastures of the adjoining dairy farm illustrate the juxtaposition of industry and agriculture once common in the northwest. A long pile of quarry waste north of the road across from the pit has weathered enough to support some stunted vegetation. Trucks haul the dolomite to the New England Lime Company plant to process for use in pharmaceuticals, cosmetics, fertilizers, and building materials.

2.0 miles. Junction with Route 7. Turn right.

2.1 miles. Lawrence Tavern (National Register) on left.

2.2 miles. Junction with Route 44; turn left.

2.3 miles. Cross the former Connecticut Western Railroad.

2.4 mile. Canaan Village, railroad station on the left. The village of Canaan owes its origin to the completion of the Housatonic Railroad in 1842. William Adam, grandson of Samuel Forbes, sold farmland along the tracks for a new community. The center of commercial activity moved away from the Blackberry River to the new town built along the railroad. The Canaan depot was one of the finest surviving 19<sup>th</sup> century-railroad buildings in Connecticut until partially destroyed by fire several years ago.

2.6 miles. Route 44 makes a sharp turn to the left here.

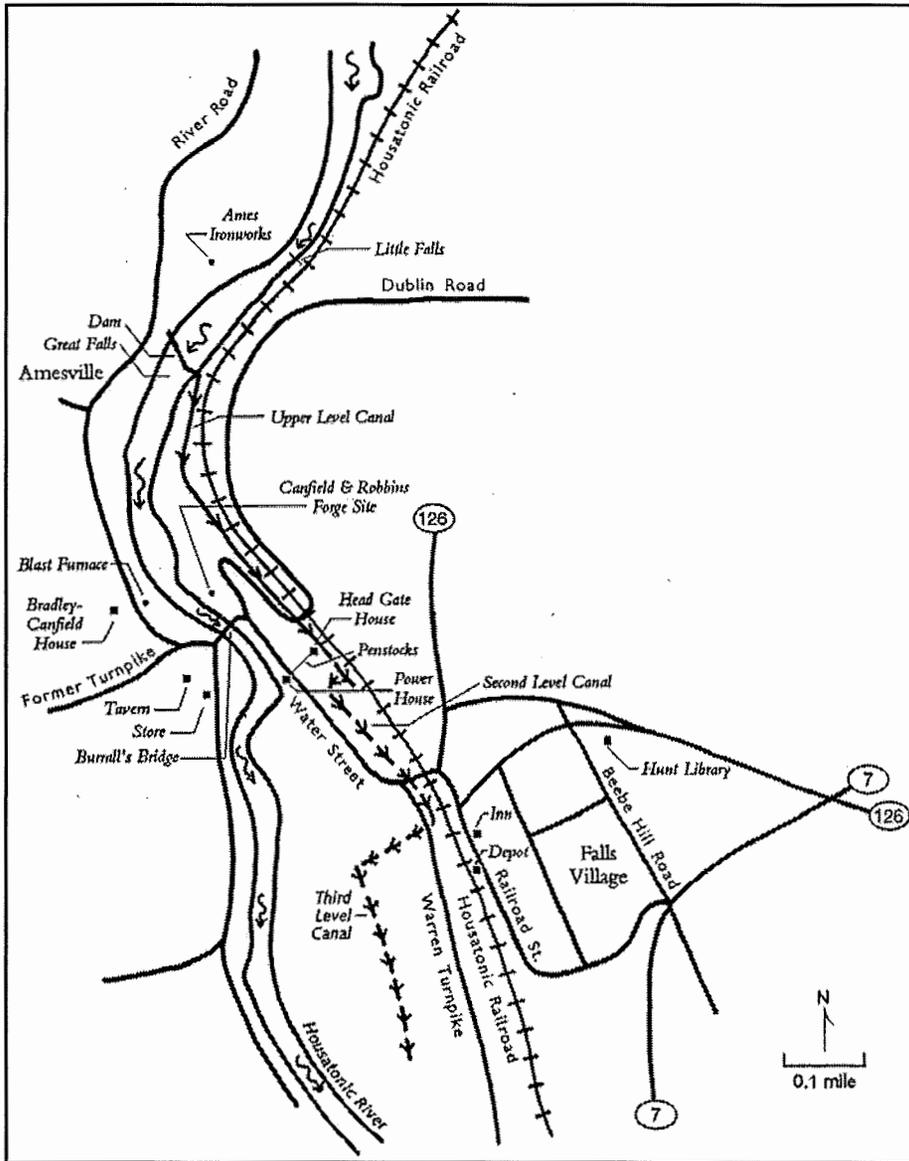
4.8 miles. Bridge over the Housatonic River.

5.6 miles. Turn left on Housatonic River Road. This is a dirt road, and follows the route of the Warren Turnpike. In the early years of the republic Connecticut relied on the traditional twice-yearly call out of all men between sixteen and sixty for make and mend roads. Frustrated with the failure of these amateur builders to make roads that could serve the growing agricultural exports and manufactures of the state, the legislature turned to privatization. State charters empowered proprietors to build and maintain roads, and collect tolls from travelers. The turnpikes were Connecticut's first regulated public utilities. The legislature chartered the Warren Turnpike Company in 1809 to build from Kent to Falls Village, and later to the Massachusetts line. Turnpike builders aimed at straight roads with low gradients.

8.6 miles. Look for the parking area marked by steel gates on the east side of the road. This is a favorite site for launching canoes; it may be necessary to park along the road.

**STOP 2. Amersville.** (10:30 AM; Figures 3-4; 30 minutes)

The Little Falls at the Housatonic River water gap and the Great Falls below it offered some of the finest water power sites in northwest Connecticut. Several early mills perched at the top of the Great Falls, built by proprietors who lacked capital to fully utilize the power potential available.



**Fig. 3. Industrial and Archaeological sites at Amesville and Falls Village.**

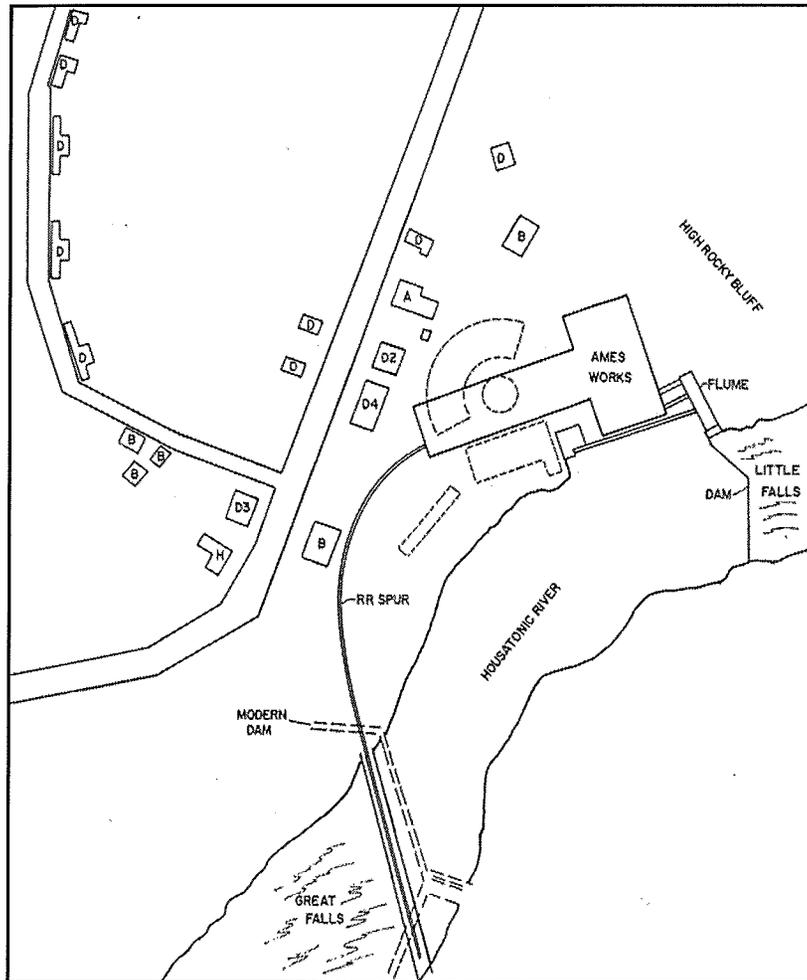
In 1832 John Eddy, Horatio Ames, and Leonard Kinsley bought land for a new ironworks where they would make wrought iron in wood-fired puddling furnaces, then a novel technique. They soon had their works producing wagon and railroad axles, crowbars, and other iron tools. A dam at the Little Falls diverted water to the wheels that drive the hammers and other works machinery. They got the wood fuel for their puddling furnaces by floating logs down the river from Massachusetts. A boom stretched across the river at the top of the Great Falls trapped the incoming logs. By 1836 Oliver Ames, who had a large shovel works in North Easton, Massachusetts, owned the ironworks, and had Horatio, his son, installed as proprietor.

The Housatonic Railroad, completed through Connecticut in 1842, ran a spur line over a bridge across the top of the falls to reach the Ames works from its tracks on the east side of the river (Figure 4). The direct rail connection enabled Ames to bring in mineral coal for his furnaces and ship out heavy forgings. By 1850 the Ames Iron Works was valued at \$150,000, and was doing \$180,000 worth of business a year. It works employed eighty men, many of whom lived nearby in houses financed by Ames. The community, eventually called Amesville, included a school, the Housatonic House hotel, and a store.

Ames further enlarged the forge so that by 1857 it had two double and four single puddling furnaces, five heating furnaces for forgings, another for tires, four swedging fires, two Nasmyth steam hammers of 5 and 2.5 tons (said to be as large as any in the United States) and six heavy, water-powered trip hammers for large forgings. Specially-built shaping machines rolled tires for railway locomotives. In 1856 the works made about 800 tons of railroad iron, steamboat shafts, and similar forgings. It employed two hundred artisans in the main shop of 250 x 80 feet and a second of 132 x 80 feet.

Ames undertook manufacture of large, wrought-iron cannon that attracted national attention during the Civil War. The Ames works made at least six fifty-pounders (later bored to eighty-pounders) and fifteen seven-inch guns.

Government tests showed the Ames guns to be superior artillery. However, they cost a dollar per pound, ten times the cost of cast iron guns. Because of the high cost, Ames never got enough orders to pay for the special machinery he had bought for the cannon project. He had neglected other work while he concentrated on the cannon. By 1863 he was in serious financial difficulties.



**Figure 4. Archaeological features at Ameshville showing dwelling houses, D, the hotel, H, barns, B, and Horatio Ames house, A. Dashed lines show the Housatonic Railroad's round house and related buildings.**

Directly after Horatio Ames's death in 1871 his brother sold the forge property to the Housatonic Railroad for its repair shops. The railroad sold off the stock of iron on hand, sent the largest steam hammer, "Thor," to the American Silver Steel Works in Bridgeport, and built its round house on the

site of the main forge shop. It built a reservoir and a gas generator (for illumination) on the hill behind the works. As the ironworkers left, railroad employees moved into their homes.

When the New York, New Haven & Hartford Railroad leased the Housatonic line in 1893, it gradually transferred locomotive repair work out of Falls Village. By 1904 all the shop buildings were vacant. The Connecticut Power Company bought the property in 1913 and razed the remaining buildings since the dam it would build at the crest of the Great Falls would flood the site of the forge and railroad shops. Remains of the former industrial community include a particularly well-preserved set of houses on Puddlers Lane built in Greek Revival style for artisans at the Ames works. Cellar holes of Ameshville's hotel, Horatio Ames's house and numerous dwelling can be found along River Road, Figure 4. The head gates of the canal that carries water to the hydropower plant below are on the east side of the river at the end of the present dam. Anchor bolts for the former railway bridge across the river can be seen just below the dam. A large mass of iron left from the Ames works rests near the crest of the falls.

(11:00 AM) Return to cars.

9.1 miles. Site of Bradley Canfield blast furnace (on left, not visible) and Canfield house (on right). Benjamin Silliman, who visited Bradley's furnace in 1817, described it as about twenty-five feet high and ten feet square, with

blast supplied by water-powered bellows. The remains of the furnace and scattered slag stand beside the old Appalachian Trail next to the river with the wall that supported the charging bridge near the road above.

9.2 miles. Junction with Falls Mountain Road (former Salisbury and Canaan Turnpike, Silliman's route to Lakeville).

9.2 miles. Junction with Dugway Road at Burrall's bridge. Bear left across the one-lane bridge, turn right on Water Street. The 1909 truss bridge was ingeniously saved for continued service with heavier loads in 1974 by inserting a steel arch into the truss and rebuilding the deck.

9.4 miles. Public parking area just past the Falls Village hydropower station. Walk up dirt road to power canal; return for inspection of the hydropower station. Entrance to the power station is restricted to those who have signed up and have a photo driver's license.

**STOP 3. Housatonic Hydropower.** (11:10 AM; Figure 3; 1 hour 30 minutes) Eighteenth-century entrepreneurs built saw, grist, and paper mills near the top of the falls, and a bloomery forge at the base. The return of prosperity after the panic of 1837 and the completion of the railroad inaugurated a wave of industrial expansion in Canaan. Lee Canfield and Samuel Robbins took over an old forge below the falls. Here they fined pig iron smelted in their blast furnace across the river to make made gun iron, railway axles and other forged products. The forge stood north of Burrall's Bridge on the east bank of the river, its site marked by slag deposits in the river bank. The large, white house across the bridge (now restored and enlarged) was at one time the home of one of the forge artisans, and was later used as a store.

In 1845 Canfield and Robbins thought they had the resources and expertise they needed to fully utilize the Housatonic's power at the Great Falls for manufacturing. They visualized an industrial community at Falls Village similar to Lowell, Lawrence, or Holyoke Massachusetts. Canfield and Robbins organized the Water Power Company with capital of \$200,000 to dig a channel at the top of the falls that would divert water into a 2,100-foot-long upper-level canal. Factories were to draw water from this canal and discharge it into a parallel, second-level canal thirty feet below. Additional factories were then to use the thirty-foot drop to the third level canal, and a final thirty-foot drop to river level, thereby realizing the full power potential available in stages that existing water wheels could easily handle.

In 1859 Canfield died, and William H. Barnum (partner in the Barnum-Richardson firm of ironmakers and later a United States Senator) bought his interests. Barnum and Robbins are said to have been rival ironmasters who failed to cooperate in advancing the water power project. Although they fixed the leaks in the canals, no customers came despite the proprietors' intermittent sales efforts over the next forty years. The power company built its canals with massive, rubble-stone walls. The wall of the second level canal is a prominent feature on the east side of Water Street. A stone set in the wall carries the legend "Water Pr. Co. E.L. Goldsmith Supt. 1851."

J.H. Roraback and other investors formed the Connecticut Power Company in 1910 to develop the power left unused by the failure of the Water Power Company's project. It hired the Stone and Webster engineering firm of Boston to design and build a hydroelectric plant. Stone and Webster started work in April 1913 by building two workers' camps on the flood plain, the "white laborers' camp" to the south of the power house site and the "Italian laborers' camp" well separated to the north on the old Canfield & Robbins forge site. By the spring of 1914 the workers had completed a dam across the top of the falls to divert water into the Water Power Company's old upper level canal (now lined and capped with concrete) to a forebay where it falls ninety-nine feet in steel penstocks to three horizontal, double-runner, central discharge turbines in the power house. Each turbine is directly coupled to a three-megawatt alternator.

The system used to deal with ice in the canal is a remarkable feature of the plant. Broken ice is diverted to the end of the canal (past the forebay), and then down a wasteway that passes through a tunnel along the south side of the power house to the river below. The spillway at the headgate house provided for dewatering the forebay is used to inject water into the wasteway to flush the ice into the river below.

When the river flow is greater than the capacity of the powerhouse turbines, excess water spills over the dam and down the falls, which are otherwise dry. When flow is below that needed for continuous running, the power company has to balance its need to generate power at times of peak demand with requirements for maintaining minimum flow in the river downstream. Two turbines running at peak efficiency at mid-day provide the best conditions for canoeing below the plant.

(12:40 pm) Return to cars.

9.5 miles. Junction of Water Street and Warren Turnpike. Turn right before the railroad bridge. (For those wishing to buy lunch 45 minutes is allowed in the schedule. Go under the bridge and turn right into Falls Village.)

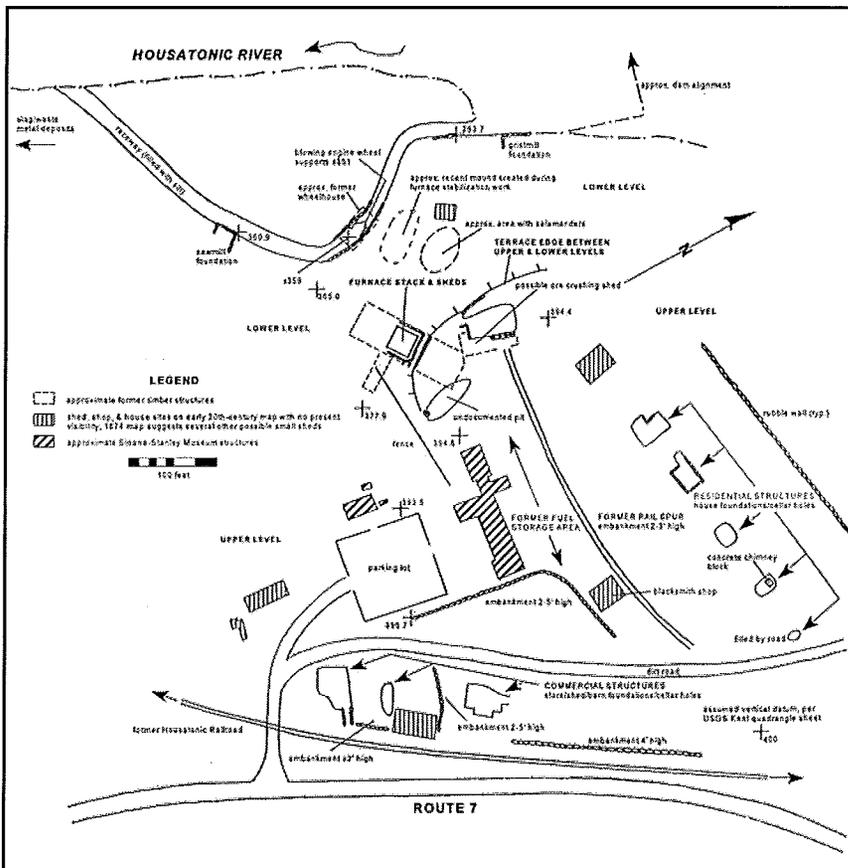
11.0 miles. Junction with Route 7. Enter the water gap through the Housatonic Highlands Massif. Follow Route 7 to Kent Furnace.

16 miles. Pass covered bridge on left. This bridge was constructed in 1864, and a secondary, queen-post truss for additional strength was added about 1887. A few years after 1939, when the bridge was posted for a five-ton limit, a twenty-ton oil truck broke through the deck. An ice jam that threatened to destroy the bridge had to be dynamited in 1961. Rather than deal with these problems the state wanted to build a high-level structure similar to the one at Cornwall Bridge downstream. Local citizens successfully resisted this scheme. In 1972 the Department of Transportation rebuilt the covered bridge with a concealed steel deck, which now carries the weight of traffic.

20 miles. The Housatonic River emerges from its water gap to enter its limestone valley at Cornwall Bridge.

24 miles. Right turn, cross railroad and park in museum lot.

**Stop 4. Kent Furnace and Museum Complex.** (2:30 PM, Figure 5) Kent blast furnace was built in 1825, a time of rapid expansion of ironmaking in Litchfield County. The first stack, twenty-eight feet high, ran on cold blast, and produced three to four tons of pig iron a day from South Kent ore and limestone flux obtained nearby. The owners enlarged the furnace in 1846, and added hot blast apparatus that allowed them to mix one part of anthracite (brought to the furnace by the Housatonic Railroad) into each three parts of their charcoal fuel.



**Figure 5. Features at Kent Furnace**

The enlarged furnace could make seven tons of iron a day. A breast wheel eighteen feet in diameter with twelve-foot long buckets provided blast air by driving two horizontal blowing tubs, each five

feet in diameter with a three-and-a-half foot stroke. Pratt's dam on the Housatonic directed water through a canal and flume to blast-engine wheel.

The owners rebuilt the furnace again in 1870. By then their customers included major manufacturers of machinery such as the Schenectady Locomotive Works in New York and the Farrell Foundry in Ansonia. The company had ten houses adjacent to the furnace bank for the men who worked at night or were needed for emergencies. John Roberts and George Bull kept a general store adjacent to the furnace. The Housatonic Railroad delivered ore from Salisbury by way of a siding to the furnace. Wagons brought ore from the South Kent mine, and hauled in the charcoal fuel made by colliers on the neighboring hills. The furnace company grew hay, oats, corn, and rye on adjacent fields, and ran its own gristmill at the dam to provide for feed all its draft animals.

When the 1892 business depression forced down the price of iron, the owners abandoned Kent furnace. Local farmers then used the charcoal sheds for drying tobacco. The furnace site was acquired by the Stanley Works for the Sloane-Stanley Museum of antique tools. The furnace remained an uncared-for ruin until the state undertook stabilization in 2002, and made the site with its surrounding land an archaeological preserve. When the restoration contractor began excavating debris that has accumulated in the furnace's casting arch in 2003, we found that the original crucible and bosh were still in place. In contrast to the Beckley reconstruction, the true interior structure of the furnace can be seen at Kent. Creation of a trail system that will lead visitors to the remains of the water power system used by the Kent furnace and its associated mills is underway.

The Sloane-Stanley Museum is adjacent to the furnace. The buildings and exhibits of the Connecticut Antique Machinery Association are north of the Sloane-Stanley Museum and the furnace grounds. The association is the sponsor of the Connecticut Museum of Mining and Mineral Science recently created by John Pawloski. John will be present to welcome visitors to the museum.

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**THE NEW QUATERNARY GEOLOGIC MAP OF CONNECTICUT AND LONG ISLAND SOUND BASIN**  
**Part 2-- Illustrated by a fieldtrip in the Connecticut River Valley**

by

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**INTRODUCTION**

The purpose of this fieldtrip is to demonstrate some of the glacial and postglacial deposits shown on the new Quaternary Geologic Map of Connecticut and Long Island Sound Basin authored by Janet Radway Stone, John P. Schafer, Elizabeth Haley London, Mary L. DiGiacomo-Cohen, Ralph S. Lewis, Woodrow B. Thompson, and Byron D. Stone. The map was printed in August (2005) and copies will be available at the NEIGC meeting. The discussion below is mostly excerpted from the Quaternary map text and includes some general information about map units across the State as well as a discussion of the chronology of ice retreat and meltwater deposition in the Connecticut River valley, which is the area we will visit on the fieldtrip. All references to "the map", sheet 1, sheet 2, description of map units, appendices, or map text in the following discussion are to the Quaternary Geologic Map (Stone and others, 2005).

The Quaternary geologic map portrays the geologic deposits and features formed in Connecticut during the Quaternary Period, which includes the Pleistocene (glacial) and Holocene (postglacial) Epochs. The Quaternary Period has been the time of development of many details of the landscape and of all the surficial deposits. At least twice in the Middle and Late Pleistocene, continental ice sheets swept across Connecticut and their effects are of pervasive importance to the present occupants of the land. The Quaternary map illustrates the geologic history and the distribution of depositional environments during the emplacement of glacial and postglacial surficial deposits and the landforms resulting from those events. A companion map, the Surficial Materials Map of Connecticut (Stone and others, 1992) emphasizes the surface and subsurface texture (grain-size distribution) of these materials. The features portrayed on the two maps are very closely related; each contributes to the interpretations of the other.

Connecticut is covered by 116 7.5-minute quadrangles, all available as U.S. Geological Survey 1:24,000-scale topographic maps with a 10-ft contour interval. Surficial geologic maps exist in various forms (either published, open-filed, or unpublished) for 98 of these quadrangles. We reviewed all 98 maps, and did reconnaissance mapping in the remaining quadrangles. An index map and a list of references to these quadrangle studies are included in Appendix 3 of the map. In the course of compiling this large body of data to create both the Surficial Materials Map and the Quaternary Geologic Map, we applied a consistent interpretive rationale; the result is that, in some cases, the original studies have been reworked or revised.

**UNITS ON THE QUATERNARY MAP**

The map units are divided into three main groups: **glacial ice-laid, glacial meltwater, and postglacial deposits**. The postglacial deposits, formed by various processes after the recession of the last ice sheet, constitute ten map units. The glacial ice-laid deposits, poorly sorted materials deposited more or less directly by the ice sheets, constitute three units. The 13 glacial ice-laid and postglacial units are Statewide in distribution, except for such limitations as are imposed by the geologic processes involved; for example, coastal beach and dune deposits are, of course, restricted to the coast. The glacial meltwater deposits, laid down by the great volumes of meltwater produced during the shrinkage of the last ice sheet, include 6 statewide depositional systems. The 6 systems of meltwater deposits are further differentiated into 204 units, closely restricted in geographic location and therefore also in age; these units have been given informal names based on their geographic localities. The rationale for this profusion of units is based on their scientific importance; they provide detailed information on the mode of disappearance of the last ice sheet and the depositional processes operating around its margin. A comprehensive understanding of these units has many practical applications in studies of socio-economic importance such as ground-water availability and coarse aggregate resources.

**Glacial Ice-Laid Deposits**

During the last (late Wisconsinan) glaciation (25-20 ka), a sector of the Laurentide ice sheet of northeastern North America spread across the St. Lawrence River valley and the Green and White Mountains of Vermont and New Hampshire, covered all of Connecticut, and reached its maximum extent on Long Island (Sirkin, 1982). Ice-movement

directions are indicated by striations and grooves on bedrock, drumlin axes, and, inferentially, by the positions of ice margins during retreat (fig. 1). Ice movement across the State was dominantly from north-northwest to south-southeast. The principal departure from that general trend was a prominent lobation in and adjacent to the Central Lowland. On the western side of that lobe, which probably became accentuated as the ice thinned during retreat, directions of movement were to the southwest or even to the west. Weaker lobate patterns occurred in the valleys of the Quinebaug, lower Connecticut, and Housatonic Rivers, and ice movement in westernmost Connecticut was influenced by the large lobe in the Hudson River valley. The glacial meltwater deposits so conspicuous on the map were all deposited during retreat of the late Wisconsinan ice sheet. Less is known of the earlier (probably Illinoian) glaciation recorded by the presence of a lower till. Drumlins are composed dominantly of this lower till, and their axial directions are probably partly inherited from the earlier glaciation. Glacial meltwater deposits of this earlier glaciation are rare; they evidently were eroded or buried during the late Wisconsinan glaciation. Still earlier continental glaciations, recorded by deposits in the mid-continent and confirmed by oxygen-isotope studies in ocean sediments and Greenland ice cores (Imbrie and others, 1984; Mix, 1987; Paterson and Hammer, 1987), probably also affected Connecticut, but no direct evidence has yet been found within the State.

Glacial ice-laid deposits shown on the map include till and end moraine deposits (figure 1). Two glacial tills, distinctive in character and different in age, are present in Connecticut. These tills are not shown as separate units on the map because the lower, older till (also called "drumlin till") (Stone, B.D., 1989; Melvin and others, 1992) occurs almost entirely in the subsurface; generally the lower till is at the surface today only in the floors of artificial excavations that are too small to show at this map scale. The lower till may be at the surface in the upper parts of some drumlins, but its occurrence is known only from local exposures and its extent cannot be predicted; most commonly, lower till in drumlins is mantled by thin, upper till. Numerous artificial exposures and subsurface well and test-boring data indicate that lower till constitutes the bulk of material within drumlins and other areas mapped as thick till (tt). Localities where lower till is (or has been) exposed are shown by an open diamond symbol on the map; such localities have now been identified in all parts of the State, in contrast to their limited known extent four decades ago (Schafer and Hartshorn, 1965; Pessl and Schafer, 1968; Pessl, 1971). The upper, younger till (also referred to as "surface till") is generally less than 4.6 m (15 ft) thick and constitutes most of the surficial material in areas mapped as thin till (t). Localities where the two tills are (or have been) exposed in superposition are shown on the map by a solid diamond symbol.

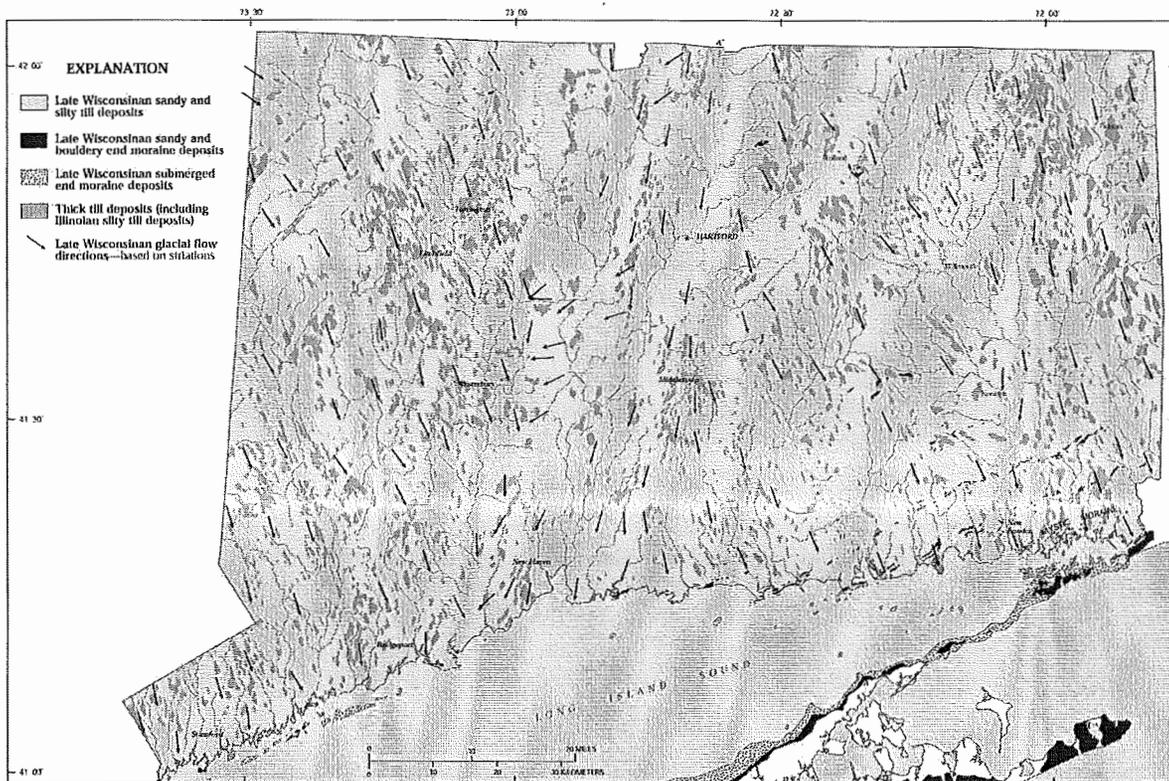


Figure 1. Distribution of till and end moraine deposits in Connecticut

Recessional end moraines were first mapped in southeastern Connecticut by Goldsmith (1960, 1962, 1964) who defined five belts of segmented moraines. They are the Clumps, Mystic, Rocky Hollow, Ledyard, and Oxoboxo moraines, most of which extend eastward into Rhode Island (Schafer, 1961; Goldsmith, 1982). Farther west, the Old Saybrook and Madison moraines were mapped by Flint (1971, 1975); however, further field investigation revealed that the distribution of some of those moraine segments was exaggerated. We reduced their areal extent on the map and mapped separately the Hammonasset moraine in Clinton. Mapping of additional moraine segments between the two areas during the State map compilation allowed linear correlation of the Old Saybrook with the Wolf Rocks moraine (own) in North Kingston, R.I., the Hammonasset with the Ledyard moraine (hlm), and the Madison with the Oxoboxo moraine (mom). Offshore mapping of moraine segments and of sub-bottom, continuous, linear ridges of proximal lacustrine fans (lcf), deposited at the grounding line of the ice margin in glacial Lake Connecticut, allowed linear correlation of the Old Saybrook moraine along the Lordship lacustrine fan ridge to the Norwalk Islands moraine in southwestern Connecticut.

The moraine belts are relatively linear, but show down-ice topographic deflection where they cross valleys. Accumulations of rock debris were concentrated in the shear zone where active ice rode up over thin stagnant ice at the margin. When the shear zone remained in one position for a significant time, concentrations of debris built up within and on top of stagnant ice (Goldsmith, 1982). This material was later deposited on the land surface by ablation processes. The linear trend of the moraine belts reflects the former position of the shear zone some relatively short distance behind the more ragged margin of stagnant ice. The segmented nature and local boulder-lag character of these moraines is probably due to the action of meltwater in the marginal zone. The moraine segments are most obvious in the upland areas between valleys. In valleys, morainic material may be buried by meltwater deposits; in most places, meltwater deposition dominated in the valley and the morainic position is represented by the ice-proximal head of a morphosequence. Locally, moraine segments, which are more lobate than in upland areas, stand at the proximal heads of meltwater deposits in the valley.

The coastal Connecticut moraines are parallel to a much larger moraine belt that includes the Charlestown and Fishers Island moraines in Rhode Island and New York, and the Harbor Hill moraine on the north shore of Long Island (Sirkin, 1982). Parts of the Harbor Hill-Fishers Island-Charlestown moraine are shown on the Connecticut map because the area is included on the topographic base and this moraine provided basin closure for the containment of glacial Lake Connecticut in the Long Island Sound area.

#### Glacial Meltwater Deposits

The shrinkage of the late Wisconsin ice sheet and the retreat of its margin from south to north across Connecticut were accomplished as the ice melted faster than it was re-supplied by movement from the north. The meltwater picked up rock debris carried by the ice and deposited most of it shortly beyond the ice margin. The deposits are sorted, stratified layers of gravel, sand, silt, and clay; these sediments accumulated in streams and lakes, large and small, that were fed by the meltwater. Because meltwater largely flowed in valleys during deglaciation, meltwater deposits are concentrated in those valleys and in many places are more than 30 m (100 ft) in thickness. A drift thickness map (see sheet 2 of the map) shows that the thickest glacial deposits in Connecticut are in the deepest parts of bedrock valleys and in the Long Island Sound basin; these thick deposits are largely meltwater sediments that accumulated in glacial lakes. Figure 2 shows selected ice-margin retreat positions across Connecticut. The approximate dates (given in  $^{14}\text{C}$  years) associated with four recessional ice-margin positions are estimated from a regional array of deglaciation dates from outside of the area as well as within it (Stone and Borns, 1986; Ridge and Larsen, 1990; Stone and Ashley, 1992). Radiocarbon dates in Connecticut that are relevant to the deglacial chronology are discussed and referred to in appendix 2 and their locations are shown on the map. Positions of major glacial lakes (IL and SL depositional systems) that dominated the deglacial history of the State are shown schematically. Deposits of these lakes, as well as of the multitude of smaller glacial lakes (IP and SP depositional systems) and glaciofluvial systems (FP and FD depositional systems) shown on the map, record a detailed history of ice retreat across Connecticut.

Most of the meltwater sediments in Connecticut were deposited in or graded to large and small glacial lakes. On R.F. Flint's 1930 map of the glacial geology of Connecticut, most of the meltwater deposits are mapped as "sand and gravel deposits in local temporary lakes (dammed by ice and controlled by spillways)" (Flint, 1930). Many of the different ideas about the retreat of the last ice sheet in Connecticut were mainly efforts to explain the abundance of deltaic deposits, especially in southerly sloping valleys (Gulliver, 1900; Flint, 1930, 1932, 1934; Lougee, 1938, 1953; Lougee and Vander Pyl, 1951; Black, 1977, 1982). The map reflects our concurrence with Flint's early observation of pervasive deltaic bedding in these deposits, though not with his regional stagnation model for the formation of glacial lakes. Of the 204 map units of correlated meltwater deposits, 183 consist of deposits that were laid down in or graded to glacial lakes. These units are grouped into four types of glaciolacustrine depositional systems (major ice-dammed lakes, major sediment-dammed lakes, related series of ice-dammed ponds, and related series of sediment-dammed ponds).

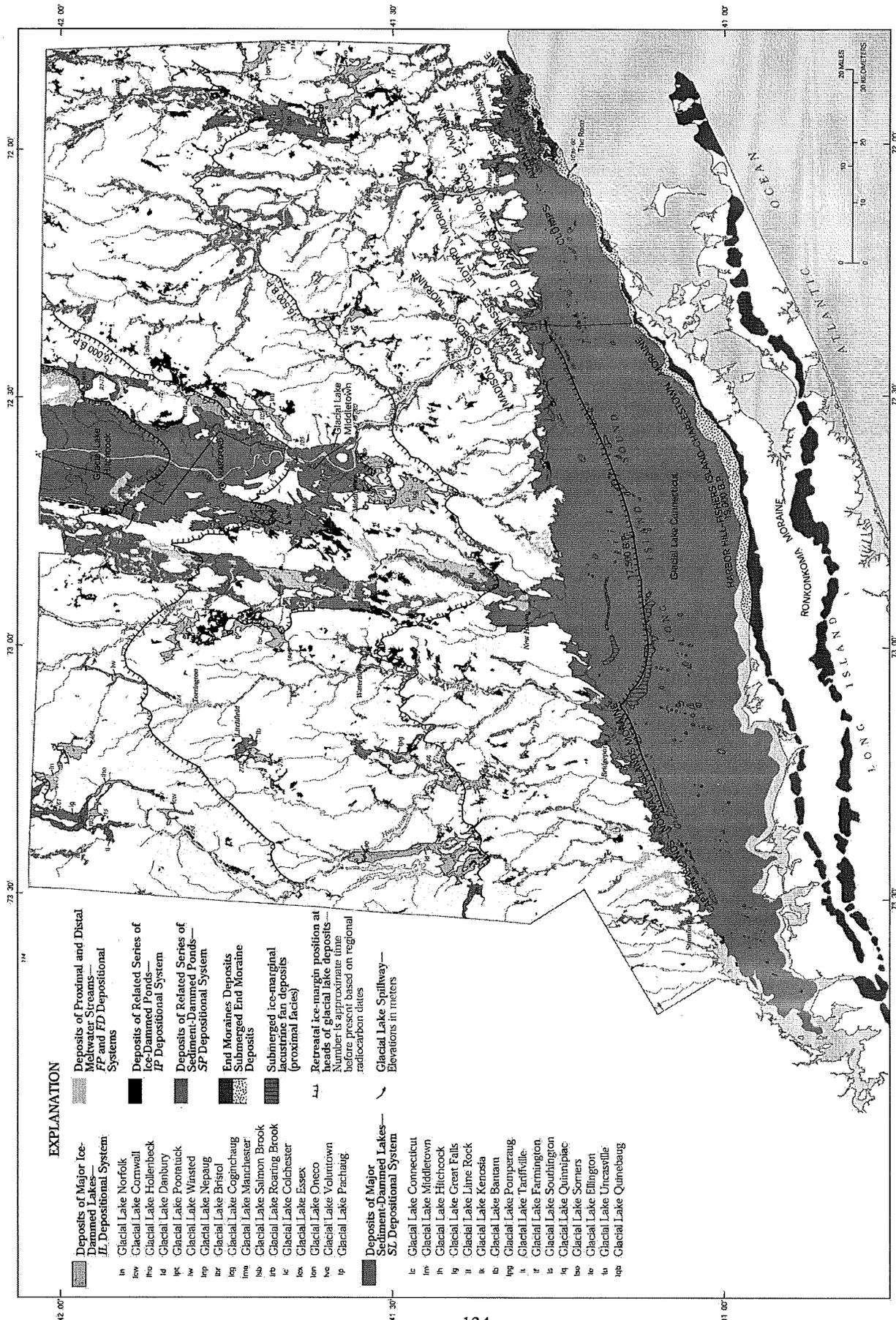


Figure 2. Late Wisconsinan meltwater deposits, end moraine deposits, and selected ice-margin retreatal positions. The distribution of the Ronkonkoma and Harbor Hill moraines on Long Island, NY is from Fuller, (1914).

We include within glacial-lake units not only lake-bottom sediments and deltaic deposits, but also the fluvial deposits laid down in tributary valleys by meltwater streams that fed deltas in the lake. This leaves only 21 of the 204 units as glaciofluvial units.

**Sedimentary Facies and Morphosequences**-- Appendix 1 of the Quaternary map text explains and illustrates the various sedimentary facies that are recognized within the meltwater deposits of the region. These facies are defined on the basis of lithic characteristics of texture and sedimentary structure and are related to specific environments of deposition along the path of meltwater flow: fluvial sediments were deposited in meltwater streams; deltaic sediments were deposited where meltwater streams entered glacial lakes; and lake-bottom sediments were deposited on the bottom of glacial lakes. Glacial sedimentary facies are combined either in facies assemblages or as single mappable bodies of sediment known as morphosequences (Koteff and Pessl, 1981). The types of morphosequences and the sedimentary facies included in them are described in Appendix 1. In general, a morphosequence is coarse grained at the glacier-proximal head and occurs in collapsed, ice-contact landforms; grain size decreases and landforms are less collapsed to noncollapsed in distal parts of the morphosequence. Morphosequences were deposited in close association with the ice margin; the surface altitude of each morphosequence was controlled by a specific base level, either a glacial lake plane or a valley knickpoint. Stratigraphic relationships between morphosequences in individual valleys provide ubiquitous evidence that these ice-marginal deposits are systematically younger from south to north. Morphosequences are the basic mappable units of meltwater deposits at 1:24,000 scale, but they are too small and too numerous to be shown as individual units at the scale of this map. Ice-margin positions at the heads of many morphosequences are shown on the map by a ticked solid line.

**Map Units**-- Each of the 204 correlated map units of meltwater deposits on this map is a group of morphosequences deposited along the same or related paths of meltwater flow. Each unit was deposited either in a single glacial lake, a related series of lakes, or along meltwater streams in a valley where no ponding occurred. The position of groups of morphosequences (map units) in the landscape further indicates the systematic northward retreat of the ice margin. Where drainage divides were parallel or oblique to the trend of the ice margin, groups of high-level deltaic sediments were deposited when paths of meltwater escape were first held to higher positions against or through uplands, and then gradually lowered as lower paths were uncovered in valleys. On the basis of stratigraphic relationships between deposits, successive retreating ice-margin positions, and changes in glacial lake levels and in paths of meltwater flow, morphosequences are grouped into map units that are chronostratigraphic in character and that define a relative chronology of ice retreat across the State (see figure 2).

**Depositional Systems**-- Six depositional systems of meltwater deposits have been identified in Connecticut as a result of regional synthesis. On the map, units are grouped by color as follows: Blues—Major ice-dammed lakes (IL), Greens—Major sediment-dammed lakes (SL), Purples—Related series of ice-dammed ponds (IP), Browns—Related series of sediment-dammed ponds (SP), Light oranges—Proximal meltwater streams (FP), Dark oranges—Distal meltwater streams (FD). Each depositional system is defined by lithostratigraphic principles and is characterized by morphosequence types, by spatial arrangements of sedimentary facies, and by typical stratigraphic relationships between individual deposits. The six depositional systems represent meltwater deposition in six paleogeographic settings that formed repeatedly in time and space, consequent to the interaction between the ice margin and the landscape over which it retreated. Ponding of meltwater occurred in nearly every valley in the State during deglaciation; as a result, most meltwater sediments were deposited in or graded to glacial lakes. Four of the six depositional systems formed in paleogeographic settings in which lakes controlled the distribution and altitude of fluvial, deltaic, and lake-bottom sediments. Lakes controlled deposition in north-draining valleys, which sloped toward the retreating ice margin. In these valleys, the ice margin impounded meltwater against opposing topography and spillways were located across the lowest points of drainage divides. The resulting water bodies are referred to as ice-dammed glacial lakes and ponds. Lakes also controlled deposition in most south-draining valleys, which sloped away from the ice margin. In these valleys, lakes were impounded behind thick, valley-filling bodies of sediment that were constructed during preceding and successive meltwater deposition in each valley. The resulting water bodies are referred to as sediment-dammed lakes and ponds. Each sediment dam was itself an ice-marginal meltwater deposit, graded to a slightly older lake in the valley. Spillways for succeeding lakes commonly were over these dams. Meltwater deposition in streams that were not tributary to glacial lakes was relatively uncommon in Connecticut. Glaciofluvial sediments were deposited in positions both proximal and distal to the ice margin.

**Glaciolacustrine Systems**-- Sediments of four glaciolacustrine systems (IL, SL, IP, and SP) were deposited in or graded to glacial lakes and ponds. Sediments deposited in lakes include delta foreset and bottomset beds, lake-bottom sediments, and local lacustrine fan sediments. Lake-bottom sediments in all map units are shown by a horizontal line pattern on the map. Sediments graded to glacial lakes include fluvial delta topset beds, delta-tributary fluvial sediments (fluvial

sediment graded to deltas), and local ice-channel sediments. Delta-tributary fluvial sediments in large glacial-lake map units are shown by a dot pattern on the map. Deposits of glaciolacustrine systems are predominantly deltaic. Altitudes of topset-foreset contacts in deltas record the paleo-water-plane altitudes of the glacial lake into which they were built. Deltas of all glacial lakes in Connecticut indicate paleo-water-plane slopes of 0.9 m/km (4.74 ft/mi) to the north-northwest. This slope is due to the glacio-isostatic tilt of the Earth's crust.

The two main types of large glacial lakes were major ice-dammed lakes (IL) and major sediment-dammed lakes (SL). The respective map units include all sediments graded to or deposited in single, relatively large, specifically named glacial lakes, some of which had several stages. These lakes existed in the wider valleys and large basins of the State. Deposits of major glacial lake systems are distinguished by several morphologic and stratigraphic characteristics: (1) deltas in each glacial lake (or lake stage) are at similar altitudes (when adjusted for glacio-isostatic tilt); (2) deltas have free fronts (that is, they prograded outward without being obstructed by earlier deposits and grade into flat-lying lake-bottom sediments); (3) lake-bottom deposits occur in front of deltas; and (4) delta-tributary fluvial deposits occur in side valleys (valleys that were tributary to the glacial lake).

The two main types of small glacial lakes were ice-dammed ponds (IP) and sediment-dammed ponds (SP). The respective map units include all sediments graded to or deposited in sequentially ponded and chronologically related series of small lakes (ponds). These small lakes existed in the narrower valleys and small upland basins of the State. Deposits of small glacial lake systems are distinguished by several morphologic and stratigraphic characteristics: (1) deltas in each map unit are at divergent altitudes; (2) deltas commonly do not have free fronts, but rather are contiguous with each other; (3) lake-bottom sediments occur only beneath the deltas, not at the surface; and (4) fluvial deposits (only in depositional system SP) occur in steeper sections of the main valley and sometimes overlie deltaic deposits.

Glaciofluvial Systems-- Sediments of two glaciofluvial systems (FP and FD) were deposited in meltwater streams that were not tributary to any glacial lake. Meltwater streams deposited ice-marginal and near-ice-marginal glaciofluvial sediments in the steeper sections of some south-draining valleys and in front of moraines; these are deposits of proximal meltwater streams (FP). Sediments of distal meltwater streams (FD) were deposited in other valleys after glacial lakes in those valleys had drained.

### Postglacial Deposits

Postglacial deposits in Connecticut include stream-terrace (st), talus (ta), dune (d), flood-plain alluvium (a), swamp (sw), salt-marsh (sm), beach (b), fluvial-estuarine channel-fill (ch), and marine delta (md, mdd) deposits; the onset of postglacial conditions was time-transgressive and began several thousand years earlier in the southern part of the State than in the northern parts.

In most of mainland Connecticut, postglacial activity consisted predominantly of incision of glacial deposits by meteoric streams along stream-terrace surfaces, followed by the establishment of flood plains at modern levels. Streams had eroded to modern flood-plain levels relatively early, in some cases before 12.0 ka (O'Leary, 1975; Stone and Randall, 1978). Postglacial winds were intense and widespread as indicated by the ubiquitous blanket of eolian sand and silt that overlies glacial sediments throughout the State and in which the modern soil is developed. The postglacial climate was severely cold for several thousand years following deglaciation. Paleobotanical studies reveal that treeless, tundra vegetation dominated by dwarf willow (*Salix herbacea*), sedges (*Cyperus*, *Carix*), and herbs and shrubs (*Dryas*, *Artemisia*), dated from earlier than 15 ka to about 13 ka, was present in the area (Davis and others, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987; Thorson and Webb, 1991). Also, wedge-shaped features with a polygonal ground pattern, interpreted as ice-wedge casts, deform eolian-sand-capped glacial sediments in numerous localities in Connecticut (Schafer and Hartshorn, 1965; Schafer, 1968; O'Leary, 1975; Stone and Ashley, 1992). These features indicate that permafrost existed locally in areas where substrate conditions were favorable to its formation. The presence of permafrost structures indicates that mean annual temperatures were below 0°C during the early postglacial time interval.

In the upper Connecticut basin, postglacial conditions were dominated by the continued existence of glacial Lake Hitchcock several thousand years after the ice margin retreated from the area. Extensive fields of eolian sand dunes formed in the treeless environment, indicating the continued effects of strong winds. Dunes are present on the relict deltaic and lake-bottom surfaces of glacial Lake Hitchcock. Dunes on deltaic and high-level lake-bottom surfaces were formed by north to north-northeasterly paleowinds; these surfaces were available as early as 15.5 ka. Dunes on stable-level lake-bottom surfaces were formed by northwesterly paleowinds; these surfaces became available at about 13.5 ka as glacial Lake Hitchcock drained. Evidence that severely cold temperatures persisted until the time of glacial Lake Hitchcock drainage exists due to the presence of hundreds of circular to subcircular, rimmed depressions (interpreted as

pingo scars) developed in the drained lakebed sediments (Stone and Ashley, 1989; Stone and others, 1991; Stone and Ashley, 1992). Paleobotanical records indicate a warming of the postglacial climate at about 12.5 ka, accompanied by reforestation of the landscape by successive spruce, pine, and hardwood forests from 12.5 to 9 ka (Davis, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987).

#### CHRONOLOGY OF ICE RETREAT AND GLACIAL LAKES IN THE CONNECTICUT RIVER VALLEY

At Middletown, the Connecticut River leaves its route through the broad Hartford Basin underlain by Mesozoic-age sedimentary rocks and enters a much narrower bedrock-walled valley through resistant Paleozoic- and Proterozoic-age metamorphic rocks of at least three different tectonic terranes-- Bronson Hill, Merrimack, and Avalon (Rodgers, 1985). From Middletown to Old Saybrook (see figures 3 and 4), the river occupies a series of north- and northwest-trending segments of a fault/fracture zone that cuts across the stratigraphic trend of the metamorphic rock units. The relatively narrow bedrock channel averages about a mile in width at its highest altitudes; the channel thalweg lies beneath the present river south of Portland except at the south end of the valley; here it lies to the west of the present river beneath Old Saybrook. Thalweg depths of the bedrock valley range from -46 m (-150 ft) altitude in Portland to -76 m (-250 ft) beneath Saybrook; the channel has been mapped offshore at depths greater than 107 m (350 ft) below sea level. Thick glacial deposits overlie bedrock through the entire length of the valley, although near vertical rock faces are exposed locally along the sides of the river. Seismic-reflection surveys reveal that glacial till fills the deepest parts of the bedrock valley in many places. Glacial meltwater deposits (stratified drift) fill the valley to maximum altitudes of about 8 m (25 ft) above sea level at Old Saybrook and Old Lyme; meltwater-deposit surfaces increase in altitude northward along the valley sides to about 15 m (50 ft) at Chester, Deep River and Hadlyme and 41 - 47 m (135 - 155 ft) at Portland. Before entrenchment by the Connecticut River, these meltwater deposits filled the valley from side to side in most places.

Deposits of Glacial Lake Connecticut along the lower reaches of the Connecticut River (lcwo) are successive ice-marginal deltaic and fluviodeltaic deposits built into the lake; this extensive glacial lake occupied the Long Island Sound Basin during the time of ice retreat from the Harbor Hill recessional moraine on Long Island (see discussion of glacial Lake Connecticut in trip A3 article). Deltas were constructed at the mouth of the Connecticut River and in its tributary valleys of the Black Hall and Lieutenant Rivers on the east side. This unit consists of deltas built at three major ice-margin positions. At the mouth of the river in Old Saybrook, an ice-marginal delta with surface altitudes of 5-8 m (15-25 ft) was built into Lake Connecticut in front of the Old Saybrook moraine. An ice-marginal delta north of South Cove in Old Saybrook and fluviodeltaic deposits in the Black Hall valley in Old Lyme at 5-8 m (15-25 ft) altitude were built from a second, non-morainal ice position. An ice-marginal delta was built in front of a segment of the Hammonasset moraine just north of the I-95 Bridge in Old Saybrook (Stop 1) and fluviodeltaic deposits were built in front of another segment of the moraine in the Lieutenant River valley in Old Lyme. These deltas have surface altitudes between 5 and 11 m (15 and 35 ft). Paleo-water-level indicators (contacts between flat-lying fluvial topset beds and dipping subaqueous foreset beds) exposed within these deltas indicate that glacial Lake Connecticut in Long Island Sound stood at a level of 3 m (10 ft) above present sea level at the time of ice retreat in the lower Connecticut River valley. Glacial Lake Connecticut slowly lowered through time as the spillway notch across the Harbor Hill moraine just west of Fishers Island (The Race) was eroded (see discussion in trip A3).

Moraine construction and deltaic deposition into glacial Lake Connecticut completely filled the lower part of the valley; further northerly ice-margin retreat from the vicinity of the I-95 bridge, resulted in meltwater deposition in smaller, successive glacial lakes that were separated from glacial Lake Connecticut. Each small lake was dammed behind deposits of the lake to the south; each lake was largely filled in with deltaic and lacustrine sediments. Deposits of this series of lakes are identified as "lower Connecticut River sediment-dammed pond deposits" on the map (units lce, lcc and lct). Deposits in the Essex-Hamburg-Deep River area (lce) are predominantly ice-marginal deltaic deposits of which only remnants remain due to downcutting by the postglacial Connecticut River. Delta surfaces are at 11 m (35 ft) altitude in the southern part of unit and 14-17 m (45-55 ft) in the northern part. Lower Connecticut River deposits in the Chester-Hadlyme area (lcc) are near-ice-marginal fluviodeltaic deposits in the Pattaconk Brook-Chester Creek valley in Chester and in the Roaring Brook-Whalebone Creek Valley in Hadlyme, and along the Connecticut River. Fluvial deposits in Chester and in Roaring Brook valley, Hadlyme grade eastward and westward respectively to deltaic surfaces on either side of the Connecticut River valley at 17-20 m (55-65 ft) in altitude.

Lower Connecticut River deposits from Tylerville to Portland (lct) (Stop 2) are predominantly ice-marginal deltaic deposits; probably more successive deltaic morphosequences exist than are indicated by ice-margin position symbols on the map. From Tylerville to north of Haddam-Middletown town line, deltaic surfaces rise from 23 to 34 m (75 ft to 110 ft) - a gradient of about 1.1 m/km (6 ft/mi), most of which can be accounted for by postglacial uplift; fluvial topset beds are generally thin, less than 3 m (10 ft) thick in this section. Deltaic surfaces at and just south of Maromas are 41-44 m (135-145 ft) and rise to 53 m (175 ft) at the northernmost ice-marginal head; fluvial topset gradients on these deposits are steeper, about 1.9 m/km (10 ft/mi), and topset beds are as much as 7.5 m (25 ft) thick. Deposition of greater thicknesses



Figure 3. A section of the Quaternary Geologic Map showing locations of Fieldtrip Stops 5-8.



of topset beds in the northern ice-marginal delta series of this unit is perhaps a result of slower retreat in this area, where the regional margin of the ice sheet was mainly retreating down opposing NW-facing slopes. The northernmost deposits in the unit, at Jobs Pond north of the Connecticut River, are as much as 75 m (250 ft) thick and block a former channel of the Connecticut River; because these deposits were not trenched by later erosion, the top of the next-to-the-last delta preserves a spillway channel at 47 m (155 ft), the same altitude as the topset/foreset contact in the last delta. The lower Connecticut River sediment-dammed pond deposits completely filled the valley from side to side; these series of deposits subsequently formed the lengthy sediment dam for glacial Lake Middletown.

Several ice-dammed lakes (IL) and series of small ice-dammed ponds (IP) glacial Lakes Essex (lex) and Colchester (lc), formed in northerly-draining tributary valleys to the lower Connecticut River. Stops 3 and 4 will illustrate one of these lake units (whd).

Deglaciation of the upper Connecticut valley was dominated by sedimentation in glacial Lakes Hitchcock and Middletown. Several ice-dammed lakes in north-draining tributary valleys of the Connecticut River valley preceded these two lakes. Glacial Lake Coginchaug was impounded in the north-draining Coginchaug River valley. The early stage of the lake (lcmd) spilled across the main drainage divide to the south; the later stage of the lake (lcmg) spilled eastward into the Connecticut valley. Deposition of the Hanging Hills unit (hh of the IP depositional system) also preceded formation of glacial Lake Middletown in Meriden in small northeast-sloping valleys which drain the dip slope of the Hanging Hills (traprock ridges) and are tributary to the Mattabeset River.

On the east side of the upper Connecticut valley, north-draining valleys contained a series of ice-dammed ponds (cd) which spilled across the upper Connecticut drainage divide. They were followed by a series of long, narrow, ice-dammed glacial lakes that formed in valleys oblique to the trend of the ice margin: glacial Lakes Roaring Brook (lrb), Salmon Brook (lsb), and Manchester (lma). These ice-dammed depositional systems formed in north-sloping tributary valleys to the Upper Connecticut basin and preceded the development of major sediment-dammed lakes in the main basin.

#### **Glacial Lake Middletown (SL)**

Glacial Lake Middletown first developed along the Connecticut River and in the Mattabeset River basin. The lake was impounded by a long mass of earlier deposits (lcc, lce, lct) in the lower Connecticut River valley at and south of The Straits; the spillway, with an initial altitude of about 40 m (130 ft) was over these deposits. Successive ice-marginal deltaic deposits were built into the lake as the ice retreated northward. When adjusted for the regionally established postglacial tilt of 0.9 m/km (4.74 ft/mi) to the N 21° W, delta topset-foreset contacts indicate that the lake slowly lowered due to erosion of its sediment dam. Glacial Lake Middletown occupied the Middletown basin in the lower Mattabeset valley and extended into the Berlin basin in the upper Mattabeset valley, as indicated by accordant delta levels, by basin geometry resulting in ice-margin positions that trend northwest-southeast, and by the extent of clays in the Berlin area. Deltas in Cromwell (lmc), Newington (lmn), and New Britain (lmw) were built contemporaneously and record lake levels at the spillway of about 34 to 35 m (110 to 115 ft).

Just north of the Cromwell deltas, deltas of the Dividend Brook deposits (db) were laid down in waters that were temporarily ponded to a higher level than glacial Lake Middletown and were controlled by the Dividend Brook spillway over Cromwell deltaic deposits (lmc); this spillway was not eroded lower than its present level of 39 m (129 ft) because of the presence of glacial Lake Middletown at its mouth.

When the ice uncovered the lower part of the divide between the Hartford basin and the Middletown-Berlin-New Britain basin, where the New Britain spillway of glacial Lake Hitchcock would later exist, glacial Lake Middletown persisted at a level high enough to spread across the divide into the Hartford basin. When the ice retreated from the north end of Cedar Mountain (Newington-Hartford town line), the Dividend Brook spillway was abandoned and glacial Lake Middletown spread eastward into the southern end of the basin later occupied by glacial Lake Hitchcock. Deltaic deposits (lmw, lme, lmg, and lmwv) as well as lake-bottom deposits (lmb) in the Hartford basin all occur at altitudes accordant with glacial Lake Middletown, but too high to have been controlled by any possible early level of the New Britain spillway. Not until glacial Lake Middletown had lowered to below 34 to 35 m (110 to 115 ft) at the divide (about 20 m (65 ft) at The Straits spillway) could the New Britain spillway come into use as the outlet for glacial Lake Hitchcock.

#### **Glacial Lake Hitchcock (SL)**

Glacial Lake Hitchcock existed in the upper Connecticut River basin in Connecticut, Massachusetts, Vermont, and New Hampshire, lengthening to at least 298 km (185 mi) as the ice retreated northward to the vicinity of Burke, Vt. The Connecticut River valley was dammed to an altitude of 46 to 49 m (150 to 160 ft) in the vicinity of Rocky Hill and Glastonbury by deposits of glacial Lake Middletown (lmc and db); this mass of stratified drift is often referred to as "the Rocky Hill dam." The spillway for glacial Lake Hitchcock was not over the dam, however, but at the lowest place across the Mattabeset River drainage divide between the Hartford basin and the Middletown-Berlin basin in New Britain. When the ice margin first retreated into the Hartford basin, north of that divide, glacial Lake Middletown water covered the later New Britain spillway location and early ice-marginal deltas in the Hartford basin were controlled by glacial

Lake Middletown. Not until glacial Lake Middletown had dropped to below 35 m (115 ft) could the New Britain spillway area emerge and glacial Lake Hitchcock exist as a separate water body; this occurred at about the time that the ice margin was at Windsor and East Windsor.

During the early life of glacial Lake Hitchcock, the New Britain spillway was eroded into till and older stratified drift so that water levels at the spillway dropped from about 35 m (115 ft) down to 25 m (82 ft) in altitude (Langer, 1977; Langer and London, 1979). In Connecticut, all ice-marginal and distal-meltwater-fed deltas, as well as one small delta built by meteoric water, record lake levels higher than the longer lived stable level. These deltas show a gradual lowering of the lake level as the ice retreated northward and the New Britain spillway was incised down to bedrock. Ice-marginal deltas in Windsor (lhhw) and East Windsor (lhhe) record 34- to 35-m (110- to 115-ft) levels at the spillway. To the north, ice-marginal deltas in Suffield (lhhr) and Enfield (lhhs) indicate 32- to 33-m (105- to 110-ft) levels at the spillway; still farther north in Suffield and Enfield, the Shea Corner (lhpsc) and Enfield (lhpm) deltaic deposits record levels just below 30 m (100 ft) at the New Britain spillway. This early phase of glacial Lake Hitchcock is recorded by ice-marginal deltas that are found well into southern Massachusetts and that were built to lake levels between 26 and 29 m (85 and 95 ft) at the spillway. This higher-than-stable-level phase of the lake is referred to as the "Connecticut Phase" (Koteff and others, 1988). It is important to note that deepening of the spillway channel was controlled by conditions 48 to 64 km (30 to 40 mi) to the south (the New Britain spillway was an independent control for lake levels). The base level for waters exiting the spillway was controlled by downcutting in the lower Connecticut River valley and by lowering levels of glacial Lake Connecticut in the Long Island Sound basin. The Rocky Hill dam area was glacio-isostatically depressed about 44 m (145 ft) and the New Britain spillway area was depressed about 50 m (165 ft) (more than the area at the mouth of the Connecticut River). In order for the New Britain spillway to lower by 10 m (33 ft) during the early phase of the lake, glacial Lake Connecticut had to have already lowered to below -25 m (-82 ft) in altitude.

Delta levels in Massachusetts indicate that a stable lake level, 25 m (82 ft) in altitude, had been reached by the time the ice margin had retreated to just north of the Chicopee River valley; regional correlation of <sup>14</sup>C dates (Stone and Borns, 1986) place the ice front in this position at about 15 ka. The 25-m (82-ft) level indicates that the water flowing through the spillway was about 7 m (24 ft) deep because its bedrock floor today is at about 18 m (58 ft) in altitude. Altitudes of topset-foreset contacts of ice-marginal deltas, from southern Massachusetts to the lake's northernmost extent, project to the stable level (25 m (82 ft) at the New Britain spillway) on a straight line, which is tilted up to the north-northwest at a slope of 0.9 m/km (4.74 ft/mi). The linearity of these projected delta altitudes indicates that the lake level was stable during the time of ice retreat from Chicopee, Mass., to Lyme, N.H., and that postglacial rebound of the land surface did not begin until after all ice-marginal deltas had been built, probably between 14 and 13.5 ka (Koteff and Larsen, 1989). Deltas that were not associated with the ice margin, but rather were built by meteoric water in most river valleys that entered the lake, also project to the stable lake level. In Connecticut, these include unit lhsh associated with the Hockanum River, unit lhss associated with the Scantic River, and unit lhsc, where the Farmington River constructed a large delta northeastward into the lake in the area now surrounding Bradley International Airport. The Bradley International Airport delta covers about 52 km<sup>2</sup> (20 mi<sup>2</sup>) and its entire surface (which is tilted up to the N 21°W in the amount of 0.9 m/km (4.74 ft/mi)) is graded to the stable 25-m (82-ft) level; these two facts provide evidence for the long duration of the stable level and also indicate that the lake was not affected by glacio-isostatic tilting until after nearly all of its deltas had been constructed.

It is important also to note that the New Britain spillway could not have lowered further than the 25-m (82-ft) level. This is because the 25-m (82-ft) altitude at the New Britain spillway is equivalent to a -25-m (-82-ft) altitude at the mouth of the Connecticut River when the 50 m (164 ft) of differential depression between the two localities is taken into account. The base of the channel, through which the paleo-Connecticut River carried water that spilled from glacial Lake Hitchcock, was imposed on bedrock at -27 m (-89 ft) in altitude at the mouth of the present Connecticut River east of Saybrook Point; this point was the actual control for the "Stable Phase" (Koteff and others, 1988) of glacial Lake Hitchcock. The "Stable Phase" of glacial Lake Hitchcock lasted from about 15 ka until about 13.7 ka; during this time, the southern part of the basin (south of the Holyoke Range in Massachusetts) was largely filled with deltaic and lake-bottom sediments. Preserved lake-bottom surfaces in Connecticut are at about 14 m (45 ft) in altitude in the south and 44 m (145 ft) in the north; the tilted stable-level paleo-waterplane over this area is at 19 m (63 ft) in altitude at the north edge of the Rocky Hill dam and 52 m (172 ft) at the Massachusetts border; thus, toward the end of the "Stable Phase" before the dam was breached, water depths in the lake were only 6 to 8 m (20 to 25 ft). Because the bedrock basin that contained the lake north of the Holyoke Range in Massachusetts is deeper, the lake was not filled with sediment to the extent that it was in the southern basin. North of the Holyoke Range in Massachusetts, preserved lake-bottom surfaces are at 46 m (150 ft) in altitude, and at the end of the "Stable Phase," water depth was about 46 m (150 ft).

Fluviodeltaic deposits (ft and lhf) built southeastward into the lake by the Farmington River record a "Post-stable Phase" (Koteff and others, 1988) of the lake during which levels were lower than the 25-m (82-ft) level at the New Britain spillway. A topset-foreset contact in the lhf deltaic deposits north of the Farmington River is at 39 m (127 ft); delta-surface altitudes in the same unit to the south of the river indicate slightly lower water levels. These levels project southward below the New Britain spillway level to 15 to 18 m (50 to 60 ft) in altitude at the Rocky Hill dam and record lowering of lake levels as the dam was entrenched. A preserved 17-m (55-ft) terrace inset into the Rocky Hill dam sediments on both sides of the present Connecticut River in Rocky Hill and Glastonbury records this "Post-Stable Phase" which was relatively brief in Connecticut. A  $^{14}\text{C}$  date of  $13,540 \pm 90$  B.P. (Beta-59094, CAMS-4875) on plant debris in lacustrine sands at the top of the lake-bottom section (radiocarbon-dated locality 10, Appendix 2) associated with the Farmington River deltaic deposits (lhf) establishes that the time of dam breach was about 13.5 ka.

The dam most likely was breached by headward erosion of streams on its south side, possibly by ground-water sapping and possibly aided by earthquakes generated by the initiation of postglacial rebound. Regardless of the mechanism by which the dam was breached, glacial Lake Hitchcock could not lower below stable level, much less drain, until its bed was raised by glacio-isostatic tilting. Dam breaching and initiation of isostatic rebound was required in order to establish the lower water-level altitudes recorded in the "Post-Stable Phase" Farmington River deltaic deposits (lhf). Once this process began, it proceeded rapidly as the dam was incised from just above 18 m (60 ft) in altitude (the stable level at the dam) to just above 12 m (40 ft); once this 6 m (20 ft) of lowering was accomplished, glacial Lake Hitchcock, south of the Holyoke Range, was entirely drained and the newly formed Connecticut River began to incise the lake floor (along the terraces of unit st) over the 80-km (50-mi) stretch between the Holyoke Range and the breached dam. Glacial Lake Hitchcock continued to exist north of the Holyoke Range with initial water depths of about 40 m (130 ft) (lowered from stable level by only 6 m (20 ft)); continued lowering of the lake was controlled by the rate of rebound, which made it possible for the lake bed south of the Holyoke Range to be incised.

An approximate 4,000-year life span for glacial Lake Hitchcock was indicated by Antevs (1922) through a method of correlating varves in clay pits from Hartford, Conn., to the north end of the lake basin in St. Johnsbury, Vt. This method assumes that the silt-clay varve couplets are annual summer and winter layers and that regional seasonal fluctuations affected the thickness of individual varves over the entire lake basin. Varved silts and clays of glacial Lake Hitchcock were used to construct Antevs' (1922) New England varve chronology between varve-year 3,001 and varve-year 7,000. Recently, Ridge and Larsen (1990) fit a 533-year varve section from Canoe Brook in southern Vermont into the relative varve chronology of Antevs (1922); they also placed the chronology in an absolute time frame with a 12.4 ka  $^{14}\text{C}$  date on plant debris in the Canoe Brook section at the position of varve 463 (varve NE 6150 in the Antevs chronology). Using this calibration of the varve chronology, lacustrine deposition at the south end of glacial Lake Hitchcock (varve NE 3001) began at about 15.5 ka. The early Connecticut phase was followed by the longer "Stable Phase" of the lake which lasted until about 13.5 ka (~varve NE 4800). The "Post-Stable Phase" of the lake, which lasted only briefly in Connecticut, continued for at least another 2,000 years north of the Holyoke Range until about 11.5 ka (varve NE-7500).

More recent work in varve correlation (Ridge, 2003, 2004) has further refined the calibration of Antevs' varve chronology. In the summer of 2004 Jack Ridge obtained a section of overlapping cores from the former Kelsey Ferguson Brickyard pit in South Windsor, CT for detailed study of the varve stratigraphy. A compiled varve sequence from the cores totaling 552 years was matched to Antevs' (1922) New England varve chronology (NE 3617-4168) without a single extra or missing varve in the section (fig. 5). According to the most recent calibration of the NE varve chronology (Ridge, 2003, 2004) NE 3617-4168 represents 14.65-14.15  $^{14}\text{C}$  ka (17.5-16.95 cal ka). Other varve measurements and well records in the area indicate that there are many meters of varves representing many hundreds of years beneath the measured KF pit section. Correlation with Antevs' chronology indicates that the 552 varves measured at the KF plant pit were deposited at least 600 years after deglaciation of the area and after the front of the last glacier had receded into Massachusetts.

Both the timing of the initial drainage event and the continuity of Lake Hitchcock as ice receded north of Massachusetts have been the center of controversy. Based on varve stratigraphy and  $^{14}\text{C}$  ages it appears that the impoundment of water north of the Holyoke Range in Massachusetts and into New Hampshire and Vermont was maintained after the initial failure of the Rocky Hill dam. Water from the north drained across exposed lake floor deposits in the southern basin of Lake Hitchcock that served as a spillway. The cross-bedded sands that unconformably overlie the varves at the KF Plant represent this fluvial drainage system. With the initial failure of the Rocky Hill dam varves at the KF Plant section were incised by fluvial activity as is indicated by the unconformity. However, erosion was very minor because the surface elevation of varves in the KF Plant is only about 5 m (15 ft) lower than the preserved lake floor surface 2 km to the east. Therefore the varve at the top of the KF Plant section (NE 4168) is no more than 3-6 centuries older than the drainage event. The top of the southern basin varve stratigraphy, exposed beneath areas of preserved lake floor on the west side of

the Connecticut Valley at the Matianuck Ave. site (Stone and Ashley, 1992; Stone and others, 2005) grades upward from thin clayey varves to thicker and sandier varves that represent very little time. The  $^{14}\text{C}$  date  $13,540 \pm 90$  B.P. (Beta-59094, CAMS-4875) on detrital plant debris from sand at the preserved surface of lake deposits records a time soon after the initial failure of the Rocky Hill dam. At present, the time of varves NE 4500-4800 seems to be a reasonable estimate for the time of initial failure of the Rocky Hill dam. The receding ice front would have been north of the Holyoke Range in the vicinity of Greenfield, Massachusetts at that time (Ridge, 2004).

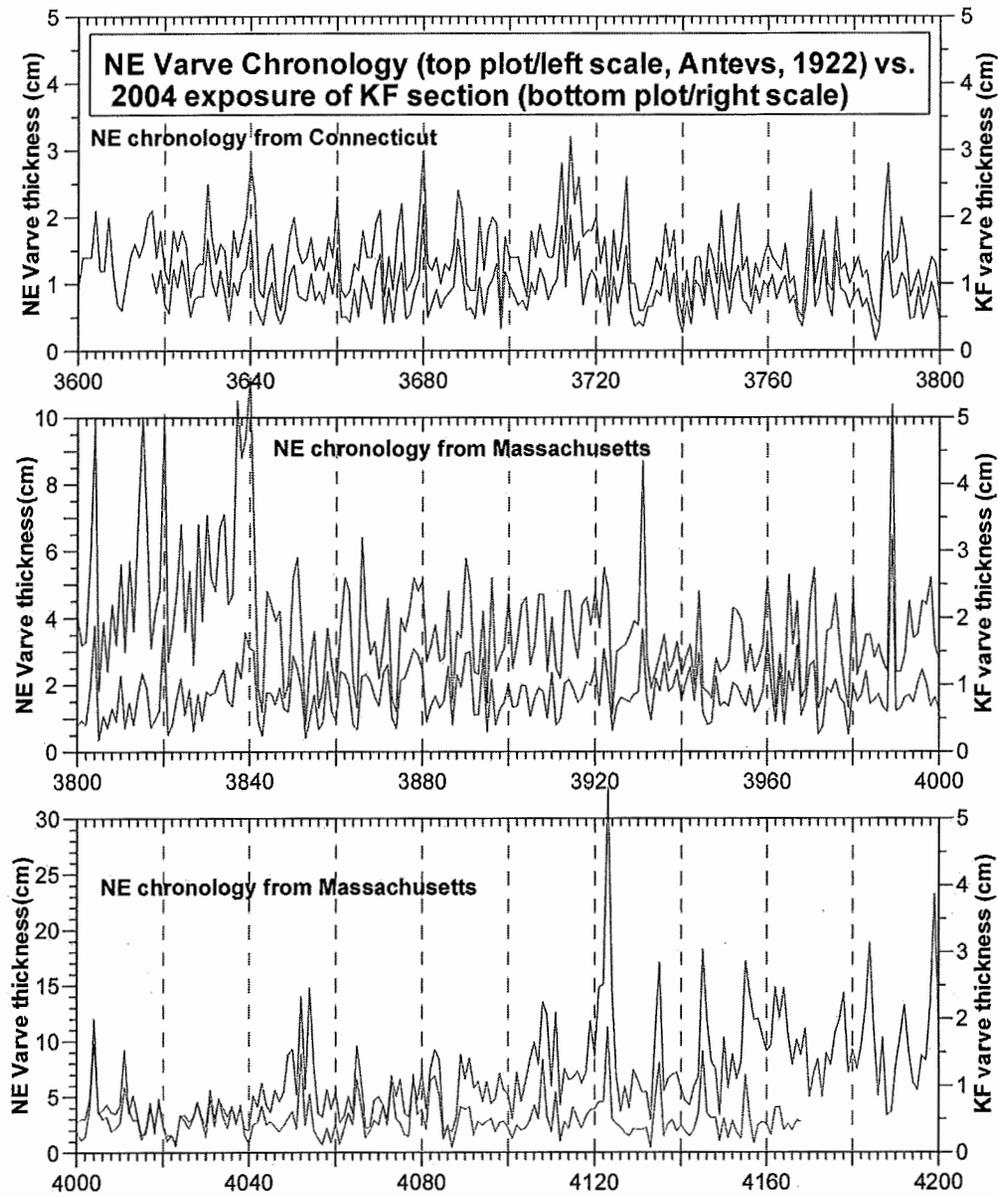


Figure 5. Match of Kelsey-Ferguson clay pit varve section with New England varve chronology (Antevs, 1922). Thickness axes for NE varves are to left and scale varies between plots. KF varve thickness scale is to right.

### ROAD LOG

The fieldtrip assembles at STOP 1 in Old Saybrook, CT at 8:00 AM. The Road Log is listed in cumulative miles beginning at Stop 1. Directions from Yale University to Stop 1 are also included. Locations of field trip stops are shown on figures 3 and 4. Future users of this road log are reminded that STOPS 2, 4, 5, 7, and 8 are on private property; it is the responsibility of visitors to request permission from the land owners to visit these locales.

Mileage and directions to Stop 1 from Yale University:

- 0.0 From Kline Geology Lab Parking Lot, turn right (south) onto Whitney Ave.
- 0.2 Turn left at Trumbull St.
- 0.4 Bear left and merge onto I-91 South, via the ramp toward I-95 New London / N.Y. City.
- 2.6 Merge onto I-95 N via exit on the LEFT toward New London.
- 29.9 Take exit 67 for CT-154 S (Old Saybrook).
- 30.0 Bear right at Middlesex Tpke. (CT-154).
- 30.1 Turn left at US-1 North (Boston Post Rd).
- 31.2 Bear right, road becomes Ferry Rd.
- 31.9 Turn right into Baldwin Bridge State Boat Launch (located under west end of I-95 bridge).

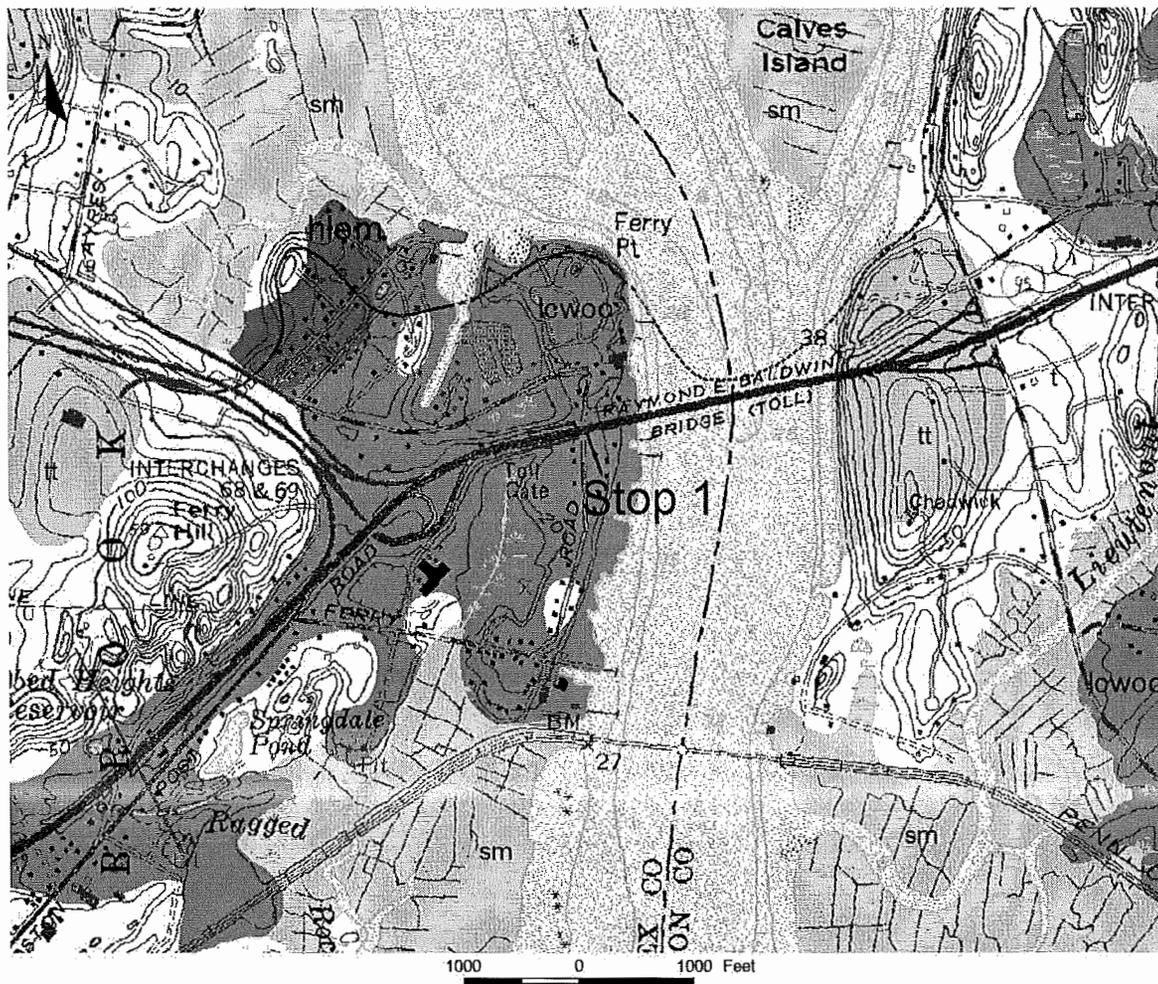


Figure 6. Topography and extent of Glacial Lake Connecticut deposits (lcwoo) and a segment of the Hammonasset-Ledyard moraine (hlem) in the vicinity of Stop 1. Heavy lines are ice margin positions.

**STOP 1. Baldwin Bridge State Boat Launch and Fishing Pier parking lot, below west end of I-95 bridge across Connecticut River, Old Saybrook (Old Lyme quadrangle).** From this vantage point, we can view the surface of ice-marginal deltaic deposits (map unit lcwoo) at altitudes of 25-35 ft in this vicinity. This deltaic deposit was built at the third in a series of successive ice-marginal positions near the mouth of the Connecticut River (fig. 4) from which deltas were built into glacial Lake Connecticut occupying the Long Island Sound Basin (see discussion for fieldtrip A3). This deposit and contemporaneous fluviodeltaic deposits on the east side of the valley in Old Lyme built in front of segments of the Hammonasset-Ledyard moraine. Together with deltas built at two earlier ice-margin positions with surface altitudes at 15-25 ft, these deposits completely blocked the lower reaches of the valley and provided a sediment dam for a series of ponding events in the lower Connecticut River valley to the north of here. At STOP 1, the flat delta-plain surface at 25-35 ft extends southerly from an ice-contact slope along its northern edge. Pebble-cobble gravel in glaciofluvial topset beds is exposed in new road and house excavations nearby. Former excavations for bridge piers across Ferry Road from the State Boat Launch area revealed a topset-foreset contact at 10-15 ft altitude. North of the delta plain, an irregular and hummocky ridge rises to 40 ft and trends northeasterly. The highest point in the ridge is a bedrock knoll; otherwise its surface is punctuated only by large glacial boulders. A former 5-m deep excavation in the ridge at the Island Cove Marina exposed sandy and bouldery till. We have mapped the ridge as a segment of moraine (map unit hlem) on the basis of that exposure, the less well-defined but similar morphology to the moraine ridge at the head of delta deposits at Cornfield Point and Fenwick farther south, and the position of this ridge along the trend of the Hammonasset-Ledyard moraine. We will drive over the delta plain, its ice marginal slope, and the moraine ridge on the way to Stop 2. Note exposures of cobble gravel along the route.

- 00.0 Exit State Boat Launch Parking area and turn right (north) on Ferry Road.
- 00.1 Turn left at stop sign onto Essex Road. Travel over the 25-ft altitude surface of unit lcwoo.
- 00.5 Turn right on Fourth Ave. Note cobble gravel exposed in bank on right. Delta plain is at 35-ft altitude here.
- 00.6 Bear left down ice-marginal slope.
- 00.7 At stop sign, continue straight on Fourth Ave. Former excavation at marina on right revealed bouldery ablation till in morainal deposits.
- 00.8 Turn left on Sunset Ave. Ascend ice-contact slope onto distal slope of segment of Hammonasset-Ledyard moraine.
- 01.1 Turn right on Essex Street.
- 01.2 Note extensive salt marshes along Connecticut River on right.
- 01.8 Turn left at sign for Rt. 9.
- 02.0 Proceed straight onto Rt. 9 entrance ramp at intersection with Rte. 154.
- 04.3 Just beyond exit 3, travel over 35-ft surface of deposits of glacial Lake Essex (IL depositional system).
- 09.2 Take exit 6 for Rt. 148, Chester.
- 09.4 Turn right at end of ramp onto Rt. 148.
- 10.1 Pass onto surface unit Ice deposits.
- 10.6 At stop sign Main St. Chester, continue straight on Rt. 148.
- 10.9 Note freshwater tidal marsh of Chester Creek along the right side of road. These rare freshwater marshes that are under tidal influence along the lower Connecticut River valley support very diverse flora including stands of wild rice and provide unique habitat for migrating bird populations; they have been designated as "Wetlands of International Significance" by the RAMSAR convention and "One of the Last Great Places" by The Nature Conservancy.
- 11.4 Turn left at stop light on Rt. 154 north. Climb back up from erosional creek level, and continue along 55-65-ft surface of unit Ice deposits.
- 13.0 Note Connecticut River to the right.
- 14.2 In this vicinity, the road follows the contact between sand and gravel deposits of unit lct to the east (right) and the till/bedrock slope to the west (left).
- 14.6 Proceed straight through light at junction with Rt. 82 east, continue along contact.
- 15.8 Turn sharply right on Ruddy Ferry Rd.
- 16.0 Turn right on pit access road.

**STOP 2. Arrigoni Bros. Pit in Tylerville section of Haddam (Deep River quadrangle).** This pit is excavated into an ice-marginal delta in the Lower Connecticut River deposits (unit lct), which are an excellent example of sediment-dammed pond deposits (SP Depositional System). In this part of the river valley the deposits are a series of shingled ice-marginal deltas. The delta plain here is above the 70-ft contour (fig. 7). From this stop to north of the Haddam-Middletown town line, delta-plain surfaces rise from 75 ft (23 m) to 110 ft (34 m), a gradient of about 6 ft/mi (1.2 m/km), most of which can be accounted for by postglacial uplift tilt. In these deltas, fluvial topset beds are generally thin less than 10 ft (3 m) thick. To the north, deltaic surfaces at and just south of Maromas are at 135-145 ft (41-44 m)

and rise to 175 ft (53 m) at the northernmost ice-marginal head (STOP 5). These delta plains define a steeper gradient of about 10 ft/mi (3 m/km), and topset beds are as much as 25 ft (8 m) thick.

The pit exposes 6 m of an upper glaciodeltaic section (fig. 8). The topset strata are typical of the sand and gravel glaciofluvial sedimentary facies (see Appendix 1 in Stone and others, 2005). These beds vary from reddish brown sand and gravel, derived from the early Mesozoic basin 14 km to the north, to yellowish light brown sand and pebble gravel derived from local metamorphic and sulphidic bedrock sources. Maximum intermediate diameter of gravel clasts is about 8 cm at a distance of 0.7 mi (1 km) from the ice-margin source at the head of the deposit (fig. 7). Cross beds in the fluvial section trend southerly to south-southwesterly. The pit wall is cut nearly perpendicular to the stream paleoflow direction so that cross-section views of wide, shallow fluvial channels and bar deposits are exposed.

Foreset strata in the pit include sandy foreset sedimentary facies in the western part of the pit, and sand and pebble gravel foreset beds in the eastern, more proximal part. The pit wall exhibits down-dip cross-section views of sets of foreset strata in both facies. Channel-filling, concave-outward sets of beds are present in foreset sets that are 4 to >10 m in width. Sandy foresets preserve curved bedding planes that merge tangentially with deltaic bottomset strata. The upper foreset beds are truncated by a planar erosional surface at the base of the yellow sandy topset strata, at an altitude of about 66 ft (20 m).

The stratigraphic significance of the deltaic section at STOP 2 is twofold. First, the pit shows a downstream coarse-to-fine distribution of foreset facies, and the extent of the delta topset plain shows a similar fining of fluvial sediments away from its correlated ice-margin position (fig. 7). Similarly, the successive elevated heads of multiple deltaic deposits, their ice-contact zones, and shingled onlap relationships among deltas demonstrate the systematic pattern of ice-margin retreat up the valley. The terraces along the river containing these deposits thus are not a single, long glaciofluvial deposit, as suggested previously (Flint, 1933). Secondly, all of the Lower Connecticut River deposits (map units lce, lcc, lct, fig. 4) record ice-marginal sedimentation in a series of sediment-dammed ponds that rise northward above the regional isostatic tilt slope, indicating rising local pond base levels over successively higher sediment dams. These deposits are the type deposits of the SP depositional system. Furthermore, these deltaic deposits are all above the levels of glacial Lake Connecticut in Long Island Sound Basin (fig. 4), which are known from deltas along the coast to New Haven to be lowering during this period of deglaciation.

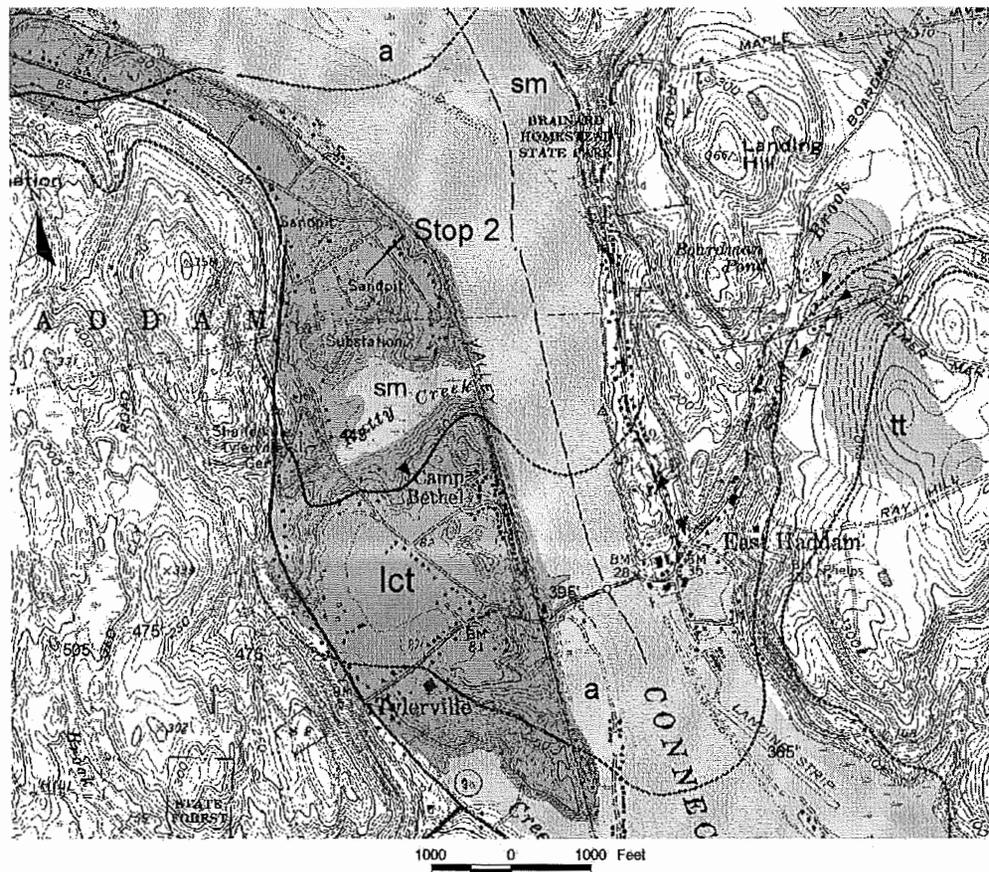


Figure 7. Topography and extent of Lower Connecticut River sediment-dammed pond deposits (lct) in the vicinity of STOP 2. Heavy lines are ice margin positions.



Figure 8. Glaciodeltaic topset and foreset strata exposed in Arrigoni Bros. pit, Lower Connecticut River deposits (unit lct) Tylerville, CT, August, 2005. Oblique view shows 3 m of sand and gravel foreset beds disconformably overlying 2 m of sand and pebble gravel foreset beds. Photo by J.R. Stone.

- 16.5 Exit Arrigoni Bros. pit and turn right on Rutty Ferry Rd. Continue along surface of lct map unit
- 17.7 Descend into erosional Mill Creek valley, climb back up onto lct surface.
- 18.0 Road crosses 100-ft contour marking an ice-margin position.
- 18.4 Turn left on Jail Hill Rd. The unit lct deposit here is lower in altitude (85–90 ft) and is the distal end of next morphosequence to the north.
- 18.5 Leave lct deposit, climb till/bedrock hill, and descend into Mill Creek Valley.
- 19.2 Bear left at stop sign over bridge
- 19.3 Bear right onto Beaver Meadow Rd (no sign).
- 20.2 Pass under Rt. 9 highway.
- 21.2 Turn left at gravel road entrance to Cockaponset State Forest (Mottland Rd.).
- 21.3 Stop and unload passengers at trailhead on right. Passengers will hike south for approximately 0.5 mile. Drivers proceed with vehicles on Mottland Rd. Road passes along eastern side of ice-channel-fill ridge.
- 22.0 Sharp right turn on dirt road.
- 22.1 Enter old pitted area, park vehicles, and drivers will hike north to meet passenger group.

**STOP 3. Hike trail along the crest of an ice-channel-fed delta deposit in the West Haddam ice-dammed pond deposits (unit whd)** (Haddam quadrangle). These deposits are a series of ice-marginal deltas deposited in several ice-dammed pond basins in small northeast-draining tributaries to the Connecticut River. Spillways for these upland valley basins range from 565 ft (172 m) down to 335 ft (102 m) across local divides (fig. 4). This series of deposits is an excellent example of the IP depositional system. Deposition of these small, high-level deltas in successively lower ice-margin positions to the northeast was due to lobation of the ice margin in the lower Connecticut River valley where the western side of this lobe retreated to the northeast. Figure 9 shows the extent of deposits in three of these north-draining upland valleys-- Beaver Meadow Brook, an unnamed tributary, and Turkey Hill Brook. STOP 3 is in the Beaver Meadow valley, where we will look at the morphology of an ice-marginal delta and its feeder channel system.

The morphology of the feature at STOP 3 includes a flat-topped delta plain at altitude 465 ft that onlaps an older ice-contact deposit to the south, and extends to ice-contact slopes on its northern and eastern sides. The sandy sediments beneath surface gravel in the plain, and inferred depth of the basin in this valley indicates that the deposit is an ice-marginal glaciodeltaic morphosequence. An ice-channel ridge, bounded by ice-contact slopes, is connected on grade (465 ft) to the delta plain, drops 60-70 ft in altitude and extends 2000 ft (610 m) to the north. The ridge has a flat-to-hummocky crest, 900 ft (275 m) long, as much as 250 ft (75 m) wide in the central part, which narrows to a sharp crest in the northern part of the ridge. As you walk along the ridge crest note its few surface boulders and its hummocky surface that includes some broad closed depressions, indicating the presence of buried ice beneath the ridge as well as on either side. Because this ridge has neither a sharp, undulatory longitudinal crest, nor a lateral sinusoidal crest, it does not conform to most definitions of an *esker*. It may be described most simply as an *ice-channel ridge*. The origin of this ridge may be debated. It may be:

- 1) a coarse *glaciofluvial* deposit collapsed downward from its original position in a wide crevasse ice channel,

- 2) a coarse *glaciodeltaic* deposit collapsed downward from its original position in a wide crevasse ice-channel,
- 3) a coarse ice-channel deposit constructed in a closed, sub- or englacial ice tunnel under high hydrostatic pressure,
- 4) a coarse ice-channel deposit constructed in an ice-tunnel and emergent fountain at the contact with the delta plain,
- 5) any of the above, eroded at the surface by water flowing to a lower spillway altitude to the east.

Origins as glaciofluvial or ice-tunnel deposits predict very coarse sediments throughout the ridge, derived from a fountain source of fluvial sediments or tributary tunnel sources of tunnel-ridge sediments. The deltaic origin predicts coarse topset and foreset facies in the ridge, derived from a tunnel or fountain source to the north.

Deltaic deposits are poorly exposed in old pit area where cars will be parked. We will continue on to Stop 4 where an active sand and gravel excavation reveals excellent internal structure of delta deposits in similar depositional setting in the Turkey Hill valley to the east (fig. 9).

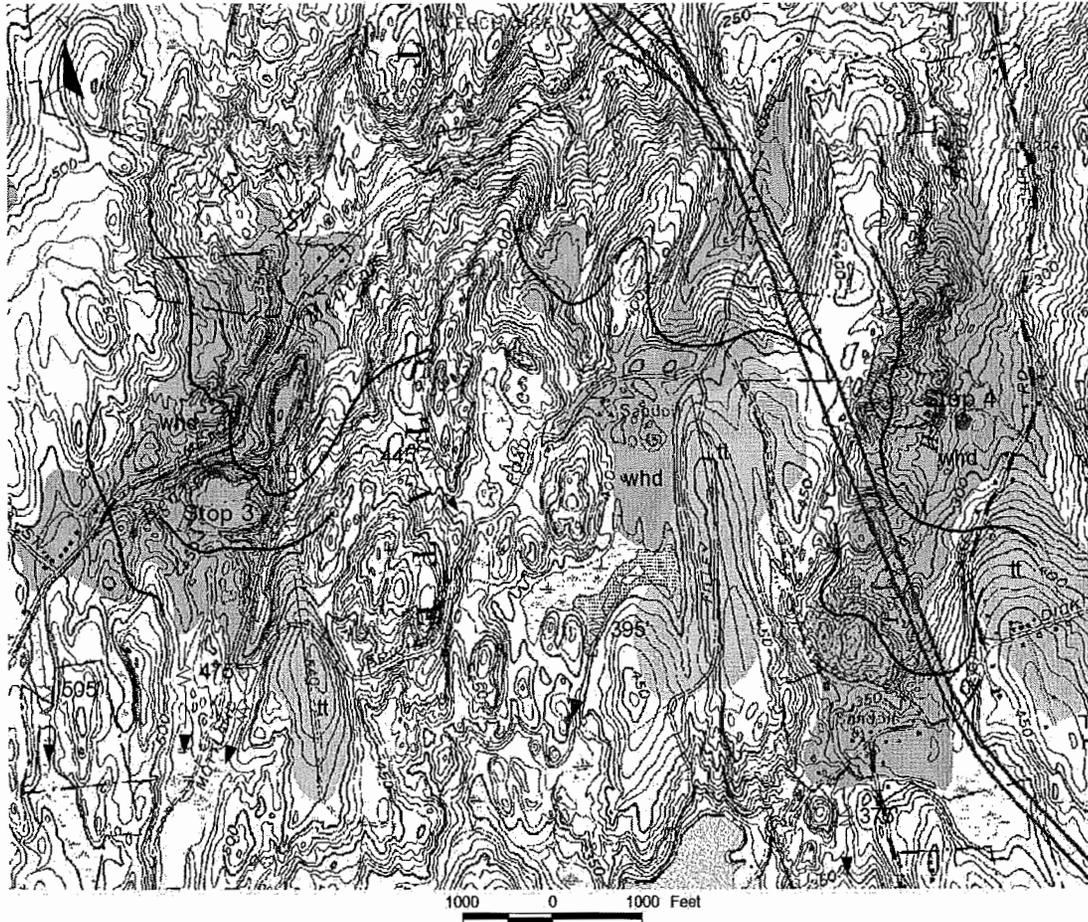


Figure 9. Topography and extent of West Haddam ice-dammed pond deposits (unit whd) in the vicinity of Stops 3 and 4. Heavy lines are ice margin positions. Arrows indicate glacial lake spillways, numbers are spillway altitude in feet.

- 22.1 Leave pit area and return to Mottland Rd.
- 22.3 Turn right on Mottland Rd.
- 22.4 Sharp left onto Jericho Rd.
- 22.5 Note spillway channel (475-ft altitude) to the right.
- 22.6 Climb onto thick till (drumlin), cross over and note smooth northern slope of drumlin.
- 22.8 Drive off drumlin into bedrock controlled topography, note numerous bedrock outcrops of gneiss/schist of the Middletown Formation that strikes north-northeast and dips steeply east in well-defined linear ridges.
- 23.6 Drive onto surface of another ice-marginal delta of unit whd with surface altitudes of 395–405 ft and graded to the 395-ft spillway (fig. 9). Note small old sand pits on right.
- 23.9 Turn left on Filley Rd (no sign).

- 24.2 Descend steep ice-contact slope. Pass under Rt. 9 highway onto paved road.  
 25.2 Turn right on Turkey Hill Rd. Note that you are now driving south up the north-draining Turkey Hill Brook valley.  
 26.1 Bear right onto Cedar Lake Rd.  
 26.4 Turn right onto pit access road.  
 26.6 Enter pit and park.

**STOP 4. WFS Earth Materials Construction Pit on Turkey Hill Road in Haddam** (Haddam quadrangle). An active pit at this location is excavated into an ice-marginal delta in the West Haddam deposits (whd, figs. 4 and 9). The small delta plain at the crest of the deposit is at 385 ft (116 m) altitude indicating its relationship to the glacial pond spillway at 365 ft (111 m) altitude to the south (fig. 9). The original landform preserved four such high pinnacles with steep ice-contact slopes along their eastern and northern sides. The morphology of flat-topped deposits in this map unit indicates multiple small deltaic deposits accumulated against the retreating ice margin in a deep, ice-dammed pond.

This pit is excavated into the upper half of the deltaic deposit. It exposes thin topset strata at the top, and a generally downward-fining, 21-m (70-ft) sequence of gravelly and sandy foreset and bottomset strata. The upper pit (fig. 10A) exposes pebble-cobble topset strata, five sets of deltaic foreset strata composed of thin beds of coarse sand and pebble gravel in planar and tangential channel-fill sets. Foreset sets have disconformable dip directions and are separated by sharp erosional surfaces. The lower level of the pit exposes sandy foreset beds in disconformable sets that erosionally truncate each other and that locally truncate collapse margins of underlying foresets and bottomsets (fig. 10B). The lower strata are deformed by large collapse monoclines and synclines marked by low-amplitude folds and normal and high-angle reverse faults in fold limbs. Irregular erosional surfaces cut the upper surfaces of these sets of beds and their deformation structures. Collapsed gravel and eroded blocks of fine sand locally are present above the upper erosional contacts. Channel-fill sandy foresets overlie the deformed and eroded lower beds. These sandy foresets include cross-cutting sets of strata, which also contain beds of pebble gravel. The patterns of collapsed deltaic strata and overlying channel-fill strata probably mimic the collapse-and-fill geometry of successive ice-marginal deltaic morphosequences in many deposits in the valleys of New England.

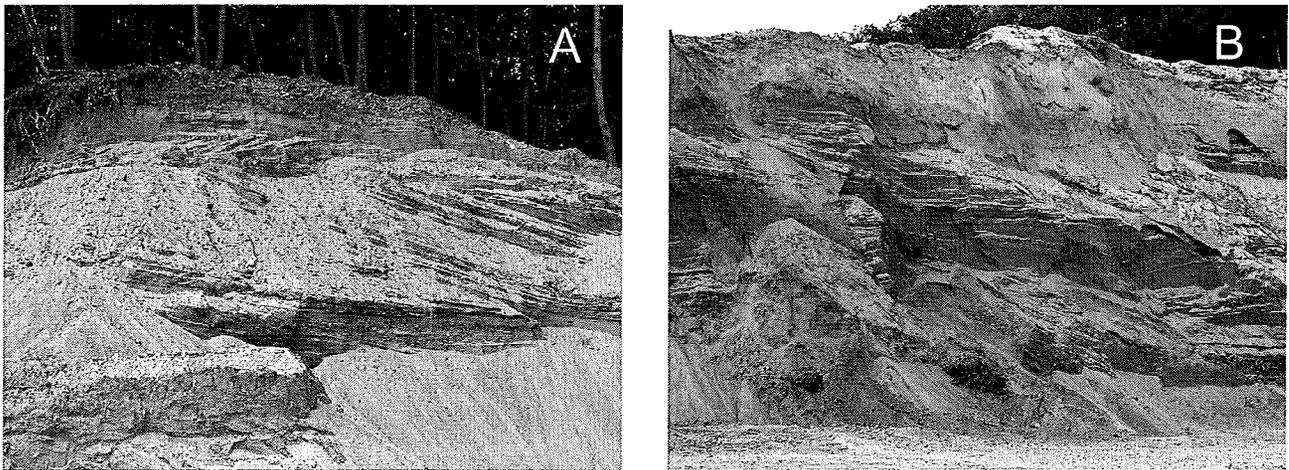


Figure 10. Glaciodeltaic deposits exposed in the WFS Earth Materials Construction Pit Haddam, August, 2005. A. Upper pit showing thin topset gravel and five sets of foreset strata, B. lower pit showing three sets of cross-cutting foreset and bottomset strata and collapse deformation features. Photos by J.R. Stone.

- 27.0 Leave pit and turn left on Cedar Lake Rd.  
 28.2 Turn left on Filley Rd.  
 28.7 Turn right on Weiss Rd.  
 29.0 Turn left on Beaver Meadow Rd.  
 29.2 Turn left on Rt. 9 South.  
 32.5 Leave Rt. 9 at exit 7 for Rt. 82.  
 35.3 Left on light on Rt. 154 North.  
 35.7 Turn right at light onto Rt. 82 East.

- 35.8 Note you are traveling along a 75-85-ft surface of unit lct.
- 36.1 Note descent to Connecticut River flood plain.
- 36.3 Cross Connecticut River at East Haddam Swing Bridge (note - if the bridge is open for boats to pass we may have to wait).
- 36.5 Sharp left to stay on Rt. 82, note famous Goodspeed Opera House on right.
- 36.6 Bear left onto Rt. 149.
- 36.8 Note you are on the west side of a spillway identified by R.J. Lougee (1953) as the outlet for Lake Hitchcock.
- 37.2 Bear left at stop sign to stay on Rt. 149. Note that you are traveling along the east side of the Connecticut River – notice bedrock outcrops on steep slopes on the right side and marshes bordering the river on left.
- 38.2 Turn left onto Johnsonville Rd.
- 38.5 Cross Moodus River – note waterfall at old mill dam to right.
- 38.7 Straight at stop sign. Note chapel on right– site of Stone wedding (circa 1975).
- 38.8 Left onto Rt. 151.
- 39.7 Note Cave Hill Resort on right. Hills above are the site of the cave which is the ‘origin’ of the famous “Moodus Noises” (earthquakes).
- 40.1 Cross Salmon River.
- 40.7 Bear left to stay on Rt. 151 toward Cobalt.
- 43.8 Straight at flashing traffic light.
- 45.8 Note you are now coming onto unit lct on left side of road (driving along the contact).
- 46.3 Proceed straight at stop light at junction with Rt. 66 (Cobalt Market on your right if you need lunch supplies).
- 46.5 Bear left at fork, bear left at stop sign onto Middle Haddam Rd.
- 46.9 Bear left to remain on Middle Haddam Rd.
- 47.3 Turn right into grassy parking area near old barn (Middlesex Land Trust Property). Park and walk across the road to private residence for **LUNCH STOP**.
- 47.3 Turn right to leave parking area, continue on Middle Haddam Rd.
- 47.8 Note we are going under brownstone tunnel for old ‘Airline Rail Road’, caution – single lane traffic.
- 50.0 Turn left at stop sign to stay on Middle Haddam Rd. Travel over the undulating surface of northern part of unit lct deposits in the vicinity of Jobs Pond (fig.11).
- 50.6 Turn right on Rt. 66.
- 51.3 Turn right onto H.E. Butler Construction Company access road. Stay to the right to get to pit area, and park.

**STOP 5. Butler Pit in Portland** (Middle Haddam quadrangle). This pit is excavated into the upper proximal part of an ice-marginal delta in the Lower Connecticut deposits (unit lct, fig. 3). At this stop we are at the north end of the same map unit as at STOP 2. Here delta-plain surface altitudes are at 175-185 ft (53-56 m) (fig. 11). Topset beds beneath these delta-plain surfaces are as much as 25 ft (8 m) thick.

The pit exposes glaciofluvial delta topset beds, collapsed and nondeformed, in a long exposure that reveals the downstream fining of gravel clast sizes from north to south (figure 12 A, B). Deeper excavations over the last 20 years have displayed gravelly upper delta foreset strata below a topset-foreset contact at about 165 ft altitude, as described by London (1985, Trip B6, STOP 10). The upper topset gravel deposits have clasts with intermediate axes of about 45 cm in the northern ice-contact edge of the deposit. Here the gravel beds are planar or indistinctly planar, with local lines of boulders at the base of beds. Clasts are rounded to well rounded; some preserve glacial striae. In the more distal part of the pit, nearly 1000 ft (305 m) to the south, maximum gravel clasts have intermediate diameters of 25 cm. Clasts are rounded to well rounded, without striae. The section contains a larger proportion of sandy and pebble-cobble beds.

The deposits here and to the south surround Jobs Pond, a scenic, compound-kettle pond that overlies the thalweg of the glacially overdeepened bedrock valley of the Connecticut River. At least two ice-marginal deltaic morphosequences extend across the buried valley here, including a fine example of an ice-channel feeder of the eastern delta (fig. 11). The deposits in this vicinity locally are >250 ft thick and originally extended across the area of the present river channel to the south. These deposits filled the valley to 165-185 ft altitude and formed the northern part of the dam for glacial Lake Middletown, deltas of which, in turn, formed the drift dam (STOP 6) for the long-lived glacial Lake Hitchcock to the north. Following drainage of the glacial lakes in the Connecticut valley and post-glacial uplift tilting, the ancestral Connecticut River was forced to flow on the lower lake-bottom plain of Lake Middletown to the west, rather than along the sediment-filled bedrock valley beneath Jobs Pond. The future course of the river and its local subsequent history was set by the nature of ice-margin retreat and valley-filling lake sedimentation.

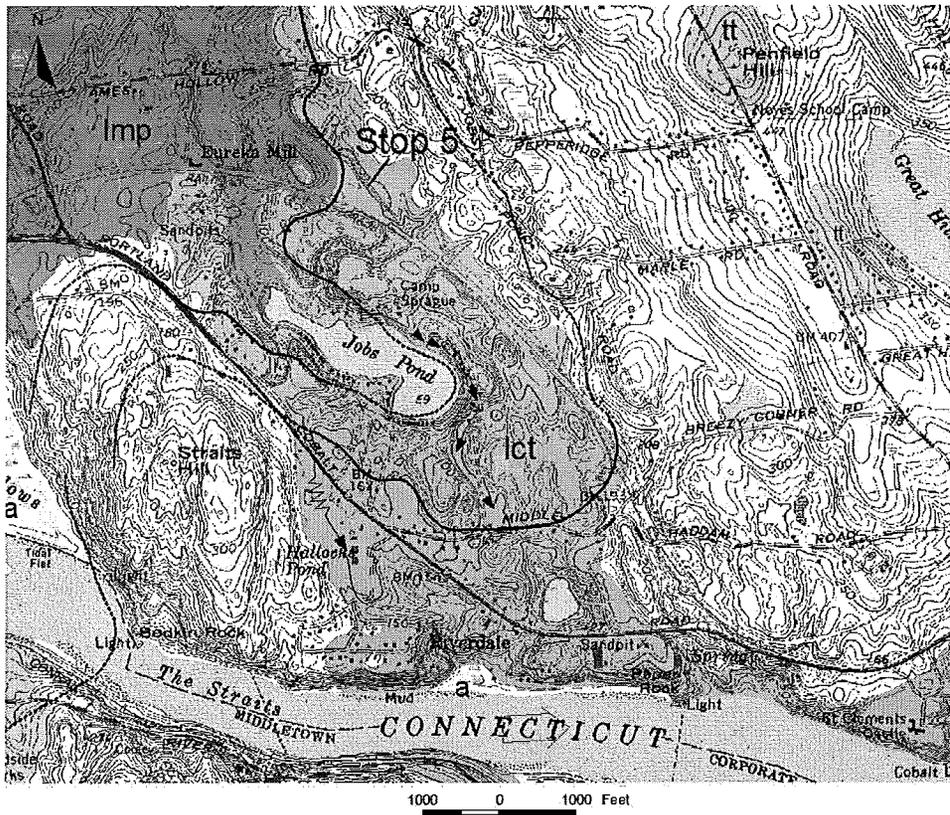


Figure 11. Topography and extent of lower Connecticut River sediment-dammed pond deposits in the vicinity of Stop 5. Heavy lines are ice margin positions. Dashed arrow heads are crest of esker feeder to lct deltas along associated ice margin position.

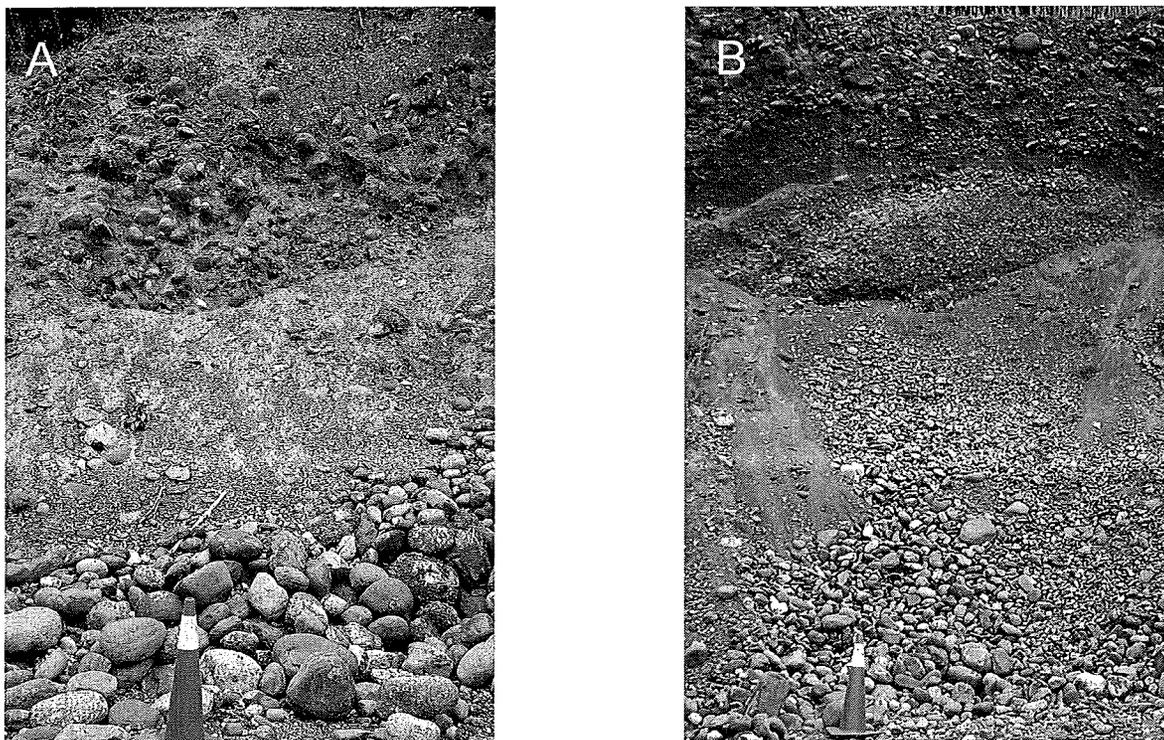


Figure 12. Glaciofluvial gravel deposits exposed in the Butler Construction Company pit, Portland, August, 2005. A. coarse boulder gravel exposed at north end of pit. B. cobble gravel exposed at south end of pit. Cone for scale is 25 in (0.6 m) high. Photos by J.R. Stone.

- 52.3 Leave Butler Pit, turn right on Rt. 66.  
 52.4 Pass the westernmost metamorphic bedrock outcrop on the left. As we descend slope toward stop light we cross the Eastern Border fault of the Hartford Basin.  
 52.5 Turn right at stop light on Rt. 17.  
 53.0 Note outcrops of Early Jurassic Portland Formation on the left.  
 53.6 Straight through light.  
 54.6 Straight through light at Fogelmarks Corner. Traveling along lake-bottom surface of glacial Lake Middletown (unit lmc).  
 54.8 Note old gravel pits on right in glacial Lake Middletown delta deposits.  
 56.2 Note we are traveling the east edge of lmc delta deposits.  
 57.4 Turn left on Old Maids Lane at the apple sign, note metamorphic outcrop on right side of Rt. 17 at the turn. We are traveling across the 175-ft surface of unit lmc delta plains (fig. 13).  
 58.0 Turn left onto gravel road (town of Glastonbury signs on right).  
 58.2 Park and assemble for view to the west across Connecticut River to the other side of the Rocky Hill dam of glacial Lake Hitchcock. Note that depending on access conditions on the day of the trip we may continue to the STOP 6 pit on foot from this spot *OR* we may continue with the driving directions listed below. Turn cars around.  
 58.4 Turn left on Old Maids Lane.  
 58.7 Descend onto 50-ft terrace cut through Rocky Hill dam.  
 59.0 Turn left on Tryon Street.  
 59.1 Turn left into Glastonbury Bulky Waste Disposal Facility for STOP 6.

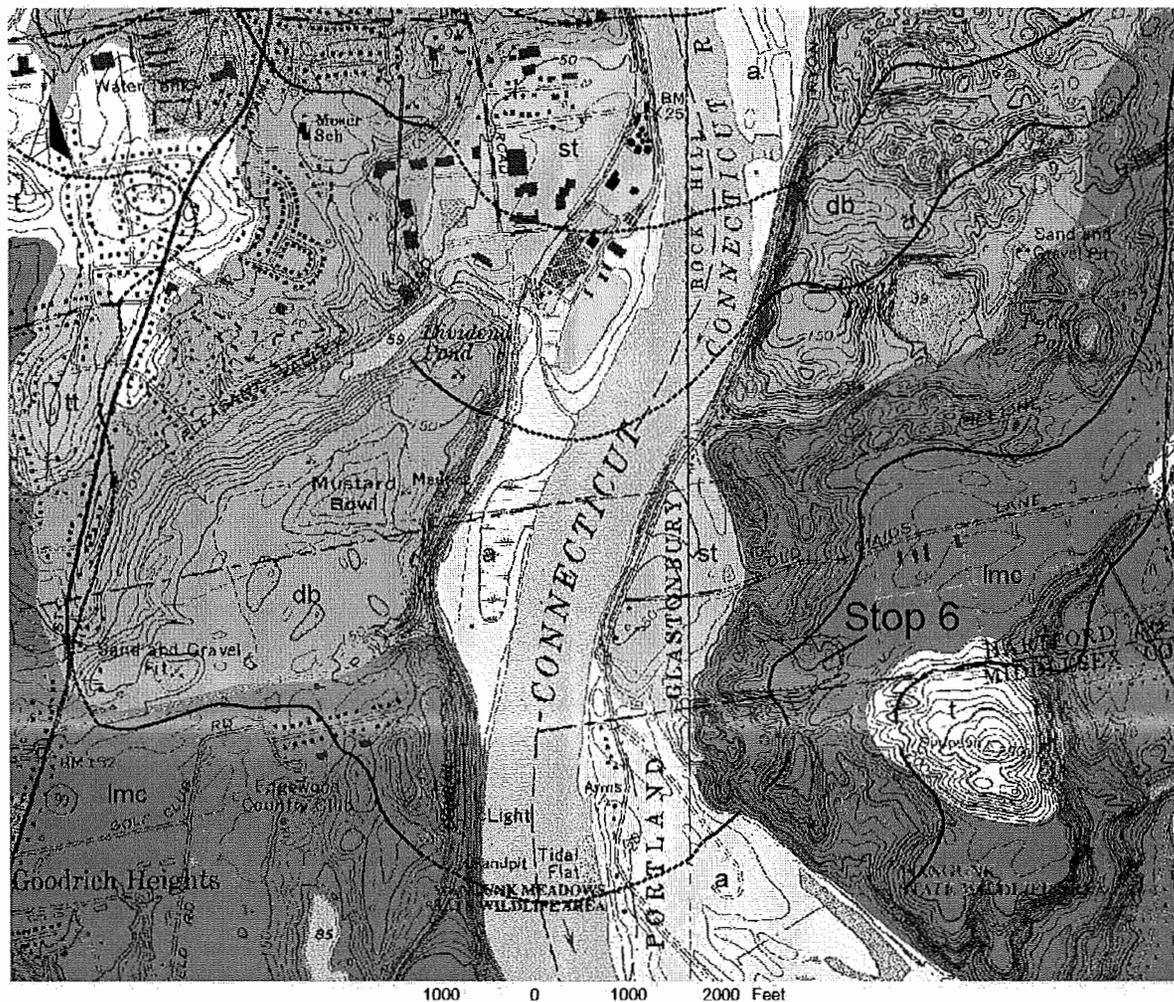


Figure 13. Topography and extent of Glacial Lake Middletown deposits and Dividend Brook sediment-dammed pond deposits in the vicinity of Stop 6. Heavy lines are ice margin positions.

**STOP 6. Glastonbury Bulky Waste Disposal Pit, Glastonbury, (Glastonbury quadrangle).** This pit is excavated into an ice-marginal delta in the Cromwell deltaic deposits of glacial Lake Middletown (map unit lmc, figs. 3, 13) SL Depositional System. In this part of the river valley the deposits are a series of extensive ice-marginal deltas with surfaces at 165 to 185 ft (50 to 56 m), having topset-foreset contacts at 149 ft (45 m), which project to spillway altitudes at 120 ft (37 m) across lct deposits to the south. Deltas in Cromwell have depositional free fronts built into open water in the glacial Lake Middletown basin. Together with Dividend Brook deposits (unit db) these deltaic sediments form a massive blockage (at 155-175 ft altitude) in a narrow part of the Connecticut River valley and constitute the sediment dam for glacial Lake Hitchcock. During the earlier life of Lake Hitchcock, water spilled through a spillway to the west (New Britain spillway, see earlier discussion), but at about 13.5 <sup>14</sup>C years ago, the Rocky Hill dam was breached and a terrace, 50 ft in altitude, cut through these deposits records this event.

At this STOP the flat delta plain is at 165 ft (50 m), but the upper exposures are cut into the 150-ft (46-m) surface of an ice-channel ridge at the juncture of walls of a compound kettle (fig. 13). Here, the upper pit exposes the coarsest, most proximal sediments (fig. 14A), which are characteristic of the coarse gravel fluvial facies and the sand and gravel ice-channel fluvial facies (appendix 1). Gravel clasts are boulder-sized, subangular to subrounded, with intermediate diameters of 40-50 cm. Even though the matrix is poorly sorted and contains fine sand and some silt, there are open-work gravel textures locally. Beds are collapsed, indicating sediment accumulation in the ice channel between huge ice blocks that later were centers of deep collapse in the compound kettle. To the west, and overlying the coarse gravel, sand and gravel foreset beds (fig. 14B) dip 25°-30° westerly into the Lake Middletown basin. The strata in this facies consist of alternating beds of coarse pebble sand and pebble-cobble gravel. Coarser beds exhibit open-work textures; bed traces thin and terminate in a down-dip direction in the pit wall. Sandy foresets contain thinly bedded sand and pebble gravel, showing local imbrication of clasts. These upper foresets along the ice-margin position for unit lmc (fig. 13) record the last depositional events at the collapsed edge of these deposits of the Rocky Hill dam of Lake Hitchcock.

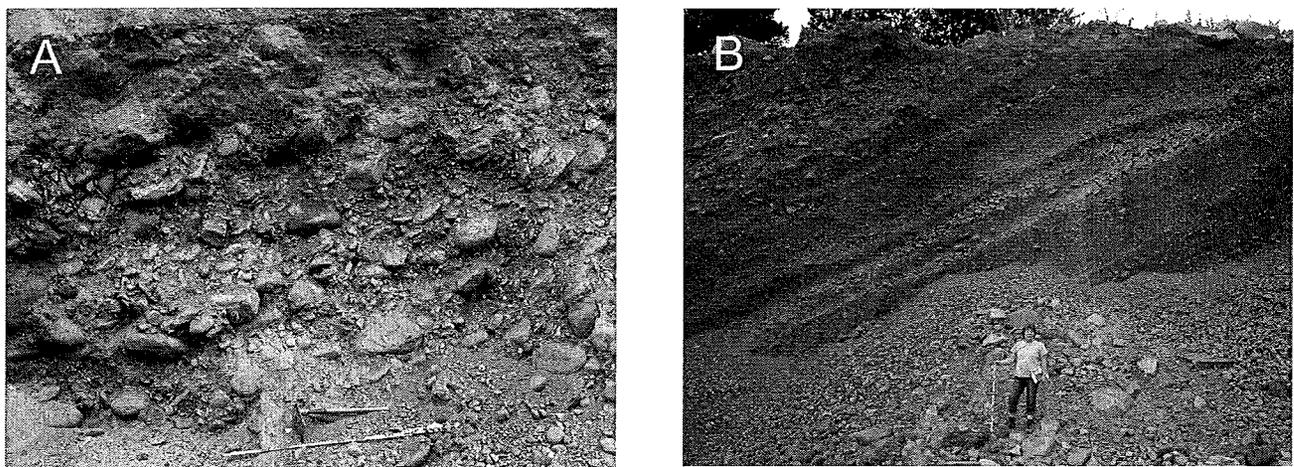


Figure 14. Ice-marginal glaciodeltaic deposits exposed in the Glastonbury Bulky Waste Site excavation, August, 2005. A. coarse ice-channel boulder gravel exposed in wall of kettle depression, B. sand and gravel deltaic foreset facies. 0.5-m shovel handle and 1-m tape for scale. Photos by B.D. Stone.

- 59.3 Leave pit area and turn right on Tryon Street. Continue to drive on 50-ft terrace.
- 59.8 Descend to modern river level.
- 60.5 Note wide flood plain of Connecticut River (15-25 ft altitude) to left and steep erosional edge of Dividend Brook deltaic deposits (unit db, also part of the Rocky Hill dam) to the right (fig. 13).
- 61.2 Turn right on Dug Rd. Note this road makes its way from the river terraces up through the db deposits to their noncollapsed surface at 185 ft.
- 62.1 Turn left onto Rt. 17.
- 62.9 Straight through lights at South Glastonbury.
- 64.2 Straight through light to stay on Rt. 17 (now 4 lanes).
- 66.0 Merge onto Rt. 2 North access road.
- 66.6 Merge onto Rt. 2 North.
- 68.7 Stay right for exits 2E and 3. Take EXIT 3 to Pitkin Street.
- 69.0 Turn right at end of ramp onto Pitkin Street.

- 69.1 Pass the USGS Connecticut Water Science Center on the right.
- 69.3 Turn left at light onto Main St., East Hartford, US Rt. 5. Through East Hartford Center travel along a 45-55 ft stream-terrace surface.
- 72.7 Pass under Rt. 291 bridge. The stream terrace surface is at 65-75 ft altitude here, also note sand dunes that reach 100 ft altitude.
- 73.5 Cross Podunk River; old clay pits to the right.
- 75.4 Turn right onto Strong Rd.
- 75.4 Immediate left into Industrial Park Rd.
- 75.7 Enter KF Plant, Redland Brick, Inc for STOP 7.

**STOP 7: KF Plant of Redland Brick Co. (formerly Kelsey Ferguson Brickyard), South Windsor, (Manchester quadrangle). NOTE!!** The clay pits at the KF plant are an active mining operation and permission is required from the plant manager for entry. The Kelsey-Ferguson clay pits have been stops on sponsored field trips in the Connecticut Valley for several decades (Hartshorn and Colton, 1967; Ashley and others, 1982; Koteff and others, 1988; Stone and Ashley, 1992). Exposed in the pits is up to 5 m of fluvial cross-bedded sand and pebbly sand unconformably overlying silt and clay varves of glacial Lake Hitchcock. Recent mining has exposed about 7 m of continuous varve section. The varves are light to medium olive gray with occasional pinkish to reddish gray tones and are 0.2-2.0 cm thick. The varves have well-defined winter clay beds and summer layers composed of multiple micrograded sandy to clayey silt beds. Summer layers also have the distinction of both starting and ending with coarse (fine to medium sand) layers that may represent overturning events at the beginning and end of the melting season.

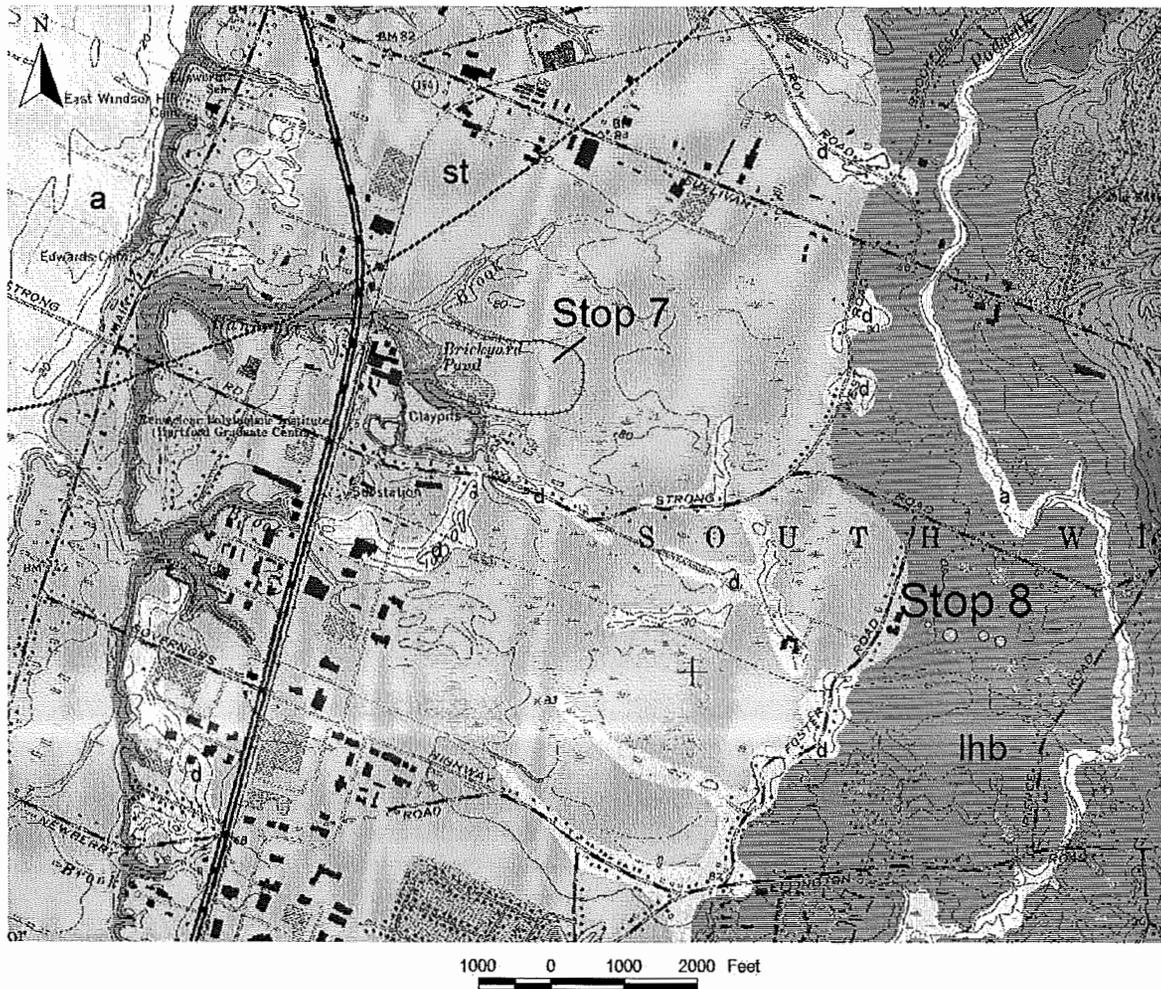


Figure 15. Topography and extent of Glacial Lake Hitchcock bottom deposits (lhb), stream terrace deposits (st), and early postglacial dunes (d) in the vicinity of Stop 7 and 8.

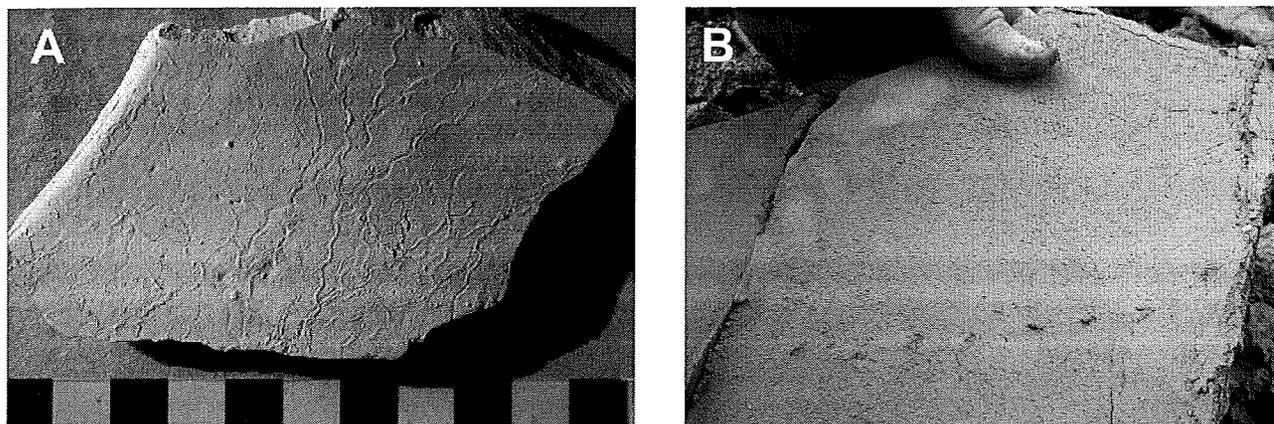


Figure 16. Trace fossils in varves at Kelsey-Ferguson clay pits. A. Sinusoidal trace (*Cochlichmus*) produced by insect larvae and nematodes. B. Evenly-spaced pits possibly produced by fish.

The varves also contain trace fossils (fig. 16), the most common of which is a small (~0.5 cm wavelength) sinusoidal trace produced by either insect larvae or nematodes (Ashley, 1972, 1975). Last summer a larger trace composed of evenly-spaced pits was discovered that may be the product of a fish (Benner and Ridge, 2004). The gray varves represent a distal glacial varve sequence dominated by sediment derived from metamorphic rock sources that flank both sides of the Connecticut Valley. Near the base of the exposed sequence the varves begin to transition downward into red, hematitic varves with increased calcium carbonate concentrations as is evident from the many red concretions scattered across the floor of the clay pit. The lower red varves are dominantly derived from glacially eroded Mesozoic rocks of the Hartford Basin that underlie the Connecticut Valley. In the summer of 2004 an exposed face on the southeastern corner of the active pit was sampled with an overlapping set of 2-ft long, 3-inch diameter PVC cores for detailed study of the varve stratigraphy. A compiled varve sequence from the cores totaling 552 years was matched to Antevs' (1922) New England varve chronology (NE 3617-4168) without a single extra or missing varve in the section (see fig. 5 and discussion in text).

- 75.9 Leave KF Plant, return to Strong Rd.
- 76.1 Turn left onto Strong Rd.
- 76.3 Straight at stop sign.
- 77.4 Bear right at stop sign, stay on Strong Rd.
- 77.5 Note - leave terrace deposits and travel onto original Lake Hitchcock bottom surface.
- 77.7 Turn right on Foster Rd.
- 78.0 Turn left into parking lot of Seventh Day Adventist Church. Park beside play ground and take grassy trail to the east for STOP 8.

**STOP 8. Rimmed depressions developed on drained lake-bottom surface of glacial Lake Hitchcock, South Windsor** (Manchester quadrangle). Enroute to this STOP, we have left the terraced lake-bottom surface into which the KF clay pit is excavated and driven onto a non-terraced lake-bottom surface of glacial Lake Hitchcock. Hundreds of circular to subcircular, rimmed depressions are found on such preserved lakebed surfaces in the southern basin. These features formed quickly after the lake drained at 13.5 ka. They are not present on terraced surfaces or the modern flood plain, indicating that the timing of their formation did not persist past terrace-building time. Based on a highway excavation through one of these features, subsurface investigations, and a depression-fill paleobotanical record dating back to 12.4 ka, these depressions have been interpreted as pingo scars (Stone and Ashley, 1989; Stone and others, 1991; Stone and Ashley, 1992). Pingos were envisioned to have formed in a harsh periglacial climate that affected the region in early postglacial time. Other evidence of harsh climate in the early postglacial time support this conclusion: the paleobotanical record of treeless, tundra vegetation, dated from >15 ka to about 13 ka (Davis and others, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987; Thorson and Webb, 1991); extensive fields of eolian sand dunes formed in the treeless environment; and wedge-shaped features with a polygonal ground pattern, interpreted as ice-wedge casts (Schafer and Hartshorn, 1965; Schafer, 1968; O'Leary, 1975; Stone and Ashley, 1992). The wedge structures indicate that permafrost existed locally in areas where substrate conditions were favorable to its formation. The presence of permafrost structures on drained Lake Hitchcock surfaces would indicate that mean annual temperatures were below 0°C during the early postlake time interval.

An alternative possibility will be discussed— that these features are water-escape structures produced by upward groundwater discharge to the glacial lake from the uplands, forcefully driven by the ~30-ft (10-m) loss of head in the basin associated with rapid lowering of the lake when the Rocky Hill Dam was breached. Formation of these features may also have been aided by the occurrence of earthquakes generated by the initiation of glacio-isostatic rebound—a time when the earth's crust in southern New England was uplifting at a rate of about 3 m per century.



Figure 17. Easterly view from a small plane of rimmed depressions in the vicinity of Stop 8. Circular features with ponded water are 30-40 m in diameter. Note circular patterns defined by trees marking rims of depressions in forested areas. Road beyond the ponds is Pierce Road (see fig. 16).

#### END OF FIELDTRIP

To Return to New Haven:

- 78.0 Leave parking lot turning left onto Foster Rd.
- 78.1 Pass under power line and travel along crest of sand dune that lies on the boundary of stream terrace to the west (right) and un-eroded lake bottom surface to the east (left).
- 78.8 Turn right on Ellington Rd.
- 78.9 Turn right at light onto Governors Highway. Travel along northwest trending arm of the parabolic dune.
- 79.2 Straight at stop sign.
- 80.2 Left at light onto US Rt. 5 South.
- 81.9 Turn right onto Rt. 291 entrance ramp.
- 84.9 Take exit 2A for I-91 South toward Hartford, continue approximately 45 miles to the New Haven area.

#### ACKNOWLEDGMENTS

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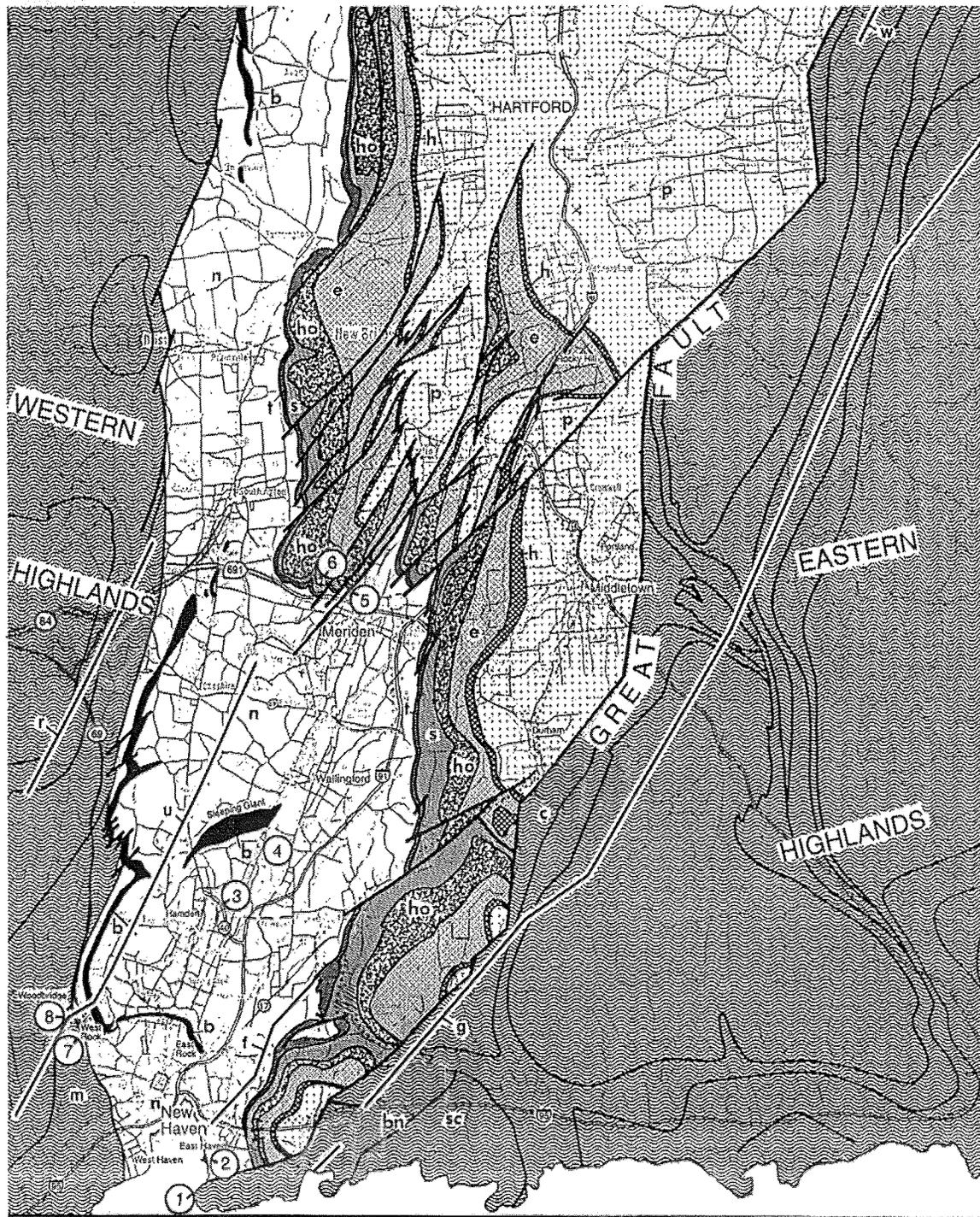
## READING THE ROCK AND LANDSCAPE RECORDS OF THE NEW HAVEN REGION

Trip Leaders: Leo J. Hickey and Copeland MacClintock

### INTRODUCTION

On this trip we will trace some of the geological events that have produced the fabric of the Hartford Basin in the vicinity of New Haven, Connecticut, and will show how this fabric accounts for the distribution of the resources, vegetation, and population centers of the area. The route starts on the crystalline rocks of the Avalonian terrain on the southeastern margin of the basin, proceeds generally northwestward through the early Mesozoic sediments forming part of the graben-fill, and terminates on the western margin of the basin on the metamorphosed sediments of the Iapetus Ocean (Figs. 1-3). The trip starts and ends on the Yale University campus at the Kline Geological Laboratory (KGL), which houses the Department of Geology and Geophysics.

Mileage X= traffic light	Description
0 X	Leave KGL parking lot. Go straight across Whitney Ave. at light onto Humphrey St.
0.7	Turn right on East St.
0.4	Crossing Ives Place road descends from the New Haven outwash to artificial fill making up the industrial area to the south and east. The fill covers what was once salt marsh and open estuary.
0.4	Turn left on Forbes Ave. The area of rubble to the left was Waterside Park, a landscaped bathing and recreation area built in the early 1890's.
0.3	Quinnipiac River and the Forbes Avenue, or "Blue Q" Bridge, the largest lift-span on the Eastern Seaboard. The north-south glacial grain and numerous inlets of southern New England made east-west travel by land very difficult. The broad reach of the Quinnipiac River, just south of its confluence with the Mill, was not spanned until the opening of a toll bridge in 1797.
0.1	Old Yale Boathouse slated for demolition with the reconstruction of the I-95 bridge to our left. Entering what used to be the community of Waterside about which a Victorian writer enthused, "The beauty of the scene cannot be described. The light hangs in a lovely radiance over the long reaches of the salt meadows..." (Brown, 1976). Change came quickly in 1900 when Standard Oil and American Steel and Wire built big facilities on the site.
0.2	Jehiel Forbes House, built 1767, on the right. This is one of the few colonial houses remaining in New Haven and the only one to have been built of stone, presumably of locally quarried New Haven Arkose. This area was the site of Jehiel Forbes's boat yard, which used timber floated down the Quinnipiac and Mill rivers. The house was shelled by the British in 1779 and a cannonball hole is reputedly still to be seen in one of its gables. The family got their household items, including a baby in a brass cauldron, out just before the British troops arrived. With the building of the adjacent Church of the Epiphany, through a Forbes Family bequest in 1904, the house became a rectory and saw the addition of the triple windows, small porch, and slate roof.
0.1	Leave artificial fill and traverse New Haven outwash valley-fill.
0.3	Diabase dike and New Haven Arkose outcrop as the road begins its ascent up the east wall of the Quinnipiac valley. This slope, though insignificant in our travels today, was another in the series of glacially accentuated obstacles to east-west wagon and foot travel in southern New England before the advent of the automobile and large earth-moving machines. This effect can be seen by looking at the route of the main line of the old New Haven Railroad whose mid-Nineteenth Century course goes on a six-mile detour up the Quinnipiac River before it crosses the old Boston Post Road two and one-half miles east of New Haven station.
1.2 X	Turn right on Townsend Ave.



GEOLOGIC MAP OF THE CENTRAL LOWLANDS (modified from Rodgers, 1985)

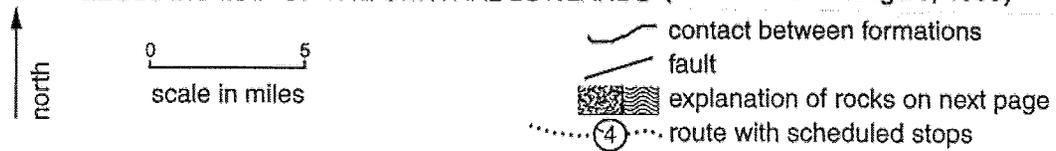


Fig. 1. Geological map of south central Connecticut.

- 1.1 On the left at 709 Townsend Avenue, Raynham. The original Federalist house was built in 1804 and rebuilt in Cottage Gothic style in 1856. Originally fields of the estate sloped down to the shore of the estuary on the right. The Townsends, New Haven merchants and bankers, have special significance to geologists as some of the original investors in Colonel Drake's Pennsylvania Rock Oil Company, which drilled the first commercial oil well in 1859 at Titusville, Pennsylvania.
- 1.1 X Turn right on Lighthouse Rd.
- 0.2 On left at 325 Lighthouse Road, the Thomas Morris House, started in 1680. Burned by the British on July 5, 1779. A rare example of one of the original colonial farmsteads in the region. Continue on Lighthouse Road, go through entrance gate to Lighthouse Point Park and bear right for .2 mi.
- 0.6 Parking lot

**STOP 1 Lighthouse Point - Eastern Highlands, New Haven. (14 MINUTES)**

This stop lies on the Avalonian terrain of the Eastern Highlands that forms the eastern margin of the Hartford Basin. The glacially sculpted bedrock here is the Lighthouse Point Granite Gneiss of probable Proterozoic age. This is composed of orthoclase and sodium-rich plagioclase feldspars, quartz, biotite, and magnetite, with locally abundant muscovite. Ordinarily this unit is well foliated but extensive shearing of the rock as a result of its proximity to the Eastern Border, or Great, Fault obscures this here. This gneiss is one of a series of granitic plutons associated with the so-called Avalonian continent that extend from here eastward through southern New England. Glacial erosion and the subsequent weathering of the siliceous rocks of the Eastern Highlands produces a hilly landscape veneered by thin, acid soils that did not support rich farms in pre-industrial New England.

To our northwest is the New Haven Harbor estuary with the New Haven skyline in the distance. This is one of the few large harbors on the southern New England coast. This combined with level, relatively rich farmlands developed on the deep, nearly circumneutral soils weathered from the Triassic and Jurassic sedimentary and igneous rocks of the Hartford Basin has made New Haven the largest port city between New York and Boston. Even today

it is the north-south axis of the Hartford basin, and not the Connecticut River that controls inland access. Thus New Haven remains the principal oil and bulk commodity-landing point for all of western New England.

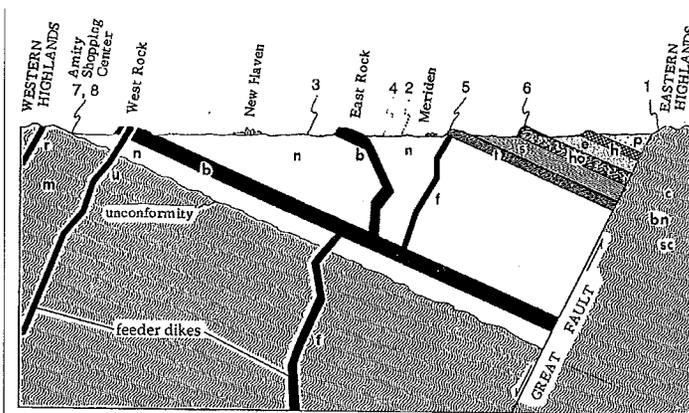


Fig. 2. Simplified west-east geologic cross section of Hartford Basin with relative stratigraphic position of numbered stops.

Fig. 3. Explanation of symbols used on figures 1 and 2.

Era	Period	Rock Units	Geography
Mesozoic	Jurassic	Portland formation (sedimentary; sandstones, shales) <b>p</b> Hampden lava (igneous; extrusive) <b>h</b> <b>C</b> Bridgeport dike (intrusive) East Berlin formation <b>e</b> Holyoke lava <b>ho</b> <b>U</b> <b>w</b> Buttress-Ware dike Shuttle Meadow formation <b>s</b> Talcott lava <b>t</b> <b>f</b> <b>g</b> Fair Haven-Higganum dike = <b>b</b> West Rock intrusion: sills, dikes, and stocks	Central Lowlands
	Triassic	<b>n</b> New Haven formation	
Paleozoic		<b>m</b> Maltby Lakes metavolcanics (metamorphic)	Western Highlands
		<b>c</b> Collins Hill schist (metamorphic)	Eastern Highlands
		<b>sc</b> Stony Creek granite (igneous; intrusive)	
		<b>bn</b> Branford gneiss (metamorphic)	

The Eastern Border Fault runs nearly east-west less than a quarter-mile to the north (Figs. 1 and 4) and glacial erosion and subsequent weathering have produced the relatively low lying terrain of the Hartford Basin. Although broad, the estuary is relatively shallow and must be dredged to maintain the ship channel. However, seismic work shows the existence of a now sediment-filled bedrock channel scoured to a depth of some 350 feet on the axis of the estuary. This was the result of lowered sea levels and glacial activity during the Pleistocene. The prominent flat-topped ridges lying behind the city are West Rock and East Rock and consist of sills of Jurassic age diabase. These have prominent seaward-facing cliffs that weather to a reddish orange and were responsible for the name "Rodeberg" or Red Hills given to the region by its European discoverer, Adrian Block in 1614.

- 0.2 Leave parking lot, return to Lighthouse Road, and make a right turn.  
 0.6 X On Lighthouse Rd., cross the nearly east-west-running Eastern Border Fault (covered) between Morris house (now on the right) and intersection with Townsend Avenue. Turn left on Townsend Ave., which runs north through Morris Cove, developed in the mid-Nineteenth Century as a summer cottage resort.  
 .8 Turn left on Fort Hale Park Rd.  
 .3 Turn left into entrance to Fort Nathan Hale and park in lot.

## STOP 2 Fort Nathan Hale, New Haven (20 MINUTES)

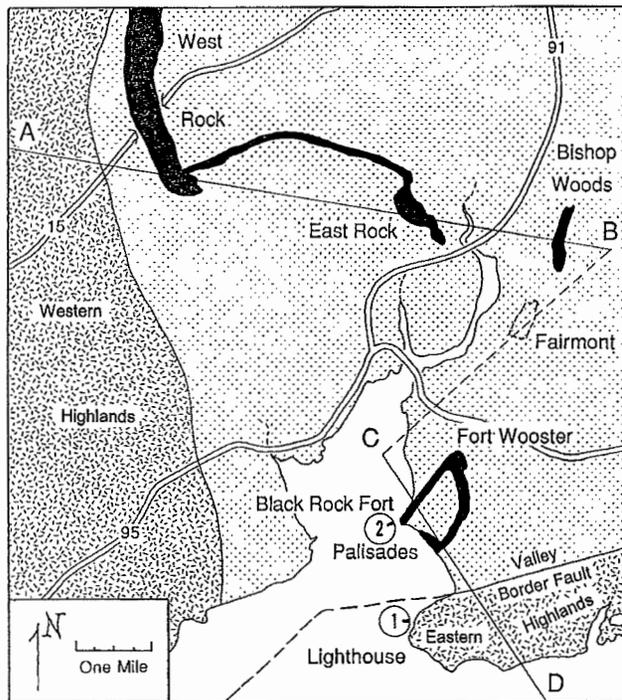


Fig. 4. Map of New Haven Harbor and surrounding diabase hills.

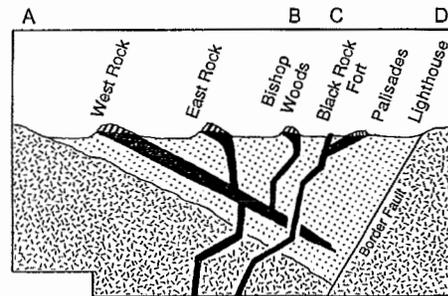


Fig. 5. Simplified geologic cross section Along line A-B-C of figure 4.

Even though much modified by human activity, a remnant of the original shoreline topography and sedimentation can be observed at this site. The parking lot lies on fill covering a former tidal salt marsh. Before human modification, salt marshes covered approximately 43 square miles of Connecticut's 98-mile shoreline. East of the parking lot the low slope lies at the inland limit of the coastal storm surge. Shoreward of the parking lot a moat and drawbridge occupy what might represent remnant tidal creeks traversing the former salt marsh, but outlets of the channels are gated and the water held at an artificially high level. Because only exceptionally high tides reach the area, the marsh is only slightly brackish, as evidenced by extensive stands of the tall grass called *Phragmites*.

Shoreward of the old tide marsh, the land rises to what was one a belt of sand dunes between the salt marsh and the beach. However, the dunes have been largely replaced by sand-banked bunkers of the second Fort Nathan Hale built in 1863. In the 1850's this was also the site of a copper smelter and saltpeter factory. Shiny black slag from the smelter is still to be found on the beach. Although this is a fill beach made up of poorly sorted sand, grit, and gravel, it does show a moderately well-developed berm marking the mean high tide line. Above it is a terraced backshore with the beginnings of aeolian dunes marking its upper margin.

The knob of dark grey rock forming the point is a diabase dike upon which stands the reconstructed Black Rock Fort, built in 1776 on the site of earlier fortifications begun in the 1680's and captured by the British on July 5, 1779. Most of the low hills to the north of this point are upheld by an interconnected system of early Jurassic dikes and sills (Figs. 4-5). In addition to serving as strong points, the diabase hills in this region were also observation and signaling posts. In the early Seventeenth Century, smoke signals from Fort Wooster Hill, the highest point along New Haven harbor's east shore, were a signal to passing Dutch ships that the Quinnipiac Indians wanted to trade.

- Go back to park entrance and turn left on Woodward Ave. (unmarked)
- 1.4 X Turn left on Main St. and immediately merge right onto I-95 south toward New Haven.
- 0.8 East Rock Sill with columnar jointing viewed from the bridge over the Quinnipiac River.
- 0.5 Merge right and Exit onto I-91 north.
- 0.5 Crossing of the Mill River. The Mill flows south on the west side of East Rock on our left. A huge surge of outwash down this stream built a large delta into the Quinnipiac valley blocking its flow and impounding Glacial Lake Quinnipiac behind it.
- 0.1 I-91 rises gently into the Fair Haven section of New Haven, which is built on the old Mill River delta.
- 1.2 Crossing the Quinnipiac River and driving along the east side of Glacial Lake Quinnipiac. This lake formed between 15,000 and 18,000 years ago and began to fill with varved clays. A series of pits located in this marshland produced most of the bricks and refractory clay for New Haven. By 2,800 years ago, a river flood plain supporting a rich, mixed coniferous and deciduous forest had replaced the lake. Subsequently, brackish and salt-water swamps followed the incursion of marine water up the course of the Quinnipiac as sea level rose.
- 1.3 View to the north of Sleeping Giant, a diabase stock.
- 1.2 Take Exit 10 and continue straight ahead on Ct. Route 40
- 2.3 Go under bridge and stop on right side immediately after first large metal light pole.

### **STOP 3 New Haven Arkose on Ct. Route 40, North Haven (20 MINUTES)**

This is an unusually large exposure of the New Haven Arkose that was cut in 1977 when Route 40 was built to connect Route 10 on the west with I-91 on the east (Fig. 6). It is yet another example of the difficulty and expense in making east-west overland connections across the bedrock grain of southern New England. The exposure on the north side of the highway is especially good, with approximately 72 m of section. After correcting for fault-duplication, the New Haven Arkose is between 1950 and 2250 m thick in the area between this cut and New Haven. Stratigraphically this outcrop lies just below the middle of the formation. The objective at this stop will be to briefly examine and interpret the lithology and sedimentology of this sequence.

The outcrop consists of broadly lenticular bodies of arkosic sandstone and conglomerate interbedded with, and often down cutting into red, sandy mudstone. The sandstone and conglomerate lenses are interpreted as channel bodies of a braided stream complex that traversed a floodplain represented by the mudstone. Trend of the channels is 256° southwestward. In terms of modern facies models this outcrop is interpreted as lying somewhere between the Donjek Type (distal alluvial fan) and Platte Type (moderate-energy braided stream).

The tops of many of the mudstone units often show the development of peds. These are vertical plains that cause the rock to break into polygons on the order of 1-2.5 mm in size and give the rock a minutely faceted appearance sometimes called "hobnail structure". They are the result of the realignment of clay crystals into a



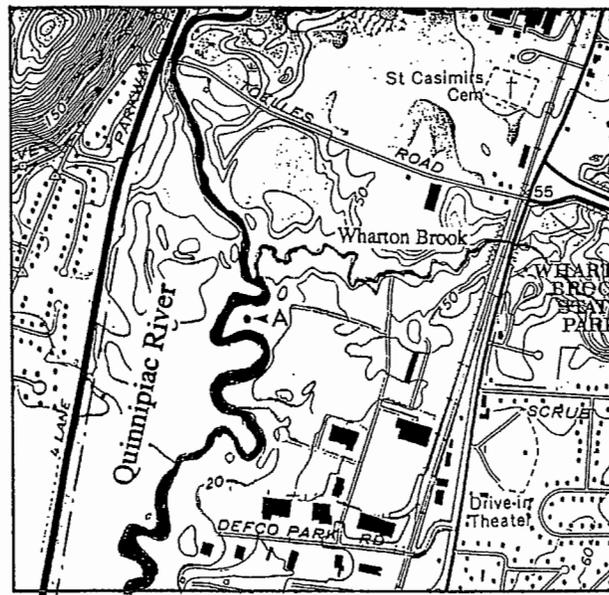
Fig. 6. Cut in New Haven Arkose along Ct. Rt. 40 looking north from the Ridge Road Bridge. Photo by William K. Sacco.

vertical plain during soil forming processes, such as biturbation by roots. Greenish grey caliche is also a prominent feature of the interchannel mudstone at this outcrop. This is present as scattered nodules that are sometimes densely spaced enough to form pavements and as rhizomorphs or root fillings. These branch downward into the interchannel sediments, sometimes to a depth of a meter. Several instances can be seen where rhizomorphs are truncated by channel scouring and where channel lenses scour down to a pavement of caliche nodules.

Rocks at this locality are interpreted as having been deposited in a rift valley under tropical, semi-arid, to moderately arid climatic conditions about 12-15 degrees paleolatitude north of the equator.

- Continue to end of Rt. 40
- 0.8 X Turn left on Rt. 10, turn around in church parking lot on the left, immediately beyond the underpass, and get back on Rt. 40 South.
- 0.9 On the left, good overall view of repetitive sedimentary packages at Stop 3.
- 1.8 Jct. I-91, proceed North on I-91.
- 2.3 X Take Exit 12 and turn right on Rt. 5
- 0.3 Rest Stop at Exxon Station (13 minutes. Achtung--Pissen Sie schnell, bitte!)
- 1.3 X Turn left on Defco Park Rd.
- 0.2 Turn right on Dodge Ave.
- 0.4 Park in lot on left at end of Dodge Ave.

**STOP 4 Wharton Brook and the Quinnipiac River, North Haven (80 MINUTES)**



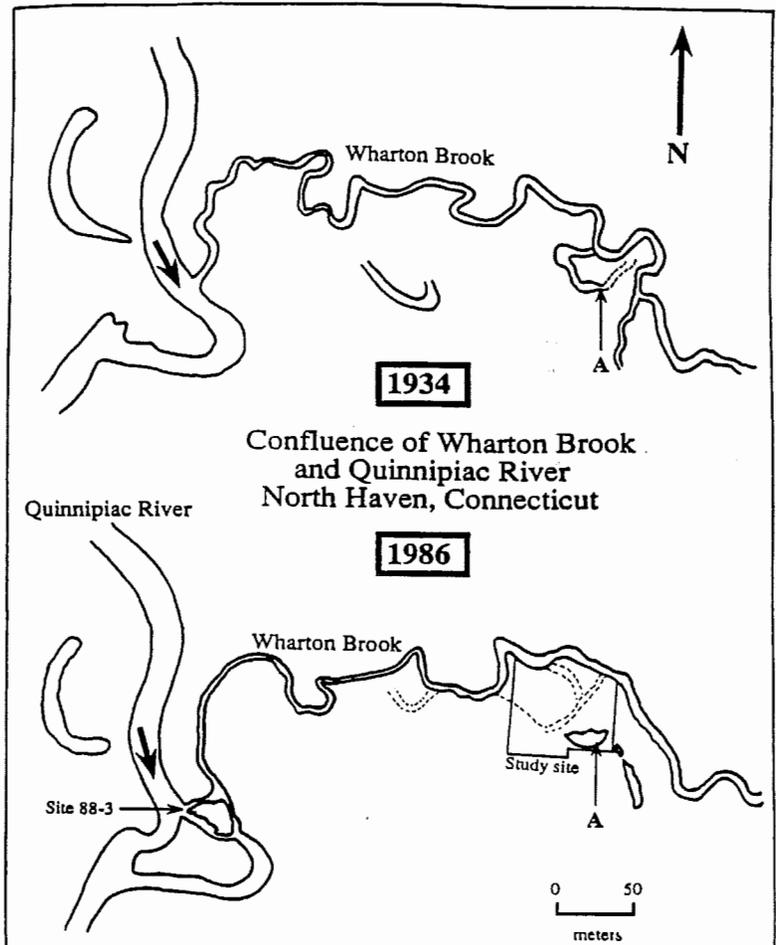
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Fig. 7. Location map for Stop 4 at Wharton Brook and the Quinnipiac River showing the topography.

This stop consists of a traverse along Wharton Brook to its confluence with the Quinnipiac River. The site is located in the meander-belt of the Quinnipiac, on a flood plain developed by the down cutting of the river into late Pleistocene and early Holocene sediments associated with Glacial Lake Quinnipiac. Two kilometers downstream the river enters tidal marsh. This location, together with the general rise in sea level over the last 5,000 years, probably means that this reach of river is at grade or even slightly aggradational. Plant remains encountered 0.35 m below the surface at point A on the map (Fig. 7) give a date of  $1570 \pm 80$  years BP.

Fig. 8. Comparison of Wharton Brook and the Quinnipiac river drawn from aerial photographs taken in 1934 and 1986. Point A is the same point on each map and can be used for orienting an overlay. Note also the location of site 88-3 at the mouth of Wharton Book. (From Johnson, 1989)

The Quinnipiac River runs some 40 miles from its source in Dead Wood Swamp, just north of the town line in Farmington, Connecticut, to New Haven harbor. The name comes from the Algonquin word meaning "Long Water Place". It was formed when outwash deposits left by the receding glacier forced a shift in the course of the ancestral Farmington River. These deposits cut the Farmington into two segments, with the upper reach reversed to flow north into the Connecticut River and the lower one becoming the Quinnipiac. Despite its short



course, the Quinnipiac receives effluent from five municipal, and two industrial sewage treatment plants and has been honored by the Environmental Protection Agency as one of the most polluted rivers in the United States.

Our traverse starts on the floodplain of Wharton Brook, a second-order drainage whose valley contains an excellent example of a lowland sugar maple forest. An examination of aerial photographs made 42 years apart shows a marked straightening of the meander course of Wharton Brook (Fig. 8), probably reflecting the transition of its drainage area from rural to suburban conditions. Among the features of interest on the Wharton Brook floodplain is a cut-off meander at Point A of Figure 8. The anaerobic sediments of the pond found in it contain a record of the forest vegetation of the immediate area since its formation around 1930. Another feature of note is the relict eastern plane tree (*Platanus occidentalis*) on the point bar of the old meander. This species cannot reproduce in a shady forest but must germinate on open, sandy soil, within a day or two of being shed.

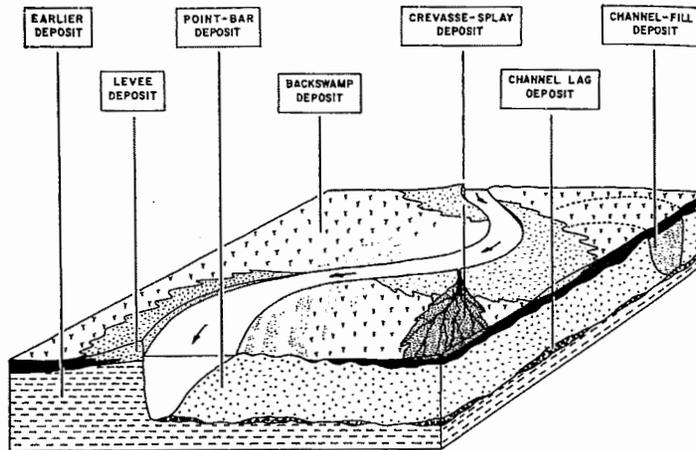


Fig. 9. Topography and lithofacies relationships in the floodplain deposits of a meandering stream illustrated by an idealized block diagram. (From Allen, 1964, reprinted by permission.)

The traverse then heads southwestward from Study Site A to the inner curve of the partially cut off meander of the Quinnipiac just downstream from Site 88-3 (Fig. 8). (For an idealized view of the riparian sedimentation see Figure 9.) As the traverse leaves the floodplain of Wharton Brook it rises to the second of two old terrace levels of the Quinnipiac. Each of these supports a distinctive association of trees. In contrast to the book floodplain, dominant trees on the more dynamic floodplain of the river are elms, white maple, box elder, ash, and sycamore.

South of Site 88-3 a meander cutoff formed sometime between 1974 and 1982. Note how strong sedimentation on the river channel has forced the mouth of Wharton Brook south of its 1934 position (Fig. 8). Diversion of some of the brook's water into the old meander has curtailed active sedimentation in its reach, although the river has built partial levees against its upper and lower ends. Note also how the dense root mat in the sandy sediment of the chute cutoff and in the banks of an island that was once the neck of the meander resists the force of river erosion.

Finally, in the cut bank of the old meander, approximately two meters of red varved clay is exposed at the base of the section. The varves are ripple bedded and the clay represents the fill of Glacial Lake Quinnipiac. Above this occurs one meter of sand and gravel representing the delta of the Muddy River, which built out over the lake sediments. The top quarter-meter is the current forest soil.

- Lunch break (27 minutes)
- Go back out Dodge Rd.
- 0.4 Turn left on Defco Park Rd.
- 0.2 Turn left on Rt. 5
- 0.4 Last stand of the pitch pine (*Pinus rigida*) on the Quinnipiac sand plain. In colonial times this tree was a major source of naval stores like pitch and turpentine.
- 0.2 X Turn right to Entrance to I-91 at Toelles Rd.
- 0.6 Go north on I-91

- 8.2 Stay right and merge onto Ct. Route 691. Stay right again and go west on Rt. 691. Then stay in middle lane.
- 3.3 X Take Exit 6 and turn left on Lewis Ave.
- 0.2 X Turn left on Kensington Ave.
- 0.6 X Turn left on Rt. 71
- 0.1 Turn right on Cold Spring Ave. at marked entrance to Target - Then go left in front of Target building
- 0.2 Park in lot along left side of building

**STOP 5 Talcott Basalt, Hanging Hills, Meriden (28 MINUTES)**

A superb exposure of the Talcott Basalt is located in a cut made for the parking lot behind the Target store at the west end of Cold Spring Avenue (Figs. 10 and 11). The Talcott is the lower of the two great basalt flows that form the Hanging Hills (Fig. 1). The sheer cliff of the upper two-thirds of the hills is the outcrop of the Holyoke basalt with the Shuttle Meadow Formation forming a swale between it and the lower terrace-forming Talcott. This formation lies in turn on the New Haven Arkose (Fig. 3).

The Triassic-Jurassic boundary is located in the uppermost New Haven Arkose within a meter of its contact. The Talcott represents the beginning of basaltic lava flows in the Hartford Basin. Interestingly, such volcanic activity, involving tholeiitic basalts of remarkably similar composition and vesicularity, began within the same 21,000 year interval at the beginning of the Jurassic Period in the Eastern North American rift basins, from the Culpepper Basin in Virginia to the Fundy Basin in Nova Scotia, (Olsen and Gore, 1989, p. 11).



Fig. 10. Pillow basalt in the lower part of the Talcott Formation in the west wall of the Target parking lot (Stop 5) in Meriden, Connecticut. Photo by L. J. Hickey.

The Talcott Basalt is about 67 m thick in this area (Hanshaw, 1968, p. 2), but the upper part is not exposed in this cut. The Talcott consists of two flows separated by 0.5 to 0.8 m of volcanic agglomerate (Fig. 12). The Target cut exposes a continuous section of approximately 47 m of Talcott Basalt, including the entire lower flow, nearly a meter of agglomerate, and the lower 12 m of the upper flow. Pillow structures and pipe vesicles are especially well developed in the lower flow (Fig. 10), as are occasional clastic dikes.



Fig. 11. Contact between the New Haven Arkose and the Talcott Basalt in the south wall of the Target parking lot (Stop 5) in Meriden, Connecticut. Thin interbeds of sandstone and mudstone making up the upper New Haven can be seen at A in the lower corner of the photograph. The gently flexed contact between the units is located at B. The pillow of basalt at C is completely surrounded by red mudstone (D), some of which is contorted.

The basal contact of the basalt can be seen on the south wall of the cut, to the left of the store building (Fig. 11). The New Haven is a thin bedded to laminated sandstone and mudstone unit. Lack of soil development and the occurrence of lamination and occasional ripple bedding point to deposition in ponded-water conditions, an interpretation that is consistent with the development of pillows in the overlying basalt. The contact itself shows the dynamics of molten basalt entering water. Bedding in the upper 20 cm of the New Haven becomes folded and convoluted, with masses of mudstone and sandstone injected into the basalt, sometimes completely surrounding and isolating individual pillows, some of which have developed thick oxidation rinds. Clastic dikes were injected several meters into the pillow layer, and rotation of vesicle tracks in some of the pillows shows that they probably rolled during emplacement of the still incoherent mass of coagulating lava. Above the pillows are some 20 m of massive basalt that becomes increasingly vesicular toward the top.

A thin agglomerate layer resting on the lower flow records a pause in volcanic activity with reworking of the volcanic terrain. A zone of weathering at the top of the agglomerate appears to represent the remains of an incipient soil. This was overridden, in turn, by the upper flow, which vaporized the water in the soil and produced prominent vesicles in its basal half-meter. In some Talcott outcrops in the Meriden area these vesicles are inclined, possibly as the basalt flowed over the land surface. Indistinct pillows characterize the lower few meters of the upper flow, but these are hard to distinguish from the spheroidal weathering that is found higher in the upper flow.

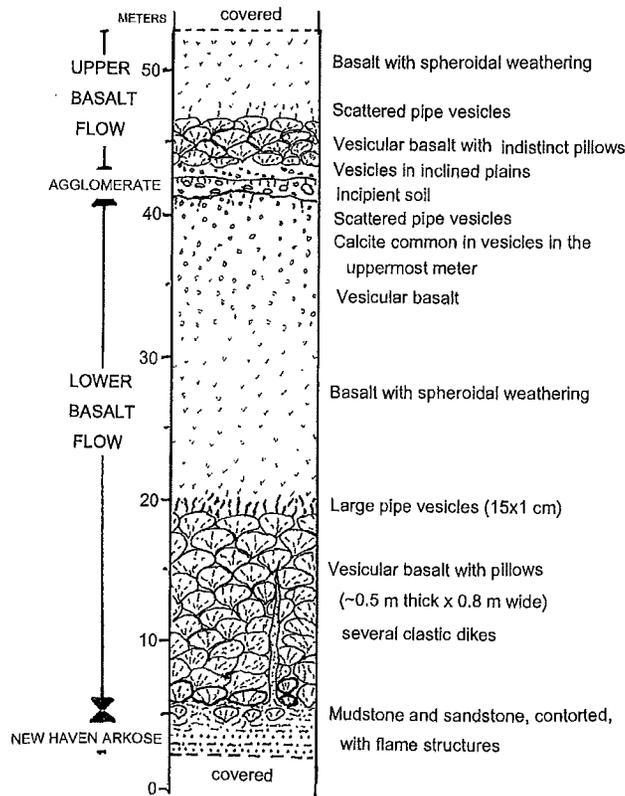


Fig. 12. Measured section of Talcott Basalt and uppermost New Haven Arkose behind Target store, Meriden

- 0.3 X Proceed back out to Rt. 71
- 0.9 X Turn right on Rt. 71
- 1.2 Turn right into Hubbard Park on Mirror Lake Drive (unmarked, but has one-way sign). Mirror lake has large fountain in center. The fields and recreation areas here are built on the terrace formed on the soft sediments of the Shuttle Meadow Formation (~100 m), a sequence of lake-margin sediments that lies between the Talcott and the Holyoke flows.
- 0.4 Turn left on Hubbard Park Dr. (unmarked) at "T" intersection - go through Castle Craig gate onto Reservoir Ave. (unmarked)
- 1.4 Turn left on West Peak Dr. (unmarked) over dam; the road now curves left to ascend to East Peak on to the dip slope of the Holyoke Basalt.
- 1.8 Park at Castle Craig parking lot on East Peak. The tower was dedicated in 1900 and was modeled after those built by the Turks along the Danube River in the 12<sup>th</sup> Century.

**STOP 6 Castle Craig and Holyoke Basalt, Meriden (20 MINUTES)**

Looking south from this vantage point on top of East Peak of the Hanging Hills in Meriden, one gets a clear view of the Central Lowlands with its basalt ridges, and the highlands marking the limits of the Hartford Basin to the west and east (Fig. 1). After deposition and tilting, the Central Lowlands rocks were subjected to normal faulting thereby repeating the stratigraphic sequence on the ground. In the Hanging Hills area there are more than 10 such normal faults resulting in a series of prominences each upheld by the Holyoke Basalt. Sleeping Giant stock to the south is one of the few instances of a large hill cutting west to east across the fabric of the lowlands that was imposed by early Mesozoic tilting and faulting, reinforced by Pleistocene glaciation.

Another Connecticut feature seen clearly from here is the so-called Uplands Peneplain of concordant summits (Bell, 1985) with its gentle slope southward of 10-25 feet per mile. The Coastal Slope Peneplain, harder to see from here, lies south of a line about 10-15 miles inland. South of this the seaward slope of the summit surface

increases abruptly to about 50 feet per mile. This break in slope is reflected in the visual difference in elevation from Castle Craig (elevation, 950 feet) for a distance of nine miles to Sleeping Giant (elevation, 700 feet), and then across the hinge line, seven miles south to East Rock (elevation, 350 feet). On a clear day East Rock is just visible to the left of the giant's feet. Three and one-half miles further south the 180-foot summit of Raynham Hill completes the ensemble.

The Holyoke Basalt is the second of three Hartford Basin tholeiitic basalt flows and dates from the early Jurassic. It is a massive unit, without the vesicles or amygdules of the lowest flow, except toward the top. The basal part has columnar jointing. Two or three brief pauses in flow activity can be seen in the upper third of the cliffs in this area. The upper surface of the Holyoke has been smoothed by glacial activity but subsequent weathering has raised veins of more resistant rock into positive relief.

The dry south- and west-facing slopes of the basaltic ridges of the Central Lowlands support a number of relict plants like cactus, yucca, prairie grasses, and blackjack oak that appear to have reached Connecticut from drier and warmer regions to the south and west during the Thermal Maximum, about 7,500-4,000 years ago (Thorson and Webb, 1992). Meriden began its history in Colonial days as a stopping point for foot and horseback traffic on the trail between New Haven and Hartford. However, the first wagon did not make it through the gap through the hills to the east of here until 1789.

- 3.6 Turn right on West Main St.
- 0.9 Jct. I-691, turn left onto 691 West. This segment of the route starts in the uppermost New Haven Arkose and runs down-section.
- 1.0 Outcrop of New Haven Arkose on right. Caliche is greatly reduced in the upper third of the formation as opposed to its abundance in the lower two-thirds. This reflects a climatic amelioration, probably related to the northward drift of the Hartford Basin beyond the horse latitudes during the late Triassic.
- 2.2 Jct. I-84, proceed straight west on 84. The slope to the west consists of metamorphic rocks of the Western Highlands.
- 1.1 Crossing the western boundary of the Hartford Basin, here a normal fault (covered). Begin a gradual rise into the highlands.
- 0.4 Shallow cut in the Beardsley Member of the Harrison Gneiss of Ordovician age.
- 1.5 X Take Exit 26 and turn right on Ct. Route 70.
- 0.4 Turn sharp right on Summit Rd. This appropriately named road runs on covered gneiss and schist of the Taine Mountain Formation (early Ordovician) that forms a crest along the up thrown (western) side of a normal fault.
- 3.1 X Turn left on Ct. Route 69 unmarked).
- 0.2 Pit Stop (13 minutes). (Pinkeln Sie sehr schnell, bitte!)
- 4.0 Foliated gneiss of the Pumpkin Ground Member of the Harrison Gneiss (middle? Ordovician) on the right.
- 0.7 View of West Rock sill to left.
- 0.2 Cut in schist of the Trap Falls Formation (early to middle Ordovician).
- 3.1 At this point the route begins a long descent diagonally down the eastward-tilted angular unconformity surface that lies beneath the Mesozoic sediments of the Hartford Basin to the east.
- 0.3 A long outcrop of the highly graphitic schist and phyllite of the Wepawaug Formation of Devonian and/or Silurian age occurs on the right side of the road.
- 0.6 North end of Dawson Lake and crossing the covered contact between the metamorphics and the New Haven Arkose of the Hartford Basin. West Rock sill is on the left.
- 1.3 The cornfield on the right is an indication of the relative richness of the soils of the Central Lowlands that made them the chief crop-farming region of the state.
- 3.7 X Turn right on Lucy St.
- 0.2 X Turn left on Ct. Route 63 (Amity Rd.)
- 0.1 Turn right on June St.
- 0.1 Stop in lot at end of June St.

**STOP 7 Basal Unconformity of New Haven Arkose, Woodbridge (10 MINUTES)**

Although the next two stops are trash-littered, overgrown, man-made cuts, they reveal much about the geologic history of Connecticut (Fig. 13). This June Street exposure lies on the western margin of the Hartford Basin and is located behind the Amity Shopping Center on the west side of Ct. Route 63. Here thin remnants of the basal-most beds of the 2000-meter-thick New Haven Arkose rest in depressions on the eastward-dipping erosion surface of the Maltby Lakes Metavolcanics of Ordovician age. The Triassic rocks are a coarse, pebbly conglomerate. The metavolcanics are foliated but original bedding, consisting of alternating layers of quartzite and schist rock, can be observed in many places.

Glacial striations are present on the unconformity surface (Fig. 13) and clearly show the southward movement of the ice sheet responsible for scouring out the Mesozoic valley-fill.

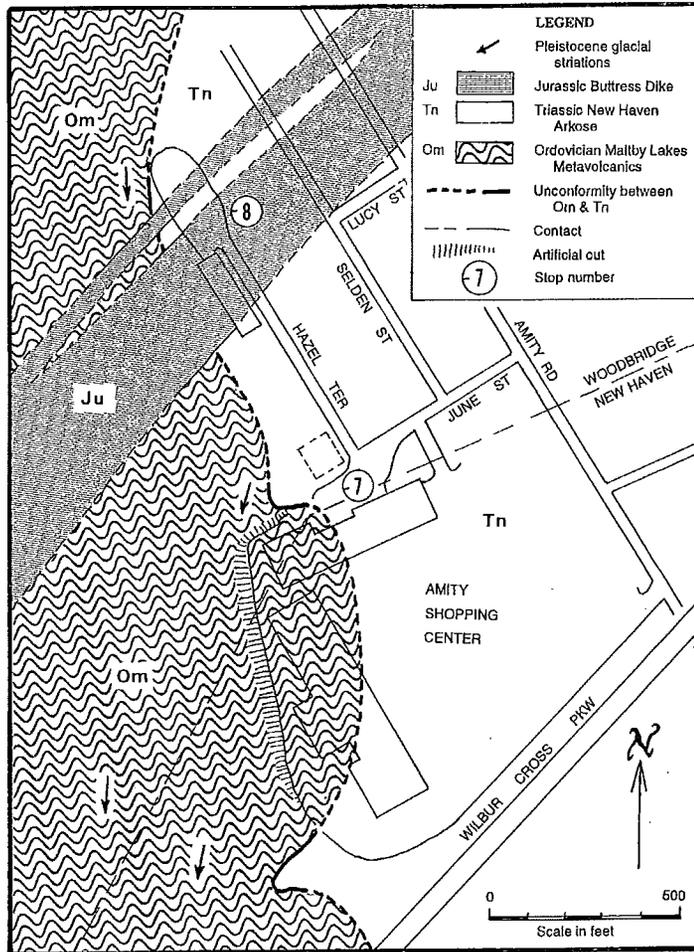


Fig. 13. Geologic map of the June Street-Hazel Terrace area (Stops 7 and 8), Woodbridge.

0.3 Walk back on June St. and turn left on Hazel St.  
Stop in lot at end of Hazel St.

**Stop 8 Hazel Street Buttress Diabase Dike, Woodbridge (18 MINUTES)**

At this one stop there is a rare, if not unique, opportunity to see the three major rock types of Connecticut in one 100-meter-long outcrop -- in this case the Ordovician Maltby Lakes Metavolcanics, the Triassic New Haven Arkose, and the Jurassic Buttress Diabase (Fig. 13). In addition, each of these rock types is polished by the last glacial push across the area. From north to south along this cut the following features can be seen. The exposure furthest north has a small patch of New Haven Arkose resting on the old erosion surface, which lies at a steeper angle here than at Stop 7. This indicates considerable relief on the unconformity surface. Proceeding southward, one sees in succession, the Maltby Lakes Metavolcanics, a thin slice of the Buttress Diabase dike, about 10 meters of New Haven, the main mass of the dike, and the continuation of the New Haven behind the south end of a large building. The irregular contact between the thin slice of the Buttress dike and the New Haven Arkose contains several small clastic dikes.

To the east, cutting the West Rock sill, the Buttress Diabase is exposed at its type locality. It is best seen in winter after a thin "shadow casting" of snow (Fig. 14).



Fig. 14. Photo, looking east from Hazel Terrace of the Buttress Diabase dike cutting the West Rock Sill. This fancied "flying buttress" is the type section for the unit. Photo by C. MacClintock.

- Go back on Hazel St.
- 0.3 Turn left on June St.
- 0.1 X Turn left on Rt. 63
- 0.1 Turn right on Lucy St.
- 0.2 Turn right on Rt. 63
- 0.2 X Turn left on Pond Lilly Rd.
- 0.5 Turn right on Valley Rd. West Rock Sill to the left.
- 1.2 X Turn left on Blake St.
- 0.7 Turn left on Osborne Ave.
- 0.1 Turn right on Goffe Ter. (which becomes Goffe St.)
- 0.6 X Turn left on Sherman Ave.
- 0.1 Turn right on Henry St.

- 0.8 X Turn right on Prospect Ave. and make an immediate left on Edwards St.  
0.3 Turn right on Whitney Ave.  
0.1 Turn right into KGL parking lot  
KGL

**End of field trip.**

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## HARTFORD BASIN CROSS SECTION - SOUTHLINGTON TO PORTLAND CT

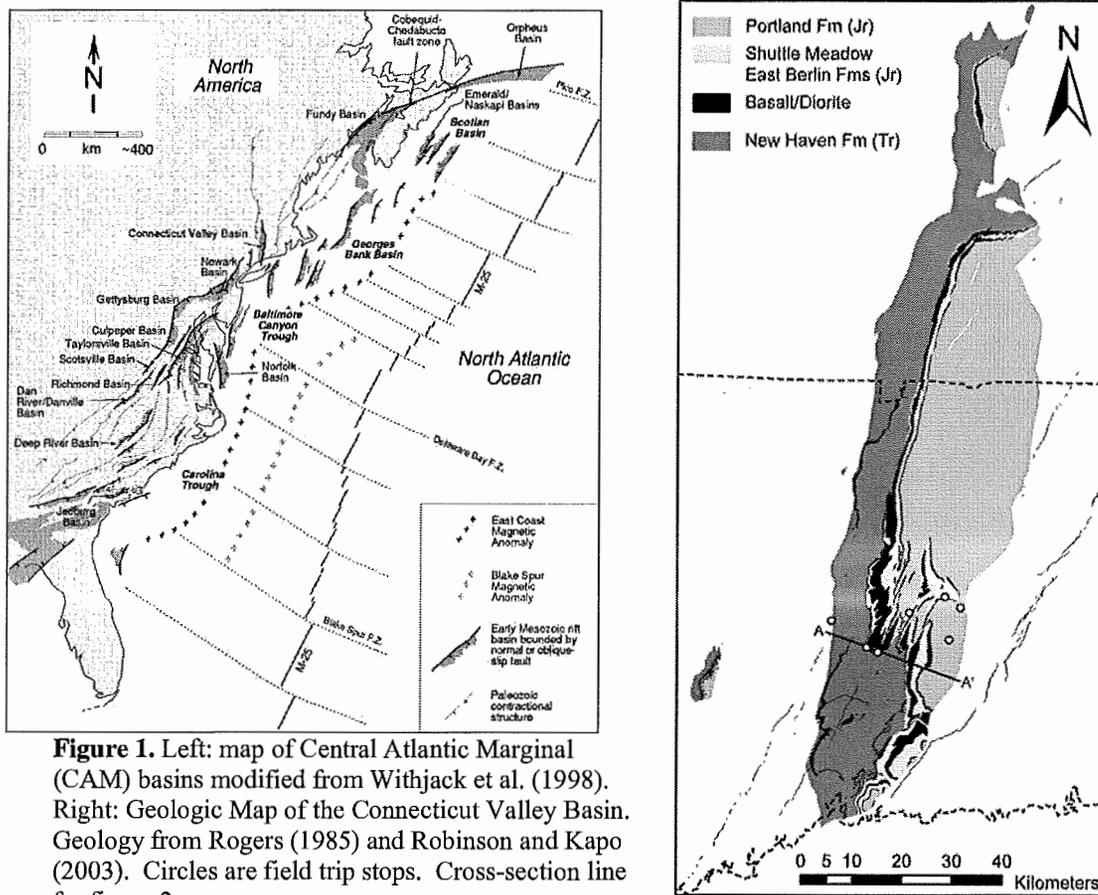
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### THE HARTFORD BASIN

#### Introduction

The Hartford Basin is one of the more intensively studied of the Central Atlantic Marginal basins (CAM) formed during the breakup of the Pangean supercontinent (Fig. 1, left). The basin has received much attention due in part to the relatively high quality exposures of Late Triassic through Early Jurassic sedimentary and igneous rocks. On this trip we will visit what we feel are some of the most spectacular exposures in south-central Connecticut while touching on many topics related to the development of the basin.

The geology of the basin was first mentioned by the earliest European explorers; however modern scientific investigations of the basin began in earnest in the 19<sup>th</sup> century (McDonald, 1996). At this time Silliman, Hitchcock and Dana began to develop and test some of the fundamental theories of modern geology using the rocks of the basin (McDonald, 1996). Early mapping by Percival (Percival, 1842) and Davis (1898) defined the structural geometry of the basin and provided the foundation for much subsequent work. We refer interested readers to MacDonald (1996) for a more extensive review of this period as well as a valuable bibliography of all works from 1681-1996. Interest in the basin continued into the early 20<sup>th</sup> century with insightful papers by Barrell (1915), Longwell (1922), Russel (1922), Wheeler (1939).



The plate tectonic revolution and oil and gas industry interest lead to a resurgence in research in the basin in the late 20<sup>th</sup> century that continues to this day. The basin serves as a natural laboratory for understanding the Late Triassic Early Jurassic break up of Pangea (e.g. de Boer and Clifton [Clifford], 1988; Schlische, 1993) as well as fundamental processes of rift-related sedimentation (e.g. Olsen, 1986), magmatism (e.g. Philpotts, 1992), and structure (e.g. Wise, 1992). Many of the more recent results have been summarized in two sets of edited volumes on the CAM rift basins (LeTourneau and Olsen, 2003; Manspeizer, 1988).

In this guide we provide brief overviews of the stratigraphy, magmatism, structure, and timing of the Hartford basin. We focus primarily on topics that we believe remain unresolved or controversial and may thus be the topics of future research in the basin. The five stops and 2 additional sites described in the guide present an opportunity to investigate and discuss these topics in the field.

### **Stratigraphy**

The sediments of the southern Hartford basin are subdivided into four formations separated from one another by basalt flows (Fig. 1, right). The stratigraphically lowest formation is the New Haven which consists of red to buff conglomerate, sandstone, and mudstone that unconformably overlies the Paleozoic metamorphic basement (Olsen, 1997, and references therein) and locally reaches thicknesses of up to ~2000m. These rocks are interpreted to have been deposited primarily in a fluvial environment (Hubert et al., 1978) between ~218 and 202 Ma. The New Haven formation is overlain by the Talcott basalt flow (~75-m thick), Shuttle Meadow formation (~100-m), Holyoke Basalt flow (~200-m), East Berlin formation (~170-m), and the Hampden Basalt flow (~50-m). The sedimentary units in this sequence consist of interbedded lacustrine and fluvial sequences made up of gray to black mudstone, sandstone, and dolomite, and red mudstone and sandstone, respectively. This entire sequence was deposited between ~202 and ~196 years (or less, see discussion in section on magmatism below). Above the Hampden Basalt lies the Portland Formation (~2000-m preserved). The Portland Formation ranges in grain size from mudstone to coarse conglomerate with the largest clasts exceeding 1-m in length (LeTourneau, 1987). The sediments in the Portland Formation coarsen toward the eastern border fault and are interpreted as lacustrine, fluvial, and debris flow deposits.

Van Houten (1962) first noted a pronounced cyclicity in the sediments of the early Mesozoic Newark Supergroup which has been further described and quantified by Olsen and coworkers (see Olsen, 1997 and references therein and contribution A4 this volume). Where lacustrine beds are present this cyclicity is marked by a three stage sequence 1) lake transgression indicated by calcareous siltstones, 2) lake high stand indicated by thinly laminated calcareous claystone and siltstone with high organic content, and 3) lake regression to lowstand deposits indicated by abundant desiccation cracks (Olsen, 1986). In the Newark basin this cyclicity has periods of 5.9, 10.5, 25.2, 32, and 96 m corresponding to time periods of ~25 ky, ~44 ky, ~100 ky, ~133 ky and 400 ky (Olsen, 1986, 1997). These periodicities correspond closely with Milankovitch cycles of ~20 ky, ~40 ky, ~100 ky, 413 ky and suggest that the cyclicity is strongly controlled by orbital parameters and associated climatic variation (Olsen, 1986, 1997).

The work of Olsen and coworkers has illustrated the strong climactic control on sedimentation, however, the tectonic record recorded in the sediments remains largely untapped. Clasts of the Portland Formation conglomerates record a footwall unroofing sequence including low-grade metamorphic rocks which no longer outcrop in the nearby eastern highlands. Sedimentary composition may also provide the key to answering other ongoing questions such as the original extent of the basalts (see discussion below). Furthermore, careful observations of sediment thickness may help to unravel the timing of various structures within the basin.

### **Magmatism**

The intrusive and extrusive rocks of the basin and surrounding highlands appear to have formed in three distinct magmatic events ca. 200 Ma. Philpotts (1992) correlates the Talcott, Holyoke, and Hampden flows with the Higganum, Buttress, and Bridgeport dikes (from oldest to youngest). These correlations are based on the bulk chemistry and petrography of the flows and dikes. In addition the extensive intrusive

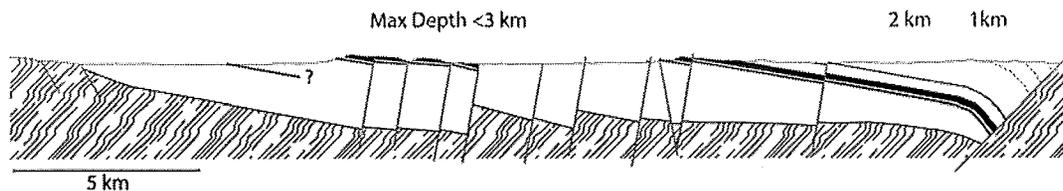
sills and laccoliths of West Rock, the Barndoor Hills, and the Sleeping Giant have all been correlated with the Talcott event. Based on the cyclicity of sedimentation mentioned above Philpotts (1992) suggests that ~138 ky and ~345 ky separate the Holyoke event from the Talcott and Hamden events, respectively. Paleomagnetic data, however suggest a longer time span between dikes and flows with the Buttress and Ware dikes possibly postdating the flows and intruding at ~175 Ma (de Boer, 1968; McEnroe and Brown, 2000). Although Philpotts' dike-flow model fits the geochemical data well it is not consistent with all the available data for instance anisotropy of magnetic susceptibility data (de Boer et al., In Press; Lindsey, 1995) suggests the presence of multiple feeders. The southern Talcott was likely fed by the Fairhaven dike, but the northern Talcott appears to have been fed by a source from the western basin, most likely associated with the Barndoor Hills.

The extent of the lava flows has also been a topic of debate. McHone (1996) has suggested that the flows may have extended across much of eastern North America rather than being confined to the preserved basins. Huber, however, cites evidence from the Pomperaug basin suggesting that the flow histories are not correlative across the intervening highlands (Huber, 1997). Detailed investigations of the composition of intervening and overlying sediments may be able to help resolve this debate as well as further study of the Pomperaug Basin (See Contribution C3, this volume).

The extent of the sills along the western margin of the basin is also enigmatic. Geophysical evidence indicates that the West Rock – Barndoor sill complex is continuous in the subsurface. The sills thus form a ~60 km long belt of intrusions that intruded at <~1 km depth (bedding perpendicular distance to the Talcott flow on the Southington/Meriden Quadrangles). Individual sills of this extent at such shallow depth are considered unlikely based on our understanding of the mechanics of sill intrusion (Jackson and Pollard, 1988). The sills are thus likely composite intrusions.

### Structure

The gross structure of the Hartford Basin is an asymmetric graben or half graben with a master west-dipping normal fault on the eastern border and a smaller (?) east-dipping fault on the western border (Fig. 2). The beds within the graben are generally tilted to the east with dips typically exceeding 5°, however in detail the dips form distinct domains, inconsistent with simple tilting in a half-graben (Wise, 1992). Bedding dips increase near the margins of the basin creating what have been interpreted as normal and reverse drag along the western and eastern margins, respectively (Wise, 1992). Geophysical studies indicate that the basin is not a typical half-graben deepening to the east, but is rather more of a bathtub shape with the deepest portions near the center of the basin (Wenk, 1984). How this top of basement geometry is achieved while honoring the observed surface geometry is at present a matter of interpretation. Proposed models include a series of abandoned normal faults creating “rider blocks” – slivers of basement that are now in the hanging wall of the master normal fault (see Schlische, 2003, Fig. 4.2 section D-D’) or an originally gently westward dipping basement surface below the New Haven Formation that was subsequently faulted during continued rifting (see Wise, 1992, fig. 2, section C-C’). Additional geophysical studies will be needed to differentiate between these models and better image the basin geometry.



**Figure 2.** Sketch cross section across the Hartford Basin. See figure 1 A-A' for location. Wavy fill pattern is Paleozoic basement. Bold black lines are basalts and intrusive diorites (at western end). Basin depths are derived from Wenck (1984). Note cumulative throw across intrabasinal faults is ~1.5 km.

The Hartford Basin is part of the larger Connecticut Valley basin which also includes the Deerfield and Northfield basins. The Hartford Basin itself is divided into a series of downwarps separated by intervening

antiformal regions called the Middletown block and Gaillard and Saltonstall grabens (Wise, 1992). The synformal warps are associated with recesses or concavities in the fault surface along strike, while the anticlines are associated with salients or convex portions of the fault (Wheeler, 1939). These folds suggest significant slip variation along strike, a feature which should be recorded in the thickness of the sedimentary and volcanic units if these folds existed throughout the development of the basin. It remains to be determined whether the variation in slip represents a variation in total extensional strain or whether the strain is accommodated by other structures within the basin. The Basin is also cut by a large intrabasinal fault with ~700-m of throw near Hartford (Chang, 1968; Ellefsen et al., 1990).

In addition to the undulatory shape of the border faults the faults are also often stair-stepped with north-northeast striking segments connected by more northeast trending segments. It has been suggested that the more northerly trend is inherited from the pre-existing structure in the metamorphic rocks while the northeast trend is more consistent with the regional stress inferred from the Jurassic dike trends (Fig. 1, right) (Clifton, 1987; de Boer and Clifton [Clifford], 1988). In some cases it can be seen that the northeast trending set cross-cuts the north-northeast set. The northeast set is most clearly developed in the vicinity of the Hanging Hills of Meriden. Although pre-existing structure appears to have played an important role in the development of the rift basin there have been few detailed studies of this process.

Perhaps one of the largest unresolved questions regarding basin structure is the total offset on the boundary faults and thus the maximum burial depth of the basinal sediments. Thermal maturity studies (Pratt et al., 1988) of organic-rich lacustrine sediments suggest that the Portland Formation did not reach peak thermal maturity (maximum temperature of <90°C) while the East Berlin and Shuttle Meadow Formations are at peak maturity (90-130°C depending on time of burial). These results suggest that erosion has at most removed 1.5 km of material from the top of the Portland formation (Pratt et al., 1988). Parnell et al. (1998) found fluid inclusion homogenization temperatures ranging from 85-96°C for syntectonic veins within the East Berlin Fm. From Turner's Falls, MA in general agreement with the thermal maturity studies. In contrast Philpotts and Martello (1986) have suggested up to 10 km of throw along the eastern border fault based on reconstruction of the Fairhaven and Higganum dikes. Roden-Tice and Wintsch (2002) have also suggested higher maximum temperatures and thus deeper burial depths based on the resetting of zircon fission track ages. Zircon fission track ages are younger than 200 Ma for all but the youngest Portland Formation suggesting that most of the basin sediments have reached temperatures of > 200-240°C. Roden Tice and Wintsch (2002) point out that the discrepancy between the temperatures suggested by fission track ages and thermal maturity results may be due to over-estimation of the zircon closure temperature or mechanisms that retard thermal maturity of organic matter in the lacustrine sediments. Additional thermochronology and further investigations of diagenesis and metamorphism within basinal rocks may help to resolve this debate.

### **Tectonic Phases and Timing**

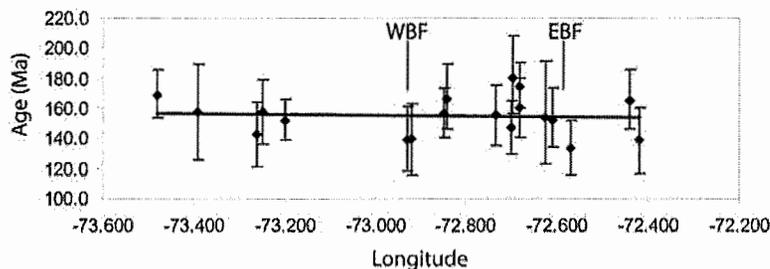
The classic model of CAM basin development is a rifting phase associated with normal faulting, sedimentation, and volcanism, followed by a drifting phase associated with contraction, often reactivating previously extensional structures (Withjack et al., 1995; Withjack et al., 1998). The history of the Hartford basin appears to generally fit this model however detailed study has revealed additional phases.

The earliest phases of basin formation are not well understood due to subsequent burial and overprinting. The basin may have begun as a sag basin (Hubert et al., 1992) in which sediment was derived from all margins. Coarse New Haven Formation sediments along the Western Border Fault and intrabasinal fault (Chang, 1968; Ellefsen et al., 1990) suggest that these structures were active during early development of the basin. During this time period sedimentation matched or exceeded subsidence so that sedimentation was dominated by fluvial environments. During the early Jurassic subsidence exceeded sedimentation and closed lake basins formed within the rift. This time period is coincident with magmatism. The early history of the Eastern Border Fault is unknown due to burial by younger sediments, however coarse clastic wedges associated with alluvial fans and debris flows suggests that the fault was active during the Early Jurassic to early Middle Jurassic deposition of the upper Portland Fm.

Subsequent to deposition of the Portland Fm. (early Middle Jurassic) basin sediments were tilted toward the east and cut by a series of northeast-trending normal faults. These structures deform all preserved units and cross-cut the pre-existing structural grain, so that their last activity is clearly post early Middle Jurassic, however the timing of their first movement is unknown. No evidence has been found for syn-depositional thickening of units within the basin. The absence of younger strata makes dating of the post early Middle Jurassic history difficult for the Hartford basin. Post rift sediment (largely unfaulted) was deposited in offshore basins (George's Bank, Scotian Basins) during the late Early Jurassic to Middle Jurassic (Withjack et al., 1998). In the Hartford Basin two phases of post-rift deformation have been described, a north-south contractional phase (shifting) and a later northwest-southeast contraction (drifting) (de Boer, 1992). Evidence for these phases comes from abundant faults with strike-slip slickenlines. In central New England the shifting phase is found in Jurassic dikes, but is absent from Early Cretaceous dikes suggesting that North-South compression occurred during the Jurassic (Manning and Deboer, 1989). Deformation associated with both shifting and drifting phases is found within Early Cretaceous dikes. In the offshore Orpheus graben Whithjack et al. (1998) found that post-rift deformation occurred before or during the early Cretaceous.

Roden-Tice and Wintsch (2002) recently questioned this entire chronology based on their fission track results. As mentioned above they found that Apatite and Zircon (except in the youngest Portland Fm) fission track ages were reset to Mesozoic ages both within the basin and in the footwalls of the border faults. Along the Eastern Border Fault footwall ages are younger than hanging wall ages. Roden-Tice and Wintsch (2002) used this evidence to argue that normal fault "*displacement [was] younger than the youngest fission track ages of  $\leq 100$  Ma (Late Cretaceous). Thus, the age of the graben structure of the Hartford Basin is Cretaceous, and this structure cannot be cited as evidence that these basins are Early Mesozoic "rift" basins.*"

The fission track data is clearly provocative and must be explained by any reasonable model of the basin. However the many lines of evidence cited above indicate that the basin was active during the early Mesozoic. We believe that multiple models for the fission track data must be explored and that the ideal model for basin development will honor all of the available data sets. Significant topography likely existed at the end of the rifting period and the fission track ages may record the differential unroofing of this landscape. A cross-section of the fission track data across the southern Hartford Basin (Fig. 3) shows that with the exception of 2 data points all data could be fit within error by a straight line suggesting slight post-rift differential unroofing. Additional fission track ages and use of other middle to low temperature thermochronology should help in the development of a better model for basin development.



**Figure 3.** Graph of apatite fission track ages across the southern Hartford Basin (41.44-41.68 deg. Latitude). Error bars are  $2\sigma$ . Line is best least squares fit. WBF and EBF are Western and Eastern Border faults, respectively. Data from Roden-Tice and Wintsch (2002).

## ROAD LOG

### Mileage

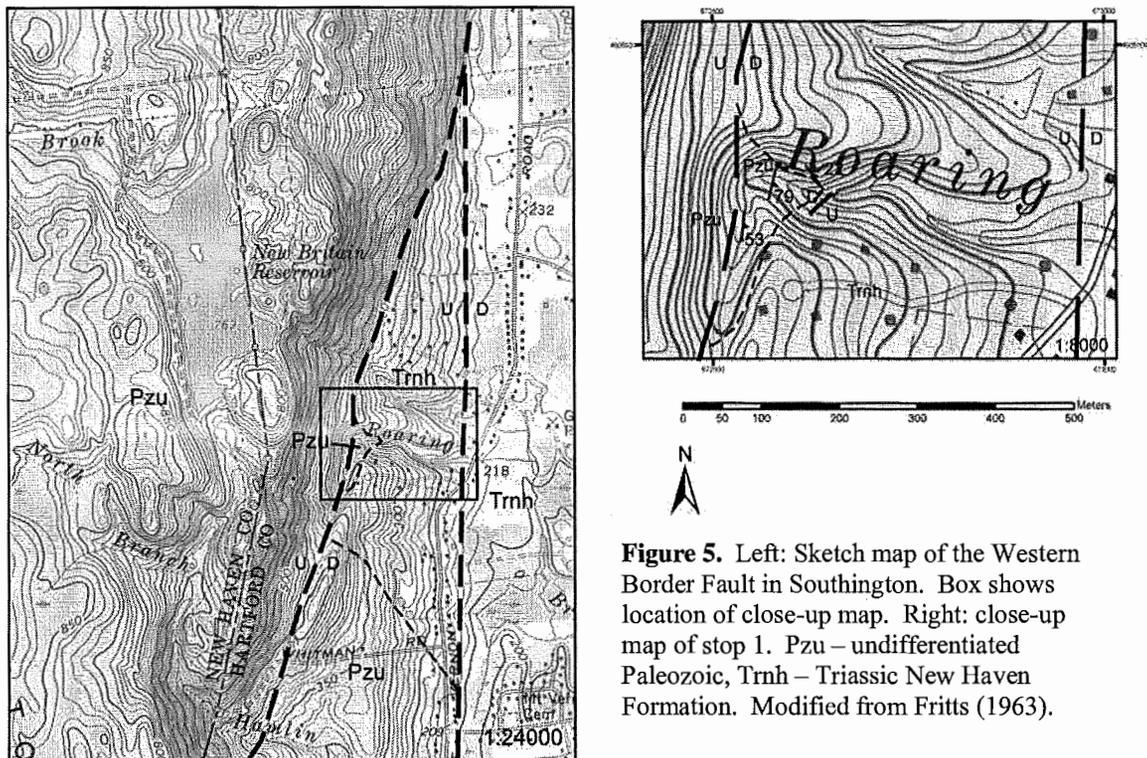
- 0.0 Start at the entrance to the Wesleyan parking lot on the west side of Vine Street, immediately south of the tennis courts. Head north on Vine St.
- 0.2 At light turn left onto CT 66 west.
- 3.7 Middletown Reservoirs on either side of road. These reservoirs are located stratigraphically above the Holyoke basalt.
- 4.1 Outcrop of Holyoke basalt.

- 5.1 Profile of basalt flow to south, entablature to north, flow is offset by a late stage fault near this location. CT 66 becomes I-691.
- 8.2 New Haven formation. View to west of 2 prominent benches – the Talcott and Holyoke basalt flows.
- 9.6 Leached (reduced) beds within New Haven Formation. This phenomenon is associated with U and Cu mineralization.
- 12.9 Take Exit 3 to CT 10.
- 13.1 Right turn on CT 10 (north).
- 13.4 Straight at light (Do not follow CT 10).
- 13.5 left onto CT 322.
- 14.5 Western border fault scarp straight ahead.
- 15.3 Right on Marion Ave.
- 16.0 Left on Mt. Vernon Rd (at large black arrow pointing right).
- 18.4 Left on Roaring Brook Rd.
- 18.6 Park at Culdesac.

**STOP 1. Basal (Triassic) unconformity and western border fault. (60 minutes)**

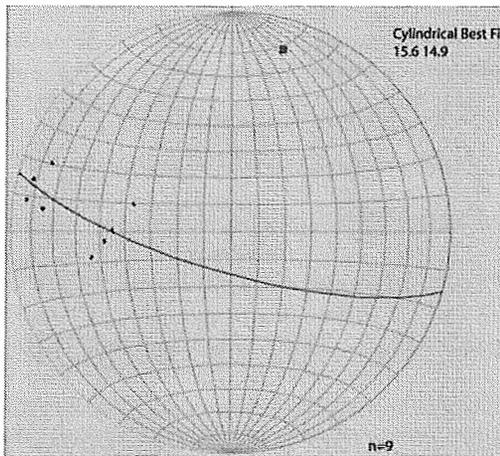
**NOTE: This stop is on private property. Permission must be granted by owner before trespassing. Please respect private property rights so that this stop remains accessible to future groups.**

This location is perhaps the most spectacular outcrop of the basal unconformity between the Paleozoic metamorphic basement and the Triassic New Haven Formation. The unconformity is preserved in this location due to the presence of two intersecting faults (Fig. 4, left) – the present-day border fault which strikes more northerly and a second normal fault in the footwall of the border fault which strikes more northeasterly. This geometry leads to an intermediate level of exposure where the unconformity is exposed at the present-day surface. The unconformity dips to the east-northeast and overlying beds strike NW and dip  $\sim 20^\circ$  NE (Fig. 4, right). The New Haven Formation here is a coarse arkose with clasts of feldspar, quartzite, and schist. The maximum clast size exceeds 5 cm. The sediments show fining upward sequences and channel geometries indicative of deposition in a fluvial environment.



**Figure 5.** Left: Sketch map of the Western Border Fault in Southington. Box shows location of close-up map. Right: close-up map of stop 1. Pzu – undifferentiated Paleozoic, Trnh – Triassic New Haven Formation. Modified from Fritts (1963).

The metamorphic rocks below the unconformity are mapped as the Southington Mountain Formation of Fritts (1964). It is a medium to fine-grained silvery mica schist composed of quartz, muscovite, biotite, oligoclase, and garnet, with layers rich in staurolite and/or kyanite. The foliation at the unconformity location is steeply dipping toward the east (strike NNE dip  $\sim 80^\circ$  SE), however immediately upstream the foliation is more flat lying suggesting map-scale folding with hinge lines trending north-northeast (Fig. 5). There is a prominent crenulation that is nearly orthogonal to the map scale folding trending SW and plunging  $\sim 30^\circ$ . Other features of interest are the pegmatites ( $355 \pm 5$  Ma.) and quartz veins that are largely foliation parallel, but cross-cut foliation just below the unconformity. One pegmatite in the stream bed shows possible duplication indicative of strike-slip (?) faulting. There is abundant evidence for hydrothermal activity along the Western Border Fault including silicification of sediments and localized mineralization. The most significant of these sites are the copper deposits near Bristol, CT north of stop 1.



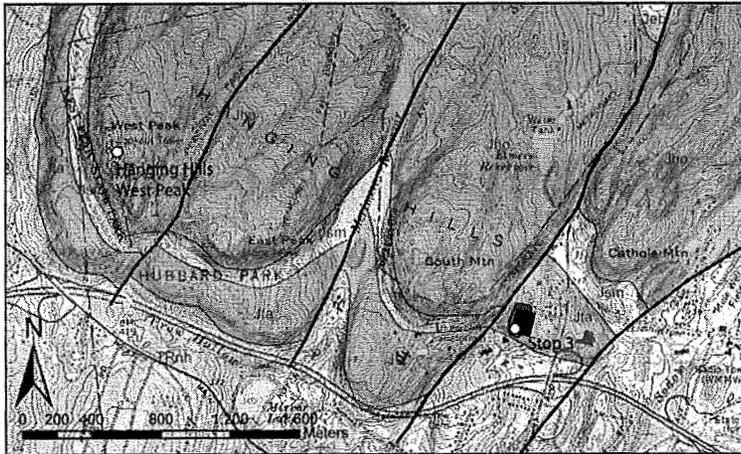
**Figure 5.** Equal area plot of foliation from Stop 1. Great circle is cylindrical best fit to data.

Downstream of the unconformity outcrop is a second prominent outcrop of the New Haven Formation. Weathering of the arkose has created an alcove here. The back of the alcove is a fault surface striking  $227^\circ$  SW and dipping  $76^\circ$  NW. This surface contains two sets of oblique-slip striations. The older set has a rake of  $41^\circ$  in a northeasterly direction. This set is overprinted by a set that has a rake of  $28^\circ$  SW. This fault is isolated so its relative timing cannot be determined; however it may be associated with the transition from rifting to drifting (shifting phase) associated with opening of the Atlantic.

- 18.6 Retrace route down Roaring Brook Rd.
- 18.9 Right on Mt. Vernon Rd.
- 21.3 Right on Marion Ave.
- 21.9 Right on CT 322.
- 23.7 Right on CT 10 south.
- 24.3 Left on to I-691.
- 28.1 Exit 5 – CT 71 – Chamberlain Hwy.
- 28.3 Left on CT 71.
- 28.5 Left into Target parking lot.
- 28.7 Park at southwest corner of lot to south of Target store.

## **STOP 2. Pillow basalts and basal contact of the Talcott Basalt. (20 minutes)**

Recent construction of a “big-box” store here has revealed a beautiful exposure of the basal contact of the Talcott basalt (Fig. 6). The basalt overlies the upper New Haven Formation which has graded beds of  $\sim 10$  cm thickness. The lower portion of the flow displays excellent pillow basalts. Sediments of the underlying New Haven fm. have locally been fluidized and injected between the pillows creating beautiful soft-sediment deformation features.



**Figure 6.** Geologic Map of Hanging Hills region. Black box is approximate location of Target store. Gray units are Talcott (Jta) and Holyoke (Jho) basalts. Geology from Rogers (1985).

**Optional Stop. West Peak of Hanging Hills.** An excellent location for a basin-wide overview.

- 28.7 Return to Target store entrance.
- 28.8 Left on CT 71.
- 29.0 Right on Kensington Ave.
- 29.5 Right on Lewis St. (To I-691). Continue under 691.
- 29.9 Left onto I-691 on ramp.
- 32.3 Left Exit 11 – I-91 north.
- 33.3 Talcott basalt.
- 38.0 Exit 21 – CT 372.
- 38.1 Right onto 372 west (Talcott basalt).
- 40.3 Hamden contact with East Berlin Fm.
- 40.4 Lacustrine highstand beds.
- 40.7 Right to US 5/CT 15 (at second light). Under CT 9.
- 40.8 Left to US 5/CT 15, CT 9 south.
- 41.0 light at US 5/CT 15. Stay in center lane to go to CT 9.
- 41.1 Park on right side of CT 9 on ramp. Pull clear off pavement.

**STOP 3. Lacustrine cyclostratigraphy of the East Berlin Formation.** (45 minutes)

At this stop we will have the chance to look first hand at lacustrine sequences within the East Berlin Formation (Fig. 7). The major lake sequences at this stop are ~32-m apart in a bed-perpendicular direction. This separation is consistent with Olsen's ~133 my period. Higher-frequency lake deposits are absent in this road cut, but on Rt. 372 just up the hill the closest spacing between lacustrine beds is only ~12m. Beds here dip ~15 degrees toward the southeast. Look for excellent exposures of mudcracks and ripples.

The black lacustrine beds here show evidence of deformation including cm- scale folding and slickensides. Folding at this location is consistent with down-dip motion while slickenlines on calcite veins indicate subsequent northeastward slip (Wilder, 1998). This phenomenon has been reported from other localities within the Hartford Basin (notably at Turner's Falls, MA). Parnell et al. (1998) argue that movement occurred in the present down-dip direction and that the lacustrine beds thus acted as a décollement during burial, tilting, and hydrocarbon formation. On a warm day the rocks here even smell like oil.

- 41.1 Continue onto CT 9 south.
- 48.6 Exit 16 right to CT 66 toward Portland CT.
- 48.9 Arrigoni Bridge (completed in 1938). Stay in left lane.
- 49.7 Left on Silver St., immediately after Hess Station.
- 49.8 Overlook to main Portland quarry.
- 49.9 Park on grass near corner of Silver St. and Brownstone Ave.



**Figure 7.** Geologic Map of the eastern Hartford Basin near Middletown. Hampden basalt (Jha) is colored light gray. Paleozoic basement rocks are filled with wavy pattern. Other formations of interest – Holyoke basalt (Jho), East Berlin Formation (Jeb), Portland Formation (Jp). Contours of aeromagnetic data from Daniels and Snyder (2004). Bold dashed line is our interpretation of the likely continuation of the Hampden basalt near stop 5. Structural data from this study, Eaton and Rosenfeld (1972) and Lehmann (1955). Geology modified from Rogers (1985).

**Optional Stop. Dinosaur State Park.** A spectacular trackway in the East Berlin Fm.

**STOP 4. Portland Brownstone Quarries.** (45 minutes)

The Portland Brownstone Quarries are a National Historic Landmark. At this location the Portland Formation is nearly flat-lying with widely spaced joints (Fig. 7). Many of the beds are greater than 1-m thick. These properties as well as the proximity to the Connecticut River made this an ideal site for brownstone quarrying. Quarrying began almost as soon as European settlers arrived and continued until the 1930's (Guinness, 2003). Peak production of 850,000-1,000,000 cubic feet of building stone was reached in the 1890's with stone shipped all over the east coast and even to San Francisco (Guinness, 2003). The oldest buildings on the Wesleyan University campus are built from Portland brownstone and funds from the Town quarry were used to help establish the school. Most recently a quarry has been re-established at the northeast corner of the quarry site. The present-day quarry specializes in restoration as well as new construction (<http://www.brownstonequarry.com/>).

- 49.9 Turn around and return to CT 66 up Silver St.
- 50.3 Left on to CT 66/17, stay in left lane to continue on 17A north.
- 53.3 Left on CT 17 (Glastonbury Tpk.) at 4-way stop.
- 55.6 Pegmatite (~255 Ma.) on right.
- 56.0 Left on Old Maids Rd. Metamorphic rocks – see description below.
- 56.4 Left onto unnamed quarry access rd. The road has been traversing the lake-bottom sediments of glacial lake Hitchcock.
- 57.2 Park at quarry floor.

**STOP 5. Coarse conglomerates of the upper Portland Formation.** (45 minutes)

The Eastern Border fault is located to the east of this location (Fig. 7) near the turn onto Old Maids Rd. (mi. 56.0). Footwall rocks are amphibolites of the Collins Hill (Rogers, 1985) or Hebron Fm. (Snyder, 1970). The foliation here dips steeply to the west (strike  $185^\circ$ , dip  $80^\circ$ ).

Coarse conglomerates of the Portland formation directly overlie lacustrine shales at this locality. The coarse deposits at this location are matrix-supported conglomerates with many clasts exceeding 50 cm in their maximum dimension. The boulders and cobbles have little to no preferred orientation and are concentrated near the tops of the beds. These characteristics are typical of debris flow deposits of the Portland Formation (LeTourneau, 1987). Deposition of these sediments directly onto lacustrine mudstones suggests deposition near the toe of an alluvial fan.

The beds at this location strike toward the southeast  $107^\circ$  and dip toward the southwest  $40^\circ$ . This unusual orientation is due to the location on the southern limb of the Rocky hill anticline that separates the Middletown block from the main Hartford Basin. The nose of the fold was omitted on the state geologic map (Rogers, 1985), however aeromagnetic data show a clear anomaly continuing across the river which we interpret as the likely continuation of the Hampden basalt (Fig. 7). Interestingly, the anomaly pattern suggests a significant right-lateral strike-slip component of displacement on the mapped northeast-striking fault.

Numerous small faults cut through the area. Some of these are nicely exposed cutting the conglomerate bedding surface while larger faults cut between outcrops leading to a repeat of the conglomerate/mudstone contact. Exposed faults strike south-southeast and northeast with both westerly and easterly dips (Wilder, 1998). At least one fault plane has striations that have a rake of  $\sim 60^\circ$  to the south-southeast indicating oblique (normal/right-lateral) slip.

- 57.2 Turn around and return to Old Maids Rd.
- 58.0 Right on Old Maids Rd.
- 58.4 Right on CT 17 (south).
- 61.1 Continue straight on CT 17 at 4-way stop with CT 17A.
- 62.0 Proceed straight at 4-way stop.
- 62.3 Portland Fm. Conglomerate dipping steeply toward Eastern Border Fault.
- 63.1 Right on CT 66 at light. Footwall metamorphic rocks are exposed on right immediately before the light.
- 63.6 Steeply dipping red beds behind hardware store.
- 65.1 left to Middletown.

- 65.5 Arrigoni Bridge. Stay right.
- 66.0 Right onto Spring St.
- 66.2 Left onto High St.
- 66.6 Straight at light with CT 66.
- 67.0 Right on Church St.
- 67.4 Right on Vine St.
- 67.6 Turn into Wesleyan parking lot.

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# BEDROCK GEOLOGY OF THE NEW MILFORD QUADRANGLE, CONNECTICUT

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## INTRODUCTION

This trip will focus on the results of recently published mapping and geochronology studies in the New Milford quadrangle, western Connecticut (Walsh, 2003; Walsh and others, 2004). It has been 30 years since Leo Hall and his University of Massachusetts students (Hall and others, 1975) last led an NEIGC field trip through the New Milford area. This trip will visit significant outcrops of Mesoproterozoic basement rocks and their lower Paleozoic cover west of Cameron's Line as well as allochthonous rocks on both sides of the line. Visits to Mesoproterozoic bedrock exposures will include the oldest dated rock in Connecticut, a  $1311 \pm 7$  Ma granite gneiss, and several exposures of syn-tectonic migmatite and granite providing ca. 1.05 Ga (Ottawan) ages. Exposures of the Neoproterozoic to lower Paleozoic cover rocks in the autochthon and allochthon will be highlighted. Stops will also include visits to Ordovician plutonic rocks on both sides of Cameron's Line. For a more complete discussion of the geology of the New Milford area see Walsh (2003).

## GEOLOGIC SETTING

Bedrock in the New Milford quadrangle consists of upper amphibolite facies Mesoproterozoic basement gneisses (Hall and others, 1975; Walsh, 2003) unconformably overlain by amphibolite facies metasedimentary and metavolcanic rocks (Hall and others, 1975; Rodgers, 1985; Walsh) (Fig. 1). The basement gneisses are unconformably overlain by the Neoproterozoic to Cambrian Dalton Formation and Cambrian to Ordovician Stockbridge Formation (Rodgers, 1985) (Fig. 1). The Neoproterozoic-Ordovician cover sequence is best preserved in the north-central part of the New Milford quadrangle (Fig. 1). On the west side of the New Milford massif (Fig. 1), the Ordovician Walloomsac Formation unconformably overlies both the Dalton Formation and the Precambrian gneisses (Fig. 1). Here, the Stockbridge Formation was eroded away prior to the deposition of the Walloomsac Formation along a Middle Ordovician unconformity during the Taconic Orogeny (Zen, 1967; Rodgers, 1971, 1985; Jacobi, 1981). Late Middle to early Late Ordovician fossils constrain the age of the Walloomsac Formation in New York and Massachusetts (Potter, 1972; Ratcliffe, 1974; Finney, 1986; Ratcliffe and others, 1999). Allochthonous rocks of the Neoproterozoic to Cambrian Manhattan Schist (after Rodgers, 1985) are tectonically above the basement rocks and the autochthonous cover sequence (Fig. 1). The source, or root zone, of the allochthonous rocks lies east of or is concealed by Cameron's Line; a tectonic contact that separates basement and autochthonous platform sequence rocks to the west from ocean-rise sequence rocks with fragments of oceanic crust to the east (Rodgers, 1985; Stanley and Ratcliffe, 1985). The rocks east of Cameron's Line include metasedimentary rocks of the Cambrian to Ordovician Rowe and Ratlum Mountain Schists and the Ordovician Brookfield Gneiss (Fig. 1) (Rodgers, 1985; Walsh, 2003). If the allochthonous rocks in the central and western part of the New Milford quadrangle truly do root from beneath Cameron's Line and the Walloomsac here is time-correlative with the dated rocks in New York, then the maximum age of movement along Cameron's Line is given by the Blackriveran age (ca. 454 Ma) of the Walloomsac Formation (Ratcliffe and others, 1999) which lies tectonically beneath the allochthonous Manhattan Schist and is inferred to be above the Stockbridge Formation east of the New Milford Formation. The minimum age of movement along Cameron's Line is less well constrained. Sevigny and Hanson (1995) reported a composite intrusive age of  $453 \pm 3$  Ma from three plutons of the Brookfield Gneiss east of Cameron's Line, and Walsh and others (2004) obtained an age of  $453 \pm 6$  Ma from a leucogranite dike that cuts the diorite phase of the Brookfield Gneiss. The Brookfield Gneiss is syn-kinematic with respect to the D2 structures, and the dominant mylonitic foliation along Cameron's Line and in the rocks on both sides of the fault is the D2 fabric, suggesting that the fault was active during the D2 event (ca. 453 Ma). Thus, the paleontologic and geochronologic data suggest that early transport of the allochthons in this area occurred after about 454 Ma, and may have been completed by about 453 Ma, the time of Brookfield emplacement and development of early D2 fabric. Sevigny and Hanson (1995) also report a post-kinematic titanite age of  $443 \pm 3$  Ma from the Brookfield Gneiss, and a post-kinematic pegmatite age at  $445 \pm 1.5$  Ma, but report younger concordant ages of  $435 \pm 3$  and  $438 \pm 2$  Ma for xenotime from granite with blastomylonitic fabric west of Cameron's Line, suggesting that D2 deformation, if

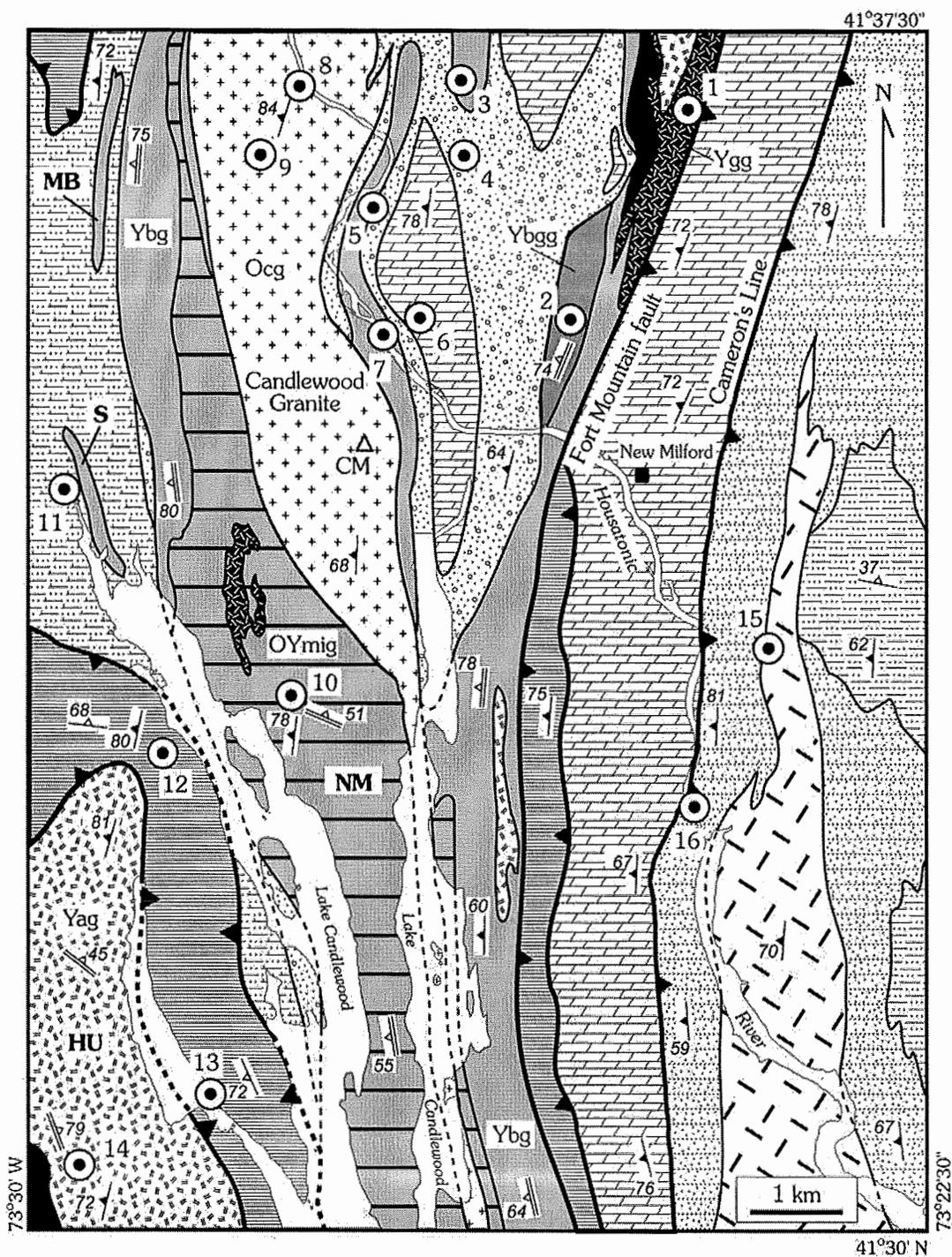


Figure 1. (On this and next page) Generalized geologic map of the 7.5-minute New Milford quadrangle, Connecticut showing field trip stops. Modified with permission from Walsh and others (2004) and based on new mapping by Walsh (2003). Labeled massifs include: Hudson Highlands (HU), New Milford massif (NM), Sherman inlier (S), and Morrissey Brook inlier (MB). U-Pb SHRIMP zircon ages on next page in Ma from Walsh and others (2004) shown in brackets []. CM = Candlewood Mountain.

DESCRIPTION OF MAP UNITS

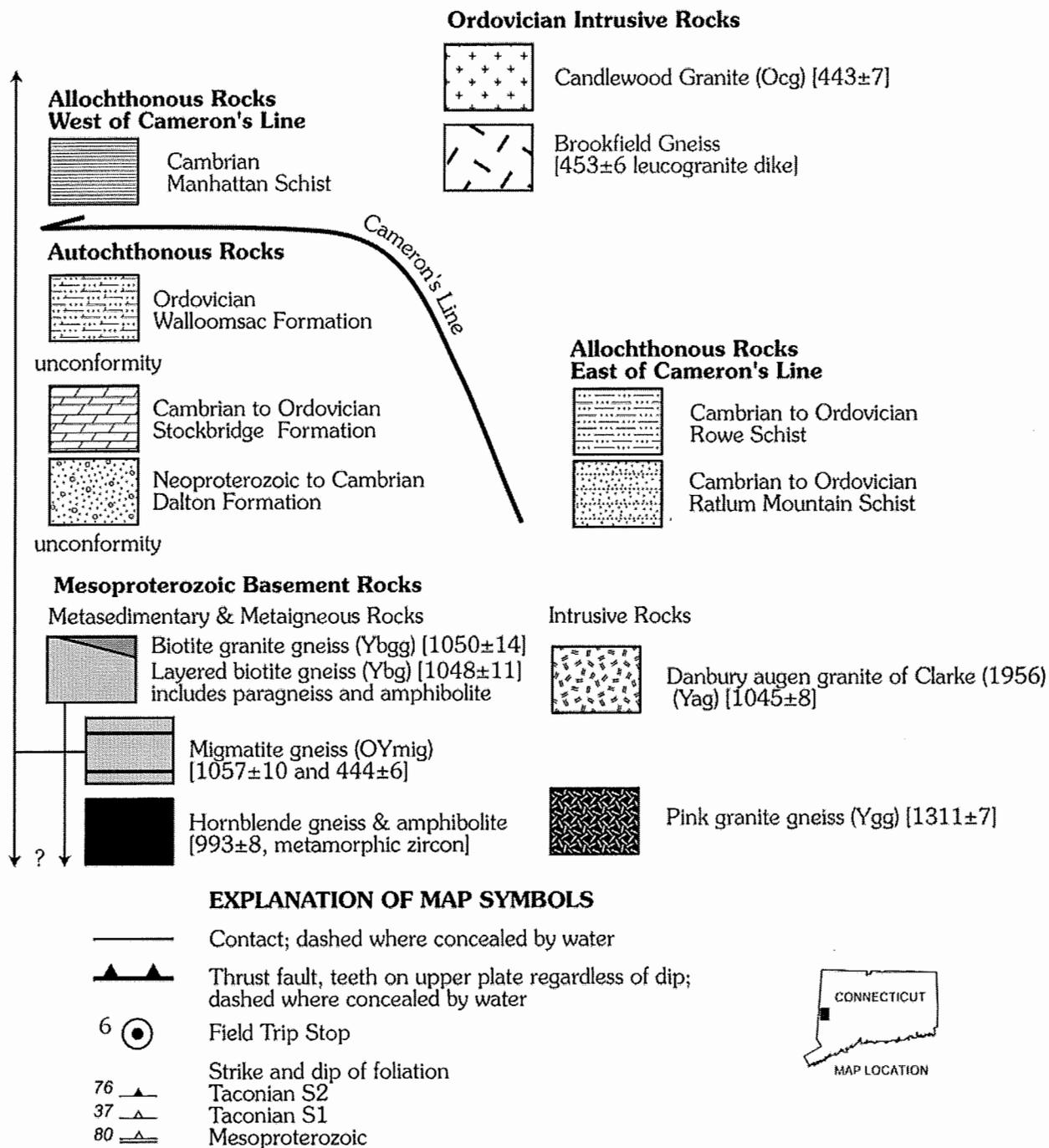


Figure 1. (Continued).

related to continued movement in the vicinity of or along Cameron's Line, may have continued into the Silurian. Currently, the Cameron's Line fault and the dominant foliation in the surrounding rocks are overturned and dip steeply to the west due to later Acadian folding (Rodgers, 1985; Panish, 1992; Sevigny and Hanson, 1995; Walsh, 2003).

### PREVIOUS WORK

Apart from the State geologic map by Rodgers (1985), no geologic maps had been published that cover the area of the 7.5-minute New Milford quadrangle. Previous mapping in the vicinity includes work to the north by Dana (1977) in the Lake Waramaug area of the New Preston quadrangle and Jackson (1980) in the Kent quadrangle; to the northeast, Gates and Bradley (1952) mapped the New Preston quadrangle; to the south, Clarke (1958) mapped the Danbury quadrangle; to the east, Gates (1959) mapped the Roxbury quadrangle; and to the southeast, Stanley and Caldwell (1976) mapped the Newtown quadrangle. Sperandio (1974) mapped the Brookfield Gneiss in the southwestern part of the New Milford quadrangle. Rodgers (1985) compiled unpublished manuscript maps of parts of the New Milford quadrangle by K.G. Caldwell and G.V. Carroll and the Connecticut part of the Pawling, New York quadrangle by R.A. Jackson for the State map. The map of the 7.5-minute Poughquag quadrangle, New York, two quadrangles to the west, represents the closest recently published mapping in the eastern Hudson Highlands (Ratcliffe and Burton, 1990).

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### ROAD LOG

**STARTING POINT.** The trip will start at the parking lot of the "Big Y" supermarket at the junction of U.S. Routes 202 and 7 in New Milford, CT. The start time is 8:30 AM on Saturday, October 1<sup>st</sup>, 2005. Please bring a lunch and a hardhat with you; hardhats are required for Stop 6 at the ASI quarry in the Stockbridge Formation. Most stops will be close to the road, but there is limited parking at a number of stops, so please try to consolidate into fewer vehicles. After Stop 14, we will pass by the "Big Y" if anyone wants to pick up their vehicles for the last two stops. Stop coordinates are given in UTM meters in NAD27 datum Zone 18. Minerals in rock descriptions are listed in order of increasing abundance.

#### Mileage

- 0.0 Big Y parking lot at the junction of U.S. Routes 202 and 7 in New Milford, CT.  
Turn right out of the parking lot.  
Turn left at the traffic light onto Route 202 East.
- 0.1 Cross bridge over Housatonic River.
- 0.3 Turn left at traffic light onto Railroad Street.
- 0.6 Bear slight right onto Wellsville Avenue.
- 1.5 View of Mt. Tom at 12 o'clock. The summit of Mt. Tom is underlain by augen granite gneiss.
- 2.2 Bear right onto Papermill Road
- 2.9 Small roadcut of pink granite gneiss (Ygg) on the west side of the road (left).
- 3.5 Park on the right side of the road for Stop 1.

#### **STOP 1. Pink granite gneiss (Ygg) on the Aspetuck River (632710 E 4608250 N).**

Sample NM174 of Walsh and others (2004). The rock is a well-foliated, tan weathering, light-pink biotite-quartz-plagioclase-microcline gneiss that has a generally uniform texture and modal composition of a granite. Chemically it is a peraluminous calc-alkaline granite with a volcanic arc signature (Walsh, 2003). The SHRIMP U-Pb age on zircon reported by Walsh and others (2004) is  $1311 \pm 7$  Ma, making it the oldest well-dated rock in Connecticut. Cross-cutting relations with the adjacent rock units were not observed during mapping, but the pink granite gneiss does contain mappable belts of amphibolite and biotite-quartz-plagioclase paragneiss that are interpreted as xenoliths. These xenoliths are comparable to the amphibolite and hornblende gneiss and layered

biotite gneiss shown in contact with the pink granite gneiss in Figure 1. The pink granite gneiss, therefore, is believed to have intruded as yet undated, and presumably older, paragneiss during the early stages of the Elzevirian Orogeny. We will see the paragneiss at later stops, especially Stop 7.

Although limited ages were available to early workers, both Clarke (1958) and Hall and others (1975) considered this granite gneiss to be among the oldest granitic rocks in the eastern Hudson Highlands. Dana (1977) and Jackson (1980) interpreted the pink granite gneiss unit as the youngest basement gneiss and stated that it cut all other rocks in the basement. We will see later at Stop 10 that both Mesoproterozoic and Ordovician pink granite gneisses occur in the area.

At the outcrop, the granite gneiss contains a well-developed planar gneissosity that strikes NNE and dips steeply to the west. An older, isoclinally folded and transposed foliation is locally visible with steeply plunging down-dip fold axes. The older foliation is a relict Mesoproterozoic (YS1) gneissosity, and the dominant foliation is a composite planar fabric that is both parallel to the regional Paleozoic S2 foliation and the second-generation Mesoproterozoic gneissosity (YS2). The S2 fabric is very well-developed in this area due to the proximity to the Fort Mountain fault, a Paleozoic mylonite zone. The dated rock was collected from just below the hemlock tree near the plunge pool.

#### Mileage

- 3.5 Carefully make a U-turn and head south on Papermill Road.
- 4.7 Turn left at stop sign onto Wellsville Avenue.
- 5.3 Turn right onto Wells Road.
- 5.8 Turn left onto Aspetuck Ridge Road (no street sign)
- 5.9 Turn left into the entrance for the Beacon Reel Company
- 6.0 Park on the right near the roadcut for Stop 2.

#### STOP 2. Biotite granitic gneiss (Yb<sub>gg</sub>) at Beacon Reel Company (631300E 4605440N).

Sample NM628 of Walsh and others (2004). The rock is a well-foliated, moderate to poorly layered, gray biotite-microcline-quartz-plagioclase granitic gneiss. The SHRIMP U-Pb age on zircon is  $1050 \pm 14$  Ma; a good Ottawa age. Chemically, the dated rock is a quartz monzonite. The homogeneous character of this outcrop is typical of the rock unit mapped as biotite granitic gneiss, Yb<sub>gg</sub> (Walsh, 2003) and suggests that the rock is an orthogneiss and that the protolith of this rock may have been intrusive, although an extrusive origin cannot be ruled out. This outcrop occurs in a belt of rocks that has been mapped separately as Yb<sub>gg</sub> from other belts of a more heterogeneous rock unit (Yb<sub>g</sub>) that contains both paragneiss and orthogneiss. We will see examples of the Yb<sub>g</sub> map unit later at Stops 3, 5 and 7.

At the outcrop, the gneiss contains a very well-developed planar gneissosity that strikes NNE and dips steeply to the west. An older, isoclinally folded and transposed foliation is locally visible with steeply plunging down-dip fold axes. The older foliation is a relict Mesoproterozoic (YS1) gneissosity, and the dominant foliation is a composite planar fabric that is both parallel to the regional Paleozoic S2 foliation and the Mesoproterozoic YS2 gneissosity. Similar to Stop 1, the S2 fabric is very well-developed in this area due to the proximity to the Fort Mountain fault, a Paleozoic D2 mylonite zone.

#### Mileage

- 6.0 Proceed east, make a U-turn in the parking lot of the Beacon Reel Company, return to entrance.
- 6.3 Turn right on Aspetuck Ridge Road.
- 7.7 Turn left on Long Mountain Road.
- 8.5 Small roadcuts of Dalton Formation schist at S-turn.
- 8.6 Carefully park on the right side of the road under the power lines for Stop 3. There is limited parking here.

#### STOP 3. Sills of layered biotite granite gneiss in Yb<sub>g</sub> on Long Mountain (630010E 4608420N).

Sample NM772 of Walsh and others (2004). Proceed uphill (east) under the power lines to the outcrops just downhill of the first set of power poles. The sampled rock is a well-foliated, 30-cm-thick layer of biotite-quartz-microcline-plagioclase gneiss with the normative composition of granite (Walsh, 2003). The biotite granite gneiss here is bounded by 20-cm and 10-cm-thick layers of amphibolite. The SHRIMP U-Pb age on zircon is  $1048 \pm 11$  Ma, and is the same, within uncertainty, as the age of sample NM628 of  $1050 \pm 14$ . The sample layer is interpreted as an intrusive sill or migmatitic leucosome within the layered biotite gneiss unit, Yb<sub>g</sub>.

Along the road where you parked, there is a new roadcut that exhibits a well developed gneissosity with thin, foliation-parallel, cm-scale migmatitic leucosomes with layers and boudins of amphibolite. Keep these migmatitic leucosomes in mind for comparison with the rocks at Stop 10. The heterogeneity seen at these outcrops is typical of the layered biotite gneiss unit Ybg (Walsh, 2003). The dominant planar fabric at the roadcut is the regional Mesoproterozoic YS2 gneissosity, which is deformed by steeply dipping, NNE striking Paleozoic F2 folds, with moderate to steep NNE plunges. Zircons separated from a sample of amphibolite from an outcrop near Stop 1 (Walsh and others, 2004, sample NM159) consisted almost entirely of metamorphic cores and rims which yielded a SHRIMP U-Pb age of  $993 \pm 8$  Ma, indicative of terminal Grenville metamorphism which reached lower granulite facies conditions (Dallmeyer and Dodd, 1971; Hall and others, 1975). A single rounded zircon-core yielded a SHRIMP U-Pb age of about 1355 Ma, suggesting that either the protolith of the amphibolite was a volcanic rock or that the cores are related to an older (ca. 1355 Ga) pyroxene granulite (?) grade metamorphism.

#### Mileage

- 8.6 Carefully make a U-turn at Stop 3 and go south on Long Mountain Road.
- 9.3 Turn sharp right onto Rooster Tail Hollow (also known as Bennett Road).
- 9.7 Park on right for roadcuts at Stop 4.

#### STOP 4. Dalton Formation sillimanite-biotite schist (629970E 4607750N) (Optional).

If time permits, we will make a quick stop here to look at the pelitic schist unit within the Dalton Formation. The rock here is a gray to dark-gray, rusty weathering, coarse-grained,  $\pm$ chlorite-muscovite-garnet-sillimanite-k feldspar-biotite-quartz-plagioclase schist. Sillimanite is fibrolitic and occurs as porphyroblastic knots (up to 2 cm) that show replacement by quartz and muscovite. The “knots” of fibrolite, quartz, and muscovite are characteristic of the Dalton Formation. Interlayered quartz-feldspar granofels are locally present in the pelitic unit and are more abundant near the base of the Dalton, but are not seen here – we will see the granofels at Stop 5. The dominant foliation at this outcrop is the S2 schistosity which strikes approximately N30°E and dips vertically or steeply to the west.

West of Cameron’s Line, peak Paleozoic metamorphism in the cover rocks produced sillimanite - K-feldspar assemblages. Hames and others (1991) dated hornblende from the Taconian staurolite zone two quadrangles to the north by  $^{40}\text{Ar}/^{39}\text{Ar}$  and obtained an age of approximately 445 Ma that they interpret as growth ages during Taconian metamorphism. This age closely agrees with ages from the syn-tectonic Candlewood Granite ( $443 \pm 7$  Ma, Stop 9) and the granite leucosomes in the basement migmatite gneiss ( $444 \pm 6$  Ma, Stop 10) (Walsh and others, 2004) suggesting that peak metamorphism, melting, and intrusion were contemporaneous during Paleozoic D2. Subsequent Acadian metamorphism produced fibrolite-muscovite assemblages during the Devonian (Hames and others, 1991). East of Cameron’s Line, mineral assemblages reflect staurolite-kyanite grade Acadian metamorphism, which Hames and others (1991) dated just to the north at about 390 to 400 Ma.

#### Mileage

- 9.7 Continue west on Rooster Tail Hollow (aka Bennett Road).
- 10.2 Bear left staying on Rooster Tail Hollow (aka Bennett Road).
- 10.5 Cross railroad tracks and turn left onto River Road (no street sign).
- 11.1 Park on right near the telephone-pole guardrail for Stop 5.

#### STOP 5. Basement – cover contact in the Housatonic Riverbelt Greenway (629190E 4605590N).

This stop shows the rocks at the basement-cover contact. The basement here consists of biotite-quartz-plagioclase gneiss (Ybg) with lenses of amphibolite and pegmatite. The cover rocks are the quartzofeldspathic granofels of the Dalton Formation. The contact is located approximately 50 m west of the guardrail along an ATV trail. Outcrops just west of the guardrail and approximately 50 m south of the guardrail consist of rusty weathering granofels with characteristic muscovite-quartz-fibrolite knots. The granofels locally contains small (< 0.5 cm), rare quartz and feldspar pebbles near the basement – cover contact. The Dalton Formation gets more pelitic up-section (like at Stop 4), and near the contact with the overlying Stockbridge Formation, may contain rare pods or boudins of calc-silicate rock. In the New Milford quadrangle, no cover rocks were mapped above the basement but below the Dalton Formation, such as the “Ned Mountain formation” of Brock (1993).

#### Mileage

- 11.1 Continue south on River Road.

- 13.2 On you left you will get a brief view of the ASI quarry in the Stockbridge Formation.  
 13.5 Turn left at stop sign onto Boardman Road.  
 13.6 Turn left into entrance to Advance Stone Incorporated (ASI) for Stop 6.

**STOP 6. Stockbridge Formation at ASI quarry (629425E 4605780N) (Hardhats required).**

The ASI quarry owned by Robert Kovacs is quite extensive and even includes an old tunnel to one of the old quarry sites up the hill. Due to time constraints, we will only visit the north end of the quarry near the entrance. According to Robert Kovacs, the cuts visible at the north end of the quarry were made for dimension stone by Valley Marble and Slate for renovations at the Old State House in Hartford. Most of the material extracted at the quarry today is used for septic system sand and road base.

The Stockbridge Formation consists largely of light-gray and white, chalky white weathering, massive to thickly bedded, weakly to moderately foliated, tremolite-dolomite marble. It locally contains thin lenses of coarse-grained sphene-quartz-phlogopite-diopside-tremolite calc-silicate rock, thin beds of light-gray and white calcite marble and rare, rusty weathering calcareous schist. The quarry is located in massive dolomite marble, and locally contains layers of calcite marble, or calcite-bearing zones within the largely dolomitic marble. The foliation in the marble strikes north-south and dips steeply to the west. The dominance of dolomite marble in the New Milford quadrangle suggests that the Stockbridge Formation here is correlative with the lowermost, or Cambrian, part of the formation regionally (Zen, 1966; Hall, 1968; Ratcliffe, 1974; Ratcliffe and others, 1993).

**Mileage**

- 13.6 Make a U-turn and head back to quarry entrance on Boardman Road.  
 13.7 Turn right on Boardman Road.  
 13.8 Park to the right, just before the railroad tracks, at Boardman Bridge. Walk across the railroad tracks and cross under the new bridge on the south side for Stop 7.

**STOP 7. Layered biotite gneiss (Ybg) at Boardman Bridge (629190E 4605590N).**

This outcrop is a good example of the layered biotite gneiss unit (Ybg) where it is believed to consist largely of paragneiss with boudins or lenses of amphibolite. The main rock type is well-foliated, gray, biotite-k feldspar-quartz-plagioclase gneiss. The gneiss has a well developed Mesoproterozoic gneissosity (YS2) with isoclinal folds (YF2) that deform a relict older gneissosity (YS1). The gneissosity and YF2 folds are deformed by upright NE striking and steeply dipping Paleozoic F2 folds. This rock type is the major rock type within the Ybg map unit, and this paragneiss is most likely the oldest rock in the area – older than ca. 1311 Ma. The Ybg map unit contains younger biotite granitic gneiss layers that are ca. 1050 Ma. (ie. Stop 3) that could not be mapped separately, except for the Ybgg unit. Thus, the Ybg map unit is a composite paragneiss and orthogneiss. Later at stop 10, we will see that it is also extensively migmatized.

**Mileage**

- 13.8 Proceed west on Boardman Road across the bridge over the Housatonic River.  
 13.9 Turn right at stop sign onto Route 7 North.  
 16.5 Turn left into gravel parking area for Stop 8.

**STOP 8. Tory's Cave (627990E 4608480N).**

Tory's Cave is located in layered calcite marble that is mapped as Stockbridge Formation. The sign at the cave states that it is, "The only marble solution cave open to the public in Connecticut." The cave is reportedly the third largest in Connecticut, and measures approximately 250 feet in total length (Yale Speleological Society, 1963). Just uphill from the cave entrance, there are outcrops of pelitic schist assigned to the Dalton Formation. The marble at the cave and the surrounding Dalton Formation rocks are part of a large map-scale xenolith within the main mass of the Candlewood Granite pluton. Here, we are in the center of the pluton which measures approximately 2 km wide at this latitude. If time permits, you can walk about 500 m NW from the parking area along a gravel road to an abandoned quarry in the granite. In the quarry walls you will see many screens or xenoliths of Dalton Formation within the granite. The presence of calcite marble here, and not dolomite marble like at Stop 6, suggests that this xenolith may have come from a higher section of the Stockbridge Formation than the marble to the east of our current location. It is also possible that the schist around the marble in this area could be Walloomsac Formation, and that the xenolith preserves part of the Stockbridge-Walloomsac contact. Without contiguous belts of schist and marble, it is difficult to determine whether the pelitic rocks here are Dalton or Walloomsac because compositionally

they are very similar at this metamorphic grade. At the cave entrance, notice the F2 folds with NS striking, steeply dipping axial surfaces and moderately south-plunging fold axes – the entrance follows the trend of the fold axes.

#### Mileage

- 16.5 Turn right out of the parking area, traveling south on Route 7.
- 17.9 Turn right onto Squash Hollow Road.
- 18.5 Turn left onto Elena Drive.
- 18.6 Park for roadcuts at Stop 9.

#### STOP 9. Candlewood Granite (627990E 4608480N).

Sample NM706 of Walsh and others (2004). This rock is a medium- to light-gray, tan- or white-weathering, moderately to well-foliated, fine- to medium-grained, muscovite-biotite-quartz-plagioclase-microcline granite. The large pluton underlying Candlewood Mountain, Pine Knob, and Boardman Mountain was formally named the Candlewood Granite by Walsh (2003). At this stop, and locally at other places, the granite is porphyritic with phenocrysts (up to 1 cm) of microcline. Xenoliths at this stop resemble rocks of the Dalton Formation. The Candlewood Granite intrudes both cover and basement rocks west of Cameron's Line. The SHRIMP U-Pb age on zircon from this locality is  $443 \pm 7$  Ma, and the monazite age of  $445 \pm 9$  Ma agrees with the zircon age (Walsh and others, 2004). The Candlewood Granite, and smaller granite bodies and dikes, are considered syn-tectonic with the D2 fabric in the area because they post-date some F2 folds in the country rock, yet contain F2 folds of folded igneous flow foliation. The granites have the S2 foliation as the dominant fabric, and generally intrude along the axial surface of F2 folds, sub-parallel to the regional trend of the S2 foliation. A possible explanation for the presence of D2 fabrics both in and out of the granites may be that the D2 structures developed within the intrusion represent a stress field before and during intrusion, and that D2 structures in the country rock experienced post-intrusion tightening within the same stress field. The age of the granite from this locality also agrees with the age of regional Taconian metamorphism of 445 Ma (Hames and others, 1991) and with the age of injected migmatitic leucosomes in the nearby basement rock dated at  $444 \pm 6$  Ma (Walsh and others, 2004). Similar relationships occur near the granite at Yale Farm which intrudes the southwest Berkshire massif in South Sandisfield, Massachusetts just north of Hames and others' (1991) Canaan Mountain samples. At Yale Farm, Zartman and others (1986) report a lower intercept U-Pb zircon age from the granite of  $430 \pm 10$  Ma. The granite intrudes a fault between the basement gneiss and the Canaan Mountain Formation. A titanite age from the granite of ca. 445 Ma confirms Taconian sillimanite grade metamorphism.

#### Mileage

- 18.6 Make a U-turn, and then turn right at the stop sign onto Squash Hollow Road.
- 19.2 Turn right at stop sign onto Route 7 South.
- 20.1 Turn right at traffic light onto Route 37 South.
- 20.2 Turn left onto Candlewood Mountain Road. The pavement outcrop at this road junction is Candlewood Granite. This is also where the Candlewood Mountain Trail crosses Route 37.
- 21.1 Turn right onto Bullymuck Road.
- 21.8 Turn left at stop sign onto Hubbard Mountain Road.
- 23.2 Stop at gate to Candlewood Lake Estates. Consolidate vehicles at this point for Stop 10. We need to get into as few vehicles as possible for the next stop. There is a small parking area to the left (east) of the gate. We will return this way after Stop 10. Proceed through gate.
- 23.4 Turn right onto Eagle's Nest Road.
- 23.6 Park in cul-de-sac for Stop 10.

#### STOP 10. Migmatite gneiss at Eagle's Nest Road (628070E 4601410N).

Samples NM576 A and B of Walsh and others (2004). These outcrops are located around a new house that did not exist when mapping was conducted in 2000. At the time, only the building lot existed, and it was simply referred to as Lot #4. The current owner tells me that his previous house was destroyed by a fire. While the remaining outcrops offer spectacular examples of the migmatite gneiss, the sampled outcrops and photographs for Figures 4 and 5 in Walsh and others (2004) are, unfortunately, located under the new house.

At the outcrop on the north side of the house, you will see stromatic migmatite with relict YS1 gneissosity deformed by Mesoproterozoic YF2 folds, which are in turn deformed by upright Paleozoic Taconian F2 folds. At the back of the house, an outcrop exhibits both gray stromatic migmatite and younger injected pink granitic leucosomes that post-date all the Mesoproterozoic fabrics, but are synchronous with the F2 folds. A sample (NM576B) of the gray stromatic migmatite yielded a SHRIMP U-Pb age on zircon of  $1057 \pm 10$  Ma, and a sample of the younger pink granite (sample NM576B) yielded a SHRIMP U-Pb zircon age of  $444 \pm 6$  Ma (Walsh and others, 2004). It was possible to unravel the complexity of the zircons from these samples only with the ion probe (SHRIMP). The ages of the zircons in the migmatite gneiss yields a mixed age and the rock in fact has two ages. In high grade terrains this complexity is probably widespread leaving the interpretation of protolith ages in such rocks difficult. In as much as the older age migmatite is Ottawan, the question of initial protolith of the migmatite is unresolved. Two inherited cores in the younger pink granite, however, yielded  $^{206}\text{Pb}/^{238}\text{U}$  ages of ca. 1392 and 1296 Ma. Enjoy the view!

#### Mileage

- 23.6 Make a U-turn in the cul-de-sac and carefully head back down the hill on Eagle's Nest Road.
- 23.8 Turn left on Hubbell Mountain Road.
- 24.0 Pick-up vehicles at gate, and continue north on Hubbell Mountain Road.
- 26.5 Turn left at stop sign onto Route 37 South.
- 29.2 Roadcuts of Walloomsac Formation schist.
- 29.5 Continue straight through junction with Route 39.
- 29.6 Small roadcuts of Walloomsac Formation calcite marble.
- 29.7 Turn left onto Saw Mill Road.
- 29.8 Turn left into Veterans Park for Stop 11.

#### STOP 11. Walloomsac Formation at Veterans Park in downtown Sherman (625440E 4603780N).

From the parking lot, walk over to the outcrop near the gazebo and flag pole. The Walloomsac Formation is part of the autochthon, and locally rests unconformably on top of either the Dalton Formation or the basement in the western part of the quadrangle. The Walloomsac Formation consists of a two map units here, schist and calcite marble. The schist is a dark-gray to black, rusty weathering, coarse-grained, chlorite-garnet-muscovite-sillimanite-k-feldspar-biotite-quartz-plagioclase schist. Sillimanite is fibrolitic and locally occurs as porphyroblastic knots replaced by quartz and muscovite. Locally the schist contains muscovite-biotite-quartz-plagioclase granofels, and discontinuous pods and lenses of calc-silicate rock generally consisting of quartz, plagioclase, epidote, diopside, and hornblende. The Walloomsac Formation also contains mappable layers of light-gray to light yellowish white, rusty orange- to tan-weathering, moderately foliated to massive, quartz-tremolite-diopside-phlogopite-calcite marble in layers ranging from less than 1 meter to tens of meters thick. The calcite marble is not exposed at the gazebo outcrop, but blocks of the marble can be seen in the gazebo's foundation. If time permits, you can walk about 50 m upstream from the bridge on Mill Brook near the entrance to the park to an outcrop of marble. Another nearby outcrop of calcite marble is located in downtown Sherman on Route 37 (see Road Log).

Structurally, the dominant foliation at this outcrop is the Paleozoic S1 schistosity. S1 is deformed by NNW striking, steeply west-dipping F2 folds with steeply NW plunging fold axes. While the general trend of the S1 foliation at this outcrop is east-west, the orientation is relatively anomalous in this belt of Walloomsac Formation because the calcite marbles trace generally north-south, sub-parallel to the regional strike of S2. The S1 fabric in this area may be related to emplacement of the allochthons, because D1 produced a penetrative fabric in both the autochthonous cover rocks and the allochthon, and both sequences are deformed by the D2 fabric. The character of D1 fabrics in the autochthon and the allochthon differs, however, as relict isoclinal F1 folds are present throughout the allochthon but were not observed in the autochthon in this area. The allochthonous Manhattan Schist (Stops 12 and 13) contains amphibolites whose distribution is controlled by the D1 fabric, whereas the calcite marbles in the autochthonous Walloomsac Formation are truncated by S1. These findings suggest that D1 fabrics developed diachronously across the area.

#### Mileage

- 29.8 Make a U-turn and turn right onto Saw Mill Road out of Veterans Park.
- 29.9 Turn left at stop sign onto Route 37 South.
- 30.3 Bear left onto Route 39 South.
- 31.9 Cozier Hill Road on right.

- 32.2 Leach Hollow Road on right.  
 32.3 Carefully pull off to the right and park for Stop 12.

**STOP 12. Manhattan Schist amphibolite (626540E 4600900N) (Optional).**

The roadcut at this optional stop shows typical Manhattan Schist amphibolite. Route 39 is narrow and there is limited parking here, and the size of the today's group will dictate whether we can safely make this stop. The amphibolite at these roadcuts forms one of several 10- to 150-m-wide belts of previously unmapped amphibolite within the Briggs Hill slice, a large allochthon that rests either on Walloomsac Formation or directly on basement. The distribution of the amphibolites follows the folded S1 schistosity along the length of the peninsula between Squantz Pond and Lake Candlewood, north through Briggs Hill and west into the adjacent Pawling quadrangle. The Briggs Hill slice probably extends southward into the Danbury quadrangle where Clarke (1958) mapped several belts of amphibolite from Candlewood Isle to Danbury Bay. Rodgers (1985), however, ended the Briggs Hill slice north of Candlewood Isle and placed the amphibolite-bearing rocks in the Dalton Formation on the State map which is not typical for the Dalton Formation, suggesting that the distribution of allochthonous rocks in the Danbury quadrangle needs to be re-examined.

**Mileage**

- 32.3 Continue south on Route 39.  
 34.9 Pull off and park to the right across from Inglenook Road for Stop 13.

**STOP 13. Manhattan Schist on Squantz Pond (627400E 4596620N).**

Walk around the fence behind the mailboxes down to the water's edge for an outcrop of Manhattan Schist. The roadcut along Route 39 also shows typical Manhattan Schist (of Rodgers, 1985), but the shoulder is narrow and traffic is dangerous along Route 39. The schist consists of gray to dark-gray, rusty weathering, very coarse grained  $\pm$ sillimanite $\pm$ garnet-muscovite-biotite-quartz-plagioclase schist with abundant quartz and k-feldspar leucosomes, pegmatite, and quartz veins. Early, rootless, F1 isoclinal folds are visible down near the water's edge. The S1 schistosity is also deformed by outcrop-scale open to tight, N20°E striking F2 folds -- a large one is visible near the concrete steps.

**Mileage**

- 34.9 Continue south on Route 39.  
 35.0 Cross causeway over Squantz Pond.  
 35.2 Turn right onto Beaver Bog Road (no sign). You will see a sign for Squantz Pond State Park.  
 36.2 Turn right onto Rocky Hill Road.  
 36.9 Turn left onto Mountain Laurel Drive.  
 37.1 Turn right onto Laurelwood Drive.  
 37.3 Turn around in cul-de-sac  
 37.4 Park on right across from mailbox at #8 for Stop 14.

**STOP 14. Danbury augen granite in the Hudson Highlands massif (625570E 4596030N).**

Sample NM100 of Walsh and others (2004). Clarke (1958) named the Danbury augen granite for exposures on the west side of Lake Candlewood in the Danbury quadrangle just south of this locality. The rock is a coarse-grained, weakly to moderately well-foliated  $\pm$ hornblende-biotite-quartz-plagioclase-microcline granite gneiss with deformed phenocrysts of microcline up to 4 cm across and 10 cm long. Chemically the rock is a peraluminous calc-alkaline granite, with a within-plate-granite signature (Walsh, 2003). The Danbury augen granite does not possess the older (YS1) gneissosity seen rarely in the other basement gneisses, but xenoliths within the augen granite do contain the older fabric suggesting that the augen granite entirely post-dates the earlier deformation event. The Danbury augen granite is variably deformed by the dominant Mesoproterozoic foliation (YS2). Contacts with the adjacent rock units are not exposed, but the augen granite contains xenoliths of amphibolite, hornblende gneiss, layered biotite gneiss, and pink granite gneiss. The augen granite occurs in three separate belts in the New Milford quadrangle; the two eastern belts are located in the New Milford massif and exhibit a greater degree of Taconian D2 overprint due to deformation along the Fort Mountain fault. Taconian F2 folds seen at Stop 13 deform the YS2 gneissosity into open to tight folds and strike NNE and dip steeply west. The SHRIMP U-Pb age on zircon is 1045  $\pm$  8 Ma (Walsh and others, 2004). The Danbury augen granite is interpreted as a late to syn-tectonic Ottawan YD2 intrusion because it post-dates YD1 deformation, contains the YS2 gneissosity, and closely agrees or is slightly

younger than Ottawaan migmatization (ca. 1057 Ma) and widespread development of granitic intrusions (ca. 1050 Ma).

#### Mileage

- 37.4 Continue south on Laurelwood Drive.
- 37.5 Turn left at stop sign onto Mountain Laurel Drive.
- 37.6 Turn right at stop sign onto Rocky Hill Road.
- 38.3 Turn left at stop sign onto Beaver Bog Road.
- 39.3 Turn left at stop sign onto Route 39 North.
- 44.4 Bear right at stop sign onto Route 37 North.
- 44.8 Continue straight on Route 37 North at junction with Route 39 North.
- 46.9 View of Candlewood Mountain at 12 o'clock.
- 47.9 Turn right at stop sign onto Route 7 South.
- 50.5 Big Y supermarket.  
Turn left at light onto Route 202 East and Route 67 South.
- 51.0 New Milford village green on left.
- 51.1 Bear right onto Route 67 South.
- 52.2 Turn right onto Town Farm Road.
- 52.5 Turn right onto Meredith Lane.
- 52.7 Park on right for Stop 15.

#### STOP 15. Brookfield Gneiss (633640E 4601920N).

Sample NM687 of Walsh and others (2004). The Brookfield Gneiss crops out as a large pluton that extends southward into the Danbury quadrangle east of Cameron's Line. Clarke (1958) and Sperandio (1974) recognized several phases within the Brookfield Gneiss that range from granite to diorite. The Brookfield Gneiss intrudes previously deformed allochthonous Cambrian to Ordovician metasedimentary and metavolcanic rocks of the Rowe and Ratlum Mountain Schists. The new mapping in the New Milford quadrangle (Walsh, 2003) and previous work by Sevigny and Hanson (1995) indicate that the Brookfield post-dates isoclinal folds in the allochthonous rocks. Walsh (2003) indicates that the pluton cuts two generations of isoclinal folds, both of which are presumed to be of Taconian origin (F1 and F2). Similar to the Candlewood Granite, the Brookfield Gneiss appears to be a syn-tectonic pluton that intruded during the development of the second-generation folds in the host rocks. A conventional composite U-Pb zircon age of  $453 \pm 3$  Ma was first reported from three dioritic samples of the Brookfield Gneiss (Sevigny and Hanson, 1995). At this stop you will see fine-grained, well-foliated aplite dikes intruding the diorite phase of the Brookfield Gneiss. The sampled leucocratic granite dike yielded a SHRIMP U-Pb age on zircon of  $453 \pm 6$  Ma (Walsh and others, 2004), which is the same as the composite age reported by Sevigny and Hanson. The aplite dikes are also considered syn-D2 because they intrude the Brookfield Gneiss parallel to the S2 schistosity yet contain the S2 foliation as the most conspicuous planar fabric. The leucogranite dike has a depleted REE chemistry that is distinct from rocks west of Cameron's Line (Walsh, 2003), and has a volcanic arc to within plate signature, unlike the Ordovician granitic rocks to the west which clearly intrude Laurentian basement.

#### Mileage

- 52.7 Continue uphill on Meredith Lane and make a U-turn at the cul-de-sac.
- 53.1 Cul-de-sac
- 53.6 Turn right at stop sign onto Town Farm Road.
- 54.2 Stay straight onto Hine Hill Road. Town Farm Road goes left.
- 54.9 Turn left at stop sign at bottom of hill onto Grove Street (no street sign).
- 55.2 Turn left onto Lovers Leap Road (no street sign) before bridge over Housatonic River.
- 55.3 Park at gate and old bridge for entrance to Lovers Leap State Park at Stop 16.

#### STOP 16. Ratlum Mountain Schist at Lovers Leap State Park (632850E 4600200N).

From the parking area, walk under the old bridge abutment on the east side of the river. Rocks at this stop include biotite schist and amphibolite of the allochthonous Ratlum Mountain Schist. The schist is silvery gray to medium- or dark-gray, tan- to light-gray weathering, medium grained,  $\pm$ staurolite $\pm$ kyanite-garnet-muscovite-biotite-plagioclase-quartz schist that locally contains thin (up to 30 cm) interlayered muscovite-biotite-plagioclase-quartz granofels and amphibolite. The amphibolite layers are mapped separately, and a small outcrop of amphibolite crops

out a few meters south of the east end of the old bridge, at the top of the steep slope to the river. If time permits, walk south on the footpath along the east side of the gorge. Just before you reach the sharp turn to the east, you will see larger outcrops of amphibolite uphill along the east side of the trail. This amphibolite traces up hill to the northeast where it crops out on the western slopes and summit of the small un-named hill, and widens to approximately 200 m. Where the main trail turns east, follow the smaller foot path down to the water's edge where the gorge opens up into Lake Lillinonah. Well-layered biotite schist and granofels are exposed at the point across from Lovers Leap (633030E 4599800N).

The dominant foliation at these outcrops is the S2 schistosity. Well-developed down-dip L2 lineations and rootless F2 folds are evident in the outcrops under the east side of the old bridge. The planar S2 fabric here is the dominant foliation in most rocks east of Cameron's Line, which traces parallel to the regional S2 fabric along the entire length of the New Milford quadrangle. Locally, rootless F1 folds can be found within the dominant S2 fabric within these rocks. Late, Acadian F3 folds with variable strikes and gently dipping axial surfaces, locally deform the S2 foliation, and are probably related to dome-stage folds which increase towards the east.

### End of Trip

Directions back to Route 7.

### Mileage

- 55.3 Make a U-turn, and turn left at stop sign onto Pumpkin Hill Road / Grove Street (no street sign).
- 55.4 Cross new bridge over Housatonic River / Lake Lillinonah.
- 55.9 Go straight at four-way stop sign onto Still River Road.
- 56.2 Junction with U.S. Route 7, turn left at traffic light for I-84 and points south.

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# GIANT STAUROLITE PORPHYROBLASTS IN THE BOLTON SYNCLINE: TECTONOMETAMORPHIC IMPLICATIONS

by

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## INTRODUCTION

In 1992 we hosted an NEIGC field trip dedicated to staurolite porphyroblasts in the Bolton syncline. At that time, a lively controversy was taking place - whether or not porphyroblasts rotate (Bell et al. 1992, Passchier et al. 1992). Our studies of the staurolite porphyroblasts in the Bolton syncline favored the interpretation that they had rotated during their growth (Busa and Gray 1992a). This claim did not sit well with the group who were strongly contending against porphyroblast rotation, and they soon after came to Bolton Notch, collected some samples, and provided an interpretation of the structural and metamorphic history Bolton syncline, including their claim that the staurolite porphyroblasts did not rotate (Bell et al. 1997, Hickey and Bell 1999).

This field trip shall include visits to some of the same outcrops in the Bolton syncline used by us and Bell and others both to observe interesting external forms of staurolite, and to provide the backdrop for discussions of porphyroblast rotation vs. non-rotation, and the validity of their FIA method. Also, we shall explore connections between staurolite orientation, morphology, size, and style of local deformation in the syncline.

## THE BOLTON SYNCLINE

The Bolton syncline in Connecticut is located between 2 to 10 km east of the Eastern Border fault of the Mesozoic Hartford Basin (Figure 1). The rocks comprising the Bolton syncline are the Clough Quartzite, the Fitch Formation, and the Littleton Formation (Bolton Group, Rodgers 1985). It is mapped as a tightly folded syncline overturned to the east. Structurally overlying the syncline to the west is the Glastonbury Gneiss. Aitken (1955) reports the Glastonbury Gneiss - Bolton Group contact is conformable at the southwest portion of the Rockville quadrangle, but the contact becomes angularly unconformable in the northern portion of the quadrangle - we have not found such a contact. The contact between the Clough and the Ordovician Collins Hill Formation (structurally underlying the syncline) can be observed at Bolton Notch (Figure 2). The Bolton syncline can be traced from Cobalt, Connecticut (approximately 25 km south-southwest of Bolton Notch) along a north-northwesterly direction to Monson, Massachusetts, "forming apparently the most extensive continuous Primary mountain range in the State" (Percival 1842). The width of the syncline averages approximately 1 km, being the widest in the Bolton area (2 km), and the most narrow in West Stafford (1/2 km).

The structural and metamorphic history of the Bolton syncline is complex and poorly understood. Even the timing of the staurolite metamorphic event is controversial. Rosenfeld and Eaton (1985) suggested that staurolite growth post-dated the formation of the syncline, whereas Robinson and Tucker (1982) argued that the folding post-dates the staurolite metamorphic event. Bell et al. (1997) and Hickey and Bell (1999) argue for several stages of staurolite growth during and after the formation of the syncline. As will be seen in this report, these arguments are in conflict with the structural data inferred from the staurolite porphyroblasts from the visits herein.

## STAUROLITE PORPHYROBLASTS

The morphology of the staurolite porphyroblasts in the Bolton syncline is variable. Among the distinct habits found are euhedral to subhedral prisms, large (typically greater than 10 cm) millimeter-thick {010} pinacoidal lozenge-shaped plates, and splayed, fan-shaped staurolite porphyroblasts (Busa and Gray, 1992b). The staurolite porphyroblasts occur along with 1-2 mm diameter euhedral garnet in pelitic schists of the Littleton Formation, and weather in positive relief forming conspicuous stubby prisms and, in places on the west limb, splayed fan-shaped groupings.

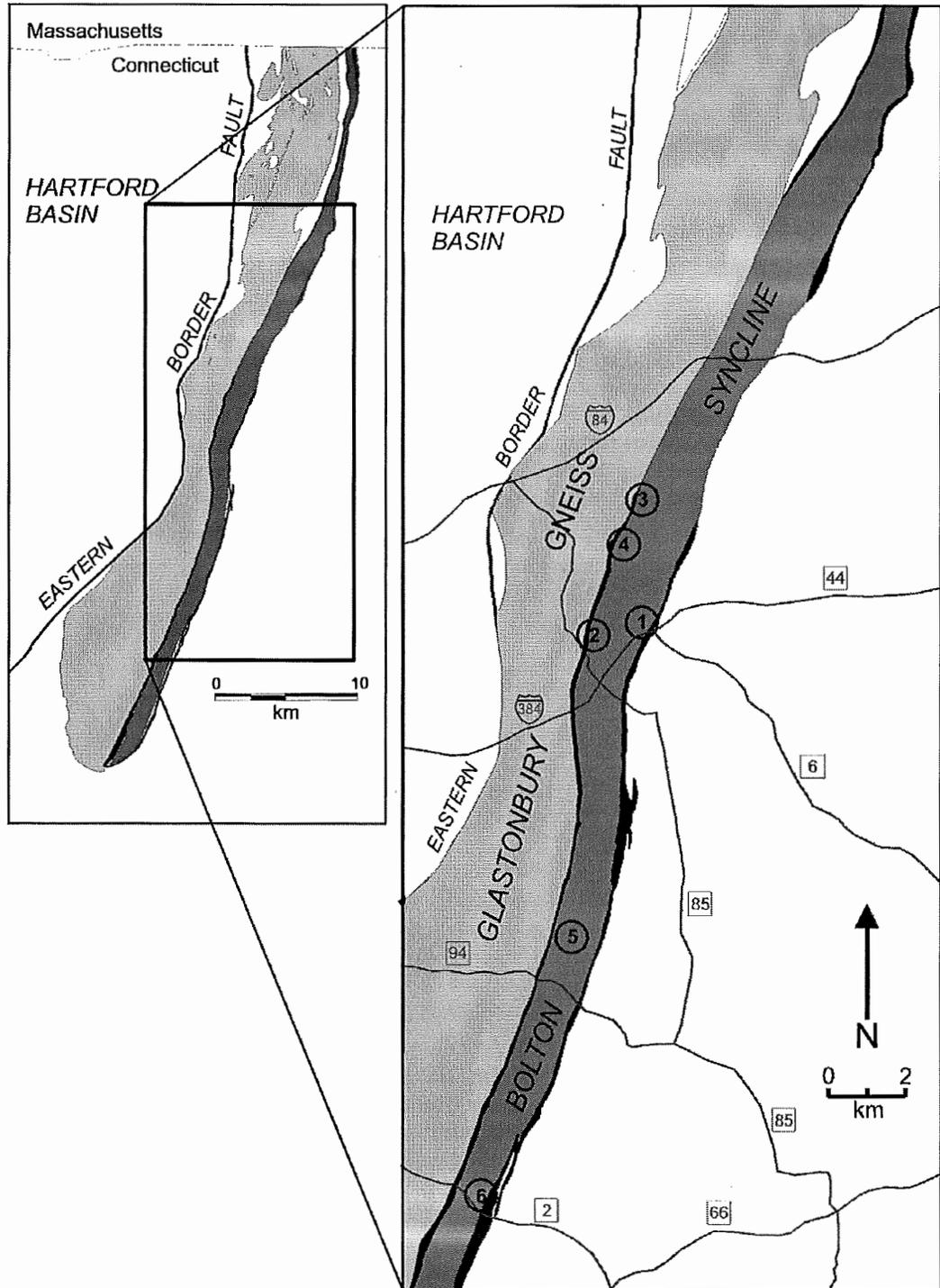
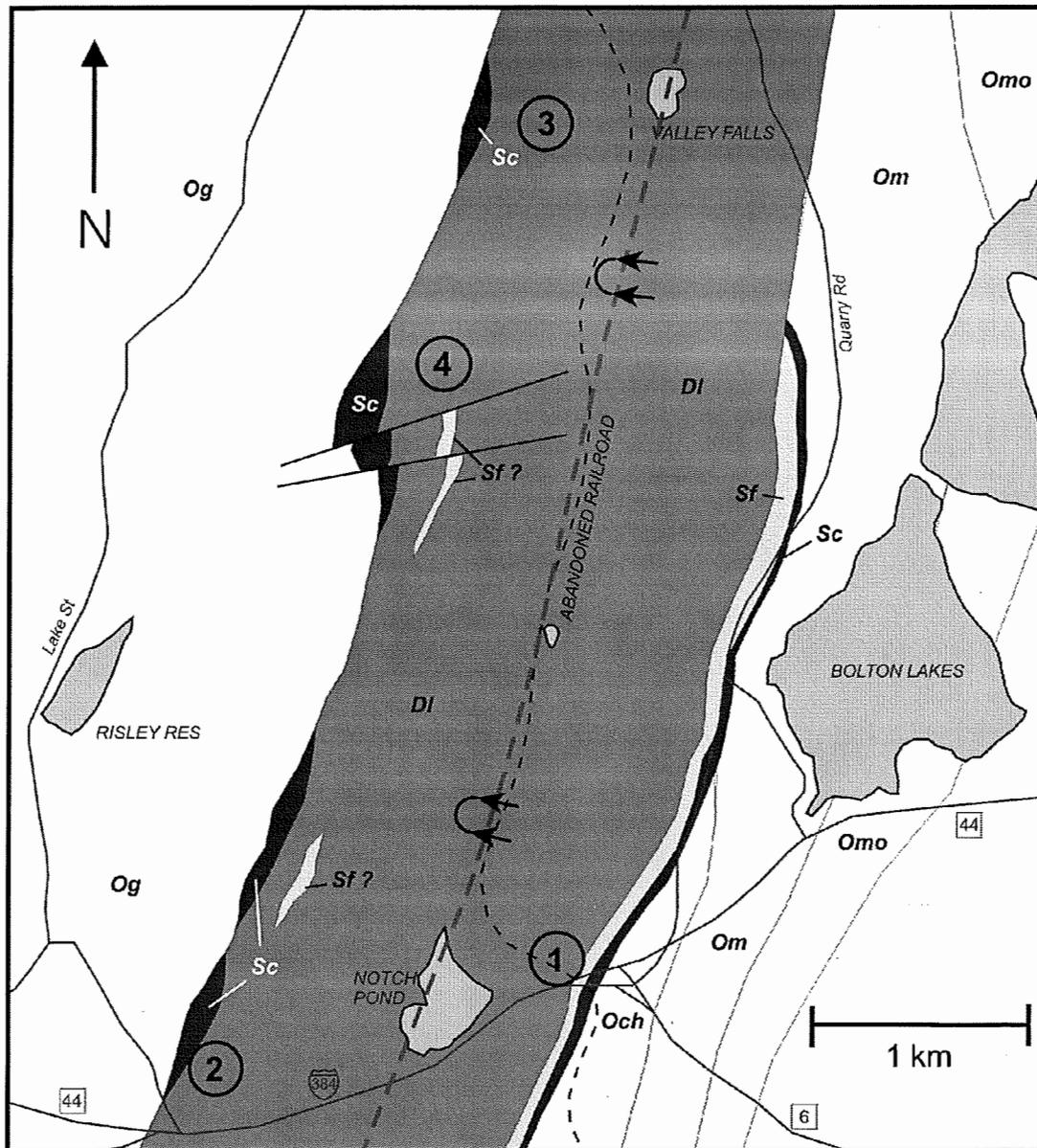


Figure 1. Location of Bolton syncline, and Glastonbury Gneiss in eastern Connecticut. Rocks immediately east of the Bolton syncline (not shown) are Ordovician metasediments and metavolcanics. Enlarged view shows locations of field stops as numbers in circles. Progression of field trip will be in order from STOPS 1 through 6. (Map after Rodgers 1985).



Legend

- |   |   |  |   |
|---|---|--|---|
| <span style="border: 1px solid black; padding: 2px;">DI</span> Littleton Fm       | Bolton Group  | <br>Trace of synclinal axis  |   |
| <span style="border: 1px solid black; padding: 2px;">Sf</span> Fitch Fm           |   |  |   |
| <span style="border: 1px solid black; padding: 2px;">Sc</span> Clough Quartzite   |   |  |   |
| <span style="border: 1px solid black; padding: 2px;">Og</span> Glastonbury Gneiss | <span style="border: 1px solid black; padding: 2px;">Och</span> Collins Hill Fm | <span style="border: 1px solid black; padding: 2px;">Om</span> Middletown Fm | <span style="border: 1px solid black; padding: 2px;">Omo</span> Monson Gneiss |

Figure 2. Geological map of a portion of the Bolton syncline (modified after Rodgers 1985). Bolton Notch is located at STOP 1; Box Mountain is along the west limb of the syncline stretching north-northeasterly from STOP 2 to STOP 3. STOPS 5 and 6 are south of this map (see Figure 1).

The large size of the staurolite porphyroblasts allow one to perform relatively simple techniques to measure their orientations in the rock and make detailed thin-sections to study the inclusion trails. It has been our experience that thin sections cut perpendicular to the long dimension (*c* axis) of the staurolite porphyroblasts shows the inclusion trails to be the simplest and most straight-forward to interpret. This is in contrast to the confusing and complex inclusion trails seen staurolite intercepted in random sections through the rock.

### Inclusions and Inclusion Trails

Staurolite porphyroblasts preserve an earlier foliation ( $S_i$ ) as distinct inclusion trails, which predate the present, dominant rock foliation  $S_e$ . Although  $S_i$  in some staurolites can be seen to be a crenulation of an earlier foliation we see no compelling evidence within any of the Bolton staurolite porphyroblasts for the  $S_3$ ,  $S_4$  and  $S_5$  inclusion trails identified by Hickey and Bell (1999).

Trails of inclusions are a common feature and are preserved in the interior of all staurolite porphyroblasts. Two distinct types of trails are observed: one is a relict of an overgrown fabric; the other is produced by some type of segregation at growth faces. Inclusion trails such as these have been described by Harker (1939), Anderson (1984), and Rice and Mitchell (1991). Rice and Mitchell (1991) introduced the term "type 1" inclusions for those representing a captured foliation, and "type 2" for those related to a growth phenomena. In the staurolite porphyroblasts, the type 1 inclusions are defined by discoid-shaped quartz blebs, planar layers of black carbonaceous material, and ilmenite platelets. The size of the type 1 quartz blebs and spacing of the planar zones of fine carbonaceous material probably reflect the grain size and scale of the overgrown foliation. The type 2 inclusions are defined by even finer discontinuous tubular shaped quartz grains. Typically these trails are oriented perpendicular to a crystal face. The ilmenite inclusions appear to have been inert and are associated exclusively with type 1 inclusion trails and are the most reliable markers of not only the trace, but as they are plates, the true orientation of a relict foliation ( $S_i$ ) (Figure 3).

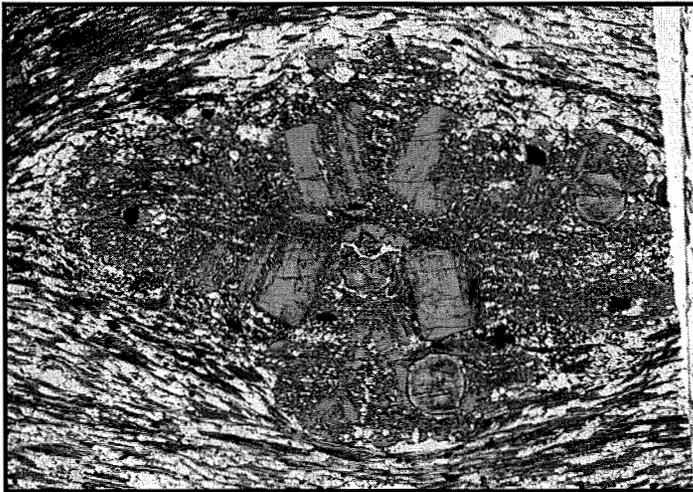


Figure 3. Type 1 and type 2 inclusion trails in staurolite porphyroblast. East-west oriented inclusions are type 1; north-south radiating inclusions are type 2. Ilmenite plates follow the type 1 inclusion trails. Width of section 12 mm. East limb Bolton syncline, approximately 1 km south of Bolton Notch.

In some staurolite porphyroblasts from the west limb, the type 1 inclusion trails bend abruptly into near parallelism with the external foliation ( $S_e$ ) at the porphyroblast margins. This sharp curvature of the trails is preserved in texturally distinct euhedral marginal zones of staurolite, characterized by a fine dusting of graphite (Figure 4). These graphite-rich overgrowth rims capturing an abrupt curvature of the rim  $S_i$  are also observed in the fan-shaped staurolite porphyroblasts. In the field, these staurolite have finely preserved, glassy crystal faces. Interestingly, such overgrowths are rare in staurolite porphyroblasts on the eastern limb of the Bolton syncline (Busa and Gray 2002).

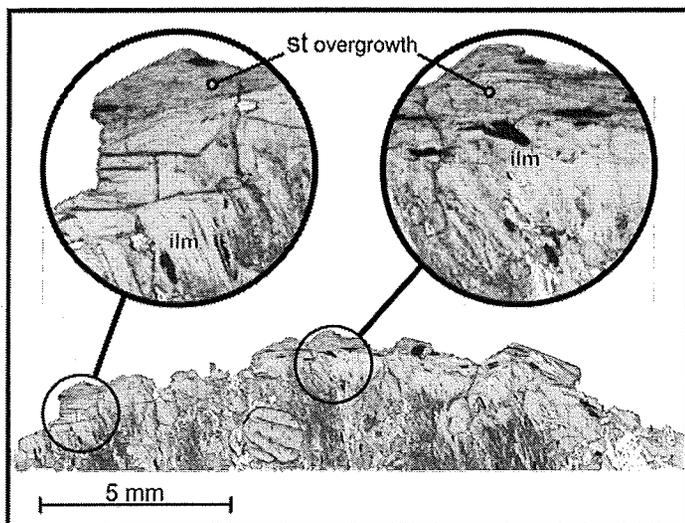


Figure 4. Sharp deflections of type 1 inclusion trails preserved within marginal staurolite (st) overgrowths characterized by dispersed inclusions of fine carbonaceous material (insert). The larger black 0.5 x 0.05 mm, plate-like inclusions are ilmenite (ilm).

What is the nature of the captured foliation ( $S_i$ )? The type 1 inclusion trails in staurolite range from being a coarse, crenulated  $S_{i-1}$  consisting of quartz blebs and ilmenite platelets, as observed in the staurolite porphyroblasts from Marlborough (southern end of the syncline, Figure 5a) to moderately coarse discoid shaped quartz and ilmenite plates (Figure 5b) observed in staurolite from various locations in the syncline. At some locations, the type 1  $S_i$  consists of a dusting of very fine carbonaceous material (graphite) and quartz, representative of a much finer foliation (slaty cleavage) in the rock (Figure 5c). Still in other places, staurolite traps a fine to coarse planar  $S_i$  consisting of fine carbonaceous material and ilmenite plates (Figure 5d). All sections in Figure 5 are cut perpendicular to the staurolite  $c$  axis; inclusion trails in oblique sections are confusing and complex (Figure 6).

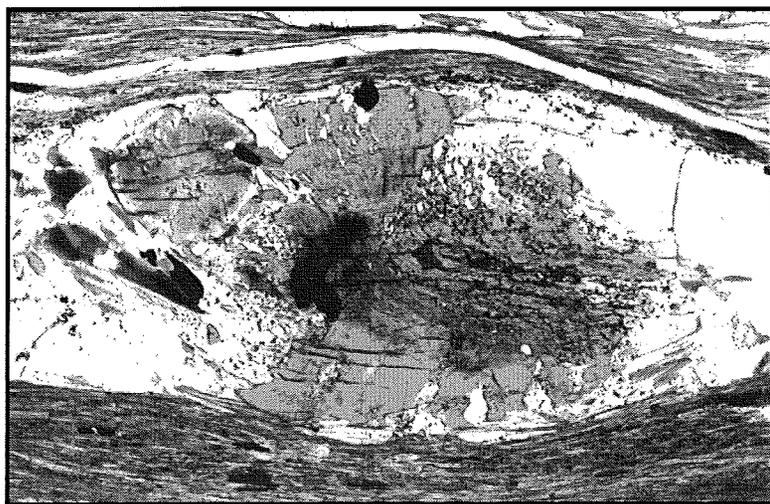


Figure 6. Vertical section of Littleton schist, intersecting staurolite porphyroblast at an oblique angle. Inclusion trails are confusing and difficult to interpret.

**Sigmoidal-Shaped Inclusion Trails.** Sigmoidal shaped  $S_i$  have been found in staurolite porphyroblasts with rotations relative to  $S_e$  up to 130 degrees. Trails are best defined and most distinct in  $c$  axis sections because  $S_i$  surfaces, even in sigmoids, are perpendicular to the  $c$  axis. Serial sections through four differently-oriented staurolite porphyroblasts show the amount of turn, as defined by the curvature of the sigmoidal shaped inclusion trails, increases with distance from the nucleation point in the staurolite (Busa and Gray 1992a). These observations suggest staurolite nucleated with its long dimension ( $c$  axis) in a foliation and rolled along its  $c$  axis as it grew. The above observations hold regardless of the staurolite's orientation in the rock. This also argues against there being a penetrative foliation intersection axis (FIA) in the rock (Bell et al. 1995). Rather, the staurolite preserves its own three dimensional sigmoidal-shaped inclusion trail, parallel to its  $c$  axis.

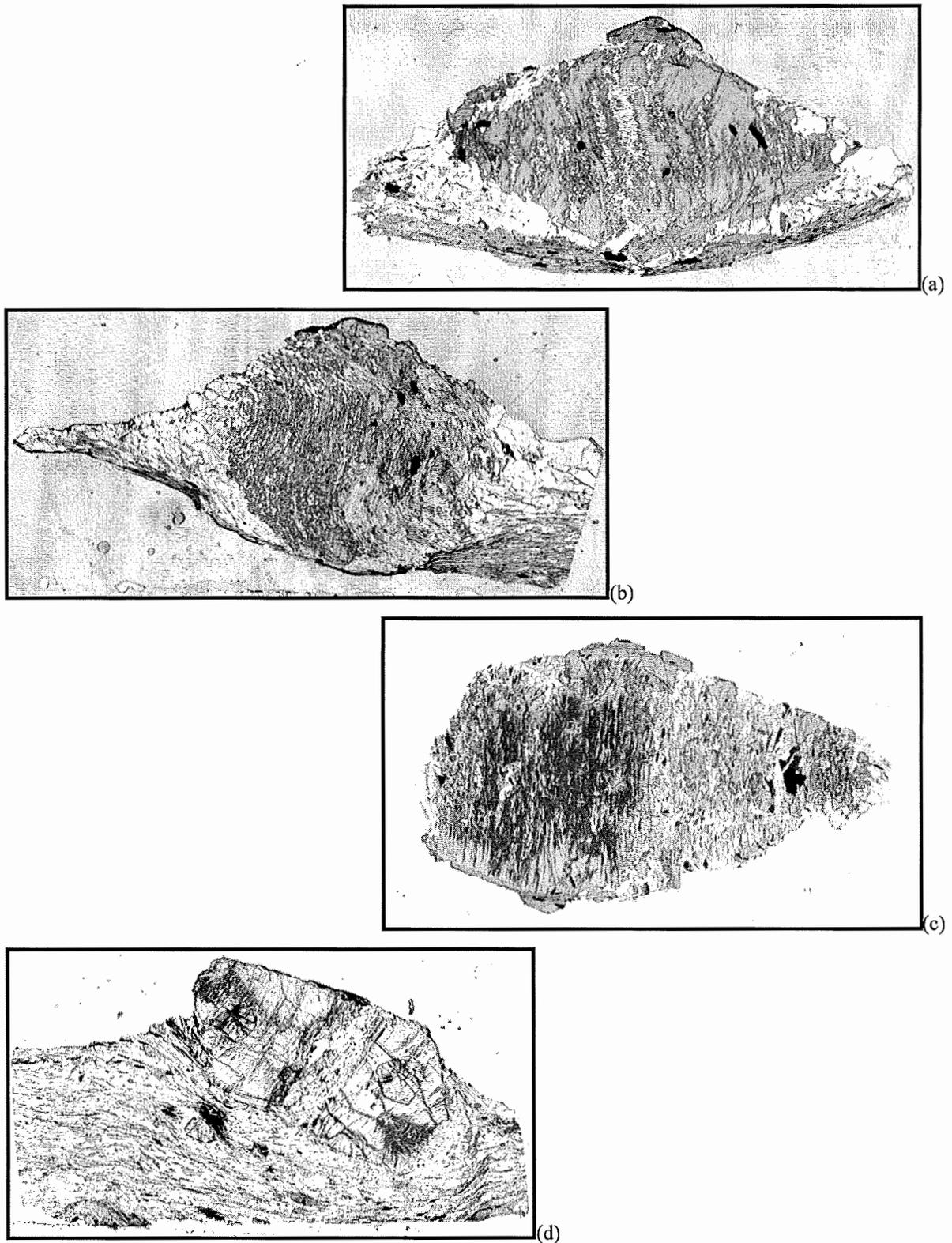


Figure 5. Types of captured foliations in staurolite porphyroblasts from the Bolton syncline. (a) a coarsely crenulated schistosity ( $S_{I-1}$ ), (b) coarse quartz blebs and discoids, (c) a very fine slaty cleavage, and (d) planar trails.

**Orientation of  $S_i$ : Crystallographic Reference Frame.** The angular relationship between the planar or discoid type 1 core inclusion trail surfaces and the crystallographic orientations of staurolite prisms and fans were documented by the U-stage measurements. The angle between the  $c$  axis and the core inclusion planes are compared in Figure 7. In both prismatic and fan-shaped staurolite porphyroblasts, the  $c$  axis consistently lies at a small angle to  $S_i$ , suggesting that staurolite preferentially nucleated with its  $c$  axis in the plane of a pre-existing foliation. The fans however differ from the prismatic porphyroblasts in that their  $b$  axis is generally perpendicular to  $S_i$  (Figure 8). The prismatic porphyroblasts seem to have nucleated with crystallographic  $b$  at all angles to  $S_i$ . It appears that only the staurolites with the included  $S_i$  nearly parallel to the  $\{010\}$  cleavage had the potential to develop into fans.

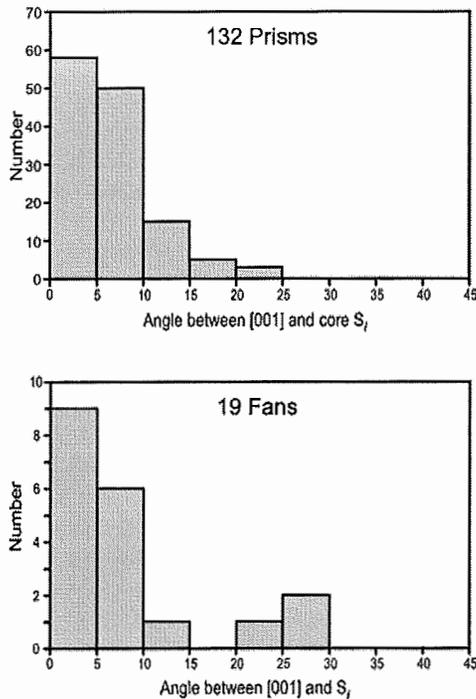


Figure 7. Histograms of 132 prismatic staurolites and 19 fans showing the distribution of the angle between crystallographic  $c$  axes and the core  $S_i$  inclusion surface. The fact that the angle is generally less than 15-20 degrees suggests staurolite in the Bolton syncline nucleated with its  $c$  axes in the plane of the early  $S_i$  foliation.

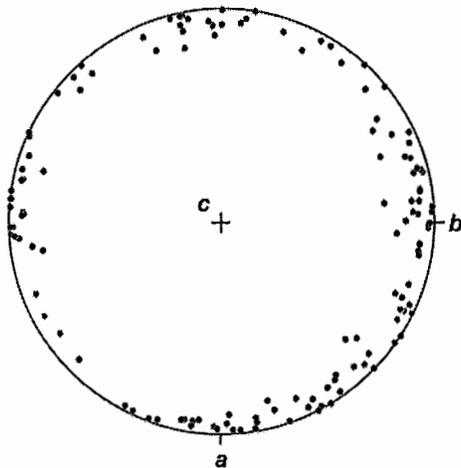


Figure 8. Poles to core  $S_i$  inclusion surfaces in 132 prismatic staurolite porphyroblasts from the Bolton syncline.  $S_i$  poles are plotted relative to staurolite's crystallographic directions  $a$ ,  $b$ , and  $c$ . Note that in the staurolite prisms, the  $S_i$  poles define a girdle about  $c$  with minor maximum near  $a$  and  $b$ , whereas the inclusion surfaces in the fans tend to be approximately parallel to  $(010)$  with their poles concentrated close to  $b$ .

**Staurolite Orientations and  $S_i$  Asymmetries**

There is quite a range of staurolite orientations in the foliation. On the east limb of the Bolton syncline at Bolton Notch (STOP 1) the long dimension of staurolite porphyroblasts ( $c$  axes) define a crude preferred orientation as shown in Figure 9a. Orientations of staurolite  $c$  axes at STOPS 2, 3, 4 and 6 are shown on the left-hand

stereoplots as Figures 9b, c, d, and e, respectively. The crude west-northwesterly preferred orientation of staurolite shown in Figure 9a is observed at all locations within a 1 km radius at Bolton Notch. A north-south preferred orientation of staurolite *c* axes has been measured at locations on the west limb at Box Mountain (Figures 9b and 9d), the exception being at STOP 3 (Figure 9c).

The stereoplots on the right-hand side of Figure 9 show the cyclographic projection of the staurolite's *c* axis, and the asymmetry, as defined by the sense of rotation from the curved or sigmoidal shaped  $S_i$  in type 1 inclusion trails. On the east limb at Bolton Notch, the switch in asymmetry is very tightly defined (Figure 9a). At the north end of Box Mountain (STOP 4 Figure 9d) some staurolite contain planar  $S_i$  with no discernible asymmetry and the switch in the asymmetry is more diffuse.

If these asymmetries are generated by staurolite rotation about their *c* axes, the direction staurolite does not roll is the transport direction (aka Hansen pole, Hansen 1971). Figure 10 is a schematic illustration of the relationship between the orientations of the prismatic and fan-shaped staurolite porphyroblasts in the Box Mountain schists, just prior to the late stage deformation responsible for the dominant external foliation  $S_e$ . As they grew, the prismatic porphyroblasts, which had nucleated with their long axes in an early fine foliation ( $S_i$ ), rotated in response to a simple shear component parallel to  $S_i$ . Figure 9 shows there is a relationship between the Hansen pole, as defined by  $S_i$  asymmetries, and the long axes of the prismatic staurolite porphyroblasts.

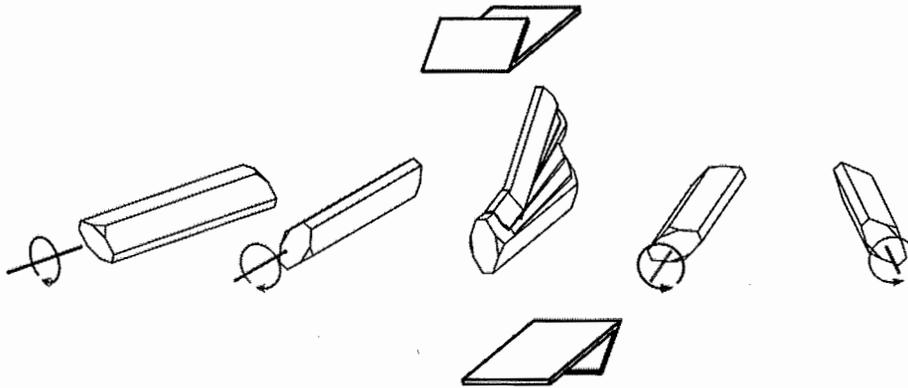


Figure 10. Orientations of prismatic and fan-shaped staurolite porphyroblasts prior to the deformation responsible for the dominant external foliation  $S_e$ . As they grew the prismatic porphyroblasts rotated in response to a simple shear component parallel to  $S_i$ . Prisms to the right of the transport direction (the Hansen pole) rotated about their long axes in a counterclockwise sense, while those on the left rotated clockwise. Porphyroblasts lying parallel to the transport direction, with their *b* axis roughly perpendicular to the shear plane formed fans rather than rotating about their long dimension.

### FAN-SHAPED STAUROLITE PORPHYROBLASTS

In addition to the common prismatic habit, staurolite in the Bolton Notch area (Box Mountain) also occurs as fan-shaped aggregates of radiating, centimeter sized prismatic crystals. Although their inclusion trails are neither sigmoidal nor curved, the internal and crystallographic features of these fans are consistent with shear-induced rotation of the individual crystals and thus provide an independent line of evidence that the staurolite porphyroblasts in the Bolton syncline rotated during growth.

Well-developed staurolite fans are rare and are recognized only at a few localities on the western overturned limb of the syncline. The fan outcrop at STOP 4 is located approximately 0.5 km from the western contact of the syncline with the Glastonbury Gneiss at Box Mountain (Figure 2 - UTM Zone 18 coordinates 4631638N, 711508E NAD 83). Figure 11 is a schematic illustration of a typical Box Mountain fan viewed in sections parallel and perpendicular to the fan plane. Inclusion trails within the fans radiate from the apex and, although not strictly crystallographic, are roughly perpendicular to the face of the fan. As drawn, the fan in Figure 11 opened in a clockwise direction as shown by the arrow. The edges of a fan, shown by arrows, subtend the apical angle. In this illustration the fan plane lies at an angle of -20 degrees from the  $+b$  crystallographic direction. The rotation axis

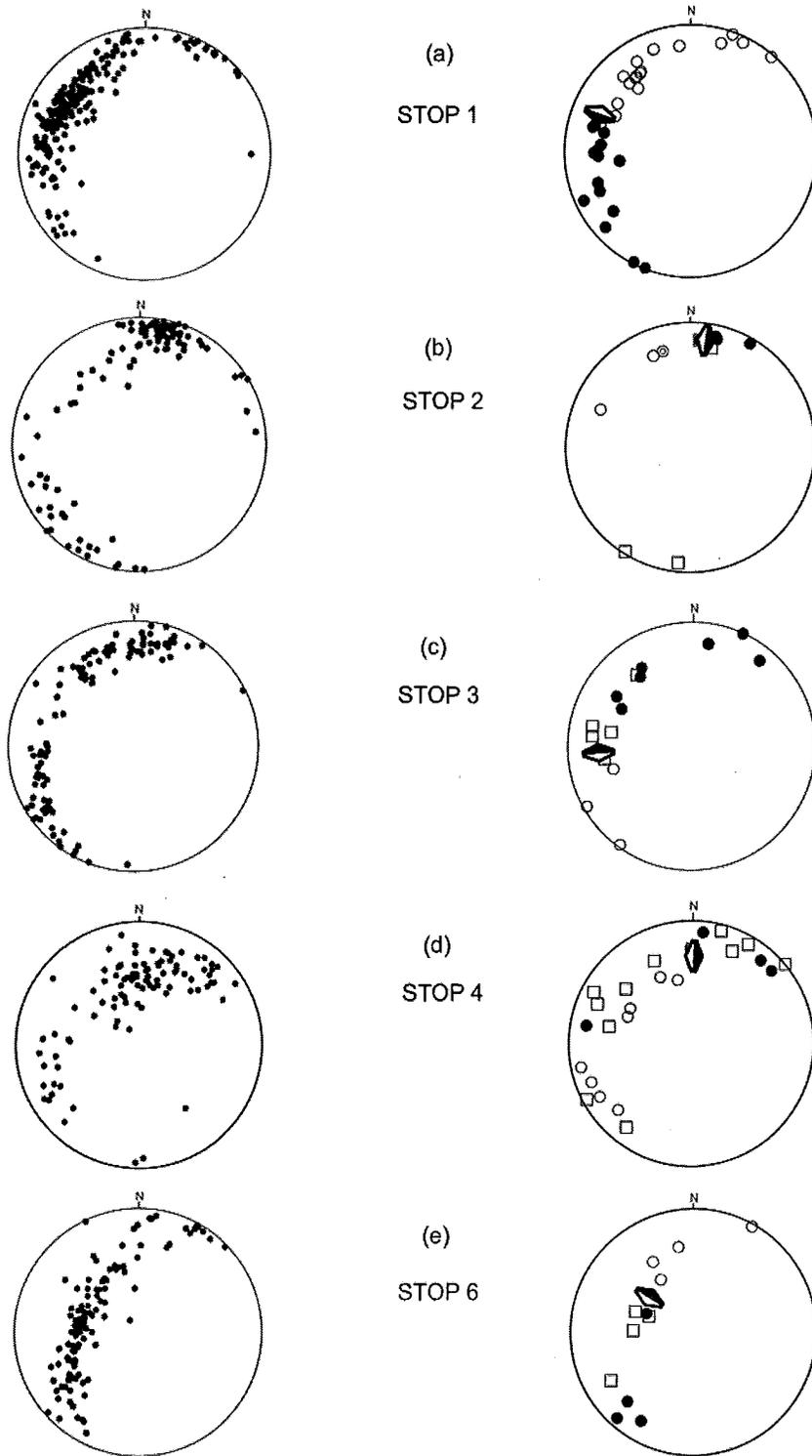


Figure 9. Orientations of staurolite porphyroblast's *c* axes (left-hand stereoplots) and asymmetries of core  $S_i$  (right-hand stereoplots). Solid circles –  $S_i$  with clockwise rotational sense; open circles –  $S_i$  with counterclockwise sense; open squares – planar trails with no discernible asymmetry. Transport direction (aka Hansen pole), as defined in Figure 10, is plotted as black and white staurolite symbol. (a) east limb at Bolton Notch, (b) west limb at Bolton Notch, (c) west limb at Boulder Ridge Park, (d) west limb at Fan outcrop, and (e) east limb at Marlborough. Lower hemisphere equal area projections.

corresponding to the 'fan-law' is also shown. This axis normal to the plane of fanning is also perpendicular to [001] in all the blades of the fan, and is controlled by the  $S_i$ .

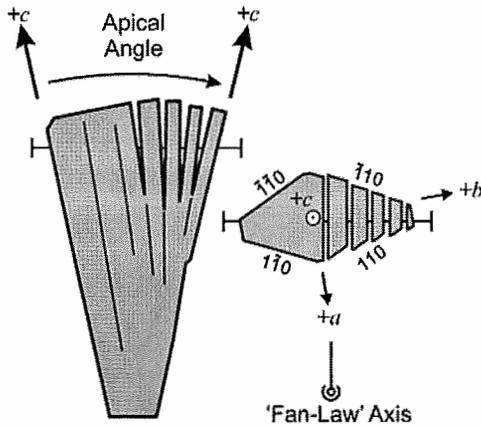


Figure 11. Schematic illustration of a typical Box Mountain fan. Defining the direction of  $+c$  (away from the fan apex) fixes the directions of  $+a$  and  $+b$  relative to the fan grouping. The edges of a fan, shown by arrows, subtend the apical angle.

The orientation of 40 fans and 110 prismatic staurolite porphyroblasts were measured in the field on the fan outcrop. Poles to the planar surface of staurolite fans define a single maximum at a small angle to the foliation pole (Figure 12). The long axes of the prismatic staurolite porphyroblasts at the fan outcrop on Box Mountain form a maxima at  $010 \rightarrow 20$  (Figure 9d - STOP 4). The prismatic staurolite porphyroblasts described in STOP 2 on Box Mountain (Figure 9b) show a similar distribution.

Poles to the fan surfaces shown in Figure 12a define only one aspect of their three dimensional orientation. A complete description of the fan geometry requires defining the vector orientation of both fan edges (Figure 11). Field measurements of fan orientations are represented on the stereographic projections in Figure 12b by the great circle arcs joining the poles to their edges. Although the opening direction of many of the fans is evident in thin sections it is difficult to deduce from the external morphology alone. The few fans in which a macroscopic blade imbrication is well defined, show a consistently counterclockwise sense of opening.

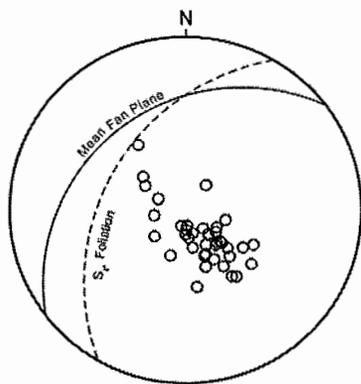
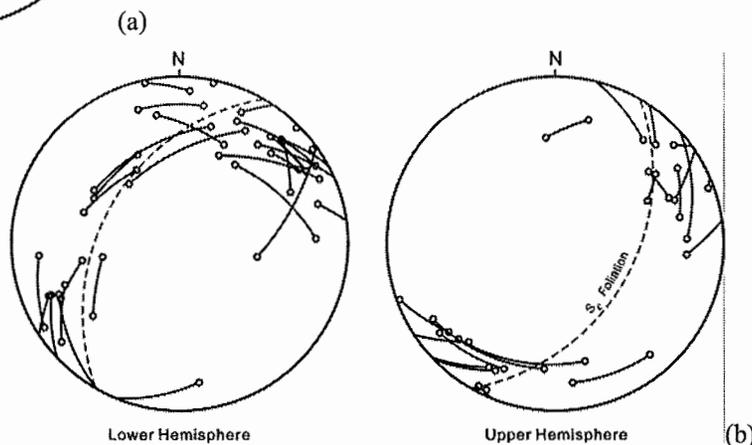


Figure 12. (a) Stereographic projection of poles ( $\circ$ ) to the fan surfaces of Box Mountain fans. Also plotted are the great circles corresponding to the cyclographic projections of the mean fan plane (solid line) and the  $S_e$  foliation (dashed line). (b) Stereographic projections of the poles ( $\circ$ ) to edges of individual fans. The great circle arcs which join the two edge poles of each fan correspond to the apical angle. As fan edges (defined in Figure 11) are true vectors, both lower and upper hemisphere projections are required to show the complete angular distribution of the apical arcs. Also plotted is the great circle (dashed) corresponding to the cyclographic projection of the  $S_e$  foliation.



The plane of fanning is not a crystallographic plane, but is related most directly to the orientation of  $S_i$  (Figure 13a). The fact that in the fans  $S_i$  is commonly subparallel to the {010} cleavage confuses the issue. However, the angle between the cleavage and the fan plane is much more variable than the angle between  $S_i$  and the fan plane (Figure 13b).

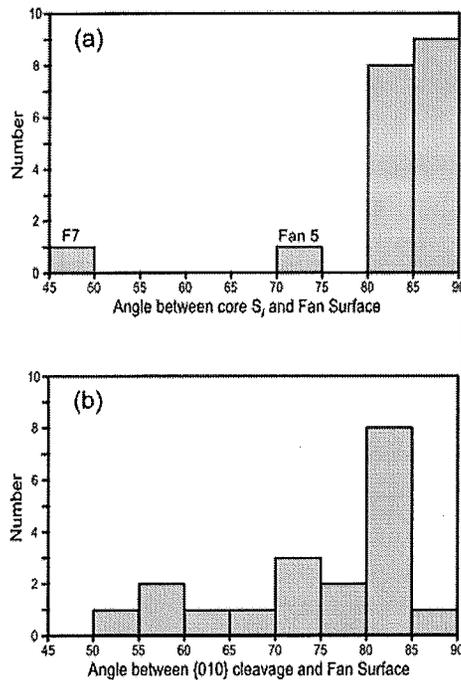


Figure 13. Histograms of 19 Box Mountain fans showing the angles between: (a) core  $S_i$  and the fan surface; and (b) (010) and the fan surface.

The amount of rotational opening recorded by fans (apical angle – Figure 11) on Box Mountain range from a few degrees to a maximum of 70 degrees. In the same outcrop, prismatic staurolite porphyroblasts containing sigmoidal shaped inclusion trails imply relative rotations of up to 40 degrees (Figure 14). Based on our model (Busa and Gray 1992a) prismatic staurolite porphyroblasts in the Bolton syncline appear to have locally accommodated the shear by a rigid rotation about their long axis. The response of the staurolite fans on the other hand was to fracture and be wedged apart in a semi-brittle fashion (Figure 10). The difference is probably due to size and initial orientation of the porphyroblasts relative to the distribution of shear in the matrix. If the shear was homogeneously distributed, the maximum rotation of prismatic staurolites lying parallel to the shear plane and perpendicular to the shear direction would be roughly one-half of the total amount of shear (Jeffrey 1922). On the other hand, the maximum apical opening of the staurolite fans oriented perpendicular to the shear plane would be on the order of the total shear (Schmidt 1918). The fact that the fans show a maximum apical angle of 70 degrees and the maximum rotation recorded by sigmoidal trails is on the order of 40 degrees is consistent with both types of porphyroblasts undergoing the same amount of shear-induced rotation.

Figure 10 illustrates our concept of the kinematic behavior of the staurolite prisms and fans undergoing shear during their growth. The prisms' long dimensions lie close to the  $S_i$  shear plane in which they nucleated. Prisms having long axes to the right of the transport direction 'roll' about that axis in a counterclockwise sense, whereas those to the left of the transport direction rotate clockwise. Fans however splay and open in a counterclockwise direction about an axis which lies in the shear plane, perpendicular to the transport direction. As a result, one would expect the pole to the surface of the average fan to lie within the girdle formed by the poles to the prisms' long dimensions and 90 degrees from the Hansen pole. That is not presently the case for the fans on Box Mountain, where the pole to the mean fan plane is nearly 90 degrees from the girdle formed by the poles to the  $c$  axes of the prism shaped porphyroblasts. Both prisms and fans lie subparallel to  $S_e$ , the foliation produced by the late stage, post core staurolite deformation.

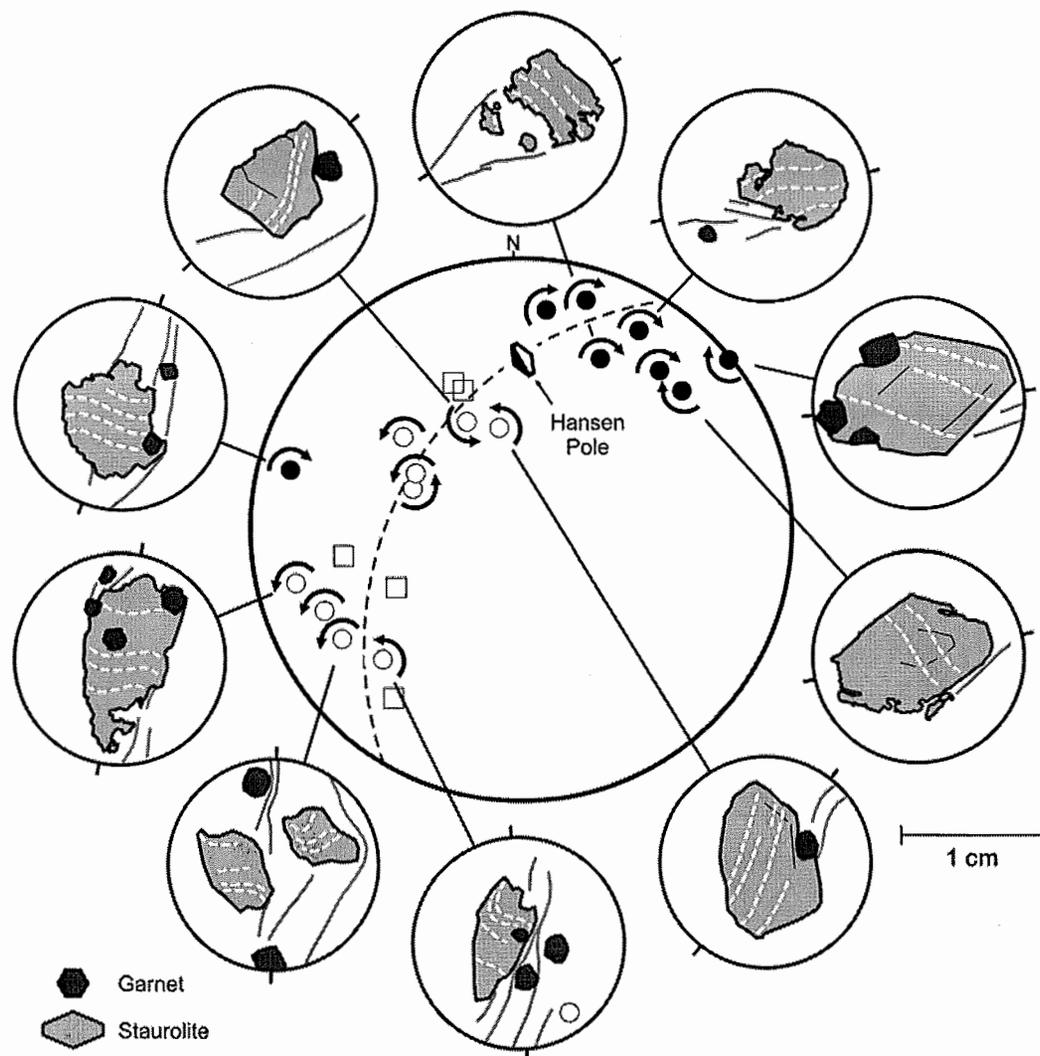


Figure 14. Lower hemisphere stereographic projection of the  $c$ -axes of sectioned prismatic staurolite porphyroblasts from Box Mountain. Crystal drawings are traced from centrally located (001) thin sections. Tick marks on the perimeter of the crystal drawings define the orientation of the trace of the average external foliation ( $S_e$ ). The rotational sense of the porphyroblast, suggested by the curvature of the sigmoidal-shaped inclusion trails (dashed lines) is shown by a curved arrow associated with the pole to each crystal  $c$  axis (open circle - counterclockwise, closed circle - clockwise, open square - no rotation). The location of the fitted Hansen pole (the switch from clockwise to counterclockwise porphyroblast rotation in the plane of staurolite prisms) is shown by the black and white staurolite symbol.

### LOCAL TECTONOMETAMORPHIC IMPLICATIONS

The orientations of staurolite  $c$  axes and sigmoidal shaped core inclusion trails from Bolton Notch are consistent with nucleation of staurolite in an early, fine foliation followed locally by a top-to-the-west sense of shear-induced rotation on the east limb. Subsequent eastward-directed backfolding of the Glastonbury Gneiss produced the Bolton syncline (Rosenfeld and Eaton 1985). Field observations suggest the external foliation in the Littleton Formation was caused by flattening which, on the east limb, post-dated the staurolite growth, resulting in muscovitization of staurolite adjacent to chocolate tablet boudins of pegmatite and truncation of the inclusion trails by the external foliation. Along the west limb of the syncline however, north-south directed stretching reoriented the staurolite, roughly parallel to the trend of the Bolton syncline and parallel to the strong biotite lineation in the adjacent

Glastonbury Gneiss. This interpretation is in conflict with Wintsch et al. (2003) who interpret the north-south lineation and preferred orientation of staurolite in the Littleton schist as an older structure, and the northwest preferred orientation on the east limb as younger. (The orientation of staurolite porphyroblasts at the north end of Box Mountain (STOP 3 Figure 9c) is inconsistent with simple overturning and southward extrusion and beckons additional work.) Thermal effects of the backfolding and extrusion of the Glastonbury Gneiss produced staurolite overgrowths on the west limb during the development of the new foliation and post-tectonic biotite porphyroblasts. The staurolite overgrowths may reflect a local metamorphic unconformity.

**Foliation Intersection Axis (FIA)**

Because of the consistency of their staurolite FIA across the Bolton syncline Hickey and Bell (1999) report that, "Consistent orientation of foliation intersection axis... [across the Bolton syncline]... suggests that the porphyroblasts did not rotate relative to geographic coordinates...". We attempted to reproduce their observations using their FIA method (Bell et al. 1997), but only few staurolite were intercepted in our vertical sections, and their asymmetries were difficult to decipher (Figure 6). In order to increase the number of staurolite intercepted in vertical sections, we compiled our data on the asymmetries of staurolite from the east and west limbs, assembled each into virtual rock blocks, and computed the staurolite asymmetries. Our computed data shows the transition from clockwise to counterclockwise asymmetry to be gradual (Figure 15). Although our staurolite Hansen pole computed on the east limb is similar to Bell's (i.e., 90 degrees from his staurolite FIA), the west limb is not.

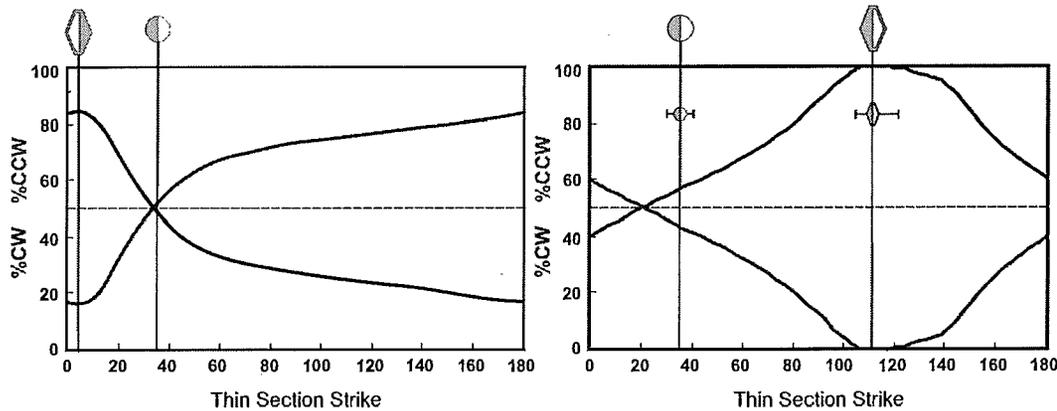
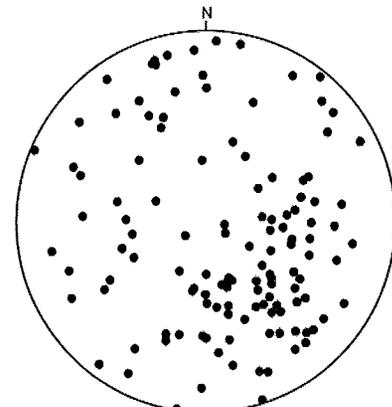


Figure 15. Staurolite FIA in Bolton syncline. The staurolite FIA at 035 degrees of Bell et al. (1997) is noted by the shaded circle; our transport direction (Hansen pole) is 90 degrees from the FIA and is shown by the shaded staurolite symbol. (a) West Limb and (b) East Limb of the Bolton syncline. Bell claims there is a sharp transition between 100% clockwise to 100% counterclockwise asymmetry in the staurolite porphyroblasts in vertical sections of the rock; we show it to be diffuse.

Bell et al. (1997) and Hickey and Bell (1999) conclude staurolite remained fixed relative to some kinematic reference frame and overgrew a succession of developing foliations all with a strike parallel to the 035 foliation intersection axis (FIA), during and after the formation of the Bolton syncline. One would therefore expect the poles to the type 1  $S_i$  inclusion surfaces captured by the staurolite porphyroblasts to define a girdle about the staurolite FIA. Our U-stage measurements of type 1 core  $S_i$  inclusion trail surfaces in staurolite porphyroblasts, when plotted in a geographic reference frame, show a broad distribution with no girdle (Figure 16).

Figure 16. Stereoplot showing geographic orientation of core  $S_i$  surfaces in staurolite porphyroblasts in the Bolton syncline.



### Garnet Porphyroblasts

Although we have just begun examining the garnet porphyroblasts in the Littleton schist, the following can be noted. At the north end of Box Mountain (STOP 4), garnets are euhedral and their mean size and size distribution are identical inside staurolite and in the matrix. About 15 km south, in Marlborough (STOP 6), the garnets have corroded edges. Interestingly, it is here we also see the largest staurolite overgrowths. Could the second stage of staurolite growth (and post-tectonic biotite porphyroblasts) be at the expense of garnet breakdown in this portion of the syncline? Additionally, the garnet porphyroblasts do contain inclusions – most are very small type 2 inclusions, particularly abundant at sector zone boundaries. Some contain type 1 trails which are a very fine quartz and carbonaceous foliation. More work on the garnet is needed.

### ROAD LOG

Bolton Notch, Connecticut is located approximately 20 miles east of Hartford. From New Haven, take I-91 north to I-84 east. Stay on I-84 east until you pass through East Hartford and into Manchester. Take right-hand exit onto I-384. Follow I-384 east to exit 5 (Route 85). Take a left at the end of the exit ramp onto Route 85 and follow for 0.0 miles to the intersection with Route 44 (Mobil gas station on right corner). Take a right onto Route 44 and go 0.0 miles east. Take a left-hand turn across Route 44 west into the Park and Ride lot adjacent to the state maintenance yard. The field trip begins 9:00 AM, Saturday October 1st. We will assemble at the Park and Ride parking lot at Bolton Notch, and finish in the Park and Ride parking lot in Marlborough.

#### Mileage

- 0.0 Left (2nd stop sign) onto Route 6/44 east.
- 0.8 Go east (quickly get onto left lane) on Route 6/44 towards Coventry, Mansfield. Make sharp left turn **before** stop light back towards Route 6/44 west.
- 1.2 Take sharp right to Notch Pond boat launch.
- 1.3 Park at Notch Pond.

**STOP 1. Bolton Notch: Syntectonic Staurolite.** (90 MINUTES) Park at the south end of Notch Pond. Follow path east to culvert under junction of I-384 and Routes 6 and 44. The contact between the Clough quartzite and the Ordovician gneisses can be seen on the south side of the culvert. The stratigraphy of the Clough quartzite is important, since it is duplicated on the west limb of the syncline at Box Mountain. The correlation of the Clough stratigraphy is the best evidence for the overturned nature of the syncline, since structures and other stratigraphic evidence is ambiguous. The lower contact of the Clough Quartzite is exposed on the north side of the culvert, 20 m from the eastern portal. Although the lower part of the Clough is a white pebbly quartzite, its base is marked by a half-meter thick muscovite-rich quartz schist. Is the contact a simple unconformity or is it a major fault zone? A complete section through the Clough Quartzite and the Fitch Formation is exposed in the darkness of the culvert. Most of the interesting features of these formations are found along the cliffs just north of this culvert and are noted in the following paragraphs.

Near its base the Clough is typically characterized by deformed quartz pebbles, 1 to 3 mm sized magnetite grains, and the virtual absence of garnet. The pebbles and magnetite disappear a few meters up in the section and garnet makes its appearance as large, centimeter-sized flat, disk-shaped crystals confined to the thin muscovite schist lamellae. The average thickness of the quartzite beds decreases systematically from 20 to 30 cm at the base to less than a centimeter near the top of the unit. The total thickness of the Clough quartzite in the Notch is 12 meters.

Ten meters of a homogeneous fine grained muscovite-biotite-quartz gneiss overlies the Clough. The gneiss was formerly quarried for flags and grindstones as it splits readily into exceptionally uniform large planar slabs two or more meters across. Quarries north and south of the Notch were active between 1812 and the late 1880s. Flagstones from Bolton were used for paving streets in Hartford, New York, Washington, Philadelphia, and Baltimore. Grindstones weighing several tons were shipped to Boston, New York, and Albany.

Finely laminated, differentially weathered calcareous rocks of the Fitch Formation conformably overlie the non-calcareous gneiss. A distinctive light grey calc-silicate layer marks its lower boundary. Alternating thin lamellae of a brown weathering calc-silicate and a non-calcareous biotite gneiss make up most of the 10 m thickness of the unit. Towards the top of the section, thick massive light green actinolite marble layers make an appearance. Meter size cavities are developed along some joints and small faults. The largest of these extends some 10 m into the cliff. A number of colorful 'legends' are associated with this cave which is known locally as 'Squaw Cave'.

Upon exiting the culvert, heading northwest, we encounter the silvery grey pelitic schists of the Littleton Formation. A dark green and white mottled amphibolite skarn 1 cm to 1 m thick is developed along the contact. Unlike its lower contact, the transition from calc-silicate to pelitic schist is exceedingly sharp and abrupt. Staurolite crystals in the Littleton are only about 5 mm long at the base, but they rapidly increase in size upward in the section. We will walk north-northwest (upward in section) out of the culvert to the base of the spectacular cliffs at the Notch. The Littleton here contains large anhedral staurolite porphyroblasts in abundance. The sigmoidal inclusion trails in differently-oriented crystals from this locality were studied in serial sections (Busa and Gray 1992a). The amount and sense of rotation implied by these trails suggests the porphyroblasts grew while the matrix was undergoing simple shear with top-to-the-west slip. Against the backdrop of this striking porphyroblastic schist we will review the controversy of whether staurolite rotated or not. We will point out the features displayed by the inclusion trails, the sector zoning, and preferred orientations of these giant porphyroblasts that lead us to the conclusion that they rotated as they grew.

We will climb approximately 250 m up the dip slope of the Littleton Formation through a well-marked trail. Along the way, note random float of silvery grey Littleton with small euhedral staurolite porphyroblasts. The crest of this ridge is capped by a large pegmatite body. The overlying schist was plucked away by ice so that the topography of the pegmatite outcrop mimics the form of its upper contact surface. The pegmatite has been boudinaged in a 'chocolate tablet' fashion on a grand scale. The individual boudins are several meters across. The thinned down necks between the blocks form an orthogonal system of distinct furrows across the surface of the outcrop. Quartz veins up to a half-meter wide fill rifts that formed along some of the more pronounced furrows. While on this outcrop it is also worth noting that the staurolite porphyroblasts within a few decimeters of this pegmatite are pseudomorphed by muscovite. This one simple observation firmly establishes the relative ages of staurolite metamorphism, the pegmatite emplacement, and the boudinage (which may be related to the Glastonbury Gneiss and overturning of the Bolton syncline).

From the top of the cliff it is evident that the valley of the Hop River heads off to the southeast, directly away from the Notch. The Notch itself was probably a water gap for a periglacial ancestor of today's Hop River. A small fault with a horizontal displacement of less than 10 m which runs along the base of the cliff through the Notch, probably controlled the location of the gap. To the west, the skyline of Hartford and the trap ridges of the Connecticut River valley are visible. The north-south ridge closest to us is Box Mountain, the opposite limb of the Bolton syncline.

Now follow the trail 250 m back down the dip slope of the staurolite schist to the abandoned Hartford and Providence railroad. Proceed 100 m north along the railbed into the railway cuts. Staurolite schists and well-bedded, fine-grained grey quartzites are complexly folded together here. Isoclinal fold noses are preserved in large quartzite boudins which swim in a sea of schist. Locally the quartzite splits into remarkable paper-thin bedding parallel sheets, a meter or more across. Randomly-oriented muscovite pseudomorphs (after staurolite? or kyanite?) up to half a meter long, 5 mm wide, and less than 1 mm thick ornament the surfaces of some sheets. Follow the railbed south to the path leading to the south end of Notch Pond and our starting point.

#### Mileage

- 1.4 Right onto Routes 6/44 West. Take first exit. Pass park and ride.
- 2.3 Stop light. Take right onto Cider Mill Road (Route 85).
- 2.4 Park on opposite side of road (careful – on-coming traffic at bend in road).

**STOP 2. Box Mountain I: Syntectonic Staurolite.** (30 MINUTES) Carefully cross road and enter into wooded area where there has been unfortunate dumping of household debris. Follow crude path north passing a small swamp and outcroppings of Glastonbury Gneiss. The Glastonbury Gneiss is a plagioclase gneiss with a strong lineation defined by centimeter-size biotite flakes and overlies the Clough here. The contact is not visible, in fact, to our knowledge, it is not exposed anywhere in Connecticut.

Walking eastward we encounter the Clough Quartzite. The Clough here is on the overturned limb of the Bolton syncline. The stratigraphy noted on the right-side-up limb at Bolton Notch is inverted here and can be matched in most detail. The lowest quartzites are pebbly, magnetite rich, and thick bedded. The top of the Clough is marked by a dark colored graphitic black quartzite. Similar graphitic zones are present at all outcrops of the Clough on the overturned limb of the syncline between here and Great Hill. On the northwest dipping ledges is a north-south trending fold axes in the Clough.

Walk east across a dry valley and observe the silvery-grey pelitic schists of the Littleton Formation. Large anhedral staurolite porphyroblasts are found on the moderately steeply-dipping outcrops. The schist is more graphite-rich than those on the east limb at the Notch. The staurolite porphyroblasts' c axes have a preferred orientation at 015, parallel to the biotite lineation in the nearby Glastonbury Gneiss (Figure 9b).

Continue eastward, cross the dirt road and examine the rocks on the next ridge. On the crest of the ridge is a minor fold. Continue eastward across the dry valley and observe the staurolite schist on the hillside. This rock type continues eastward to the crest of the hill where it is cut by a pegmatite.

#### Mileage

- 2.4 Return to vehicles, turn and go north on Cider Mill Road (Route 85).
- 2.9 Stop sign. Turn right onto Lake Street.
- 3.4 Vernon town line.
- 4.4 Stop sign. Turn right onto Tunnel Road.
- 5.0 Turn right onto Risley Road.
- 5.5 Stop sign. Turn right onto Boulder Crest Road.
- 5.8 Stop and park at cul de sac near gate for Boulder Ridge Park.

**STOP 3. Box Mountain II: Syntectonic and Flattened Staurolite.** (60 MINUTES) Walk south across soccer field into woods. Outcrops of Littleton schist contain euhedral to subhedral staurolite porphyroblasts. Some of the strike-parallel staurolite porphyroblasts are boudinaged. In these pull-apart zones, an aureole of muscovite is seen, absent of graphite. Similar graphite-free dilation zones with new muscovite, quartz and biotite is seen in some of the fan-shaped staurolite porphyroblasts at our next STOP 4. The  $S_1$  asymmetry observed at this location (Figure 9c) is top to the east, unlike top to the south seen at other west limb locations on Box Mountain. Also seen on these outcrops are splayed porphyroblasts of staurolite, some looking like fans. More well-developed fans will be seen and their significance discussed at our next stop, just south of this location.

Walk east along blue and orange marked trail. The grey fine-grained micaceous quartzite in these outcrops near the top of steep cliffs (BE CAREFUL) splits easily into large thin sheets. The surfaces of some sheets are covered with giant lozenge-shaped staurolite porphyroblasts. The crystals are typically no thicker than 1 mm but range in size up to 30 cm in length and 10 cm in width. Crystals are randomly oriented within the bedding plane and all are boudinaged in a chocolate tablet fashion along their long and short dimensions. Fractures are filled with muscovite. The flattening was apparently isotropic as no orientation dependence of the boudinages is evident. As the porphyroblasts grew into each other with no change in width and no obvious 'zone of depletion' material transport could not have been rate limiting. Perhaps the extraordinary size of these crystals is related to an absence of appropriate nuclei.

Follow the blue trail back to the athletic field and to parked vehicles on Boulder Crest Lane. Turn left onto Rainbow Trail. Note the northwest dipping Clough Quartzite outcrops along the right side (north) of Rainbow Trail, and the pelitic schists on the left side of the road. Boulders and walls of Clough Quartzite can be seen on property west of Rainbow Trail. Loop around Rainbow Trail and back out to Boulder Crest Lane to Tunnel Road.

#### Mileage

- 5.8 Leave Boulder Ridge Park. Turn left onto Rainbow Trail.
- 5.9 Note outcrops of Clough Quartzite on your right (north) on front yards of residential properties and Littleton schist on your left (south). Stay left, then bear right at bottom of hill around to the other side of Rainbow Trail.
- 6.4 Stop sign. Turn left onto Boulder Crest Road. Follow Boulder Crest Road to stop sign. Turn left onto Tallwood Drive.
- 6.7 Stop sign. Turn right onto Willow Stream Road.
- 6.9 Stop sign. Turn left onto Tunnel Road.
- 7.4 Turn left onto Echo Drive.
- 8.0 Turn left onto Echo Ridge Drive.
- 8.3 Stop at cul de sac. Visit outcrop on front yard of 119 Echo Ridge Drive. A post-staurolite folding event has crenulated the schist and has broken staurolite porphyroblasts. Similar to STOP 3, some staurolite have been pulled apart. This may be dependent on the orientation of the staurolite long dimension during flattening and extension. An earlier folded surface can be seen at this outcrop. This surface is defined by quartz-rich zones and finer grained schist. Splayed and incipient fan-shaped staurolite porphyroblasts can

be found on the outcrop surfaces. At our next stop, we will visit an outcrop containing many fan-shaped staurolite porphyroblasts. Leave cul de sac.

- 8.7 Turn right onto Echo Drive.
- 9.2 Left onto Tunnel Road.
- 9.3 Left onto Scott Drive. Follow all the way to top of hill. Bear right (Cross Drive) and go to top of hill.
- 9.9 Park on Brent Drive area. Walk up middle driveway and turn left into woods.

**STOP 4. Box Mountain III: Staurolite Fans and Box Mountain Quarry.** (90 MINUTES) Walk up middle long driveway then bear left into woods along light foot trail. Walk about 130 m to trail intersection. Follow path to left for 300 meters. Outcrops of Littleton schist along the way contain 1 to 2 cm long anhedral staurolite porphyroblasts. The trail also passes small outcrops of various calc-silicates and muscovite-biotite gneisses. Are these exposures of the Fitch Formation on the overturned limb or are they part of the Littleton? Staurolite schists outcrop on both sides of the calc-silicate rocks, but the Clough Quartzite is only a few hundred meters away. Leave the trail at its high point in a col and ascend the hill on its north side. The summit is marked by '789' of the USGS topographic map. Staurolite porphyroblasts several centimeters long are remarkably abundant in outcrops on hill top. Triangular fan-shaped porphyroblasts formed of radiating groupings or splays of curved staurolite crystals are common. The morphology of the porphyroblast aggregates is consistent with shear-induced semi brittle breakage during growth. Small growing fragments break and rotate independently until rigidly intergrown with the rest of the porphyroblast. Why on the eastern limb the rotating staurolite porphyroblasts grew as a single crystals while on the west they continually fractured is unclear. Perhaps rate of deformation, size of porphyroblast, scale of foliation, scale of simple shear zones are factors.

On our way back we will examine muscovite pseudomorphs after staurolite in the aureole of a large pegmatite boudin on the south side of the col near where we left the trail. In the same general area staurolite schists adjacent to large boudinages quartz veins are converted to a banded tourmaline-quartz schist.

Before returning to the trail to our parked cars we will bear left (south) and approach the west end of the Box Mountain quarry. Most of the Clough Quartzite quarried here is used locally for patios, fireplaces, and stone walls. Selected thin slabs with large pink garnets are marketed nation-wide as a veneer stone. The stone is used as a facing on one national chain of gas stations and it adorns many buildings in the Midwest. Disseminated arsenopyrite causes some discoloration problems. The yellowish-green stain which colors the exposed surfaces of some slabs is the mineral scorodite ( $\text{FeAsO}_4 \cdot 2\text{H}_2\text{O}$ ), a weathering product of the arsenopyrite. With time, the green color turns to rust as the scorodite weathers to limonite. The stratigraphically highest quartzites in the quarry seem to be the most susceptible to scorodite discoloration. Angry customers whose expensive retaining walls turned an ugly green less than a year after they were constructed have made the quarrymen very wary of the upper quartzite.

Euhedral honey yellow staurolite occur along with large flat pink garnet in thin schistose interbeds between the pure quartzite layers. The total section is 35 m thick. The top of the Clough is marked by a tectonic mélangé of dark colored graphitic black quartzite (seen at STOP 2) and black garnet-biotite schist similar to 'Diamond Reef' in the Notch (Shepard 1837). The deformation here however is much more intense, and the layering is completely transposed. Similar graphite zones are present at all outcrops of the Clough on the overturned limb of the syncline between here and the southern terminus of the syncline in Cobalt.

The view from the crest of the hill overlooking the quarry is spectacular. On a clear day the trap ridges of the Hartford Basin are plainly visible. The Sleeping Giant of Hamden and the Hanging Hills of Meriden are on the southwest horizon and Avon and Talcott Mountains on the northwest. Follow trail north to residential driveway and return to vehicles.

#### Mileage

- 10.0 Turn right onto Vine Drive.
- 10.1 Turn left onto Scott Drive.
- 10.4 Stop sign. Turn left onto Tunnel Road.
- 10.8 Stop sign. Turn left onto Lake Street. (Note stonewalls and many houses are faced with Clough Quartzite).
- 11.7 Bolton town line.
- 12.3 Stop sign. Bear right onto Lake Street.
- 12.9 Stop sign. Turn right onto Middle Turnpike East. Quickly get into left lane.
- 13.0 Turn left into Shady Glen Restaurant.

#### LUNCH

**Mileage**

- 13.0 Turn right onto Middle Turnpike East.
- 14.0 Stop light. Turn right onto Route 85 South.
- 14.4 Overpass Interstate 384.
- 15.3 Make very sharp right onto Birch Mountain Road Extension. (Note Littleton schist on right side of road).
- 16.0 Stop sign. Go straight across Camp Meeting Road and continue on Birch Mountain Road.
- 17.3 Follow hill, pass Tinker Pond Road.
- 17.5 Turn left onto Villa Louisa Road.
- 18.0 Villa Louisa Banquet Facility/Restaurant
- 18.8 Park on right side of road near metal posts and gate.

**STOP 5. Birch Mountain.** (30 MINUTES) Walk southwest into woods. Outcrops of staurolite schist contain larger garnets than previously seen. Along the east side of these outcropping of schist are 5 to 10 cm diameter knobby clots consisting of intergrown staurolite, garnet, and graphite. They weather in positive relief, and may represent a boudinaged pelitic-rich zone (layer?). The staurolite and garnet in these clots are tightly clumped and the staurolite is euhedral and smaller than the larger staurolite porphyroblasts we have been seeing at other locations.

Common in this area of the syncline are staurolite porphyroblasts exhibiting very small needle-like splays of staurolite from the ends of the main porphyroblast. Surrounding the staurolite is a lighter, muscovite-rich aureole, free of graphite. These 'broccoli-shaped' splays may be related to a late-staurolite growth during flattening and boudinage.

Further west into the woods is the Shenipsit Trail. Avid hikers refer to 'garnet ledge' where pea-sized gemmy garnets can be found. The size of garnet varies with location ranging between 1/2-2 mm in diameter. However, at the STOP 4 outcrop preliminary measurements of the size of garnet is amazingly consistent (2 mm diameter), both in the matrix and included in staurolite porphyroblasts. Did garnet not participate in the reaction to form much of the staurolite?

**Mileage**

- 18.8 Leave STOP 5 and go south on Villa Louisa Road (Birch Mountain Road). Note the remains of apple orchards on left, overtaken by residential properties.
- 20.3 Radio towers on right.
- 20.5 Stop sign. Turn left onto Route 94.
- 20.7 Turn right onto Marlborough Road.
- 21.0 Stop sign. Stay straight and follow Marlborough Road down long hill into Blackledge River Valley.
- 22.6 Stop sign. Bear right.
- 23.2 Stop sign. Turn right onto West Street.
- 24.0 Stop sign. Go straight through stop sign.
- 24.7 Turn right onto Planeta Drive.
- 24.9 Stop sign. Go straight onto East Glastonbury Fish and Game Club property.
- 25.1 Park on left.

**STOP 6. Glastonbury Fish and Game Club (SPC 83 coordinates: 1068666E, 799301N).** (60 MINUTES) We shall walk north into the woods close to Route 2 westbound. The first few ridges are the Clough Quartzite, followed by ridges of Fitch Formation. Folding is nicely preserved in the Fitch, and its pitted, furrowed surface is easily recognized in the field. As we climb up the steeper ridges, we encounter the silvery-grey Littleton schist. At the top of the locally highest ridge, the surface of the outcrop is loaded with anhedral, stubby staurolite porphyroblasts. We will walk north-northeasterly along strike and visit many of the small ridges of staurolite schist. Some of the staurolite are euhedral. This is made possible by the late-staurolite overgrowths. Staurolite porphyroblasts from this location in the syncline have the largest overgrowths. If you are lucky, you may be able to see, in a hand sample perpendicular to the *c* axis, the near orthogonal bend of the *S*<sub>1</sub> into the staurolite overgrowth, which are nearly always parallel to the external foliation. Does overgrowth occur along the entire length of the porphyroblast or is it oblique? At this area of the syncline where the staurolite overgrowths are the largest, garnet porphyroblasts have corroded edges. Are the overgrowths a result of a garnet break-down reaction to grow more staurolite?

In the 1985 NEIGC field trip, Rosenfeld and Eaton (1985) reported that the strike parallel elongate staurolite porphyroblasts in this area of the syncline showed snowball microstructure and east-side-up sense of shear. This is consistent with our data at this location (Figure 9e STOP 6).

**TRIP ENDS****Mileage (Back to Yale)**

- 25.5 Stop sign. Turn right onto West Street.  
 25.6 Turn right onto Route 2 west (or)  
 25.7 Turn left onto North Main Street Junction of North Main Street and Route 66.  
 27.5 Turn right and follow Route 66 to Middletown.

**ACKNOWLEDGEMENTS**

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# SEDIMENT DYNAMICS OF THE BRANFORD RIVER ESTUARY

By

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## INTRODUCTION

The Connecticut shoreline is marked by numerous drowned stream valleys that now act as long, narrow estuaries. Sediment behavior in these estuaries is quite complicated. Sources include both upland watersheds and Long Island Sound. In the estuary, sediment can go through cycles of deposition and resuspension followed by either export or redeposition plus burial. The long term pattern seems to be burial at a rate close to relative sea level rise (Benoit et al. 1999; Rozan and Benoit 2001) so that the estuaries are close to steady-state in terms of their bathymetry. Humans interfere with this pattern in many ways, perhaps most dramatically through periodic dredging to maintain navigation channels.

Sediment transport, deposition, resuspension, and burial in estuaries is important for several reasons. These processes determine the fate of a number of natural and anthropogenic chemical substances that are delivered to the estuary from the ocean, atmosphere, and land. The biologically rich estuarine environment is important because it provides soft bottom habitat and nurseries for shellfish and finfish. Many salt marshes are also located here and are strongly influenced by sedimentary processes. Estuaries also provide a number of ecosystem services, including providing sheltered anchorages, navigation, and recreation, that are affected by sedimentation. Estuarine sediments are also the repository for considerable amounts of particle-reactive and hydrophobic contaminants.

Short-lived radionuclides can be useful for studying the dynamic behavior of sediments in estuaries. They are often added as discrete pulses during storms and can be followed for days or weeks until the next pulse input starts a new cycle. The relatively short half-lives match the timescale of processes occurring in the estuaries. The known radioactive decay rates can be used to provide age and rate information. In this field trip we will focus on one such radionuclide,  $^7\text{Be}$ , while considering several others ( $^{234}\text{Th}$ ,  $^{210}\text{Pb}$ , and  $^{137}\text{Cs}$ ).

## BERYLLIUM-7 AS A SEDIMENT TRACER

$^7\text{Be}$  is a useful tracer for short-term sediment dynamics in aquatic systems thanks to its short half-life of 53.3 days (Dominik et al. 1987) and its rapid association with particulate matter.  $^7\text{Be}$  is produced in the atmosphere by cosmic-ray spallation of nitrogen and oxygen (Feng et al. 1999). Although it is predominantly produced in the stratosphere, the bulk of the  $^7\text{Be}$  that is delivered to the earth's surface is produced in the troposphere. This is because the residence time of beryllium in the stratosphere is much longer than the life of the radionuclide (Dutkiewicz and Husain 1985).

Direct atmospheric deposition has been shown to be the dominant source of  $^7\text{Be}$  in estuarine systems studied to date (Olsen et al. 1986; Dibb and Rice 1989). Direct atmospheric deposition is responsible for greater than 90% of the total  $^7\text{Be}$  inputs into Chesapeake Bay, while the transport of beryllium into or out of the Bay is relatively insignificant (Dibb and Rice 1989). The relatively low contribution by watersheds of  $^7\text{Be}$  to estuaries in past studies is a result of beryllium's adsorption to vegetation and soil in the watershed for a period significantly longer than its half-life, and the small ratio of watershed to estuarine area (Olsen et al. 1986). When  $^7\text{Be}$  is first deposited from the atmosphere, it is solubilized then quickly rescavenged by particles (Olsen et al. 1986; Dibb and Rice 1989). Studies have shown that the distribution coefficient ( $K_d$ ) for  $^7\text{Be}$  in estuarine and coastal waters is relatively large (approximately 105, Dibb and Rice 1989; Baskaran and Santschi 1993). The residence time of  $^7\text{Be}$  in Hudson River waters ranges from less than 1 day to 13 days, with seasonal variability (Feng et al. 1999). Greater than 80% of the  $^7\text{Be}$  in Chesapeake Bay is found in the sediments, while less than 20% of the  $^7\text{Be}$  is in the water column (Dibb and Rice 1989; Dibb and Rice 1989) as expected from beryllium's very strong affinity for particles. Also, 74-86% of the total atmospheric deposition of  $^7\text{Be}$  during a rainfall became associated with particles within one hour in Galveston Bay, Texas (Baskaran and Santschi 1993). Aggregation and settling of particles resulted in removal of 70% of the  $^7\text{Be}$  from the water column in less than one day (Baskaran and Santschi 1993; Fitzgerald et al. 2001).

### FIELD TRIP FRAMEWORK

The several stops of this fieldtrip illustrate the sources, sinks, and various transport and redistribution processes experienced by  $^7\text{Be}$  and what we can infer about short-term sediment dynamics. The framework for the fieldtrip is construction of a mass balance for  $^7\text{Be}$ . As we make stops related to reservoirs and processes in the following diagram, we will add amounts and fluxes of  $^7\text{Be}$  measured over one seven day period last year.

## Branford River $^7\text{Be}$ Mass Balance

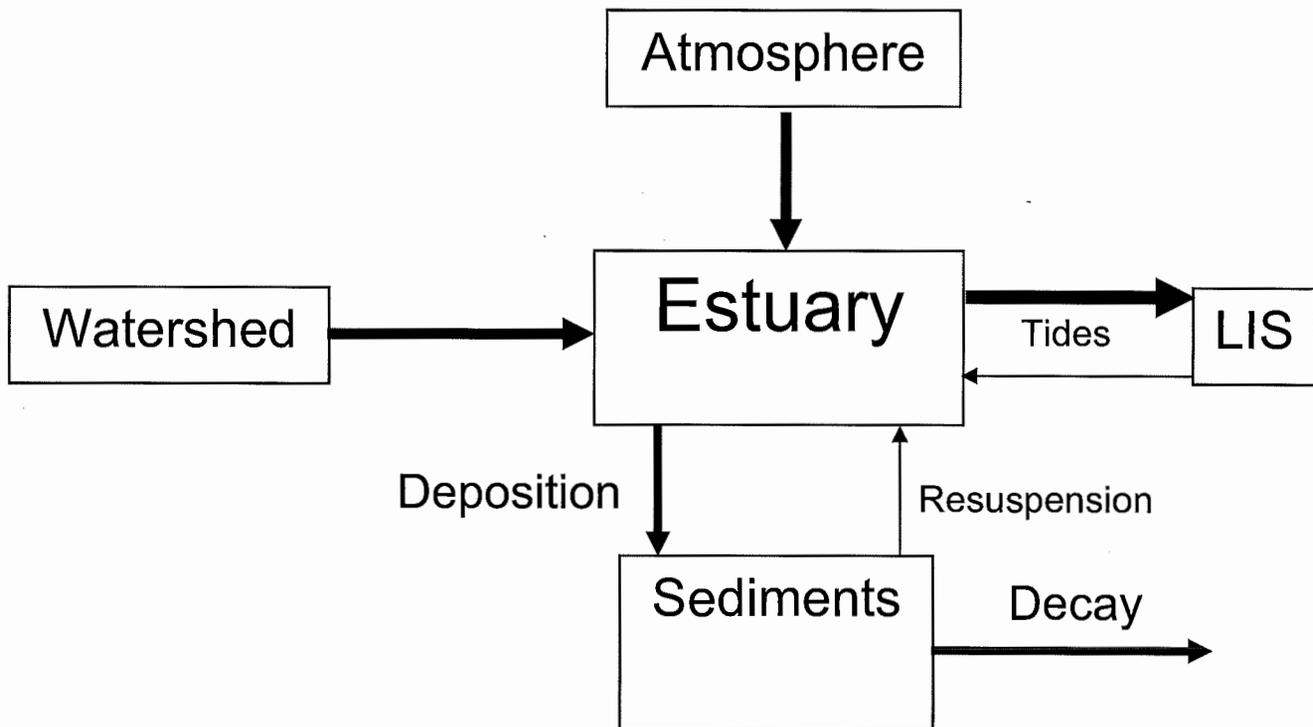


Figure 1: Schematic diagram of the  $^7\text{Be}$  cycle in a Connecticut coastal estuary. Since  $^7\text{Be}$  rapidly adsorbs to particles, this is close to the cycle for fine sediment.

### ROAD LOG

The trip begins at the Kline Geology Building, Whitney Avenue, New Haven, Connecticut. Entrance to its parking lot is on Whitney Avenue at the end of Humphrey Street. Meet on the steps of the lab facing the parking lot (north side of building). Our first site is a precipitation collector on the roof, to which we will walk.

#### FIRST SITE: Atmospheric deposition

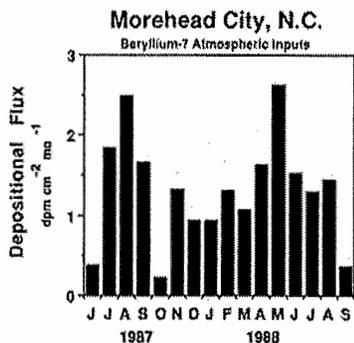
##### Precipitation collector on the roof of the Environmental Science Center

Because  $^7\text{Be}$  that actually reaches the earth's surface is produced mainly in the troposphere, the depositional flux is influenced by weather and location. It tends to increase with latitude and with solar activity (Kaste et al. 2002). Each year, there tends to be greater deposition in spring and summer than fall and winter. Most  $^7\text{Be}$  is



Figure 2: Topographic map of the Branford River estuary. Scale of map as printed is 1:29,000. Stop 1 is between Boston Post Road and I-95. Stop 2 is near the track of Branford High School (off East Main Street). Stop 3 is near the intersection of Montowese Avenue with the Amtrak line. Stop 4 is a boat launch off Goodsell Point Road (the continuation of Stannard Avenue beyond Harbor Street). Stop 5 is Parker Memorial Park, the highland between Branford Point and Lindsey Cove.

delivered to the surface with wet deposition, so the timing and magnitude of rainstorms has a strong influence on the amount deposited. Total annual fluxes vary by as much as a factor of 7 from one location to another, and individual storms at a single location can vary by at least a factor of 20. This means that most applications of  $^7\text{Be}$  as a tracer requires direct measurement of the local flux. This collector on the roof of the Environmental Science Center has been used in the past to measure the  $^7\text{Be}$  flux. For one continuous collection period of 267 d we found the deposition of  $^7\text{Be}$  to be  $290 \text{ mBq/cm}^2\text{y}$ , or  $0.79 \text{ mBq/cm}^2\text{d}$ . This is adequate to maintain an average inventory of  $61 \text{ mBq/cm}^2$  if the only loss were through radioactive decay. Weather willing, we will take a sample of precipitation to leave on the gamma counter in the analytical lab. For the seven-day period corresponding to our mass balance, the atmospheric flux directly to the surface of the estuary was  $49 \times 10^6 \text{ Bq}$  (49 MBq). This corresponds to  $0.86 \text{ mBq/cm}^2\text{d}$ , very close to the long-term average in this area.



#### Analytical laboratory, 340 Kline Geology Building

This lab is a multi-user facility containing instrumentation for analysis of environmental samples, especially water, sediments, soils, and tissues. It is a joint venture of the Environment School -- who provide the instruments and a full-time lab manager -- and the Department of Geology and Geophysics -- who provide the space. We will quickly encapsulate our rain sample and put it on the counter for gamma measurement. There should be a clear  $^7\text{Be}$  peak when we return at the end of the field trip.

#### Mileage and directions to Stop 1 from Kline Geology

- 0.00 Leave parking lot and turn left onto Whitney Avenue
- 0.44 Turn right onto Willow Street
- 1.14 Turn left onto entrance to I-91 south. Once on the ramp take the right hand fork. As soon as you get on the highway, begin to make your way to the left lane. **You have only 1.1 mile to get into the extreme left hand lane for a left exit!**
- 2.50 Take the I-95 north exit (a left exit)
- 8.02 Proceed on I-95 north to Exit 54
- 8.11 At the end of the ramp turn right onto Cedar Street
- 8.28 At the light, turn left onto North main Street, US 1
- 9.59 Turn left onto Mill Plain Road
- 9.66 Turn right after small cemetery into parking lot of condominium complex, driving to far back left (NE) corner

#### STOP 1: Head of tides for the Branford River estuary

Just north of this site is the end of the tidal portion of the Branford River. Upstream from this location, the watershed has an area of  $57 \text{ km}^2$ , with an additional  $6 \text{ km}^2$  between here and Long Island Sound. The estuary itself has an area of  $0.82 \text{ km}^2$ . Based on measurements of  $^7\text{Be}$  in river water and USGS gauging data, we calculated a riverine flux of 49 MBq during the seven-day study period. Coincidentally, this is identical to the direct atmospheric deposition to the estuary's surface. This relatively large contribution from the watershed (compared to other estuarine studies) probably reflects the large ratio of watershed:estuary areas (70:1). The riverine flux represents only 1.4% of the  $^7\text{Be}$  landing on the watershed, consistent with previous findings elsewhere. The  $^7\text{Be}$  that is transported by the river is a combination of direct deposition to its surface and erosion of soil particles that

received atmospheric  $^7\text{Be}$  in the past few months. Several investigators have used  $^7\text{Be}$  as a tracer of eroded soils and have compared sources from different depths in the soil profile by using atmospherically deposited radionuclides of differing half-lives (Wallbrink and Murray 1993; Wallbrink and Murray 1996; Matisoff et al. 2002) {Bonniwell, 1999 #3870}.

- 0.00 Leave parking lot and turn left
- 0.13 Turn right onto Boston Post Road/Route 1
- 0.38 Bear left at light onto East Main Street
- 0.48 Turn left into the parking lot of Branford High School. Keep to the left and drive to the end of the track. Park and walk along the end of the track to the river.

### STOP 2: Upper Branford River Estuary

$^7\text{Be}$  added to the estuary from the atmosphere and watershed is either already associated with particles, or is quickly scavenged by them. Residence time of  $^7\text{Be}$  in the water column of lakes and estuaries usually falls in the range from a day to two weeks. We measured total  $^7\text{Be}$  in the water column of the Branford River estuary over the course of three rainstorms. Removal occurred exponentially, with an e-fold removal rate constant of  $0.96 \pm 0.04$  for the three storms (Fig. 3). Adding inputs from precipitation and the watershed to  $^7\text{Be}$  pre-existing in the water column produced a good match with exponential removal during and after the storm (Fig. 3). This removal reflects the combination of scavenging of any dissolved  $^7\text{Be}$  and subsequent sedimentation, as well as removal by other processes. Clearly, scavenging and sedimentation are rapid in this system.

$^7\text{Be}$  levels in the water column before, during, and after the storm studied in July 2004 fell in the range from 10 to 84 MBq, with highest levels occurring towards the end of rainfall, and lower levels before and after. (These values can be placed in the "Estuary" compartment of the mass balance.)

### Storm Series 2: 27 - 30 July 04

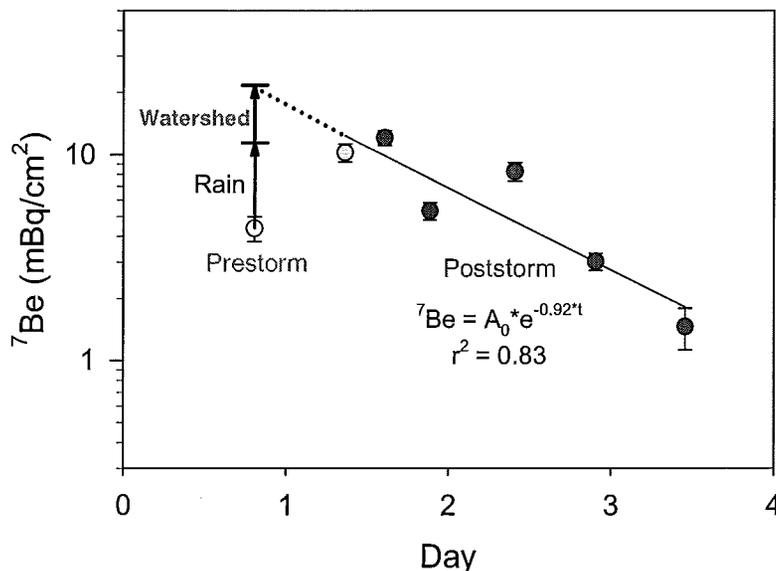


Figure 3: Removal of  $^7\text{Be}$  from the water column of the Branford River estuary during a storm in July 2004. Two storms in November 2002, data not shown, had similar removal rate constants.

- 0.00 Leave parking lot and turn left onto East Main Street
- 0.60 Turn left onto Montowese Avenue
- 1.10 Immediately after passing under the railroad tracks, turn right into the small parking lot.

**STOP 3: Middle Branford River estuary**

This stop is near the top of the navigable portion of the river. Sediment samples were collected from the length of the estuary and at several cross sections to evaluate spatial variability and distribution of sedimentation. Two synoptic sets of samples were collected in July 2003 and June 2004. Results show that there is great variability along the length of the estuary, and sometimes across it (Fig. 4). For all 53 samples the relative standard deviation is fully 61%. This indicates that scavenging and deposition of  $^7\text{Be}$  is uneven in space, or that surficial sediments are rapidly redistributed unevenly in the estuary. There was also a substantial difference in the  $^7\text{Be}$  distribution in surface sediments in the samples collected approximately a year apart. The part of the river seen at this stop had the most consistent  $^7\text{Be}$  levels from one year to the next. Replicate measurements within a few meters of each other have a lower variability, but surprisingly, even these showed considerable changeability on apparently flat and featureless bottoms. For example, in the part of the river visible here, five samples collected simultaneously over a distance of less than 2 m had a relative standard deviation of 20%. The total inventory in sediments during the period of the mass balance was 364 MBq, which can be added to the "Sediment" compartment of the diagram. Radioactive decay of this standing stock of  $^7\text{Be}$  over seven days amounts to 32 MBq, the "Decay" flux on the mass balance.

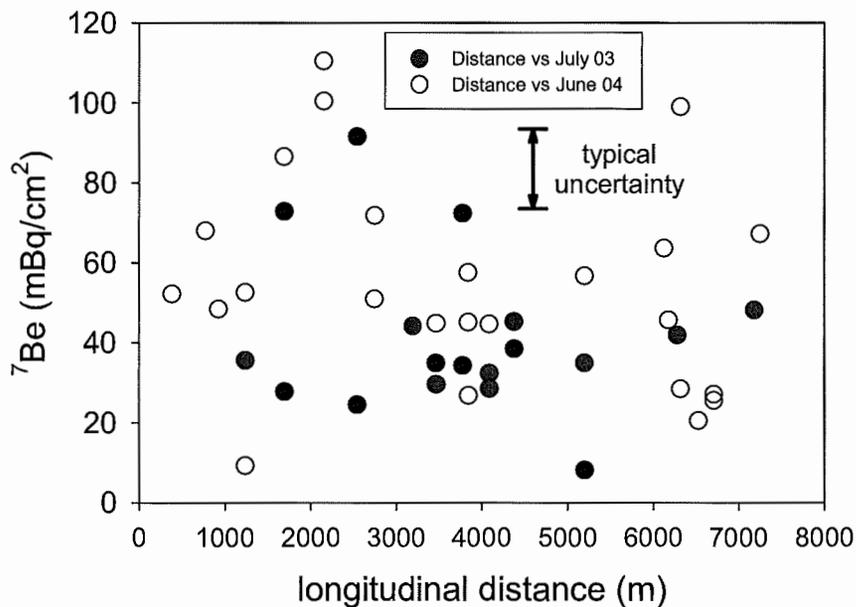


Figure 4: Synoptic distribution of  $^7\text{Be}$  in surficial sediments of the Branford River estuary. Note large variation with distance downstream, between years, and sometimes in short distances across the river at a single distance downstream. Note also that variability is much greater than the measurement uncertainty.

- 0.0 Leave parking lot and turn left onto Montowese Avenue
- 0.16 Turn left onto Meadow Street
- 0.59 Bear left, continuing on Meadow Street
- 0.72 Turn left onto Maple Street and pass over the tracks
- 1.01 Turn left onto Harbor Street. Notice the intensive use of the river by marinas in this section.
- 1.50 Turn left onto Goodsell Point Road
- 1.88 Turn right into marina

**STOP 4: Lower Branford River estuary**

(This portion of the river is used intensively by marinas.) To better understand medium-term variability of  $^7\text{Be}$  inventories in sediments, we collected surface samples repeatedly between rainstorms for a period of three months. Part of the data record is shown in Fig. 5. Within measurement uncertainty, sedimentary  $^7\text{Be}$  could be explained by loss through radioactive decay and gain from deposition of atmospheric  $^7\text{Be}$ . Additional loss, perhaps associated with high winds and above normal sediment resuspension, occurred around day 36 of the sample series. Based on this consistency, we believe that net deposition to sediments in the period of the mass balance was close to loss through decay. The “Deposition” flux of the mass balance is therefore 32 MBq.

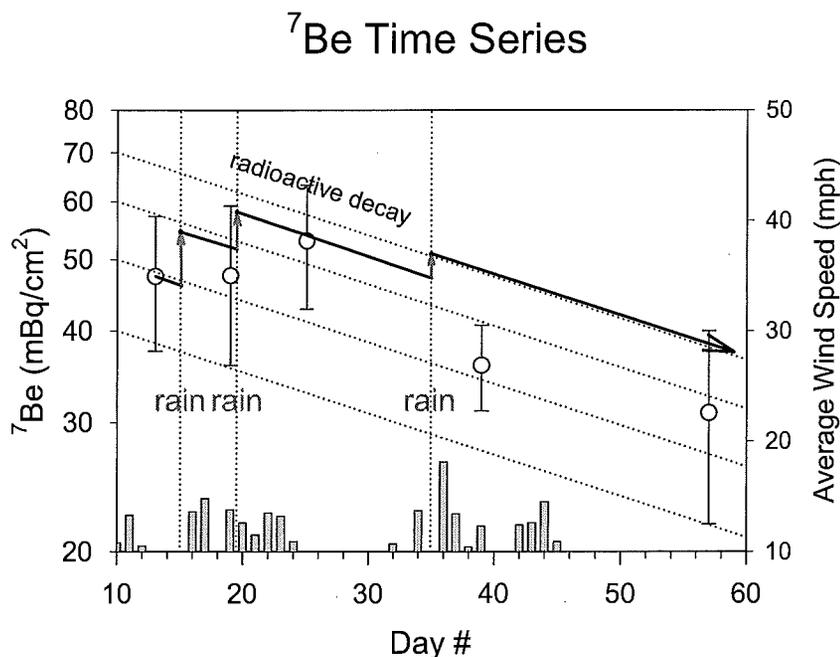


Figure 5: Comparison of  $^7\text{Be}$  inventory in bottom sediment to loss by radioactive decay and gain from atmospheric input.  $^7\text{Be}$  in sediments seems to be nearly in steady state between these gain and loss terms.

- 0.00 Leaving the parking lot turn left onto Goodsell Point Road
- 0.38 Turn left onto Harbor Street
- 0.54 Turn right into the parking lot of Parker Memorial Park. Walk to the beach.

### STOP 5: Long Island Sound

Here the Branford River enters Long Island Sound. Direct measurement of  $^7\text{Be}$  flux would be extremely difficult because of the large variations with each tidal cycle. By difference between inputs from rain and the watershed (49 + 49 MBq) and loss to sediment deposition (32 MBq), the net exchange with Long Island Sound is a loss of 66 MBq for our seven day period (“Tides” flux of the mass balance). The overall budget thus indicates roughly equal inputs from the atmosphere and the watershed, with roughly twice as much lost to tidal flushing as deposition within the estuary.

### RETURN TO KLINE GEOLOGY LABORATORY

- 0.00 Leave parking lot and turn left onto Harbor Street.
- 0.65 Turn right onto Maple Street
- 0.93 Passing over railroad tracks, Maple becomes Kirkham Street
- 1.18 Turn right onto Main Street
- 1.33 With church on your right, turn left onto Cedar Street
- 1.81 Cross Route 1, staying on Cedar
- 2.06 Take on ramp for I-95 south (you can also get on I-95 north if you are headed in that direction)

- 7.46 At the end of the Quinnipiac River bridge, take right exit for I-91 north
- 8.29 Take exit 3 for Trumbull Street
- 8.82 At end of long ramp, stay straight through light for Trumbull Street
- 8.96 Turn right onto Whitney Avenue
- 9.31 Turn left into parking lot for Kline Geology Laboratory

**OPTIONAL STOP 6: Analytical Lab**

Anyone interested can meet me on the steps of Kline Lab and we can check the rain sample in the gamma counter in the analytical lab.

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## **A Visit to the North Branford Trap-Rock Quarry Operated by Tilcon Connecticut Inc.**

by

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Frank T. Lane, Tilcon Connecticut Inc., North Branford, CT 06471-2001

### **INTRODUCTION**

The North Branford quarry, operated by Tilcon of Connecticut Inc., is one of the largest trap-rock quarries in the world. It supplies crushed stone for roadbeds, asphalt, and concrete aggregate for use in Connecticut and New York States. The eastern face of the quarry, which purports to be the longest face of any trap-rock quarry in the world, provides a spectacular 8-km long vertical exposure through the central part of the Holyoke flood-basalt flow, the middle and thickest of the three Mesozoic basalt units in the Hartford Basin. Because of its great thickness, the Holyoke flow cooled slowly and was able to differentiate to produce horizontal sheets of ferrodiorite and granophyre (granite) in its central part. This exposure provides an ideal site to witness the products of igneous differentiation where we have a clear understanding of the differentiation mechanisms with none of the complications normally associated with deep-seated plutonic bodies.

The excursion will look at the operation areas of a trap-rock quarry, and explanations will be given of the flow of rock from the freshly blasted face through the crushing and sorting stages to its final shipment. We will look at the structures and distribution of rock types within the Holyoke flow in the vertical quarry walls from a safe distance and have a chance to sample the various rock types in a collection of large boulders in the center of the quarry floor. During the visit no one is allowed near the quarry walls. These are extremely dangerous because loose blocks are continuously falling from them.

### **HISTORY OF THE NORTH BRANFORD QUARRY**

Despite the present importance of the North Branford quarry, it began as a minor business venture of a Mr. Louis Fisk, who in the late 1890s owned a horse-trotting racetrack in Branford, Connecticut (Reigeluth, 1962). To facilitate access to the racetrack, he built a three-mile-long railway connecting it to the New York, New Haven and Hartford Railroad. At the same time, early environmentalists became concerned about the deleterious effect of quarrying trap rock from the Palisades sill on the west bank of the Hudson River. In 1900 the Palisades Interstate Park Commission was formed, and all but one of the trap-rock quarries were closed, much to the pleasure of the wealthy residents of New York City and Westchester County. In fact, J. P. Morgan contributed most of the \$132,500 that was required to purchase the large trap-rock quarry at Fort Lee. The closing of these quarries increased the demand for Connecticut trap rock. Mr. Fisk realized that if he extended his railway beyond the racetrack to Totoket Mountain (Indian name for Great Hill) where he owned 319 acres underlain by trap rock, the railway could be used for shipping crushed stone, as well as race enthusiasts. Although Mr. Fisk appears eventually to have lost controlling interest in the venture, the *New Haven Trap Rock Company* opened the North Branford Quarry in 1914. Although it has seen several changes in ownership since its inception, the quarry has remained in continuous operation to the present. Today, Mr. Fisk's rail bed is still the same one that transports the crushed stone from the quarry to barges on Long Island Sound, although the old steam engine and rails have been replaced with modern diesels and tracks.

When the North Branford quarry first opened in 1914 it produced 2000 tons of crushed stone per day, which was impressive at that time, but with modern techniques the quarries daily production is now ten times that amount. The properties of strength and durability that make basalt a desirable construction material also make it tough to crush. This is done at the North Branford Quarry today with a manganese steel gyratory crusher-- a single crusher handles the 20,000 tons crushed each day. Stone crushers used today in quarries and mines around the world are all descended from the jaw crusher invented in New Haven, in 1857, by Eli Whitney Blake, nephew of the famous inventor Eli Whitney (Cotton Gin, musket mass production). In 1852 Blake was appointed superintendent of a project to install a macadamized road from New Haven to Westville, a distance of several miles, along what is today Whalley Avenue. The road was to be constructed from trap rock quarried from the West Rock sill in Westville. Blake was dissatisfied with the inefficiencies of hand breaking of the stone as done in those days--mainly by prisoners--and invented a steam driven rock crusher to increase speed of output and to improve quality control. The method was simple. One sheet of steel was fixed, and a second was placed to form a V-shaped hopper with the first.

The second sheet was hinged at the top, and the bottom was attached to a steam engine that moved it back and forth about an inch. When the sheet pulled back, a stone would slip down; when the sheet moved forward, the stone was crushed. After repeated back and forth motions, any stone big enough to fit in the top of the hopper would be crushed to a size that would fall through the bottom opening between the sheets. The bottom opening was adjustable, thereby allowing control of the size of the aggregate produced. Blake stone crushers and their descendants have arguably been the most valuable invention to come from New England.

In the North Branford quarry today, the freshly blasted stone, after being dumped from 65-ton-quarry trucks into the primary gyratory crusher, begins a 2-mile-long trip on conveyor belts that take it to other crushers and sorters. The major products are crushed stone for road and rail beds, and aggregate for asphalt and concrete. Basalt is, without doubt, the most desirable stone for all of these purposes, because of its durability and density. Its durability is a direct consequence of its texture, which is determined by the way in which the basaltic magma originally crystallized. We will address this point below.

### GENERAL GEOLOGIC SETTING OF THE QUARRY

The North Branford quarry is located in the extreme southeastern part of the Hartford Mesozoic Basin, the southern end of the quarry actually terminating against the basin's eastern border fault (Figs. 1 and 2). The basin was formed during the Mesozoic as a result of the extension that eventually rifted apart Pangea to form the Atlantic

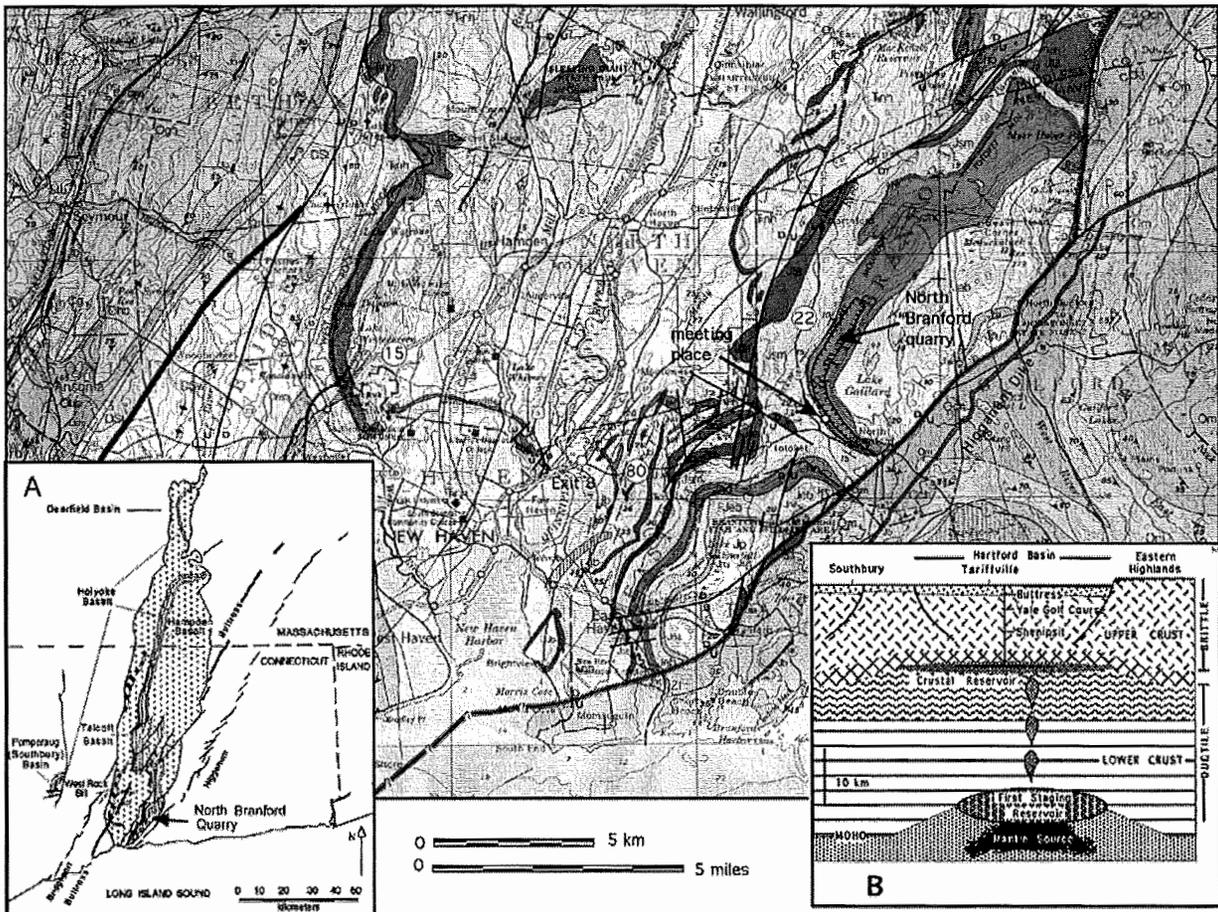


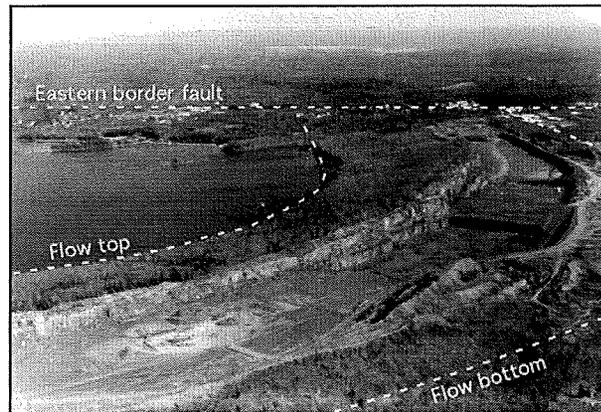
Figure 1. Geological map of the southeastern part of the Hartford Basin (from Rodgers, 1985) and location of the North Branford quarry. Inset A shows the distribution of the three volcanic units in the Hartford Basin—Talcott, Holyoke, and Hampden—and their associated dikes (Higginum, Buttress, and Bridgeport) and sills. Inset B shows a hypothetical cross section across the Hartford Basin at the time of the Holyoke eruption (Philpotts, 1998).

Ocean (Olsen, 1997, 1999). During the Triassic, the basin was filled initially with coarse arkosic sandstones (New Haven Formation), but at the beginning of the Jurassic, large lakes partly filled the basin so that in addition to alluvial fan deposits around the basin's margin, finer fluvial and lacustrine sediments were deposited in its central part. At the same time igneous activity associated with the extension began with the eruption of the Talcott basalt and intrusion of West Rock sill into the base of the basin's sedimentary fill. Most of the Talcott was erupted into a lake and consequently is pillowed. Climatic changes related to Milankovitch cycles caused the lakes to expand and contract repeatedly, leaving a record of alternating shallow (red, coarser, cracked mud silt and sand) and deep (black, finely laminated shale) water deposits. Based on these Milankovitch cycles in the ~100 m of Shuttle Meadow Formation overlying the Talcott basalt, ~138,000 years elapsed before the next episode of igneous activity (Olsen, 1984; Weems and Olsen, 1997), which was marked by the eruption of the Holyoke basalt, the rock quarried at North Branford. At the time of its eruption, no significant lakes existed in the basin because the Holyoke forms a massive flow. However, in the Deerfield Basin to the north of the Hartford Basin (Fig. 1), the base of the Holyoke is pillowed, which indicates that a lake existed in that basin. Milankovitch cycles in the ~170-m-thick East Berlin Formation overlying the Holyoke basalt indicate that ~345,000 years passed before the eruption of the Hampden basalt, the last of the igneous rocks in the Hartford Basin.

During extension, the faults bounding the eastern side of the Hartford Basin were more active than those on the western side. As a result, all rocks in the basin dip gently eastward. The three lava flows form a north-south trending ridge which is approximately in the center of the basin except at its southern end, where a series of left lateral strike-slip faults moved the lava flows to the eastern side of the basin (Fig. 1A). Differential amounts of down-drop along the eastern border fault warped the basalts into a series of gentle eastward plunging synclines and anticlines. One such anticline immediately to the south of the North Branford quarry is responsible for warping the Holyoke basalt around from a northeasterly to a northwesterly strike and having the flow truncate against the eastern border fault (Figs. 1 and 2). To the north of this, however, greater down drop on the border fault caused the Holyoke to sag into a gentle syncline, which explains the arcuate shape of the outcrop pattern and the shape of the quarry (Fig. 1).

This arcuate ridge of resistant Holyoke basalt and the less resistant overlying sedimentary rocks of the East Berlin Formation formed a natural basin to the east of the quarry, which has been dammed to form Lake Gaillard, one of the region's major reservoirs (Fig. 2). Water from the reservoir passes beneath the quarry through the Great Hill Tunnel and is delivered to a filtration plant for distribution into the New Haven area.

Figure 2. Oblique aerial view looking southeast across the North Branford quarry and the Lake Gaillard water reservoir. The small building in the lake near the southwestern shore is the water intake to the Great Hill tunnel, which passes beneath the elevated section of the quarry floor to a filtration plant on the south side of Route 80. The eastern border fault of the Hartford Basin and the upper and lower contacts of the Holyoke flow are shown with dashed lines. Long Island Sound is visible in the background. (Photograph by David Burkett)



The three volcanic units in the Hartford Basin (Talcott, Holyoke, Hampden) have distinct chemical compositions (Puffer et al., 1981; Philpotts and Reichenbach, 1985), which allow them to be correlated with the three regional diabase dikes that traverse southern New England (Philpotts and Martello, 1986). The Talcott basalt erupted from the longest and most easterly of these dikes, the Higganum (Fig. 1), which can be traced for over 750 km across Connecticut, Massachusetts, and coastal Maine into New Brunswick (McHone, 2003). The Holyoke basalt erupted from the Buttress dike, so named because of its topographic prominence in the cliff just north of the West Rock Tunnel on Route 15 where the dike cuts the West Rock sill (Fig. 1). The Hampden basalt erupted from the Bridgeport dike, the most westerly of the three dikes (Fig. 1A). Each of these dikes is up to 50 m wide, and extensive contact metamorphism and melting of the wall rocks testify to the passage of large quantities of magma, probably in turbulent flow (Philpotts and Asher, 1993). Only the feeder to the Talcott can be seen to actually

connect with its volcanic products in present exposures. The Buttress dike should connect with the Holyoke flow in Meriden (Fig. 1A), but here the base of the flow is concealed beneath thick talus deposits. However, immediately to the south of this, where the Buttress dike passes west of the Sleeping Giant Laccolith (Fig. 1), the dike contains a central zone of breccia formed where the erupted magma drained back into the dike. Bent pipe-stem vesicles at the base of the Holyoke flow in the southern part of the Hartford Basin indicate that the lava flowed away from the Buttress dike in an east-northeasterly direction until it became ponded against the escarpment of the eastern border fault, which accounts for the great thickness of the Holyoke flow in the vicinity of the North Branford quarry.

## THE HOLYOKE BASALT FLOW IN THE NORTH BRANFORD QUARRY

### General Features

The Holyoke basalt forms a single thick flow that can be traced northward for 160 km through the Hartford and Deerfield Basins and 50 km to the west through the Pomperaug Basin. With a thickness that varies from 57 m in the Pomperaug Basin to 200 m in the North Branford quarry, the total volume of this single eruption was probably in excess of 1000 km<sup>3</sup> (Philpotts, 1998). In the vicinity of Farmington, a second thinner Holyoke flow erupted soon after the first flow, but this flow is not found in other parts of the basin. The great thickness of the Holyoke flow in the North Branford Quarry is most likely due to ponding of the lava against the escarpment formed by the eastern border fault. The lower contact of the flow can be found on the western side of the quarry in the steep hillside on the east side of Route 22 (Fig. 2). The top of the flow follows the western shore of Lake Gaillard (Fig. 2), at the northern end of which, fluvial sediments of the East Berlin Formation lie directly on a vesicular flow-top breccia (Fig. 3), indicating that none of the flow was removed by erosion prior to the deposition of the sediments. These sediments were deposited from water that flowed across the surface of the flow soon after the eruption.

Many thick flood basalt flows are thought to thicken by a process known as inflation (Self et al. 1996), whereby successive pulses of magma intrude into the still molten interior of the initial flow. Flows thickened by inflation contain multiple layers of vesicles, formed when the gas escaping from each new pulse of magma accumulated beneath the thickening crust of the flow. In the case of the Holyoke flow in the North Branford quarry, however, only one period of vesicle accumulation is evident. Vesicles systematically increase in size and decrease in number downward from the surface of the flow to a depth of 25 m, below which all macroscopic vesicles disappear. Below this depth, there was sufficient time for the vesicles to rise into the top 25 m of the flow. This evidence combined with the fact that the flow has only one entablature (fractures propagating downward from the top of the flow) and one colonnade (fractures propagating upward from the bottom of the flow) indicates that the Holyoke basalt in the vicinity of the North Branford quarry cooled and solidified as a single unit in a gigantic ponded flow. Lava lakes of comparable thickness have formed in historic times in numerous Hawaiian calderas, but none has had the great lateral extent of the Holyoke flow.

Along most of the length of the east wall of the North Branford quarry, a prominent cusped boundary separates the entablature from the colonnade (Fig. 4). The entablature is formed by fractures that propagated down from the surface of the flow, and the colonnade is formed by fractures that propagated up from the bottom (Ryan and Sammis, 1978; DeGraff and Aydin, 1987). The fractures break the basalt into polygonal columns, those in the entablature typically having a smaller diameter than those in the colonnade. The columns in the entablature also commonly form clusters that radiate downward, and it is the tips of these radiating clusters that produce the cusped shape of the entablature-colonnade boundary. This boundary is a prominent feature of the Holyoke basalt throughout the Hartford Basin and can be seen in many cliff exposures, as for example on Higby Mountain and the Hanging Hills of Meriden.

In thin flows, the entablature commonly constitutes two-thirds to four-fifths of the flow, as might be expected because cooling from above is more rapid than cooling from below (DeGraff and Aydin, 1987). However, at the North Branford quarry this boundary occurs 120 m above the base, which means that the entablature occupies only two-fifths of the flow. This fact is one of the important clues concerning how this flow solidified. One cannot invoke the unreasonable condition that cooling was slower from above than from below; besides, the texture of the rock in the entablature indicates that it did indeed crystallize more rapidly than did the rock in the colonnade (see textures below).

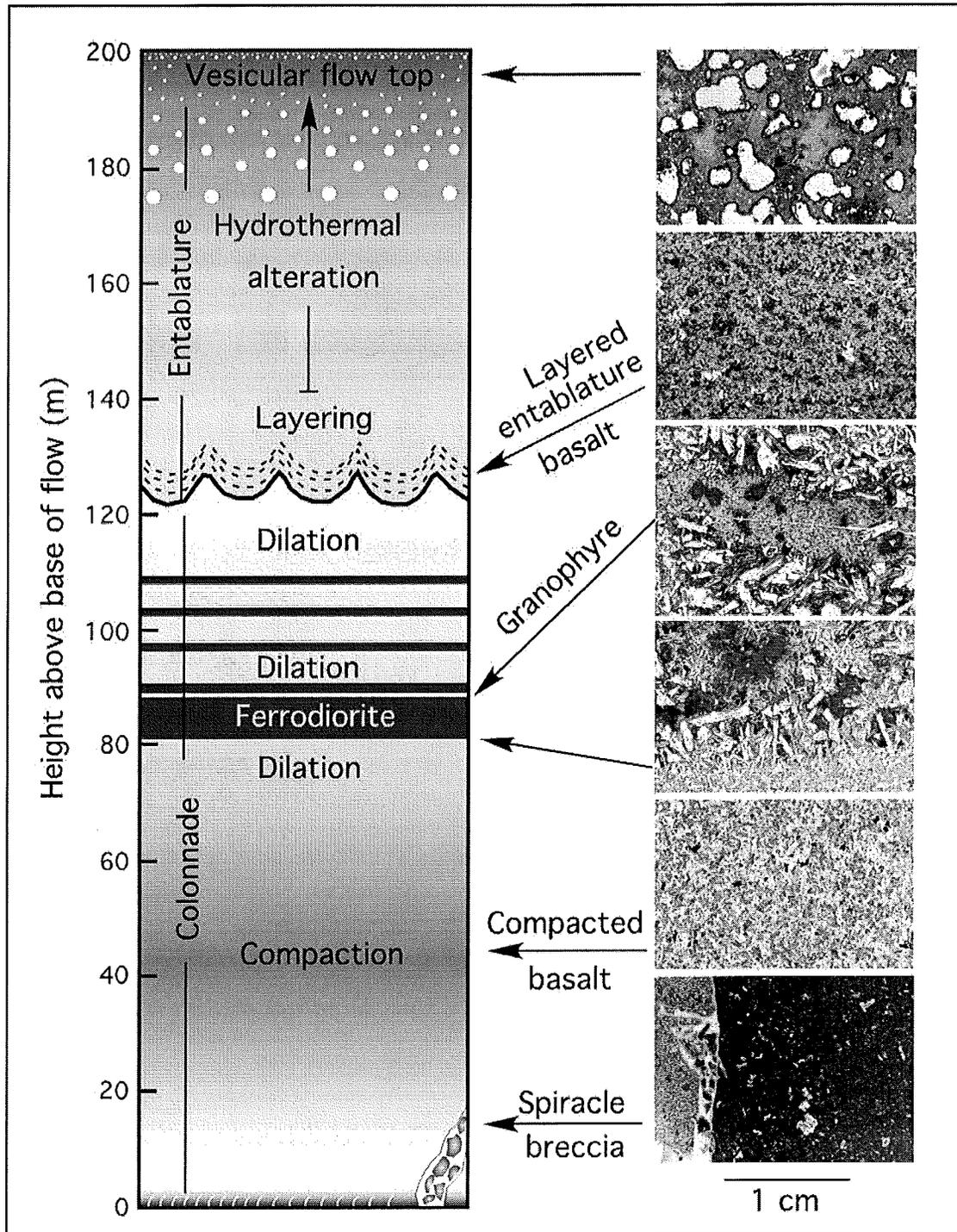


Figure 3. Photomicrographs of the different rock types in the Holyoke flow at the North Branford quarry under plane-polarized light and at a magnification given by the 1 cm scale bar. From top to bottom: Vesicular flow top with altered plagioclase phenocrysts in a devitrified matrix. Basalt of the entablature with small ophitic clusters of pyroxene and plagioclase crystals separated by dark patches of mesostasis. The ophitic clusters and patches of mesostasis form a horizontal layering. Sheet of fine-grained granophyre in the upper part of a ferrodiorite sheet. Lower contact of coarse-grained sheet of ferrodiorite with fine-grained basalt. Basalt of the colonnade showing clusters of small plagioclase crystals surrounding patches of granular pyroxene and equant magnetite grains. Spiracle breccia with fragments of crystallized basalt (left) and fragments with quenched glassy margins (right).

Upward displacement of the final temperature maximum is common in intrusive bodies (Irvine, 1970), as for example in the Palisades sill where it occurs only 50 m below the top of the 300-m-thick body (Gorring and Naslund, 1995). The upward displacement of the temperature maximum in intrusive bodies is attributed to crystal mush sinking from the roof to the floor of the body during solidification (Jaupart and Brandeis, 1986; Brandeis and Jaupart, 1986; Marsh, 1988; Bergantz and Ni, 1999). The position of the entablature-colonnade boundary in the Holyoke flow is evidence that processes similar to those taking place in intrusive bodies may also occur in thick sheets of basaltic magma solidifying on the surface of the earth.

Despite its 200-m thickness, the Holyoke flow is fine grained throughout, except for sheets of ferrodiorite in its central part, which typically are extremely coarse (~1 cm). In contrast, the 50-m-wide Buttress dike, which fed the flow, attains a 1-cm grain size within only 25 m of its contact with Mesozoic sedimentary rocks. This again provides a clue about the solidification of the Holyoke flow. For the grain size to remain fine, despite a lengthy cooling period in the center of the flow, large numbers of crystal nuclei must have formed early in the solidification of the flow. Once formed, they ensured that the final rock would have large numbers of crystals and thus be fine grained. If a liquid can form that contains few nuclei, the cooling in the central part of the flow was slow enough to grow large crystals. The coarse grain size of the sheets of ferrodiorite is evidence that such growth was possible. If material sank from the roof of the magma sheet, as is implied by the upward displacement of the entablature-colonnade boundary, large numbers of small crystals would have been continuously introduced into the body of the flow (Bergantz and Ni, 1999).

Although the basalt in the North Branford quarry appears macroscopically homogenous (except for the ferrodiorite and granophyre), significant textural changes occur with height in the flow. These textures provide important clues to processes operating during the solidification of the flow and which caused it to differentiate and produce bodies of ferrodiorite and granophyre. Figure 3 shows photomicrographs illustrating these textures from various heights in the flow.

The top of the flow, as exposed in a recent excavation at the northwest end of Lake Gaillard, consists of fist-size fragments of scoriaceous basalt containing ~30% vesicles (Fig. 3). This passes downward over a meter or two into massive vesicular basalt. The basalt contains phenocrysts of plagioclase and minor olivine in a groundmass of devitrified glass and feathery microlites of plagioclase. All of the primary minerals have been totally altered, but the texture still reveals the original makeup of the rock. The vesicles decrease in abundance to only 6% at a depth of 25 m. Below this, macroscopic vesicles abruptly disappear, but microscopic ones are present to a depth of 60 m, and dictyaxitic cavities are present throughout the flow, where they typically constitute ~4% of the rock. The dictyaxitic cavities are filled with a clay mineral that in hand specimen appears black and glassy. This material has been mistakenly identified as residual glass in many eastern North American Mesozoic basalts. At 25 m beneath the flow top, the macroscopic vesicles are variable in size but are on the scale of a centimeter and are irregularly distributed. These vesicles undoubtedly rose from deeper in the flow and were arrested in their upward migration by the downward solidifying crust. The fact that no macroscopic vesicles are present below this depth indicates that the underlying magma cleared itself of bubbles by the time the solidification front reached this depth.

Beneath the scoriaceous flow top the entire upper part of the flow down to the entablature-colonnade boundary is characterized by ophitic pyroxene-plagioclase clusters that are separated from each other by intersertal plagioclase crystals and patches of dark mesostasis. Many plagioclase crystals in the ophitic patches are cone shape and radiate from the center of the patch. The pyroxene must have grown rapidly because it has a disequilibrium sub-calcic augite composition (Philpotts and Dickson, 2000). Approximately 5% plagioclase phenocrysts are present, and small olivine crystals (now totally altered to clay) occur near the center of the pyroxene-plagioclase clusters where they are rimmed by pigeonite. The ophitic pyroxene-plagioclase clusters tend to be linked together into planes that give the rock a subtle 2-mm scale sub-horizontal layering, which is evident on polished glaciated surfaces and in thin sections (Fig. 3). The layers of pyroxene-plagioclase clusters are separated from each other by discontinuous sheets of extremely fine-grained dark mesostasis consisting of magnetite dendrites, small pyroxene spheres (crystallized immiscible iron-rich liquid), and apatite needles in a granophyric groundmass. This material formed from the liquid residue (rejected components) that built up ahead of the growing clusters of pyroxene and plagioclase. Continued growth of these clusters required that pyroxene and plagioclase components in the melt ahead of the crystallization front were able to diffuse through the residual melt zone. Once the diffusion gradient across this zone became too small, a new layer of pyroxene-plagioclase clusters would have nucleated (Philpotts and Dickson, 2002). This layering marks the shape of the downward crystallizing roof zone of the flow.

East Face of Tilcon's North Branford Quarry

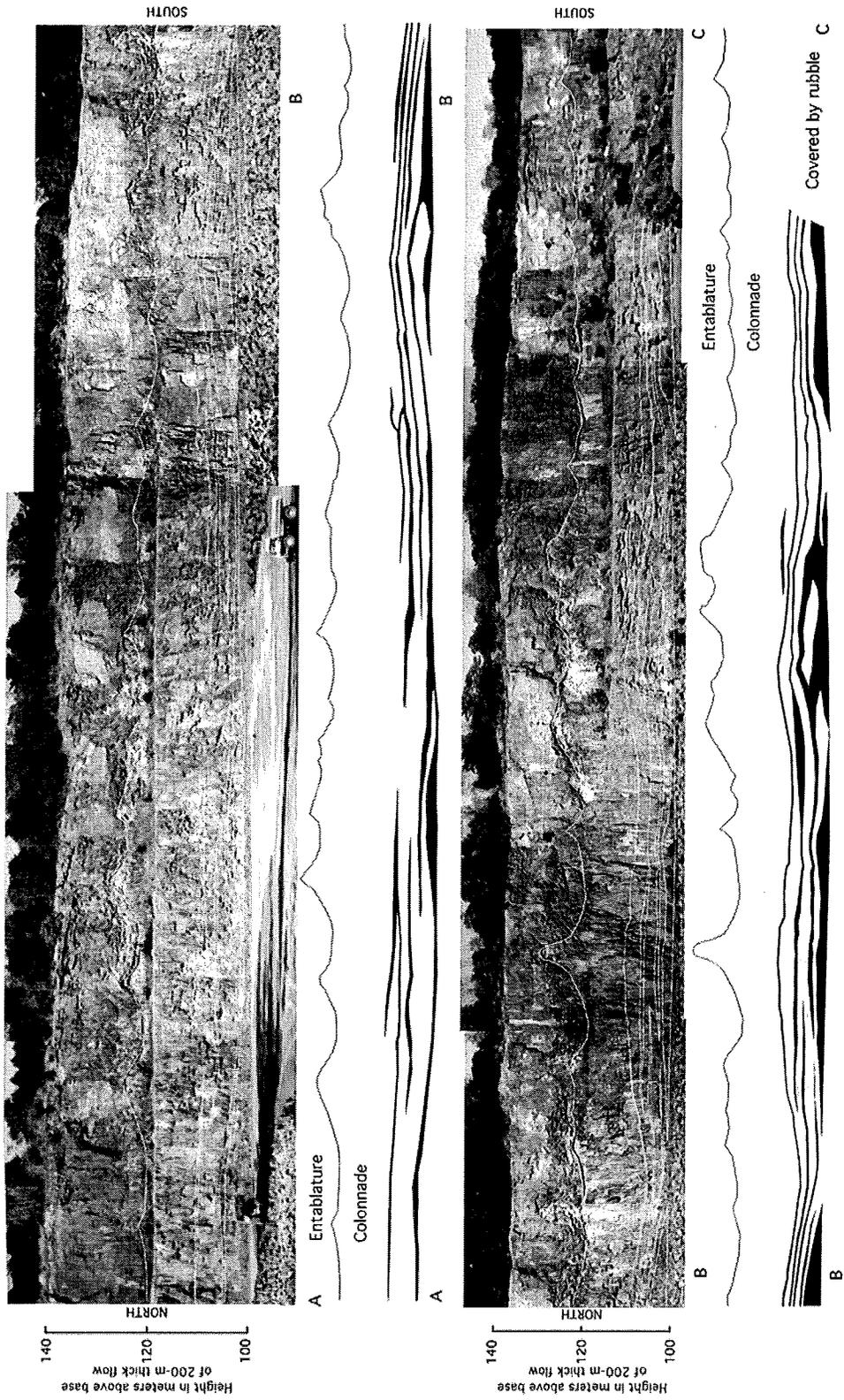


Figure 4. Photo-mosaic of part of the east wall of the North Branford quarry with boundaries between entablature and colonnade and contacts of ferrodiorite sheets highlighted in white and drawn in black below the photograph.

Of particular interest is the fact that this primary igneous layering parallels the cusped boundary between the entablature and the colonnade. This boundary is therefore more than simply the surface along which downward and upward propagating fractures met. It shows the shape of the downward solidifying roof zone of the flow. The radiating clusters of columns at the base of the entablature propagated in directions perpendicular to the crystallization front, which would have been an isothermal surface. The cusped nature of this surface is probably related to water descending through the crust along prominent fracture intersections (DeGraff, J.M., and Aydin, A., 1987). Fractures that may have acted as "cold fingers" can be seen in the center of some of the cusps on the east wall of the quarry.

On crossing the entablature-colonnade boundary, the texture of the rock changes immediately to intergranular, with plagioclase crystals clustered together into a 3-D network that wraps around isolated patches of granular pyroxene grains (Fig. 3). Magnetite forms equant grains rather than the dendritic ones found in the entablature. Quartz and alkali feldspar form a granophyric intergrowth, which almost invariably is enclosed within the plagioclase network. As in the entablature, ~5% plagioclase phenocrysts are present, but olivine is totally absent. The granular patches of pyroxene consist of augite and pigeonite that form polygonal grains with 120° grain boundary junctions. Unlike the pyroxene in the entablature, the augite and pigeonite have equilibrium compositions plotting on either side of the pyroxene solvus (Philpotts and Dickson, 2000). Each granular patch of pyroxene is about the same diameter as the pyroxene oikocrysts in the entablature, and like those crystals, many granular patches contain cone-shaped crystals of plagioclase radiating from their center. The texture of the rock in the colonnade is homogeneous and there are no prominent alignments of crystals. The rock is consequently resistant to fracture and abrasion, which explains its desirability as a construction material.

The textural difference between the entablature and colonnade in the Holyoke flow is typical of the differences found between the entablature and colonnade in other thick flood-basalt flows. Long and Wood (1986) were the first to point out that the texture of the entablature is characteristic of rapid quenching, which they believed resulted from the surface of the flow being flooded with water that descended through fractures in the crust and promoted rapid cooling. Lyle (2000) showed that fluvial sediments, which indicated that water was probably responsible for the rapid quenching of the entablature, invariably overlay flood-basalt flows with prominent entablatures. The most likely interpretation of the texture in the colonnade is that this rock was formed from material that began as a rapidly grown crystal mush in the roof zone of the flow but separated and sank as a dense plume to the floor of the sheet where it recrystallized and solidified more slowly than it would have done had it remained in the roof zone. Material that did not detach from the roof retained the quench textures of the entablature.

Between 120 m (entablature-colonnade boundary) and 80 m above the base of the flow, the fine-grained basalt contains multiple horizontal sheets of coarse-grained ferrodiorite (Figs. 3 and 4). Although the typical grain size of this rock is on the centimeter scale, individual pyroxene crystals up to several centimeters long are common. The thickness of most sheets is on the decimeter scale, but the lowest one is 10 m thick. The thinner sheets are spaced at regular intervals of ~1 m. The thicker sheets can be traced along the entire length of the east wall of the quarry (Fig. 4). Some sheets terminate because they migrate up or down to connect with overlying or underlying sheets. Rare discordant dikes connect adjoining sheets. The ferrodiorite consists of long blades of plagioclase, long twinned prisms of augite commonly with a core of pigeonite, abundant magnetite, patches of clay mineral that have probably formed from fayalitic olivine that encloses blades of tridymite, abundant granophyre containing needles of apatite, and dictyotaxitic cavities filled with a dark clay mineral, which in hand specimen has the appearance of black interstitial glass. The contacts with the adjoining basalt are sharp and show no signs of the ferrodiorite being quenched against the basalt. Instead, large crystals of plagioclase and pyroxene nucleated on the contacts and grew into the ferrodiorite sheets (Fig. 3). Many crystals of plagioclase and pyroxene are bent through as much as 90° in extreme cases.

Toward the top of many ferrodiorite sheets, thin irregular sub-horizontal sheets of fine-grained granophyre are present (Fig. 3). They have sharp contacts with the ferrodiorite, but some of the large plagioclase and pyroxene crystals in the ferrodiorite protrude into the granophyre, and some of the granophyre connects with interstitial patches of granophyre in the ferrodiorite. The granophyre is extremely fine grained and appears almost like chert in the hand specimen. It contains euhedral patches of clay mineral that were probably originally fayalite. Small augite crystals are zoned to hedenbergite compositions on their rims. Small vesicles filled with the same "black glassy" clay mineral are common. These granophyre sheets are clearly formed from a highly fractionated gas-rich residue.

Beneath the ferrodiorite sheets, the basalt appears remarkably homogeneous down to the lower contact, where a quenched margin contains bent pipe stem vesicles indicating an east-northeasterly flow. Despite the apparent homogeneity, chemical analyses and detailed fabric analysis indicate that the basalt in the lower third of the flow underwent compaction during solidification, which is indicated in Figure 3 by shading. Residual liquid expelled from the zone of compaction rose into the central part of the flow where it dilated and ruptured the crystal mush to form sheets of ferrodiorite. The evidence for the compaction is presented below.

One other rock type that is rarely encountered in the quarry is a breccia consisting of angular fragments of basalt in a white matrix consisting of quartz, calcite, and anhydrite crystals. The fragments are of interest because while some consist of the normal basalt found in the colonnade, others were clearly molten at the time of brecciation, and as a result, their margins are quenched to a glass (Fig. 3). The most likely origin for these breccias is as spiracles, that is, steam explosions from the base of the flow.

### Chemical Composition of the Holyoke Flow

Samples were collected at approximately 2-m intervals between the base and top of the Holyoke flow in the North Branford quarry and western side of Lake Gaillard. Analyses of typical samples are given in Table 1 and profiles of MgO, TiO<sub>2</sub>, Na<sub>2</sub>O, and CaO through the flow are shown in Figure 5.

Table 1. Representative analyses of Holyoke basalt from North Branford quarry, Connecticut.

Sample	Lower chill	Average of	Most compacted	Average Entablature	Ferrodiorite Segregation	Granophyre
Height (m)	4.03	0 - 140 m	46.31	118-125	90.48	105.08
SiO <sub>2</sub>	53.39	53.33	53.04	53.07	54.55	70.69
TiO <sub>2</sub>	1.04	1.03	0.81	1.04	1.67	0.55
Al <sub>2</sub> O <sub>3</sub>	14.08	14.06	15.94	14.10	13.16	11.90
Fe <sub>2</sub> O <sub>3</sub>	4.32	3.33	2.59	3.41	5.14	5.91
FeO	8.40	9.27	8.10	9.50	10.18	2.19
MnO	0.20	.21	0.18	0.21	0.22	0.11
MgO	5.81	5.46	5.85	5.55	3.32	1.19
CaO	9.19	9.76	10.02	9.90	7.77	2.92
Na <sub>2</sub> O	2.60	2.52	2.64	2.44	2.68	2.69
K <sub>2</sub> O	0.75	0.70	0.68	0.72	0.98	1.54
P <sub>2</sub> O <sub>5</sub>	0.10	0.07	0.07	0.07	0.20	0.18
Total	99.89	99.75	99.91	100.00	99.87	99.88

The Holyoke basalt is a low-titanium variant of Weigand and Ragland's (1970) high-iron quartz tholeiite, one of the common magma types in the northern Mesozoic basins of eastern North America. Its composition indicates that if it were derived from a source containing plagioclase, olivine, and augite, it would have formed at a pressure of 3.8 kbar (Yang et al., 1996). On this basis, Philpotts (1998) interpreted that the magma must have risen from a mantle source to a mid-crustal reservoir (~12 km) that formed near the brittle-ductile transition in the crust (Fig. 1 B), where it fractionated to its present composition. Similar mid-crustal reservoirs have been identified beneath the Basin and Range province (Holm, 1995), the Mackenzie dikes in Canada (Mandler and Clowes, 1997) and the Siljan Ring area in Sweden (Juhlin, 1990).

The systematic sampling through the flow indicates that it is not homogeneous. In Figure 5, the vertical dashed lines indicate the average concentrations of MgO, TiO<sub>2</sub>, Na<sub>2</sub>O, and CaO based on analyses between 0 and 140 m above the base of the flow. Above this the composition has been affected by hydrothermal alteration, as can be seen from the steady increase in Na<sub>2</sub>O and decrease in CaO between 140 m and the top of the flow. The dashed lines provide references with which to compare the composition of the rock at any given height.

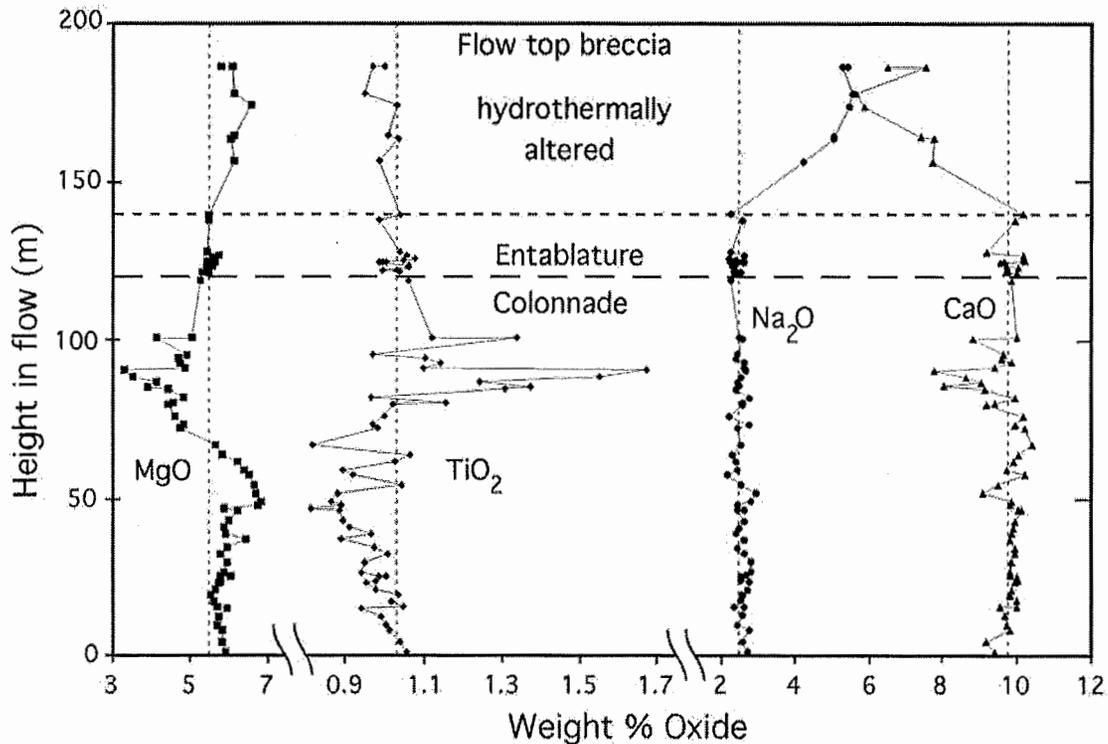


Figure 5. Chemical profiles through the Holyoke flow in the North Branford quarry.

The profiles of MgO and TiO<sub>2</sub> illustrate the behavior of compatible and incompatible elements respectively, that is, elements that enter or are rejected, respectively, by early crystallizing minerals in a basaltic magma. The profiles of these two components are mirror images of each other. While MgO steadily increases away from the base of the flow, TiO<sub>2</sub> steadily decreases. The MgO reaches a maximum and TiO<sub>2</sub> a minimum at 50 m above the base of the flow. Above this, the trend is reversed and MgO decreases and TiO<sub>2</sub> increases until at 85 m the first ferrodiorite sheet is encountered, which has significantly lower MgO and higher TiO<sub>2</sub>. Between 85 and a 120 m, the fine-grained basalt has low MgO and high TiO<sub>2</sub> relative to the average composition, with prominent spikes being analyses of ferrodiorite sheets. On crossing from the colonnade into the entablature the rock returns to the average composition, which is similar to the lower chilled margin (Table 1), and remains at this value until the zone of hydrothermal alteration is encountered at 140 m. Relatively immobile elements such as titanium are not significantly affected by the alteration and remain at about the same concentration through the upper part of the flow. Sodium and calcium, and to a lesser extent magnesium, are affected by the hydrothermal alteration, which clearly becomes more severe toward the top of the flow.

Similar profiles have been found in the Holyoke flow in Tilcon's quarry in Wallingford and in the Tariffville gorge on the Farmington River. Philpotts et al. (1996) showed in the case of the Tariffville section that this compositional variation could be explained only by invoking compaction of a crystal mush in the lower part of the flow with upward expulsion of the residual liquid. The variation cannot be explained by sinking of early crystallizing dense minerals, because this would bring about enrichment in olivine and pyroxene components but not those of plagioclase, because plagioclase was slightly buoyant in the Holyoke magma. The lower part of the flow is enriched in both pyroxene and plagioclase in precisely the proportions that these minerals crystallized from the magma. The bulk density of this pyroxene-plagioclase mixture was greater than that of the liquid, so as long as they formed an interconnected network they were able to undergo compaction and expel upward the residual liquid, which took with it the incompatible elements, such as titanium. The upward migrating liquid was trapped beneath the downward-solidifying roof zone, and as its pore pressure increased due to continued compaction below, it eventually ruptured the crystal mush, with the residual liquid filling horizontal fractures and crystallizing to ferrodiorite.

This compaction process was successfully modeled using the IRIDIUM program of Boudreau (2003), which links a version of the MELTS program of Ghiorso and Sack (1995) for calculating mineral-liquid equilibria with standard mass and heat transport equations developed by McKenzie (1984) and Shirley (1986). Boudreau and Philpotts (2002) showed that the chemical profiles through the Holyoke flow at Tariffville could form through compaction within a period of 25 years, a period well within the solidification time for a flow as thick as the Holyoke.

The chemical profiles through the Holyoke flow in the North Branford quarry are believed to have formed through the same compaction process as at Tariffville. Evidence confirming that compaction did indeed take place in the lower part of this flow is preserved in the textural anisotropy of the basalt in the colonnade. This anisotropy is revealed through quantitative image analysis of the rock's texture.

**Detailed Textural Analysis of Basalt in Colonnade**

The texture of the basalt in the colonnade of the Holyoke flow is typical of the texture in the colonnade of other thick flood-basalt flows. Figure 6 shows a 20 x 30 mm petrographic thin section of sample NB 16 from 40.6 m above the base of the flow in the North Branford quarry. This sample shows the typical clustering of plagioclase crystals around the granular patches of pyroxene crystals. X-ray CT scans of rock with the same texture from Tariffville reveal that the plagioclase crystals form a continuous 3-D network surrounding the pyroxene clusters (Philpotts et al., 1999). Melting experiments reveal that this network formed when the rock was only one-third crystallized. If this early forming network were initially isotropic, compaction would have distorted it, and its anisotropy should provide a quantitative measure of the degree of compaction. A network was drawn by tracing the centerlines along clusters of plagioclase crystals in oriented thin sections (Fig. 6). Initially (Philpotts et al., 1999), traverse lines were drawn through this network at various angles to the vertical, and the average intercept spacing with the network was measured manually as a function of traverse direction. If the network were isotropic the average intercept with the network should be the same in all directions, but if it were anisotropic the average intercept would be less in a direction of compaction. Subsequently, the measuring of the intercepts was automated and done by computer on digitized tracings of the network (Gray et al., 2003). Figure 7 shows the computer output for the network traced from sample NB 16. In the bottom graph, the average intercept of traverse lines with the network can be seen to vary smoothly from a minimum of 1.02 mm in a direction near vertical to a maximum of 1.15 mm near horizontal. In the graph above these values have been converted into percent compaction based on the assumption that the network was initially homogenous. Apparently this sample underwent ~11% compaction in a direction near vertical based on the orientation of the columns in the colonnade.

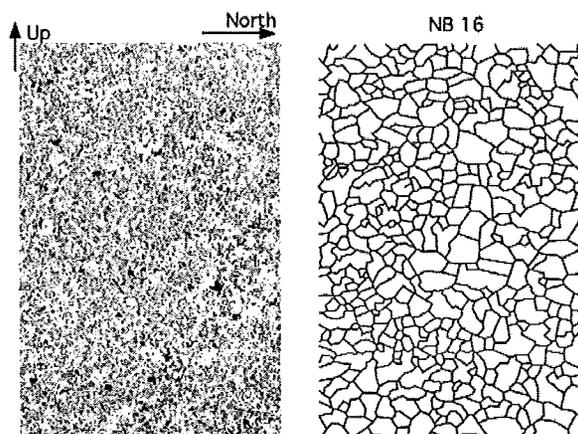


Figure 6. Scanned vertical oriented 20 x 30 mm thin section of a typical sample of basalt from the colonnade showing clusters of plagioclase crystals surrounding granular patches of pyroxene. The network of plagioclase crystals has been traced in the adjoining figure.

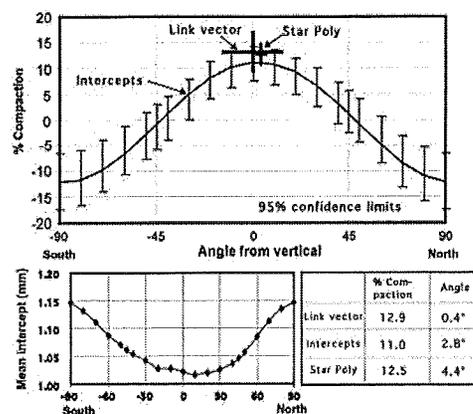


Figure 7. Variation in the length of the average intercept along traverse lines through the network shown in Figure 6 as a function of angle of traverse line (0 is vertical). Assuming these intercepts were initially the same in all directions the degree of anisotropy is used to calculate the amount and direction of compaction. Results from two other methods of calculating the anisotropy (link vector and star polygon) are also shown (see text for discussion).

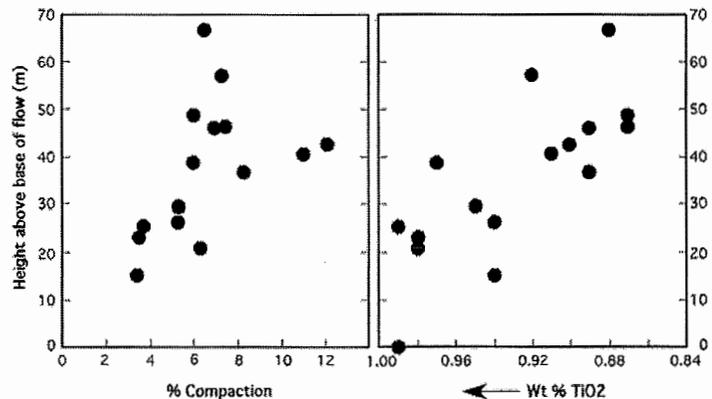
In addition to measuring the average intercept in the network, it is possible to take the individual line segments that make up the network and plot them as vectors, each line segment giving a point in an  $x$ - $y$  plot. If the network were isotropic the distribution of points around the origin in the  $x$ - $y$  plot would be spherical, whereas if the network were anisotropic the distribution would be elliptical and the magnitude and directions of the major and minor axes of the ellipse could be used to determine the magnitude and direction of compaction. The computer also performs this analysis automatically on the digitized network (Gray et al., 2003). The results for sample NB 16 are shown in Figure 7 as the “link vector” values, which indicate 12.9% compaction in a direction near vertical.

A third independent way of analyzing the anisotropy of the network is to fit an ellipse to each polygon constituting the network and then to determine a weighted average of all ellipses in the whole thin section. This analysis technique, referred to as the “star polygon” method, is also automated once the network has been digitized (Gray et al., 2003). The value obtained for the network in sample NB 16 indicates 12.5% compaction in a direction tilted  $4^\circ$  from vertical.

These three independent methods of determining the anisotropy of the network of plagioclase crystals in sample NB 16 agree well at the 95% confidence limit in both the amount and direction of compaction. If we accept the assumption that the network was initially isotropic, this sample underwent  $\sim 12\%$  compaction in a direction within a few degrees of vertical.

The textural anisotropy has been measured in samples between 15 and 68 m and generally, as the concentration of incompatible components decreases ( $\text{TiO}_2$ , for example), the degree of textural anisotropy increases (Fig. 8). This inverse relation is consistent with the crystal mush having undergone compaction and implies that the assumption was reasonable that the plagioclase crystal network was isotropic when it first formed. The textural anisotropy can therefore be used to give a quantitative physical measure of the amount of compaction that the crystal mush underwent during solidification. In the case of the Holyoke basalt in the North Branford quarry the maximum degree of compaction was  $\sim 12\%$  and this occurred between 40 and 50 m above the base of the flow.

Figure 8. Variation in the amount of compaction as measured from textural anisotropy and variation in amount of  $\text{TiO}_2$  as a function of height in the lower part of the flow. Note  $\text{TiO}_2$  increases to the left. As compaction increases,  $\text{TiO}_2$  decreases as a result of expulsion of residual liquid, which takes with it incompatible elements.



### Crystallization and Differentiation of the Holyoke Flow

The Holyoke basalt erupted from the Buttress dike, which must have tapped a mid-crustal ( $\sim 12$  km) magma reservoir (Fig. 1B), as indicated from the composition of the basalt (Yang et al. 1996). Flow in the dike must have been rapid and was probably turbulent in order to cause melting of the wall rocks (Philpotts, 1998). As a result, a large ponded flow formed in the Hartford Basin, reaching a maximum thickness of 200 m in the vicinity of the North Branford quarry, where it ponded against the eastern border fault of the basin. This huge lava lake extended as far north as the Deerfield Basin in Massachusetts and as far west as the Pomperaug Basin in Connecticut. The presence of one zone of vesicles at the top of the flow and the presence of only one entablature and colonnade indicates that it solidified as a single cooling unit.

The lava erupted across alluvial mud flats and, in the Deerfield Basin, into a lake. Where the surface of the flow is visible, it suffered no erosion before being covered by fluvial sediments. The surface of the flow was therefore probably flooded with water soon after the eruption and this would have aided in developing quench textures in the crust of the flow down to 80 m beneath the surface at the North Branford quarry. Despite the rapid

quenching from above, the boundary between the entablature and colonnade occurs at 60% of the height of the flow. Material that crystallized near the roof of the sheet of magma must have been transferred to the floor in order to account for the upward displacement of the final temperature maximum. This material would have sunk in dense plumes of crystal mush, with detachment from the roof being facilitated by the presence of alternating millimeter-scale layers of crystals and residual liquid (Philpotts and Dickson 2002). This layering shows that the cusped shape of the entablature-colonnade boundary is actually the final shape of the roof of the magma sheet. Dense plumes sinking from this roof would most likely have sunk from the bottoms of each cusp.

The dense plumes would have distributed crystals throughout the flow, ensuring that the basalt would eventually have a fine grain size, despite the great thickness of the flow. Once the abundance of crystals reached approximately 30% they formed an interconnected network (Philpotts, et al. 1998), and because the bulk density of the solids was greater than that of the residual liquid the crystal mush could undergo compaction. Because the mush was initially highly porous (Philpotts and Carroll, 1996), compaction took place rapidly at first, but once the crystallinity exceeded 40%, the permeability was reduced, the mush became stronger, and compaction ceased. The maximum amount of compaction, as measured from the anisotropy of the basalt's texture and its content of incompatible elements was ~12%.

Compaction of the crystal mush in the lower part of the flow caused residual liquid to be expelled upward. The liquid accumulated near the center of the flow where it dilated and eventually ruptured the crystal mush. Horizontal fractures opened into which the residual liquid entered by porous flow from the underlying crystal mush. Because this liquid carried few crystal nuclei, it crystallized to a coarse-grained ferrodiorite. Toward the end of the crystallization of the ferrodiorite, the residual liquid would have encountered the liquid immiscibility field, which in the entablature of the flow produced small immiscible droplets of iron-rich liquid (quenched to pyroxene spheres) in a quartzo-feldspathic host. In the more slowly cooled segregation sheets in the center of the flow, the immiscible quartzo-feldspathic liquid was able to coalesce and rise toward the top of the sheets of ferrodiorite and form thin sheets of granophyre. The iron-rich liquid would have crystallized to form some of the iron-rich pyroxene, fayalite, magnetite, and apatite, which are abundant in the ferrodiorite.

The Holyoke flow provides a very simple example of magmatic differentiation. We know that the differentiation must have occurred in situ as the lava solidified on the surface of the earth. Unlike plutonic magma chambers, no mysterious sources of different magma can be invoked, and contamination from wall rocks is ruled out. The chemical profiles and quantitative measures of textural anisotropy show that differentiation resulted from compaction of crystal mush. If this process caused a thick flood-basalt flow to differentiate as it solidified on the surface of the earth, the process must be effective in plutonic bodies where solidification times are much greater.

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#### ROAD LOG

The excursion will begin at the entrance to the North Branford quarry. The quarry can be reached from New Haven by taking I-91 north and exiting onto Route 80 East at exit 8. From I-91 travel precisely 5 miles east where you will turn north onto Route 22, and then immediately take a right into the parking lot beside Tilcon's Weighing Station and Office (Fig. 1). We will assemble at 8:45 a.m. and consolidate into the smallest number of vehicles possible. We will then drive into the quarry. The exact route to be taken will depend on the operations in the quarry that day. Remember, this is an active quarry, and it is essential that everyone stay together in a convoy. Also, at no time may anyone *go nearer a quarry wall than a quarry wall is high*. Everything can be seen and sampled from a safe distance in the center of the quarry. Please do not abuse the kindness of our hosts, who have our safety in mind.

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## A NEW LOOK AT THE STRUCTURE AND STRATIGRAPHY OF THE EARLY MESOZOIC POMPERAUG BASIN, SOUTHWESTERN CONNECTICUT

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### INTRODUCTION

#### Why a new study of the Pomperaug basin?

The Pomperaug basin is a small (4 km x 13 km) early Mesozoic rift basin located in the highlands of western Connecticut, a little over 20 km west of the western margin of the much larger Hartford basin. On the Bedrock Geological Map of Connecticut (Rodgers, 1985) the highly faulted Triassic and Jurassic sedimentary and volcanic rocks stand out in stark contrast to the early Paleozoic metamorphic and plutonic rocks surrounding them. The basin is located within the Pomperaug River watershed and the townships of Southbury and Woodbury. Concerns about preservation of drinking-water quality and adequate water flow in the river have led to the formation of a local citizen's group, the Pomperaug River Watershed Coalition (PWRC). The PWRC has supported a study by the U.S. Geological Survey (USGS) in Woodbury, where municipal water-supply wells have levels of MTBE that exceed EPA standards, as well as the present mapping project. This new mapping of the Pomperaug basin, which comprises portions of four 7.5-minute quadrangles, was begun in the fall of 2003 at the request of Connecticut State Geologist Ralph Lewis (now retired), who identified the Pomperaug basin as one of the areas in the state most in need of new geologic mapping and interpretation.

For the past decade the basin has also been the focus of ongoing studies by Phillip Huber (Minnesota State Univ.) and Peter LeTourneau (Lamont-Doherty), with the goal of a detailed sedimentological and paleoenvironmental analysis of the sedimentary rocks. From this work, they have constructed a new lithostratigraphic column for the basin that differs considerably from the interpretation depicted on the Connecticut bedrock map, but not too much from that put together a century ago by Hobbs in his report on the basin (Hobbs, 1901). In addition, J. Gregory McHone (an independent geologist) performed a brief study of the basalts of the Pomperaug basin under contract with the State Geological and Natural History Survey. McHone (2003b) correlated flow units using both thin-sections and geochemical analysis of the basalts, the latter obtained in cooperation with this USGS mapping study.

The new bedrock geologic map of the Pomperaug basin (Fig. 1) employs a topographic base digitally assembled from portions of the four quadrangles, at a scale of 1:12,000. In addition to lithologic contacts the map contains data on bedding, joints, and faults within the basin, and both ductile and brittle structural data from the surrounding crystalline rocks out to a distance of 1 to 2 km from the basin margin. The mapping of the crystalline rocks was done in an effort to better understand the crustal response to extension during rifting and formation of the basin.

#### History of previous geologic work in the basin

The first observations on the geology of the Pomperaug basin were made by Silliman (1820) and Hitchcock (1828), both of whom noted the broad similarity of the succession of sedimentary and "trap" rocks with that of the Hartford basin. In 1842, Percival portrayed a generalized distribution of rock units within the basin on the first geologic map of Connecticut, and erected the first stratigraphic nomenclature for the Mesozoic rocks of Connecticut (Percival, 1842), which persisted until Krynine (1950). Percival (1842) made a number of accurate observations of the general geology of the basin, including the arrangement of traprock ridges and division of the main basalt into a lower, massive member and an upper amygdaloidal member. He was followed in the 1880's by William Morris Davis

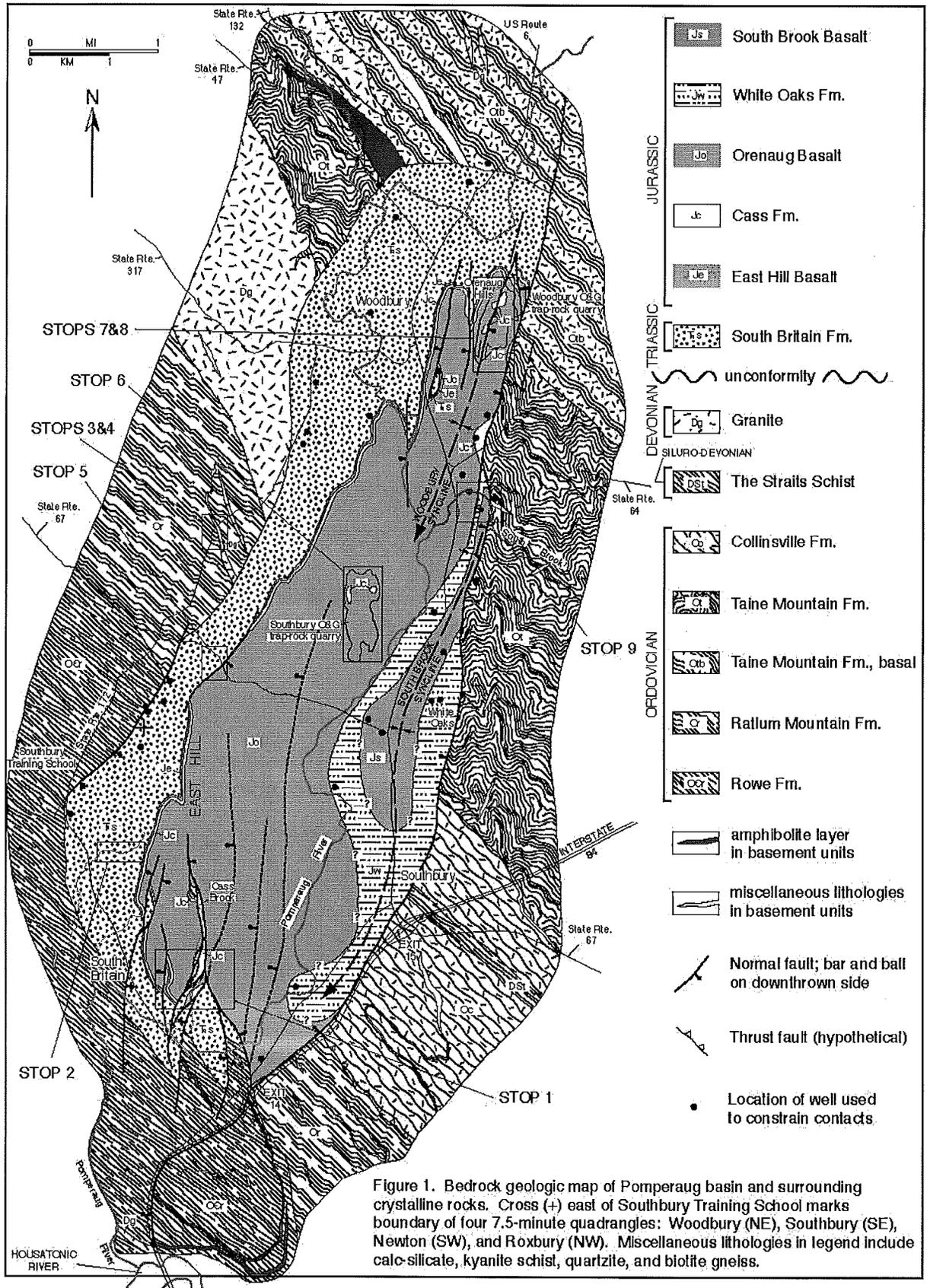


Figure 1. Bedrock geologic map of Pomperaug basin and surrounding crystalline rocks. Cross (+) east of Southbury Training School marks boundary of four 7.5-minute quadrangles: Woodbury (NE), Southbury (SE), Newton (SW), and Roxbury (NW). Miscellaneous lithologies in legend include calc-silicate, kyanite schist, quartzite, and biotite gneiss.

(1888), who concentrated on two areas of the Pomperaug basin near the south and north ends, respectively, of the basin: the hill and river exposures in and east of the village of South Britain, and the Orenaug Hills in the town of Woodbury. In South Britain, Davis mapped a relatively thick lower sandstone unit grading up to coarse arkose and conglomerate, overlain by a thin amygdaloidal basalt, a thin shale interval, and a thick basalt. He was the first to recognize that multiple traprock ridges in both the South Britain and Woodbury areas were the product of north-trending, steeply-dipping normal faults repeating the section, and that the Pomperaug basin was an east-tilted half-graben. Davis used his new block-faulting model in later studies of the Hartford and other early Mesozoic basins.

Around this time an oil exploration well was drilled in the eastern part of the basin, as reported in an article in *Scientific American* by E.O. Hovey (1890). The well went through two trap sheets separated by shale, and the total thickness of the Mesozoic section was determined to be 1,235 feet. In 1892, I. C. Russell included a generalized map of both the Hartford and Pomperaug basins in his *Correlations of the Newark System* (Russell, 1892), along with a synthesis of the "Broad Terrane" basin model. This concept envisioned all the various Newark Supergroup basins as erosional remnants of an originally expansive depositional basin, extending from Nova Scotia to Alabama, along the coast regions and as far inland as western Virginia.

William Herbert Hobbs, a geologist with the USGS, was the first to map the entire Pomperaug basin and study its geology in detail, publishing his results in the 21<sup>st</sup> Annual Report of the United States Geological Survey (Hobbs, 1901). Following on the work of Davis (1888) and the oil well report of Hovey (1890), he recognized a lower, thin, amygdaloidal "anterior basalt" and an upper, thicker "posterior basalt," separated by a thin interval of "anterior shale" that contained a limestone horizon. His report contains detailed descriptions of exposures and petrographic descriptions of the lithologies. Along with his own observations he recorded information from others who had first-hand knowledge of the geology of the area, including a local minister who had collected fish fossils from the shale, farmers who had incorporated into their stone fences pieces of fossil wood from near the base of the lower sandstone, and witnesses to the drilling of the oil exploration well—all information that would have been unrecoverable later. Hobbs reported a "posterior shale" above the upper basalt, based on the existence of an outcrop of black shale near the oil well, now gone, that was later represented by a bedding strike and dip symbol on the *Bedrock Geological Map of Connecticut* (Rodgers, 1985).

Hobbs had what we might term an extremist point of view on the role of brittle faults in the deformation of the basin. Every deviation in strike in the sedimentary rocks, every interruption of a linear bedrock ridge, and every planar outcrop face, however small, was justification for Hobbs to map a linear fault trace, every one of which extended without interruption straight across the entire basin. A number of these faults were also associated with linear arrangements of springs. Hobbs' map depicting the fault system of the Pomperaug Valley has, by his own count, over 250 faults, including faults along the entire western margin of the basin (Hobbs, 1901; Fig. 2). This complex fault network was formed, according to Hobbs, by *compressional* stresses acting in a WNW-ESE direction, following which the weakened crust subsided under its own weight (Hobbs, 1901). Today we find Hobbs' model untenable, of course—not only because the forces responsible were extensional, not compressional, but also because such a proposed fault network, with perfectly linear, closely-spaced fault traces without any branching or consolidation, is unrealistic in light of what we now know about how rocks deform. To Hobbs' credit, however, he backs up his reasoning with extensive discussion and illustrations in his 1901 report. Hobbs' tectonic interpretations led to a disagreement between him and George Otis Smith, then Geologist in charge of Geology of the United States Geological Survey, and ultimately to Hobbs' resignation from the Survey in 1906.

Between Hobbs' 1901 publication and the state-wide geologic quadrangle mapping that started in the 1950's, the only published bedrock data on the Pomperaug basin was an outcrop map by Meinzer and Sterns (1929) of the Pomperaug River basin, as part of their regional study of ground water sources. In 1956, Donald Schutz produced an unpublished bedrock map of the Pomperaug basin for his senior thesis at Yale. Although his map shows considerable detail, there is little documentation of his observations. Schutz identified a third basalt flow near the eastern border fault, and northwest-trending faults extending from the crystalline uplands across the western part of the basin; both of these features were later shown on the *Bedrock Geological Map of Connecticut* (Rodgers, 1959, 1985). In 1954, the state published Robert Gates' 1:24,000-scale bedrock geologic map of the Woodbury quadrangle, which includes the northern end of the Pomperaug basin, as part of its statewide mapping program, but his crude depiction of the early Mesozoic rocks reflected his relative lack of interest in them (Gates, 1954). The rest of the basin was mapped at 1:24,000 by Gates (1959) for his Roxbury quadrangle map, by Scott (1974) as part of the

Southbury quadrangle, and by Stanley and Caldwell (1976) for the Newtown quadrangle. Scott's map of the southeast portion of the basin is very detailed, showing three basalts within a stratigraphic section that is cut by numerous northeast-trending faults, and truncated by an east-northeast-trending fault marking the southern end of the basin. These map features were also incorporated into the newer Bedrock Geological Map of Connecticut (Rodgers, 1985).

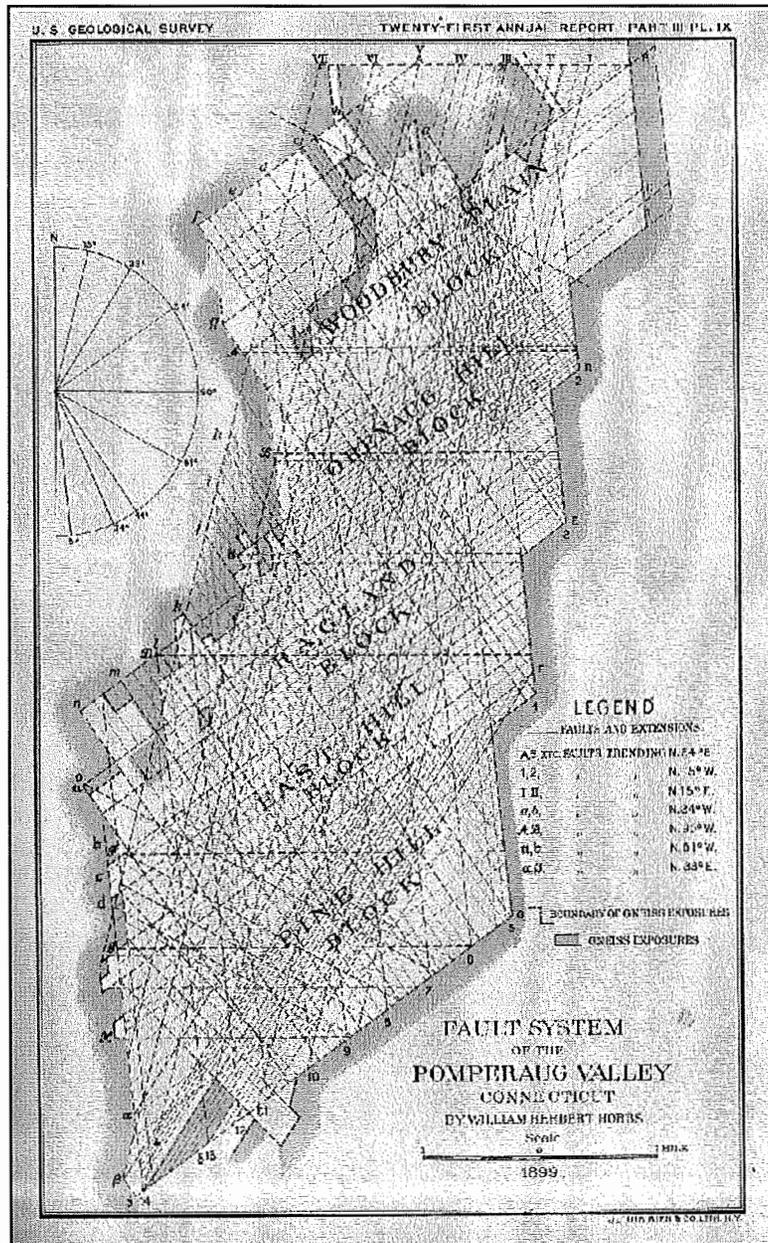


Figure 2. Hobbs' model of faulting in the Pomperaug basin, as published in the 21st Annual Report of the U.S. Geological Survey (1901).

Hubert et al. (1978) and Weddle and Hubert (1983) were the first geologists to incorporate both petrologic data and interpretations of sedimentary structures (grain lineation, cross bedding, facies analysis) into a depositional model and paleogeographic reconstruction that portrayed Hartford and Pomperaug basin stratigraphic relationships. These authors built upon an earlier modified "Broad Terrane model" advocated by Krynine (1950), who was the

first to engage in comparative petrology of the sedimentary rocks contained in both of the basins. These concepts were reviewed and discussed further by Lorenz (1987). Cornet (1977) recovered palynomorphs and megafossil plants from both basins, and McDonald (1982), Olsen et al. (1982) and Olsen (1984) collected fossil plants and fishes and discussed aspects of Pomperaug basin stratigraphy and paleogeography. Tolley (1985) suggested a stratigraphy that largely mirrored that of the Hartford basin, while Huber and McDonald (1992) provided a revised stratigraphic framework and fossil distribution for the Pomperaug basin that closely resembled Hobbs' (1901) original interpretation. LeTourneau and Huber (1997, in review) described a basin-wide eolian sand sheet occurring just below the main, ridge-forming basalt in the basin, and several papers on Hartford basin basalts (i.e. Hurtubise and Puffer, 1983; Philpotts et al., 1996) also included geochemical analyses of this flow unit.

### **Constructing a new geologic map of the basin**

New mapping of the Pomperaug basin was begun in October, 2003 and completed in April, 2005 (Fig. 1). Following the advice of LeTourneau, Huber, and McHone, Burton first visited areas of good exposure such as the Platt Farm Preserve, Cass Brook, and Red Spring, the east flank of East Hill, the O&G trap rock quarries in Woodbury and Southbury, and South Brook. Along with the use of traditional mapping tools, station and outcrop data were recorded on magnified raster images of the 1:24,000-scale topographic bases using a pocket PC with a GPS running ArcPad, and structural data were recorded on a palmtop device using Pendragon Forms. This was especially important for mapping areas such as Cass Brook (Stop 1), where closely-spaced faults juxtapose different formations over distances of a few meters to tens of meters. Thanks to the lithostratigraphy previously established by Huber and McDonald (1992) and LeTourneau and Huber (1997; in review), tracing faults and assigning senses of offset was a relatively straightforward exercise in this area. Another critical area was the trap rock quarry in Woodbury (Stop 7), which exposed a fault with a clearly demonstrable east-side-down and left-lateral sense of offset that became a model for other, less well-exposed faults in the basin. Recovery of two of Hobbs' (1901) arkose outcrops in the Orenaug Hills, and a new exposure of a basalt/sedimentary rock contact in the Woodbury quarry (Stop 8), helped outline a hanging-wall syncline near the north end of the basin. Careful examination of the exposures in South Brook (Stop 9) led to the rediscovery of a conglomerate/basalt contact first noted by Schutz (1956), and new geochemical analysis of the basalt supports the existence of a third basalt (Hampden equivalent), as shown on the Bedrock Geological Map of Connecticut (Rodgers, 1985). Many of the basalt outcrops in the basin were visited in order to obtain information on jointing. Domestic well records were compiled by Claudia Tamayo of the USGS Water Science Center in East Hartford, and lithologic logs from these records were critical in helping to constrain contacts in poorly-exposed areas of the basin.

In addition to the mapping of the basin, the metamorphic basement rocks surrounding the basin were mapped out to a distance of 1-2 km. The primary reason for this was to gauge the response of these rocks to faulting and crustal extension, to check the trace of the eastern border fault, and to help confirm or deny the existence of northwest- and east-northeast-trending faults that are shown on the state geologic map (Rodgers, 1985). This necessarily involved detailed lithologic mapping and an attempt to reconcile field descriptions of rock type with the formations shown on the state map. The new crystalline-rock mapping helped disprove the existence of most of the Mesozoic faults shown cutting the basement, retraced the eastern border fault along its southern extension, and yielded a broader perspective on Mesozoic crustal extension in the area. In addition, study of the pre-Mesozoic ductile structures in the crystalline rocks has spawned new ideas about the Paleozoic tectonic evolution of these rocks.

## **LITHOSTRATIGRAPHIC FRAMEWORK OF THE POMPERAUG BASIN**

Phillip Huber and Peter M. LeTourneau

### **Introduction**

The Pomperaug basin preserves at least ~400 m of strata and intercalated basalts that, with one exception, have been traditionally assigned the same unit names as the broadly coeval rocks of the Hartford Group that fill the nearby Hartford basin. Regardless of the persistent controversy regarding the depositional relationship of these rocks with those of the Hartford basin, Pomperaug basin strata and basalts are lithostratigraphically distinct from their Hartford basin correlatives, and merit a nomenclature of their own. We recognize five formation-rank units in the Pomperaug basin (Fig. 3) and note the possibility that the basin-fill sequence might preserve one or more

younger units whose outcrop areas, if the units exist, are obscured by thick glacial and colluvial cover and/or complicated intrabasinal structure. These strata and basalts are referred by us to the Pomperaug Group. Note, for reasons too numerous to mention here, the chronostratigraphic-based Group-rank scheme suggested by Weems and Olsen (1997) for strata contained by the collective basins of the Newark Supergroup, is not used here. However, our usage of new lithostratigraphic names should be considered informal until the criterion of publication in a widely distributed, peer-reviewed format is achieved (LeTourneau and Huber, in review). The purpose of clarifying and revising the stratigraphy of the Pomperaug basin is to establish the correct number, sequence, and map pattern of rock units used for paleogeographic and paleoclimatic analysis. With the exception of Hobbs (1901), previous work in the basin showed little agreement between the actual stratigraphy and the location of the sedimentary and volcanic rocks that fill the basin (e. g. see maps by Scott, 1974; Rodgers, 1985).

The Pomperaug Group consists of at least five formation-rank units which are, in ascending order: South Britain Formation (~250 m) which consists of the lower, Pierce Hollow (~200 m) and upper, Rattlesnake Hill (~50 m) Members; East Hill Basalt (7-10 m); Cass Formation (35 m); Orenaug Basalt (~80 m); and White Oaks Formation (60m+?) (Fig. 3). Recent mapping and geochemical work supports the presence of a third basalt flow, herein referred to as the South Brook Basalt. Details concerning the justification for new rock unit names, and the designation and description of appropriate type and reference sections, will be published by LeTourneau and Huber (in review).

### South Britain Formation

The South Britain Formation is composed of siltstone and fine to coarse-grained arkosic sandstone and pebble conglomerate that represent the oldest strata deposited in the Pomperaug basin (Upper Triassic, Norian-Rhaetian age). These strata were originally named South Britain Conglomerate by Hobbs (1901), while Krynine (1950) and subsequent workers assigned these rocks to the New Haven Arkose (e. g. Rodgers et al., 1959; Scott, 1974 (in part); Hubert et al., 1978; Weddle and Hubert, 1983; Rodgers, 1985). The South Britain Formation is approximately 250 meters thick, and at least the uppermost ~170 m are well exposed in composite at several localities. Percival (1842) discovered a locality near South Britain Village where he found "a mass of sandstone [reposed] on the primary slate," but neither we nor Hobbs (1901) could locate that outcrop. The basal ~80 m of the formation, including the unconformable and/or fault contact with crystalline rocks have been penetrated by at least two wells (Hovey, 1890; Janet Stone, pers. comm., 1996), and the contact of these strata with basement rocks at the western margin of the basin can be located with reasonable accuracy based on outcrops, topography, and well data. The top of the formation is defined by its contact with the overlying East Hill Basalt. The lower ~200 m of these strata are assigned to the Pierce Hollow Member, while the upper ~30-50 m are called the Rattlesnake Member (Fig. 3).

The lower ~80 m of the Pierce Hollow Member are not exposed. However, Hobbs (1901) excavated small areas immediately south of the preserved basin margin, along the eastern slope of what was then called Horse Hill (there are two hills near South Britain called Horse Hill, though only one of these is named on recent editions of the Southbury 7.5' quadrangle. The named feature is *not* the location of Hobbs' observations nor the location of his petrified wood locality). He found small pockets of sediment preserved within small depressions of crystalline basement, and the topographic surface represents the exhumed basin floor. This is significant as it indicates strata extended over a small area south and beyond the basin's present limits, and beyond the inferred northeast-trending "Pomperaug fault" of Scott (1974). Strata comprising the next ~50 m of the Pierce Hollow Member are well exposed at several localities. These beds consist of 1-5 m thick sequences of red-brown channel sandstones and overbank siltstones. The channels are broadly lenticular, ripple-laminated, trough crossbedded or massive sandstone bodies that usually have a thin, extrabasinal pebble or intraclast zone at their base. The immature sandstones plot well within the compositional field for arkose (Weddle and Hubert, 1983). Floodplain deposits are represented by relatively thin (1 m or less), structureless to disrupted bedded/pedoturbated siltstone. Excellent exposures of the Pierce Hollow Member are found along the Pomperaug River in South Britain; these outcrops were described by Schutz (1956), Hubert and others (1978), Weddle and Hubert (1983), and Lorenz (1987).

The next ~75 m of strata display features illustrative of classic, fining upward point bar sequences, and consist of thin pebble conglomerates (~1 m or less thick) overlain by as much as 18 m of massive, ripple cross-laminated and/or bedding disrupted siltstone. The basal conglomeritic units are noteworthy for containing abundant carbonate pebbles, some of which we believe were derived from the Stockbridge Marble, a unit that is restricted in outcrop to areas located some 12 km or more to the west and northwest. Most of the siltstone beds have disrupted bedding and

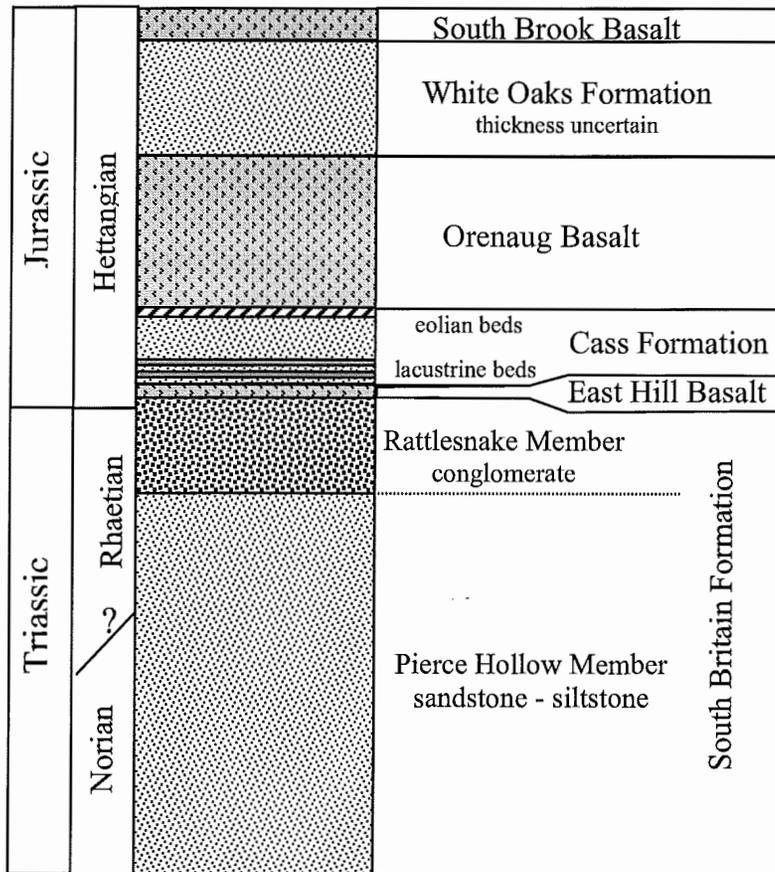


Figure 3. Stratigraphic nomenclature of the Pomperaug basin. Thickness of the White Oaks Formation and the South Brook Basalt uncertain.

display extensive, subvertical to horizontal burrowing and rhizolith structures. Further evidences of pedogenic modification are indicated by the occurrence of discrete intervals of greenish-gray mottled reduction zones and both in-situ and reworked caliche nodules (Hubert et al., 1978; Weddle and Hubert, 1983). Some horizons within these siltstones preserve primary bedding structures of both horizontal and ripple cross-laminations. Where these beds are present, they display abundant, small horizontal burrows and trails on bedding surfaces, indicative of a moderately diverse invertebrate fauna.

The upper 30-50 m of the South Britain Formation is dominated by arkosic, trough crossbedded, pebble conglomerate belonging to the Rattlesnake Member. Exposures of this unit are present across the western and southwestern margin of the basin, and at most outcrops, paleocurrents indicate derivation of sediment from source areas located west and northwest of the basin. In contrast to the lower member of the South Britain Formation, which is dominated by features indicative of well-organized stream and river systems with well-defined channels and floodplains, the upper member was deposited in high-energy braided streams or low-angle alluvial fans. The contact between the upper and lower members appears unconformable; therefore the shift from low- to high-energy fluvial environments may indicate changes in the tectonic rather than the climatic regime. The Rattlesnake Member of the South Britain Formation may be observed at a number of localities along the southwestern and western side of the basin, including outcrops described and figured by Davis (1888) at Platt Farm Park, exposures above the church parking lot in South Britain which comprise the unit's stratotype (LeTourneau and Huber, in review), and Hobbs' (1901) O. Mitchell brook section. The Triassic-Jurassic boundary is inferred to be located at, or just below the contact with the overlying East Hill Basalt.

### East Hill Basalt

The East Hill Basalt is the stratigraphically-lowest extrusive unit in the Pomperaug basin, and is named for exposures that occur near the eastern base of East Hill, just northwest of Hobbs' (1901) Red Spring locality (for details, see LeTourneau and Huber, in review). Previous names for this unit include "Anterior Basalt" (Percival, 1842; Davis, 1888; Hobbs, 1901; Krynine, 1950; Rodgers and others, 1959), "Talcott Basalt" (Hubert and others, 1978; Weddle and Hubert, 1983; Rodgers, 1985; Tolly, 1985), and "Talcott Formation" (Scott, 1974). Hobbs (1901) accurately mapped the distribution of most known outcrops of the East Hill Basalt, while Rodgers missed several critical outcrop areas on the Bedrock Geological Map of Connecticut (Rodgers, 1985). Scott (1974) mapped the entire outcrop area (Red Spring locality) that includes the 10 m-thick East Hill Basalt stratotype as "New Haven Arkose," though only 1 m or less of South Britain Formation actually crops out in the area, and no indication of this basalt unit (or the overlying Cass Formation) was included on his map.

At its type section, the East Hill Basalt consists of a 10 m-thick vesicular basalt that directly overlies the Rattlesnake Member of the South Britain Formation and is capped by basal siltstones of the overlying Cass Formation. The East Hill Basalt is a high-TiO<sub>2</sub> quartz tholeiite that is correlated with the Talcott Basalt of the Hartford basin, Orange Mountain Basalt of the Newark Basin and North Mountain Basalt of the Fundy basin on the basis of both stratigraphic position and geochemistry (McHone, 2003c). Three other complete sections of the East Hill Basalt are known in the South Britain area: (1) along Cass Brook in Platt Farm Park, just below the type section of the Cass Formation (discussed below); (2) at the southwestern margin of Platt Farm Park (Spring House locality of Hobbs (1901); and at O. Mitchell Brook, in contact with, and above, the section designated by LeTourneau and Huber (in review) as the lectostratotype of the South Britain Formation. Other scattered outcrops are common in the southwest area of the basin, and include those on Pine Hill originally mapped by Davis (1888), at Red Spring, and the east side of Rattlesnake Hill (Hobbs, 1901). A discussion of this basalt's correlation with other basin basalts follows below.

### Cass Formation

The Cass Formation overlies the East Hill Basalt, and is named for outcrops along lower Cass Brook at Platt Farm Park, Southbury, Connecticut. The Cass Formation is 35-40 m thick and composed of: 1) black and buff, laminated to massive micrite, and black and gray, laminated to microlaminated calcareous shale; 2) gray and red, laminated, ripple cross-laminated and massive to nodular siltstone; and 3) red-brown and buff, fine- to coarse-grained, moderately to poorly sorted, arkosic, litharenitic and quartzose sandstone and pebble-cobble conglomerate. Previous names used to describe Cass Formation strata include "Anterior Shale (Percival, 1842; Davis, 1888; Hobbs, 1901; Longwell, 1933; Krynine, 1950); Shuttle Meadow Formation (Rodgers et al., 1959; Hubert et al., 1978; Weddle and Hubert, 1983; Tolley, 1985; Rodgers, 1985), and Talcott Formation (Scott, 1974). The Cass Formation is basal Jurassic (lowermost Hettangian) in age and, based on the composition of its palynofloral and fossil fish assemblages, belongs to the Wassonian Land Vertebrate Faunachron of Lucas and Huber (2003). The Cass Formation thus correlates with the Shuttle Meadow Formation of the Hartford basin, the Feltville Formation of the Newark Basin, the Midland Formation of the Culpeper Basin, and the Scotts Bay and McCoy Brook Formations of the Fundy basin.

The Cass Formation stratotype comprises two outcrops that, combined, represent the lower 14 m of the formation. The base of Section "A" is located along the small, easterly bend of Cass Brook just below the footbridge, 150 m north of the Platt farmhouse (see trip log of Stop 1 for location). Section A consists of 4.8 m of strata that begin with 2 m of red, ripple cross-laminated and evenly laminated siltstone directly overlying East Hill Basalt. These beds are gradational with 2.8 m of overlying gray, laminated siltstone that passes upward into dark gray to black, finely laminated shale and laminated limestone. Immediately overlying beds are covered, but low, poorly exposed outcrops of gray siltstone occur in the bed of the stream several meters east, where Cass Brook bends to flow south. Cornet (1977) recovered a moderately diverse, *Corollina*-dominated palynoflora from the dark shales and siltstones, and these strata also contain a low diversity megafloora of *Brachyphyllum* conifer foliage and less common cycad (cf. *Otozamites* sp.) and equisetalian fragments. The black shale and limestone unit contains moderately abundant fossil fish that occur as isolated scale and dermal elements to fully-articulated specimens of *Redfieldius gracilis* and *Semiontous* sp. The limestone unit is notable for its bitumen-coated fractures (Fig. 4), and will literally bleed bitumen within several minutes upon a fresh break with a hammer. The unit certainly is a mature source rock, and perhaps provided the impetus for the failed oil well venture reported by Hovey (1890).

Section B of the Cass Formation stratotype is located ~300 m upstream from section "A" and begins with gray siltstone of the same lacustrine beds exposed at Section A that are here faulted against the Orenaug Basalt. The beds are tilted vertical at the fault contact, but rapidly splay toward horizontal within several meters laterally, where they assume strike and dip values within the regional average (strike: N25E, dip 25SE). The gray siltstone is gradational with overlying red, ripple cross laminated siltstone that is largely structureless for the next 7 m of exposure along the stream bed. The upper part of Section B is a small stream bank outcrop of red siltstone comprising a 2 m-high vertical face. Near the top of this outcrop are gray limestone nodules within a reddish-purple siltstone that occur 14 m above the base of the East Hill Basalt. Scattered float blocks suggest the presence of a bedded to massive limestone higher in the stream bank. If present, this bed would correlate with the Red Spring Limestone bed discussed below.

The Red Spring locality of Hobbs (1901) shows sparse outcrops that represent the lower 3 m of the formation, and an additional 1 m-thick interval located 15 m above the East Hill Basalt adjacent to its type section. Abundant float blocks of the laminated micrite unit are scattered throughout the area, and are likely evidence of the trenching efforts described by Hobbs (1901) in his report. At 15 m above the East Hill Basalt, a 0.5 m-thick, massive, white to buff-colored limestone is poorly exposed, and bracketed by gray and red shale. This is the limestone mentioned by Hitchcock (1828) and Hobbs (1901) that was quarried for water lime early in the 19<sup>th</sup> century. Hobbs (1901) noted the presence of fossil fishes from this limestone, and more recent collecting efforts indicate isolated squamation and dermal elements to articulated specimens of *Redfieldius* sp. and *Semionotus* sp. are moderately abundant. When dissolved in 5% acetic acid for a two week period, the limestone readily dissolves. A 2 kg sample was processed and picked for fossils which revealed abundant piscine elements as well as small, ovoid scales that might represent lacertalian reptiles (N. Fraser, pers. comm. 1997). The Red Spring section is important for its paleontological potential, and also because the entire thickness of the Cass Formation can be measured in one location with reasonable accuracy. The contact with the overlying Orenaug Basalt is concealed, but is estimated to occur 35 m above the East Hill Basalt-Cass Formation contact.

Outcrops of the middle Cass Formation are sparse and consist of small, isolated exposures in Platt Farm Park and a more lengthy, albeit overgrown, cut along a gravel fire road that lies south of and parallel to the O. Mitchell Brook section of Hobbs (1901). This location exposes 10 m of section that include a 1 m-thick gray lacustrine siltstone and shale interval that occurs some 20 m above the base of the formation. The top of the section is defined by a poorly exposed, arkosic sandstone, above which is apparently a trough crossbedded quartzose sandstone, based on loose blocks that have weathered loose from the highly vegetated slope.

The upper Cass Formation was formerly well exposed at both the Southbury and Woodbury O & G Industries quarries, and the former location displayed an 11 m-thick section dominated by diverse coarse clastics that included mid- and distal alluvial fan facies containing abundant, natural sandstone molds and casts of dinosaur and other tetrapod bones, a well-defined paleosol horizon, a pebble conglomerate that includes abundant extrabasinal clasts derived from the Stockbridge Marble, a dinosaur track-bearing horizon, and a 2.5 m-thick eolian sandsheet. A synthesis and discussion of these outcrops is included with the trip description for Stop 6).

### Orenaug Basalt

The Orenaug Basalt is herein named for the 80+ m thick basalt that is the main ridge-forming unit in the basin. The stratotype of the Orenaug Basalt is located just southeast of Woodbury Village in a town-maintained park, and consists of 10 m of cliff exposures that overlook a small pond (see LeTourneau and Huber, in review). The Orenaug basalt is a quartz tholeiite, and it has been attributed several names in the older literature including Davis (1888), Hobbs (1901), Rodgers et al. (1959), "Holyoke Basalt" (Scott, 1974; Rodgers, 1985) and "Main Basalt" (Huber and McDonald, 1992).

According to the Southbury quarry exposure, Orenaug Basalt contains at least two, and possibly three flow units, but attempts to map these individual units in the rest of the Pomperaug basin have been unsuccessful. The top of one flow is well defined by a vesicular texture and weathered horizon, and the undulating base of the succeeding flow displays pipe vesicles and at least 2 m of relief. This horizon occurs in the Southbury quarry at approximately 40 m above the Cass Formation-Orenaug Basalt contact. A possible, stratigraphically-lower flow contact occurs about 20 m above the Cass-Orenaug contact where a laterally-continuous, 0.1 m-thick zone of highly fractured

basalt and clayey mush delimits a change in lithology from black massive, highly compacted basalt with abundant silica minerals (clear quartz, amethyst, banded agate) to a dark gray, columnar basalt that contains abundant cavities filled with zeolite minerals. However, there is much less of a textural contrast at this boundary than at the overlying flow boundary. (See also discussion in section below on internal basalt stratigraphy).

### **White Oaks Formation**

The name White Oaks Formation is given for the sequence of largely covered strata that occurs above the Orenaug Basalt and below the possible third basalt unit in the Pomperaug basin, the South Brook Basalt (see below). The only currently known exposures are small ledges of conglomerate and pebbly arkose in South Brook (Stop 9), which comprise the north end of the inferred belt of White Oaks Formation (Fig. 1). The rest of the inferred belt is almost entirely covered by glacial debris, natural vegetation, and artifacts of human development such as golf courses and strip malls, and is based largely on structural considerations and interpretation of well logs (Fig. 1). Hobbs (1901) confirmed the existence of at least one lacustrine black shale bed that was located near the failed oil well venture (see well location near Pomperaug River east of East Hill in Figure 1), and Hovey (1890) documented that this drilling enterprise encountered an unknown thickness of strata before penetrating the upper of two basalt units. The White Oaks Formation is correlative with the East Berlin Formation of the Hartford basin, Turners Falls Formation of the Deerfield basin, and Towaco Formation of the Newark basin, based on its stratigraphic position between the "second" and "third" basalt flows.

We have collected large (0.5 m-thick) float blocks of organic- and carbonate-rich laminated black shale from a gravel pit located along Main Street South in Southbury that do not resemble any of the three lacustrine black shales of the Cass Formation, and therefore, are attributed to the White Oaks Formation. Blocks of a similar shale were also collected by Paul Olsen (pers. comm., 1992) in the 1970s during the construction of the Southbury Plaza shopping center. The thickness and lithologies contained by the White Oaks Formation are otherwise unknown, but a conservative minimal estimate of 50 m thick seems reasonable, based on the distribution of stratigraphically-adjacent basalts in combination with topography.

### **South Brook Basalt**

Immediately above and in contact with the arkose and conglomerate in South Brook is a highly altered, vesicular basalt that superficially resembles the East Hill Basalt (Stop 9). The basalt occupies the core of a hanging-wall syncline next to the eastern border fault (Fig. 1, Stop 9), and its exposed thickness, about 10 meters or so, is permissible for the East Hill Basalt. However, the two samples that have been analyzed from this site have a chemistry that, despite alteration of the samples, is distinct from the East Hill and Orenaug Basalts and their Hartford and Newark basin equivalents, as discussed below. We therefore infer the existence of a third basalt, possibly equivalent to the Hampden (Hartford basin) and Hook Mountain (Newark basin) Basalts, on the basis of both the chemistry and the position of this basalt at the east margin of an east-tilted half-graben, where a stratigraphically higher flow would be expected. The area underlain by this basalt is inferred to extend southward on the basis of well records and the fact that the underlying White Oaks Formation likely also extends a considerable distance southward, as discussed above (Fig. 1).

## **POMPERAUG BASIN BASALTS**

J. Gregory McHone

A major lithostratigraphic aspect of the Early Mesozoic basins of northeastern North America is the presence, age, and position of basaltic lava flows. After many years of uncertainty and confusion, the absolute ages of basalts, sills, and dikes within and surrounding the basins are now reasonably well established to be between 200 and 201 Ma, and they formed during three major volcanic events spanning about 600,000 years (Olsen and others, 1996; McHone, 1996; Olsen et al., 1996; West and McHone, 1997). In addition, basalts within the larger basins are closely correlated by chemistry, paleomagnetism, and stratigraphy, and in essence must be co-magmatic flows (Puffer et al., 1981; Hozik, 1992; Puffer, 1992). As shown in Figure 4, there are three separate basalts (some containing several flow units) within the adjacent Newark (New Jersey) and Hartford (Connecticut) basins, and other basalts in the Culpeper (Virginia) and Fundy (Nova Scotia and New Brunswick) basins are also correlated (Olsen, 1997). Two similar basalts exist in Morocco, which prior to rifting was adjacent to eastern Canada. The

Fundy basin basalt is identical to the lowest basalt that exists in other basins. Moreover, the lowest basalt in each basin is only a few meters or less above the Triassic-Jurassic boundary (Fig. 3), thus serving as an important time-stratigraphic marker for that transition and event.

Regional relationships of basins, Connecticut dikes, and basalt correlations

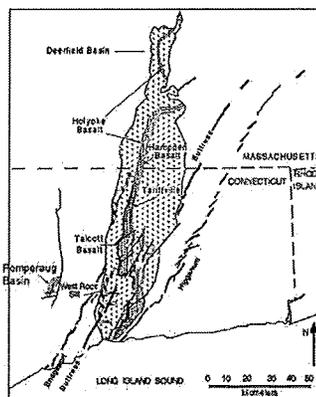
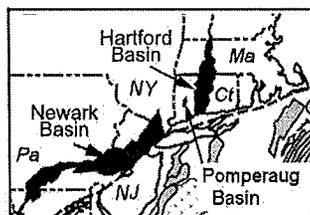
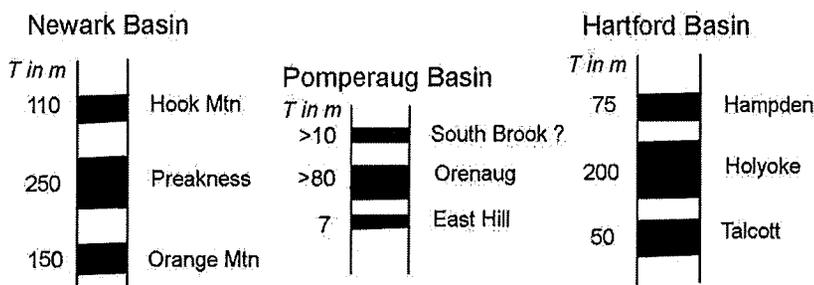


Figure 4. Lithostratigraphic correlation of basalts in the Pomperaug and adjacent basins, northeastern North America.



Despite his problems with faults and lineaments, William Hobbs (1901) was an able petrographer who made good descriptions of hand samples and thin sections of the basalts, and he listed two chemical analyses performed by the famous U.S.G.S. analytical chemist, W. F. Hillebrand. Both samples are from the Rattlesnake Hill area east of South Britain Village, but unfortunately for this and later analyses, the lower amygdaloidal basalt appears to be highly altered at all outcrops. Like Davis,

Hobbs found only two different basalt formations in the Pomperaug basin, but unlike Davis he believed that the lower, thin amygdaloidal flow did not continue more than a mile or so north of the South Britain Village along the west side of East Hill.

Hobbs examined many outcrops of the “main or posterior basalt” and concluded that it is subdivided into a lower, compact, and massive to columnar member and a higher, amygdaloidal, and weathered upper member, together “several hundred feet thick” (Hobbs, 1901, p. 45). The western belt of basalt outcrops forms the highest ridges and contains the lower more solid member, while an eastern belt of outcrops is made from the upper amygdaloidal, friable member.

An important student project was conducted by Donlon Hurtubise at Rutgers University, supervised by John Puffer (Hurtubise and Puffer, 1983). Hurtubise sampled several of the basalt outcrops mapped by Scott (1974) east of South Britain, and he followed the three-basalt system used by Scott. Chemical analyses of the lower amygdaloidal basalt, and of samples considered to be an upper amygdaloidal basalt, were not satisfactory because of intensive alteration. The original data and student report have been lost (Puffer, pers. comm. 2003) but an average of the best analyses, collected from the main “compact” or posterior basalt at Rattlesnake Hill near South Britain, is listed in their abstract (Hurtubise and Puffer, 1983). As they conclude, this basalt is a good match for the Holyoke Basalt of the Hartford basin as well as the Preakness (Second Watchung) Basalt of the Newark Basin.

Eastward from the church parking lot in South Britain (across from the general store), the west slope of Rattlesnake Hill has exposures that confirm the sequence from bottom to top of South Britain Formation, a thin amygdaloidal East Hill Basalt, gray-green to red Cass Brook Formation shale and siltstone, and thick, massive columnar Oreanaug Basalt. Philpotts and others (1996) measured a minimum thickness of the Oreanaug Basalt at Rattlesnake Hill of 57 m, including a colonnade and entablature, and they described new chemical analyses of the complete basalt section. The shale and lower amygdaloidal basalt are better exposed on-strike to the south between

Rattlesnake Hill and Sherman Hill, and especially along Cass Brook, and the same sequence occurs along the western slope of Pine Hill as well (Davis, 1888; Fig. 1).

Hobbs (1901) believed that the uppermost portion of the South Britain Formation is indurated, or hardened by mineralization related to heating from the overlying basalts, which may partially explain the resistant ridge of conglomeratic arkose exposed high up on the slope of Rattlesnake Hill. Outcrops in this area show northerly strikes and steep dips of 20° to 30° eastward. However, there are abrupt changes in the attitudes of basalts and sedimentary strata around Cass Brook, as described by Hobbs (1901) and Scott (1974). These are related to high angle faults with north-south to northeast-southwest trends and generally west-sides down (Fig. 1). In addition, strata dips become less steep (10° to 20°) toward the east to southeast. Several faults are exposed along Cass Brook, some associated with "reibungsbreccia," which as described by Hobbs (1901) as an unusual tectonic breccia formed by fault activity between solid basalt and unlithified shale and siltstone.

### Basalt Breccias

Four different types of breccias are associated with basalts within the Pomperaug basin. They have interesting differences and provide evidence concerning tectonic activity, much of which appears to be roughly contemporaneous with the emplacement of the basalts and sediments.

In the O&G Southbury Quarry, vertical clastic dikes of 40 to 60 cm width and composed of fine red sediment with inclusions of pebbles of quartz and sedimentary rocks (?) have been observed at several locations of interior quarry walls (Fig. 5 A). These may have been generated by explosive steam vents from wet sediments that were overrun by the Orenaug Basalt.

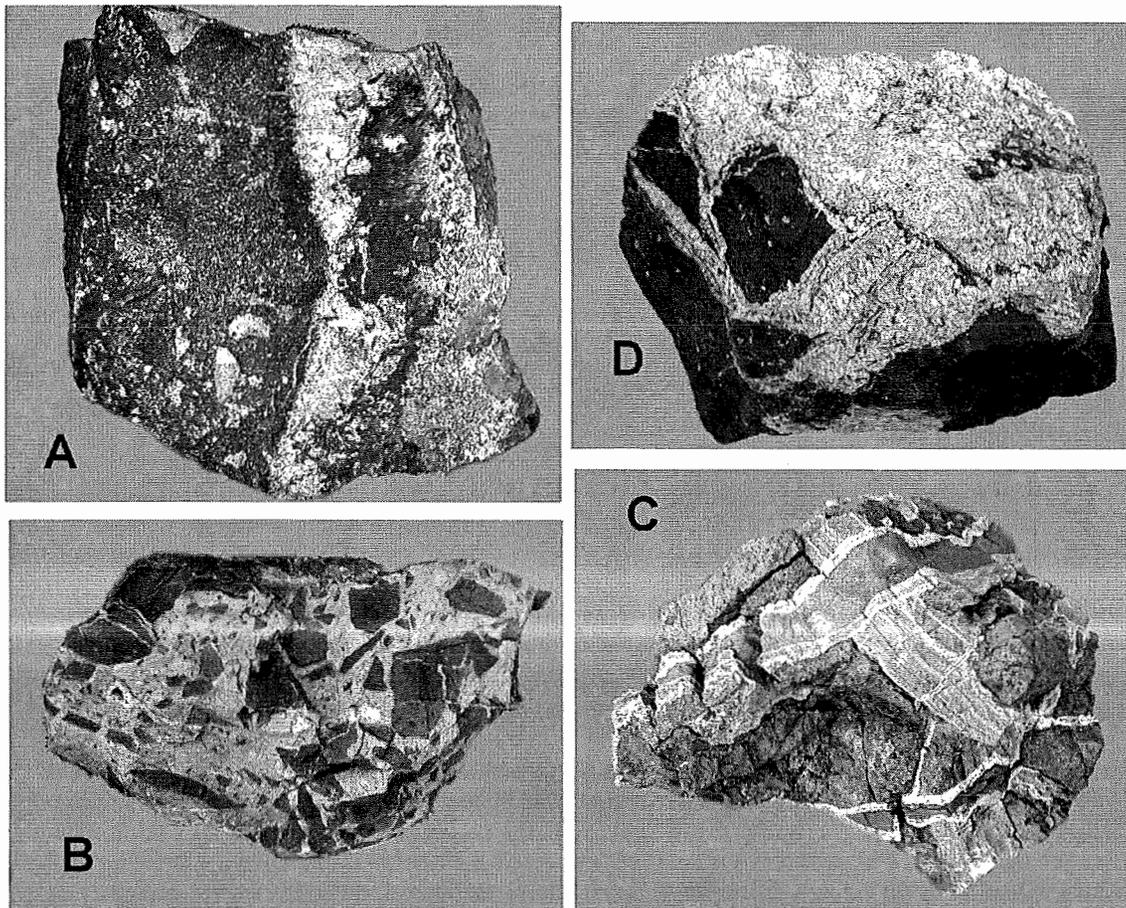


Figure 5. Photographs of four breccias associated with basalts in the Pomperaug basin.

In the O&G Park Road (Woodbury) Quarry, a vertical dike about 1 m wide of fine-grained light-gray "micritic" calcium carbonate with inclusions of basalt was observed in an E-W interior quarry wall (Fig. 5 B). The carbonate merges into calcite-lined vugs of several cm lengths adjacent to some of the angular basalt clasts.

"Reibungsbreccia" was considered by Hobbs (1901) to represent solid basalt faulted against unlithified mudstone (see also discussion at Stop 1). As can be seen in Figure 5 C, pink to gray sediment has intruded highly-weathered basalt in two or more generations of thin (1-2 cm) dikes. In all cases a thin border or rind of calcite has precipitated against the basalt. Although this is best observed at Cass Brook in the Platt Farm Preserve (Stop 1), similar breccias have been described from other localities within the basin.

Crystalline calcite surrounds basalt breccia in sub-vertical "veins" of 10 to 30 cm wide at the Southbury Quarry (Fig. 5 D). A few small pieces of dark red sedimentary (?) clasts also occur within the calcite matrix, implying a steam-driven origin like the red clastic dikes.

### Internal Basalt Stratigraphy

As expected in thick lava flows, the Orenaug Basalt has developed internal divisions or members with visibly distinct textures and colors. The most famous of these is the highly-altered prehnite-bearing member that has been a target for mineral collectors for more than a century, especially at the O&G Southbury Quarry (formerly known as the Silliman Quarry). The Southbury Quarry provides a continuous section through about 70 m of the lower part of the Orenaug Basalt. The basalts and sedimentary strata are tilted eastward about 15°. As can be seen in the images below (Fig. 6), its south-facing cuts expose boundaries that clearly define lower, middle, and upper members or flow units. The lower member is about 18 m thick and is relatively massive to columnar, except near major fractures and the basal contact zone. The middle member, about 25 m thick, is highly altered to a gray-green color and contains abundant amygdales filled by calcite, quartz, prehnite, pumpellyite, apophyllite, and other minerals described by Garabedian and others (1996). There are several lens-shaped sections in this member that might represent separate lobes or lava tongues. The upper basalt member contains finger-sized basal tube vesicles directly over the brown weathered upper surface of the middle member. This upper member is hard, dark gray, non-vesicular and massive to columnar, although it displays pyrite roses on fracture surfaces.

The base of the upper (third) member displays proof in the form of pipe vesicles that it was a separate or second Orenaug lava flow. The other boundary, between the lower and middle members, is relatively planar, although it dips with the strata to the east, and is brown-stained (Fig. 6). A possible interpretation is that this represents the boundary between a lower colonnade and entablature. It is not clear if another entablature exists over the third member as well, as the top of the Orenaug Basalt is not clearly exposed.

These Southbury Quarry observations essentially confirm the previous descriptions by Percival, Davis, and Hobbs concerning several petrographic units of the "Posterior" or Orenaug Basalt, in particular an amygdaloidal prehnite-bearing member above a massive columnar lower member. The relatively massive lower and upper members apparently form most of the ridges through the central sections of the Pomperaug basin, including the Orenaug Hills, while the lower massive member must thicken to the south, where it forms the main ridges of East Hill and Bear Hill in the western basin. The highly-altered and relatively soft middle member (containing most of the prehnite and other late minerals) might exist under several of the strike valleys eroded between basalt ridges. Because of its low durability, it has been eroded between the lower and upper basalt members, and so is an important control on the topography.

The Woodbury traprock quarry is also operated by the O&G Corporation. It cuts out part of the "eastern twin" of the Orenaug Hills in the northeastern section of the basin. The quarry operations have exposed large pavements of white sandstone beneath the basalt, while the basalt itself is highly fractured and generally weathered. Although there may be a boundary exposed between the lower and middle members, in general the flow relationships are hard to define. The upper flow (third member) that is evident in the Southbury Quarry is either not present or not exposed in the Woodbury Quarry. There are large piles of till that was bulldozed from the areas of quarry expansion, which contain abundant red siltstone cobbles and boulders. It is likely that these represent South Britain and/or Cass Brook sedimentary rocks from the northern end of the basin, which were carried into and over the quarry area by glacial actions.

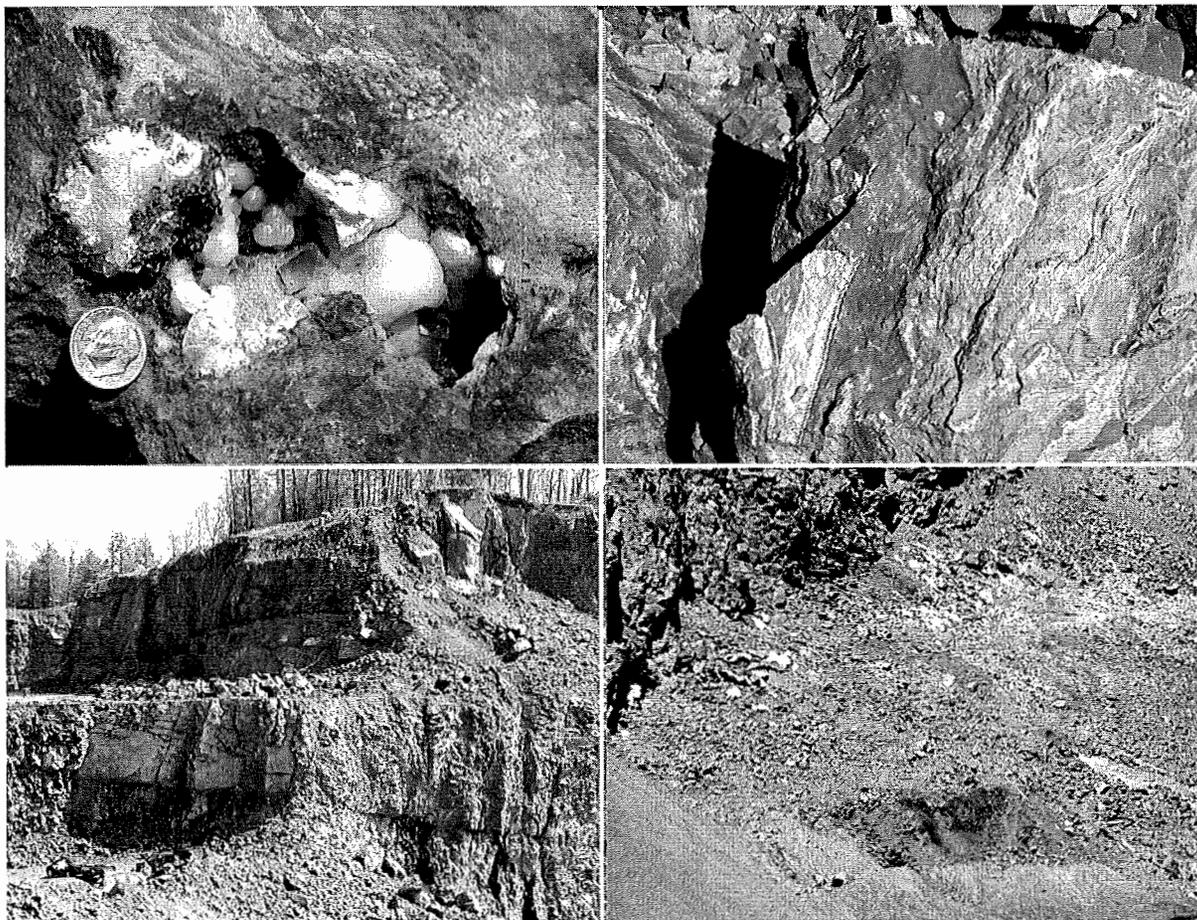


Figure 6. Images of the O&G Southbury Quarry. Upper left: prehnite “hearts” in a gas vesicle within the middle basalt member. Upper right: red siltstone and sandstone breccia dike in a fracture within the lower basalt member, possibly caused by explosive steam discharges from wet sediment beneath the basalt. Lower left: view to the north showing boundaries that define the middle member. Note the lighter tone of the middle member. Lower right: red siltstone (Cass Brook Fm.) and buff sandstone beneath the lower basalt member.

### Basalt Petrography

Petrographic thin sections were prepared for samples from basalt outcrops in the South Britain area, the traprock quarries, South Brook in Woodbury, and the Orenaug Hills. In general, hand samples of the East Hill Basalt show the highly-weathered state of this formation. All samples show abundant small (BB-sized) vesicles and amygdales filled with calcite or bluish-colored chalcedony. The basalt is generally soft and it crumbles when hammered, except for a few places not close to fractures. In some places the rock has disintegrated into “fish scales” that may be related to the shapes of gas bubbles. The basalt is very fine grained and may originally have been glassy, and only a few minerals other than amygdales are recognized. In thin section, small laths of plagioclase are abundant and well preserved, and clinopyroxene crystals can be recognized by their crystal forms (Fig. 7). No phenocrysts of orthopyroxene, such as can be found in unaltered Talcott Basalt, were recognized, although they may be present in altered forms.

In contrast to the East Hill Basalt, the upper and lower members of the Orenaug Basalt are typically massive, medium grained, and relatively unaltered. Hand samples are hard, and fresh surfaces are dark gray with small feldspar crystals visible. In thin section, the basalt is subophitic, with fresh clinopyroxenes and abundant plagioclase, and scattered phenocrysts that might be relict olivines (Fig. 8). The middle prehnite-bearing member is more altered, as expected.

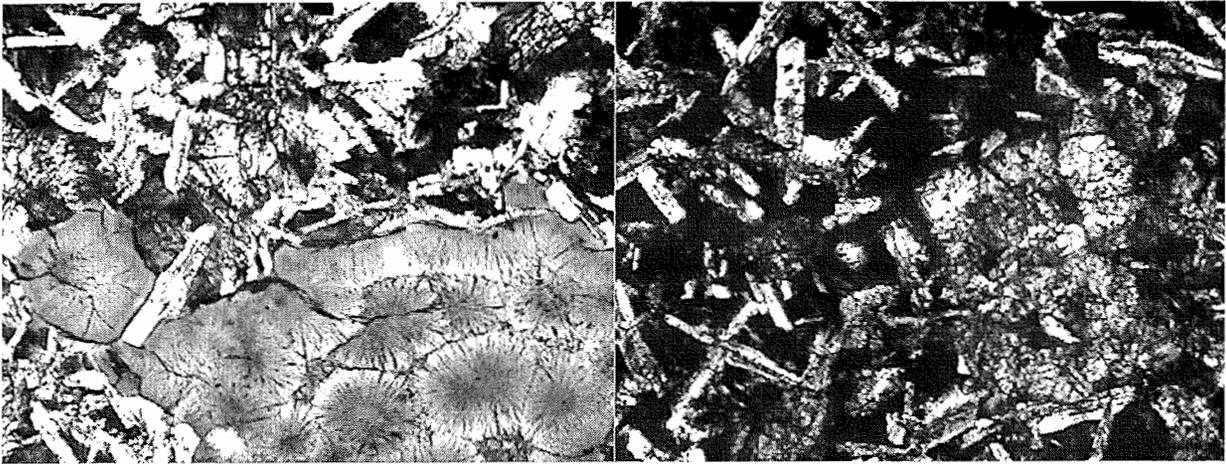


Figure 7. Thin section images (crossed-polarizers) of the East Hill Basalt at Red Spring. The left image show a chalcedony amygdale, with laths of plagioclase and altered equant grains of clinopyroxene. The right image shows plagioclase and altered pyroxene, with interstitial black material that may be devitrified glass.

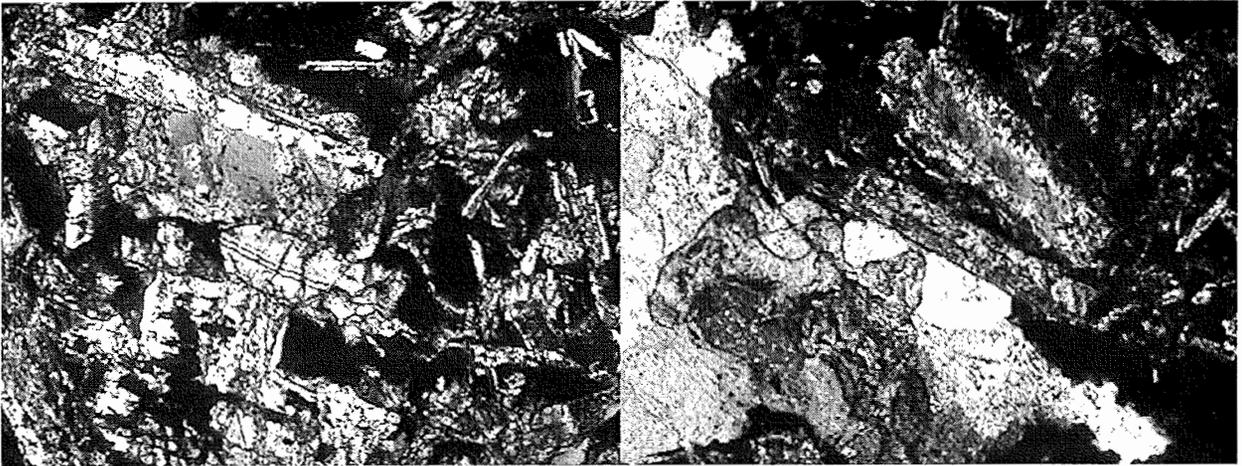


Figure 8. Thin section images of Orenaug Basalt from the Southbury quarry (crossed polarizers). Left: image of the lower member with dark patches of devitrified glass surrounded by plagioclase and clinopyroxene crystals. Right: carbonate-filled amygdales and plagioclase of the middle basalt member.

### Chemistry

As already demonstrated by Hurtubise and Puffer (1983) and Philpotts and others (1996), the Orenaug Basalt at Rattlesnake Hill in South Britain is chemically and petrologically identical to the Holyoke Basalt of the Hartford basin. As discussed above, field studies show this thick (c. 80 m) basalt to be the main ridge-forming basalt throughout the Pomperaug basin. The analyses published by Hobbs (1901), Hurtubise and Puffer (1983), and Philpotts and others (1996) all overlap on chemical diagrams with the Holyoke Basalt and also the Preakness Basalt of the Newark basin, including the commonly-used MgO - TiO<sub>2</sub> diagram (Fig. 9).

Previous to this work, the East Hill Basalt had only one published analysis (by Hobbs, 1901). At all locations, high degrees of weathering or hydrothermal alteration of this unit are shown by a high loss on ignition as well as by petrography. Alkalies, Ca, and Si are particularly affected. Hurtubise and Puffer (1981) discarded their analyses of the East Hill Basalt for this reason, using only 7 of 18 samples analyzed from all locations. Difficulties with

analyzing such highly-altered basalts results in scattered values even for relatively resistant elements, such as Ti and Zr (Figs. 9 and 10).

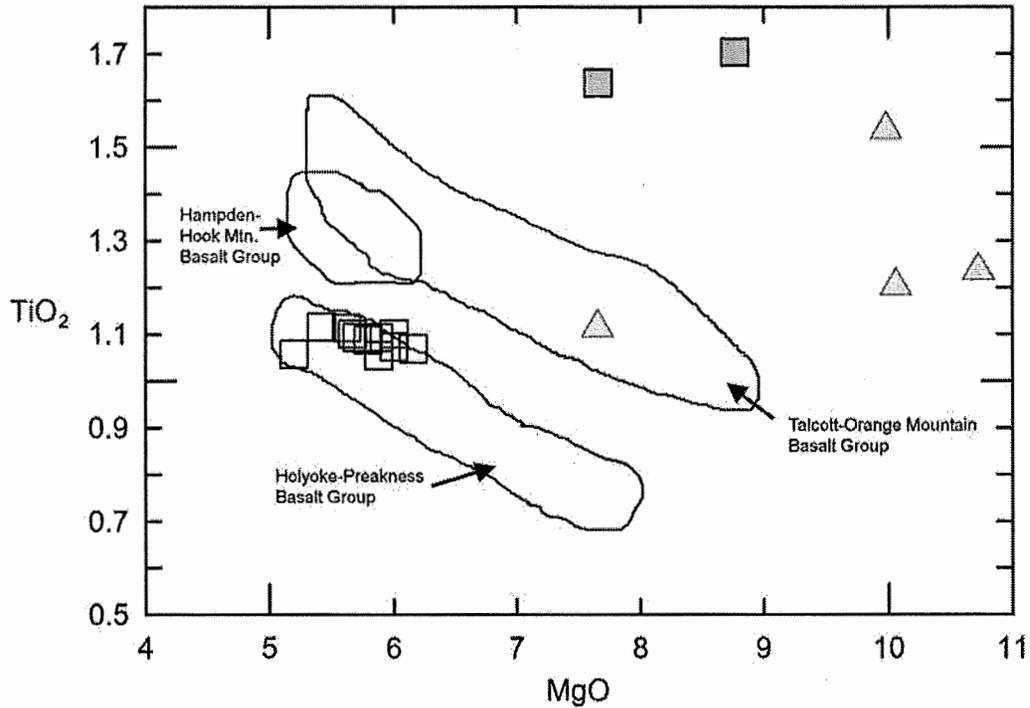


Figure 9. MgO-TiO<sub>2</sub> groups of basalts in the Hartford and Newark basins (outlines areas) relative to Pomperaug basin basalt analyses performed for this study. Triangles represent East Hill Basalt, open squares represent Oreanaug Basalt, and filled squares represent the proposed South Brook Basalt (see text).

On these and other geochemical diagrams (Figs. 9 and 10), Oreanaug Basalt samples maintain tight clustering except for modest linear indications of fractional crystallization, while the East Hill and South Brook Basalts show non-linear scatter due to alteration. However, it is apparent that the East Hill Basalt are relatively rich in Ti, Mg, Cr, and heavy rare earth elements, while the South Brook Basalt appears to have even higher Ti, as well as high Ba, Fe, V, and light rare earths relative to the other basalts. These differences most likely characterize magmas separate from the Oreanaug Basalt, and furthermore, they are similar to differences among the three basalts of the Hartford and Newark basins. Therefore, we can conclude that the Oreanaug Basalt is comagmatic with the Holyoke and Preakness Basalts of the larger basins, while the East Hill and South Brook Basalts are probably (but not definitively) the chemical equivalents of the Talcott-Orange Mountain and Hampden-Hook Mountain Basalts, respectively.



Mountain Formation, northwest of the basin. The ENE-trending faults shown in the southern basin parallel those originally drawn by Davis (1888) and later by Scott (1974) along notches in a ridge of South Britain Formation arkose that lies just east of the Platt farm buildings. While Davis' faults cannot be ruled out, their offsets, if they do indeed exist, are small (~20 m or less), and therefore are not shown on the new map. The fault drawn by Scott that truncates the south end of the basin is not compatible with Hobbs' (1901) description of trenches that expose redbeds which unconformably overlie schist on Horse Hill, south of the Pomperaug River and Scott's mapped fault.

**Eastern border fault.** The eastern border fault of the Pomperaug basin is most tightly constrained near the north end of the basin and particularly in South Brook, where a north-flowing segment of the brook follows the border fault. Here Mesozoic sedimentary and volcanic rocks to the west are separated by about 10 meters from Paleozoic igneous and metamorphic rocks to the east (Fig. 1, Stop 9). Exposures along the brook of the South Brook Basalt, the proposed third basalt in the basin, are strongly altered by secondary growth of calcite, dolomite, and other secondary minerals, and contain a couple of copper prospects. The association of faulting and mineralization is also a notable feature of the intrabasinal faults, discussed below; however, no breccia that would mark the actual fault zone itself has been found here. Immediately east of the border fault in South Brook the crystalline basement rocks are strongly jointed. The border fault is also fairly well-constrained by exposures and well records to the north of South Brook, where it passes just east of the Woodbury O&G trap rock quarry.

The remapped trace of the border fault south of the village center of Southbury is different from that mapped by Scott (1974) and shown on the state geologic map. There is no compelling reason for passing the fault through the crystalline uplands south of I-84, as Scott did. A thin, NW-trending body of amphibolite and calc-silicate can be traced from south to north across Peter Road and his mapped trace of the fault. About 1 km to the southeast a zone of northwest-striking, northeast-dipping, late (D3) cleavage extends across the mapped fault. Scott's cited stream exposure of the fault is just southwest of this zone of cleavage, but could not be recovered. His compositional justification for locating the fault could not be justified, either—namely, that abundances of staurolite and sillimanite west of the fault are higher and lower, respectively, than those east of the fault.

A topographically far more compelling location for the southern extent of the border fault is the valley between Southbury and the Housatonic River through which I-84 passes, just north of the bridge over the Housatonic. This valley likely marks an old channel of the Pomperaug River before, for reasons unclear, it was diverted to its present, more circuitous course around the south end of the Pomperaug basin. Passage of the border fault down this valley is also supported geologically: two prominent, ridge-forming amphibolite layers within the Rowe Formation, mapped by Scott (1974) and confirmed by the recent mapping, extend westward to the I-84 valley but could not be found to the west of it in the Russian Village area, despite fairly good exposures (Fig. 1). Assuming west-side-down offset across the fault, the moderately N to NE-dipping amphibolites should be offset west of the fault to positions that are south of the map area.

**Faults within the southern basin.** Assuming the stratigraphic sequence developed by Huber and McDonald (1992), and LeTourneau and Huber (1997) is correct, the best way to explain the distribution of lithologies in the basin, particularly at its south end, is through a series of small fault blocks bounded by mostly NNE-trending normal faults (Fig. 1). Faults of this general trend are exposed or well-constrained in several places within the Platt Farm Preserve (Stop 1), particularly along Cass Brook. One of the faults in Cass Brook is marked by zones of mineralized breccia that Hobbs (1901) mistakenly(?) termed "reibungsbreccia." (According to the AGI Glossary of Geology, reibungsbreccia is a synonym of "fold breccia", which is "a local tectonic breccia composed of angular fragments resulting from the sharp folding of thin-bedded, brittle rock layers between which are incompetent ductile beds" (Bates and Jackson, 1980). Given Hobbs' (1901) predilection toward faults, as discussed above, it is hard to imagine that he was thinking about folds.) In a gully formed on the hillside below a road culvert and above Cass Brook, a one to two-meter wide zone of mineralized fault breccia (the "reibungsbreccia") contains blocks of East Hill Basalt and marks a west-side-down normal-fault contact between east-dipping South Britain Formation arkose to the east and west-dipping Cass Formation siltstone to the west (Fig. 1, Stop 1). This fault can be traced southwestward to two breccia exposures in the brook, one of which likewise separates east-dipping arkose from west-dipping siltstone. To the east of Cass Brook, on a hillside, east-dipping arkose lies east of East Hill basalt, marking another northeast-trending west-side-down normal fault (Fig. 1, Stop 1). Both this fault and the fault marked by breccia merge into a NNW-trending, east-side-down normal fault that bounds a large block of Orenaug basalt to the east, and which extends northward to an area of complex geology at the north end of Cass Brook called by Hobbs (1901) Red Spring (Fig. 1).

About 0.5 km west of the south end of Cass Brook, outcrops of east-dipping South Britain arkose and overlying East Hill Basalt and Cass Brook Formation can be found juxtaposed along a north-trending, west-side-down normal fault against a ridge of Orenaug Basalt known as Rattlesnake Hill (Fig. 1, Stop 1). One kilometer farther to the northwest exposures of these lithologies in a brook outline opposing, NE-trending, east-side- and west-side-down normal faults that merge southward into a single east-side-down normal fault. These faults extend south across the Pomperaug River and offset the unconformity that marks the southern margin of the Pomperaug basin (Fig. 1). Farther still to the north and northwest, well records and outcrops indicate that two east-side-down normal faults occur along or near the western margin of the basin, one juxtaposing basement schist against South Britain arkose, and the other juxtaposing arkose against Orenaug Basalt (Fig. 1, Stop 5).

**Faults and folds within the northern basin.** Near the north end of the basin, within the Orenaug Hills in the town of Woodbury, an intrabasinal fault is outlined by a prominent topographic lineament and a (formerly) spectacular exposure in the O&G trap rock quarry (Fig. 1, Stop 7). Here pavements of a white, fine-grained, well-sorted eolian and ripple marked sandstone signifies the uppermost Cass Formation at the base of the Orenaug Basalt. This stratigraphic horizon is offset about 20 m in an east-side-down sense along a NNE-trending, steeply east-dipping normal fault which was exposed in the quarry wall (now buried), marked by a 2-m wide zone of brecciation and secondary mineralization (mostly quartz and calcite, plus zeolites). The mineralized fault face displayed outcrop-scale, gently north-plunging slickenlines which, combined with the east-side-down offset, indicate a sinistral strike-slip component of offset (Fig. 15). Three other faults are hypothesized within the Orenaug Hills, based on topography and likely lithologic offsets (Fig. 1). The westernmost one is mapped also on the basis of a well record in Woodbury that records sandstone in the shallow subsurface (Fig. 1).

In South Brook, east-dipping White Oaks Formation conglomerate is exposed just below the South Brook Basalt, about 90 m to the east of Orenaug Basalt that was recovered at 36 ft in USGS core CT-WY-87. Although this would appear to be a normal stratigraphic succession, the Cass Formation is inferred from a well record not far to the north along strike, and, farther to the north, exposures of the Cass Formation dip gently west underneath Orenaug Basalt (Fig. 1, Stop 8). The short distance between west-dipping Cass Brook and east-dipping White Oaks suggests that the intervening Orenaug Basalt is absent; to account for this missing section an east-dipping normal fault is mapped as extending southward from the left-stepping jog in the border fault shown by the well records (Fig. 1). This fault separates two *en echelon*, map-scale, hanging-wall synclines: the Woodbury syncline to the northwest, defined by bedding orientations in South Britain and Cass Brook rocks underlying the Orenaug Basalt in the Orenaug Hills (and cut by the intrabasinal faults discussed above), and the South Brook syncline to the southeast, defined by opposing bedding dips of White Oak rocks exposed in South Brook underneath South Brook Basalt (Fig. 1). The South Brook hanging-wall syncline involves mainly these two formations and is inferred to extend southward almost to the southern margin of the basin, to account for well records of sedimentary rock (White Oaks?) and basalt (South Brook?) overlying the Orenaug Basalt (Fig. 1). Hopefully this inferred distribution of younger sedimentary rock and basalt can be confirmed or disproved with future drilling and surface geophysics.

#### **Evidence for post-extensional tectonics in the basin**

Normal, dip-slip movement along the eastern border fault was obviously necessary to produce the Pomperaug basin itself, and the offsets along the intrabasinal faults can be mostly explained through normal dip-slip movement as well. However, there is evidence for post-extensional strike-slip movement on some of the faults within the basin, in the form of slickensided surfaces with gently-plunging slickenlines that probably record the last episode of fault movement. For the example of the NE-trending, slickensided fault in the Woodbury quarry (Fig. 15, Stop 7) cited above, if we assume that all of the fault movement occurred in the direction given by the slickenlines on the fault surface, which were measured to plunge northward at 27 degrees, the bedding surface in the Cass Brook Formation was displaced left-laterally about 40 meters in addition to its 20-meter east-side-down vertical displacement. An alternate explanation is that fault slippage parallel to the slickenlines occurred after most of the vertical displacement, resulting in a smaller, but still late, strike-slip component. Another NE-trending, slickensided minor fault surface, with horizontal slickenlines, occurs at the east edge of the quarry at the Cass Brook-Orenaug Basalt contact exposure (Stop 8).

A roadcut of Orenaug Basalt along Rte. 67, just east of Stop 5, exposes a NE-trending minor fault surface that has horizontal slickenlines whose asymmetric profiles indicate left-lateral movement. Along strike of this fault

about a kilometer to the southwest is a NE-trending, steeply NW-dipping fault in Orenaug Basalt with slickenlines that plunge NE 30 degrees and indicate—assuming normal fault motion—a right-lateral component of slip. More evidence for sinistral strike-slip motion can be found in Cass Brook, near the southern edge of the basin, including a NE-trending, moderately NW-dipping minor fault with slickenlines that plunge 35 degrees SW. Other slickensided faults around the southern end of the basin show near dip-slip motion, evidence that the sense of late offset on these minor faults is not uniform. Since the sense of offset associated with the slickenlines on these minor faults could not, in most cases, be determined, a reverse component of movement for some is a possibility. Similarly, although the map-scale intrabasinal faults are depicted as normal faults, some of them could be reverse faults with dips opposite those shown on the map, indicating basin inversion following extension.

### **Comparison of joint orientations in the Pomperaug basin and surrounding basement**

During the course of mapping, orientations of joints and joint sets were measured in exposures of both Pomperaug basin and surrounding crystalline basement rocks. Measurements were made on the basis of visual inspection and identification of joints with trace lengths over two meters, without the use of scanlines. A comparison of joints in Pomperaug basin rocks with those in older rocks would hopefully allow us to distinguish those joints produced during and after Mesozoic extension from those produced previously.

The results (Fig. 11) show that the most prominent joint orientation for both the Pomperaug basin rocks and the crystalline basement rocks is NNE and subvertical, with the strike of this trend in the crystalline rocks (Fig. 11D) about 14 degrees to the east of the Pomperaug trend (Fig. 11B). The Pomperaug basin joints also form a secondary peak trending WNW, orthogonal to the primary peak, and a tertiary peak trending NE (Fig. 11A, B). Less prominent joint orientations in the crystalline rocks show a more uniform distribution, among which are peaks that correspond to the secondary and tertiary peaks in the Pomperaug basin rocks (Fig. 11C, D).

The NNE major joint trend in the Pomperaug basin rocks fits well with the extensional tectonic environment during basin formation, and it is approximately parallel to the long axis of the basin. A major question is whether the NNE trend is Mesozoic in age. In his detailed study of the crystalline rocks of the New Milford 7.5-minute quadrangle, less than 20 km to the west, Walsh (2004) divided the map area into 17 domains and showed that the dominant joint trends in each domain are generally east-west and orthogonal to foliation, which has a regional strike of about N-S. In several of these domains secondary or tertiary peaks trend in a NNE direction similar to that of the crystalline rocks around the Pomperaug basin. These peaks, however, are mostly due to foliation-parallel parting joints, which are well-developed in the New Milford quadrangle (Walsh, 2004). In the Pomperaug basin area the mean foliation strike in the crystalline rocks is about due northwest (Fig. 11E), and foliation-parallel parting joints are not well developed and probably do not account for any of the principle joint trends. A principle joint trend orthogonal to foliation, similar to that found in the New Milford quadrangle, would produce a peak trending roughly due NE, or more easterly than the one determined for the Pomperaug area rocks. Therefore the NNE principle joint trend determined for the crystalline rocks is here proposed to be Mesozoic in origin as well, perhaps slightly refracted into a trend more orthogonal to the prevailing foliation. This trend is also nearly parallel to the third-most prominent trend in the Mesozoic rocks (Fig. 11B), and may be a hybrid of that as well.

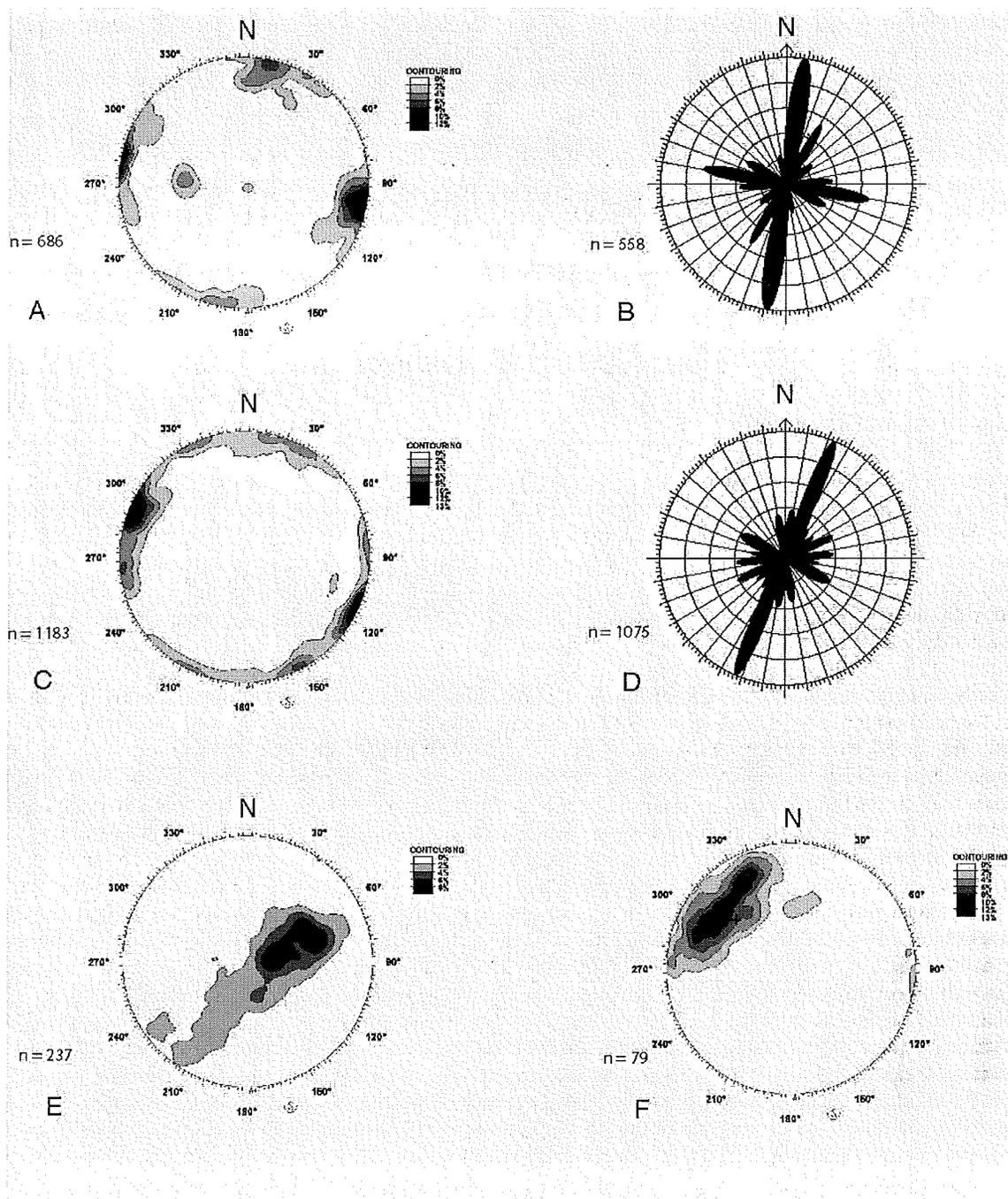


Figure 11. Contoured lower-hemisphere, equal-area plots and azimuth-frequency (rose) diagrams of structural elements in rocks of Pomperaug basin and surrounding crystalline basement rocks. Data plotted using the Structural Data Integrated System Analyzer software (DAISY v3.94) of Salvini (2004), which employs a Gaussian curve-fitting routine for the rose diagrams. A: Poles to joints in early Mesozoic rocks of Pomperaug basin. B: Rose diagram for Pomperaug basin joints with dips greater than 59 degrees. C: Poles to joints in pre-Mesozoic basement rocks. D: Rose diagram for basement joints with dips greater than 59 degrees. E: Poles to foliation in metamorphic basement rocks. F: Fold axes and mineral lineations in metamorphic rocks; heavy cross (+) marks pole to great circle fit to foliation in plot E.

Another question raised by the joint data is the origin of the WNW principle trend in the Pomperaug basin rocks, which is flanked by two lesser trends to the north and south, respectively (Fig. 11B). Such a joint trend suggests east-west compression or north-south extension; i.e. a stress field that is rotated 90 degrees from the one that produced the basin. Perhaps the presence of this principle joint trend is another indication of post-early Mesozoic compression, or inversion, of the Pomperaug basin. These joint trends appear to be present as well in the crystalline rocks, but reversed in prominence, with the northern of the three trends most dominant and the middle trend the least dominant. This disparity could be explained by the fact that the joint measurements in the crystalline rocks are nearly twice as numerous as the basin rocks, and undoubtedly reflect the influence of one or more pre-Mesozoic events in addition to the Mesozoic and post-Mesozoic (?) events. In addition, the compressive event that produced the WNW joint trend, if indeed the cause, would have had less of an effect on the crystalline rocks than the relatively weak basin rocks.

## ORIGIN OF THE POMPERAUG BASIN

### Sources for basalts and sediments

There are three models that attempt to explain the origin of the early Mesozoic Pomperaug basin: (1) it is a separate extensional basin on a par with its much larger neighbors the Hartford and Newark basins, with its own distinct depositional sources and tectonic history (Huber and McDonald, 1992; Huber, 1996); (2) it is a down-faulted remnant of the central area of a much larger, regional basin structure (a deeply eroded anticline) whose original extent is defined by the present eastern and western border faults of the Hartford and Newark basins, respectively (the Broad Terrane model of Russell, 1892; Sanders, 1960); and (3) it is an erosional outlier that represents the "feather edge" of an originally much larger Hartford basin, preserved by post-depositional downfaulting (the modified Broad Terrane model of Krynine, 1950; Hubert et al., 1978). Because of the stratigraphic and geochemical similarities between the Pomperaug basin and the Hartford basin, either variant of the Broad Terrane model has been adopted by many workers over the past 150 or more years, e.g. Percival (1842); Sanders and others (1974; 1981); Hubert and others (1979), Weddle and Hubert (1983); McHone (1996). An additional hypothesis has been put forth recently by Blevins-Walker et al. (2001) and Wintsch et al. (2003) that considers the Hartford and Pomperaug basins to be merely the preserved portions of a large, low-relief, regional foreland basin that encompassed much of southern New England during the Permian through Cretaceous, thus rejecting an extensional tectonic origin for the individual basins.

The evidence presented here suggests that another model, intermediate between the modified broad terrane and isolated basin models, might be more realistic. In our opinion, the Pomperaug basin is not entirely a local basin with local sources; nor is it merely the downfaulted remnant of a once-much larger basin. Apart from the overall gross similarity in lithostratigraphy (which itself is as much the product of regional long term, Milankovich-driven paleoclimatic fluctuations as of tectonism), two lines of evidence most strongly link the Pomperaug basin with the Hartford basin. The first is basalt geochemistry: the East Hill and Orenaug Basalts plot close to the same geochemical fields as the "first" and "second" basalts, respectively, of the Hartford and Newark basins, suggesting that the East Hill and Orenaug Basalts are in fact distal portions of the Talcott and Holyoke Basalts, respectively. However, the two geochemical analyses of the "third" basalt of the Pomperaug basin (South Brook Basalt) that we have do not fall into the fields for Hampden Basalt or the other basalts. Their consistent grouping together on various discrimination diagrams suggests that this difference could be real and not due to sample alteration, and implies, perhaps, a magma source separate from those that fed the Hartford basin basalts. After all, the basalts of the widely separated Newark and Hartford basins, despite their similarities in chemistry and timing of eruption, possibly flowed from separate vents — could not the same be argued for the third Pomperaug basalt? However, we note the absence of any vents and source dikes in Connecticut other than those identified as feeders to the main basalts of the Hartford basin (Philpotts and Martello, 1986).

Basalts of the Pomperaug basin could therefore represent a combination of distal and local sources. For instance, the eruptive centers for Talcott magma were located within the southern and east-central Hartford basin, within 30 km of the Pomperaug basin (Olsen et al., 2003; Huber et al., 2005). Conceivably, Talcott lava could have advanced westward from the Hartford basin, along the topographic low areas of the regional drainage network, as far as the Pomperaug basin. Some support for this concept comes from regional thickness data for Talcott-East Hill extrusive volcanic rocks, which are 20 m-thick or less hyaloclastic breccias in areas adjacent to the Hartford basin

eastern border fault, but comprise intercalated basalt and hyaloclastic and pyroclastic breccias up to ~300 m thick at their eruptive source areas, located only about 10 km to the west within the basin. The Talcott Basalt thins to 50 m thick at Meriden in the central Hartford basin, where it comprises two distinct basaltic flow units, and to less than 10 m thick in the Pomperaug basin where it represents a single flow unit. Potential intrabasinal sources for Holyoke lavas have not been located, and it is entirely feasible that these basalts were extruded either from now-eroded, eruptive fissure complexes developed along the western margin or west-central portion of the Hartford basin, or from fissures located in the crystalline uplands located east (or southwest) of that basin (Philpotts and Martello, 1986). Eruptive sources for the Hampden Basalt have also been located within the Hartford basin, in north-central Massachusetts, though the geochemical affinity of the Hampden Basalt with the "third" basalt of the Pomperaug basin has not yet been demonstrated -- indeed, the internal stratigraphy and regional flow indicators of the Hampden Basalt are not well-documented.

The second line of evidence that suggests at least some degree of common basin-fill source areas for the two basins is in the form of  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of detrital white mica from the coarse clastic rocks of the Hartford and Pomperaug basins. Blevins-Walker and others (2001) analyzed white micas from the New Haven Arkose in the Hartford basin and the correlative South Britain Formation, and found that Alleghanian ages dominated over Acadian ages. Since an Alleghanian thermal overprint is widespread in eastern Connecticut (Wintsch and others, 1992), but not known in western Connecticut, they interpreted the results to mean that the source for both basins was the post-collisional upland terrain of eastern Connecticut, whose erosion spread sediment westward over a broad alluvial plain, covering most of western Connecticut, prior to formation of the fault-bounded basins. Thus, the source of the first sediment for the Pomperaug basin was neither its own eastern border fault nor that of the Hartford basin--neither of which existed yet—but land east of the Hartford basin. The two Acadian ages that they did get from the detrital micas in the South Britain Formation they attribute to isolated monadnocks in the alluvial plain in western Connecticut. This hypothesis, relying heavily as it does on geochronology to overrule other, more direct lines of evidence for local sedimentation, would require for confirmation a more extensive collection of not only white mica  $^{40}\text{Ar}/^{39}\text{Ar}$  ages, but also an updated geochronology of certain rock bodies in western Connecticut, such as the small, numerous late stage pegmatites and perhaps the Nonewaug Granite, all of which occur in basement rocks not too distant from the Pomperaug basin. Until that time, however, we must at least consider the possibility that some of the sediment in the Pomperaug basin came from eastern Connecticut.

In support of the argument that the sediment in the Pomperaug basin is at least partially locally derived, the distribution of sedimentary facies in the Pomperaug basin mirrors that of larger, typical nonmarine extensional basins, with coarse-clastic, alluvial-apron sediments along the margins grading into finer-grained fluvial, playa and lacustrine facies toward the basin center. Clast compositions in the coarse-grained sediments of the Pomperaug basin suggest a local provenance. Clasts in the South Britain, Cass and White Oaks formations include those of garnet-mica schist (Fig. 12A), a lithology that is widespread around the basin in the early Paleozoic metamorphic rocks, and marble (Fig. 12B), whose only feasible source is the Cambrian Stockbridge Marble, near the western edge of the state (Rodgers, 1985). Also found in all three formations are intraclasts of shale and siltstone (Figs. 12C&D), which would appear to represent internally reworked sediment, and rocks showing lower-grade deformation or metamorphism, including strained quartz (Fig. 12E) and chlorite-quartz schist (Fig. 12F). The only source for low-grade metamorphic rocks in Connecticut at the present land surface is the Orange-Milton belt, about 20 kilometers southeast of the Pomperaug basin, which occupies a corner of the western Connecticut uplands from Long Island Sound northeast to the Hartford basin (Rodgers, 1985); it is also plausible that these clasts could have been derived more locally from less-deeply-buried metamorphic rocks that have long since been removed by erosion.

Regardless, all of these lithologies, along with the extremely elongate shape (Fig. 12C) and size (Fig. 12E) of some of the clasts, argue for at least some of the sediment sources being more proximal or in a direction different than that of the Hartford basin. The general immaturity, poor to moderate sorting and coarseness of Pomperaug basin sandstones and conglomerates also indicate local provenances. In any case, the present sedimentological, geochemical, and geochronological evidence suggests that the Pomperaug basin is not simply a down-faulted outlier of the Hartford basin, but may reflect a combination of local and regional sedimentary and igneous sources.

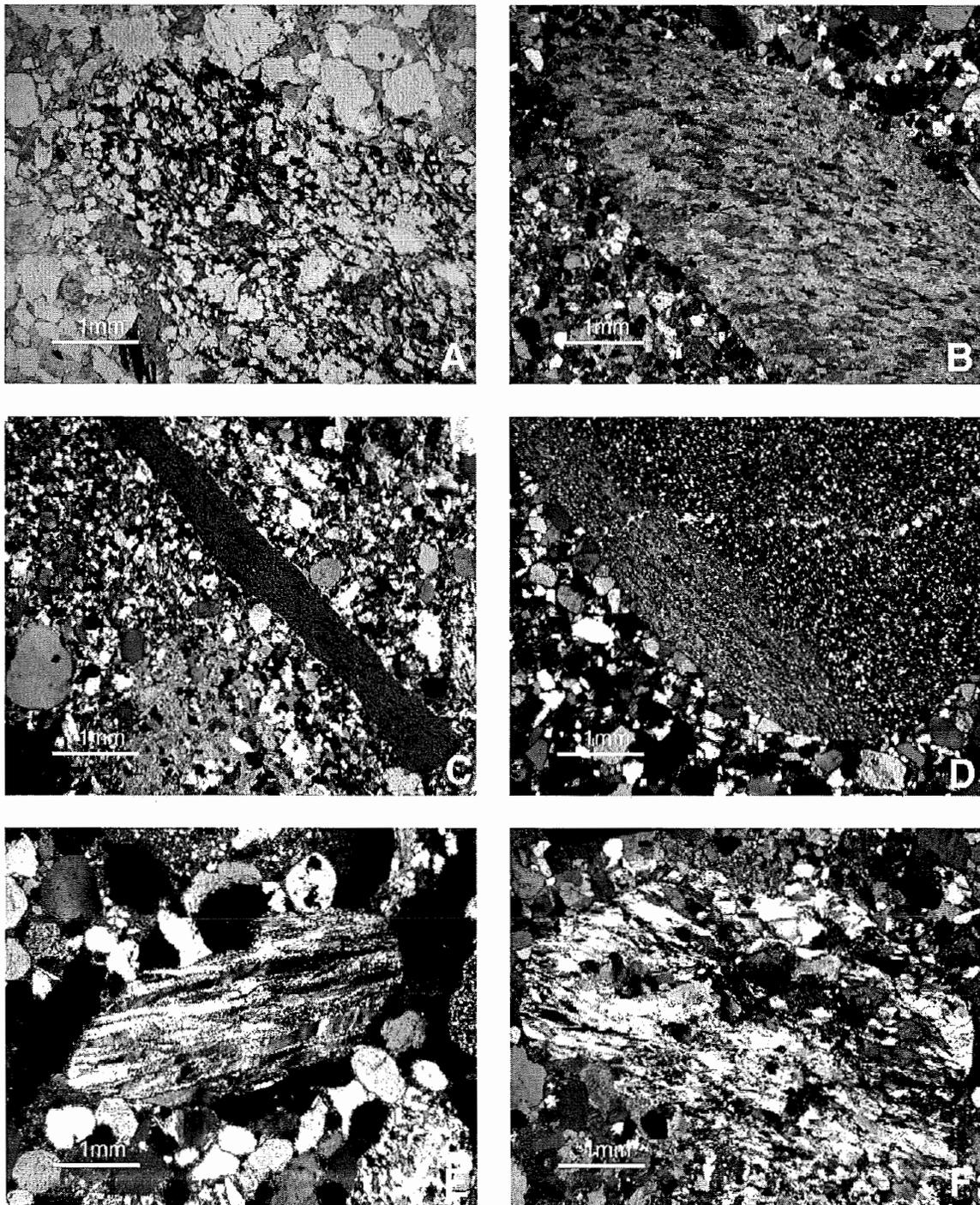


Figure 12. Clasts in conglomerate in South Britain (A) and White Oaks (B-F) formations. A: garnet-muscovite-biotite schist. B: foliated marble. C: shale chip. D: interbedded siltstone and shale. E: strained quartz. F: chlorite-quartz schist.

#### Size and location of the basin

Why is the Pomperaug basin so small? Evidence from the larger early Mesozoic basins suggest that their primary border faults formed through tectonic reactivation of pre-existing Paleozoic structures (Schlische, 1993):

e.g., the western border fault of the Newark basin formed from southeast-dipping Paleozoic thrust faults (Ratcliffe and Burton, 1985), the eastern border fault of the Hartford basin formed along the west-dipping west limb of the Bronson Hill anticlinorium (Rodgers, 1985; Wintsch and others, 2003), and the western border fault of the Culpeper and Gettysburg basins formed along the east-dipping east limb of the Blue Ridge-South Mountain anticlinorium (Southworth and others, 2000). In contrast, the NNE-trending eastern border fault of the Pomperaug basin clearly cuts across the structural grain of northwest-trending Paleozoic contacts and metamorphic foliation (Fig. 1); no known older fabrics parallel the border fault. We propose that the Pomperaug basin is so small because there were no pre-existing structures in favorable orientations that could facilitate fault development and opening of the basin—it is therefore “the exception that proves the rule” about tectonic inheritance and early Mesozoic extension. Its small size is probably also related to its position between the large Hartford and Newark basins, where most of the major regional extension would be accommodated (Olsen and Schlische, 1988). However, the late-faulting model (model 3 presented above) is harder to justify for the same positional argument: why would extensional faults develop in an unfavorable area after adjacent, much larger zones of extension are already well-established? Clearly the “why” of the Pomperaug basin is not easily answered.

### PRE-MESOZOIC ROCKS

William C. Burton

The Pomperaug basin is underlain by schists, gneisses, and granitic rocks of early Paleozoic age that were metamorphosed during the Acadian orogeny (Scott, 1974; Stanley and Caldwell, 1976; Rodgers, 1985). In the present study these rocks were mapped out to a distance of about 2 km from the basin margin. The twin goals of this portion of the study were to try to understand the nature of the response of the crystalline basement rocks to the crustal extension associated with basin formation, discussed in the previous section, and to shed new light on the Paleozoic deformational history of the rocks. Consequently, ductile structures were mapped along with the brittle structures, including several generations of foliation and folds, and mineral and intersection lineations.

#### Lithologies

Most of the basement rocks can be described as northwest-trending, several-kilometer-wide belts of metapelitic schist and gneiss, with only subtle compositional and textural differences between belts. These differences include grain size, color, nature of foliation, and the relative abundances of common minerals such as muscovite, biotite, and garnet. The map units were mapped using these criteria and then reconciled as well as possible with the map units on the Bedrock Geological Map of Connecticut (Rodgers, 1985). The resulting interpretation is shown on Fig. 1. Thin elongate, mappable bodies of amphibolite, calc-silicate rock, quartzite, and other primary lithologies were mapped within many of these units, as well as sills of foliated granite and pegmatite. The new map differs from the state geologic map in that quartz-rich Ratlum Mountain schist (Or, Fig. 1; Stop 3) extends southeast of the Pomperaug basin, through an area previously mapped by Scott as “Hartland II” (Rowe on the state map), and interlayered Taine Mountain and Collinsville formations. In the new interpretation the Taine and Collinsville formations are kept separate, with Taine Mountain (Ot, Fig. 1) entirely northeast of a belt of Collinsville Formation (Oc, Fig. 1), which forms an overturned syncline with a thin belt of The Straits Schist (Dst, Fig. 1) at its core. The Taine Mountain formation is underlain to the northeast by basal Taine Mountain (Otb, Fig. 1), which just east of the map area directly overlies the Waterbury Gneiss in the core of the Waterbury Dome, according to Rodgers (1985). An early (D1?) thrust fault is hypothesized for the contact between Collinsville Formation and Ratlum Mountain Formation, separating the sequence of rocks connected to the Waterbury Dome (Collinsville Fm./Taine Mountain Fm./Waterbury Gneiss) from Rowe and Ratlum Mountain schist. The extension of this fault northwest of the basin is instead occupied by a previously unmapped, 2-km wide body of foliated two-mica granite (Fig. 1, Stop 4). This granite, as well as the smaller granitic sills mapped northwest of the basin, is perhaps comagmatic with the intrusive body north of the basin mapped as Nonewaig Granite by Gates (1954), although the latter is actually a series of pegmatite bodies (Robert Tracy, oral comm., 2005). The metamorphic grade of all these rocks is upper amphibolite facies, as represented by kyanite-bearing mineral assemblages, as well as lesser sillimanite- and staurolite-bearing assemblages.

#### Tectonic models

As previously mentioned, the basement rocks underlying the Pomperaug basin are distributed in northwest-striking lithologic belts, despite the fact that in a very broad sense the metamorphic formations of western

Connecticut trend north-south or northeast-southwest. Within each belt are folds ranging from map scale, as outlined by amphibolite and kyanite schist layers mapped in the Rowe, Ratlum Mountain, and Collinsville formations (Fig. 1), to outcrop scale, as shown by tight to isoclinal to locally rootless folds of amphibolite and granofels layers within the schists. The thin, northwest-trending belt of The Straits Schist (Dst, Fig. 1) within the Collinsville Formation (Oc, Fig. 1) was considered by Scott (1974) to represent the nose of an early-stage, east-facing synclinal nappe that trended north-northeast across the present position of the Waterbury dome.

Also widespread in the map area, particularly in the south and west, is a prominent, gently northwest-plunging lineation, expressed in outcrop by the hinges of minor folds and on foliation surfaces by mineral elongation directions (Fig. 11F). Despite the fact that they mapped the same generation of folds and lineations in adjacent quadrangles, Scott (1974) and Stanley and Caldwell (1976) had very different ideas as to the origin of these structures. Scott (1974) thought that all of the northwest-trending structures, including contacts, dominant schistosity, fold hinges, and lineations were produced during formation of the Waterbury Gneiss dome just to the northeast, which in his model represented the fourth major phase of folding in the region, all phases of which were Acadian in age. The doming produced a "pinching" of rocks on the dome flanks or a "draping of the rocks off the dome," and resulted in the formation of these structures tangential to the dome. The lineations were of his "Period-4" generation, and accompanied northwest-trending, map-scale, upright folds (Plate 1 in Scott, 1974) that produced the alternating belts of Taine Mountain and Collinsville formations as shown on the Bedrock Geologic Map of Connecticut (Rodgers, 1985).

In contrast, Stanley and Caldwell (1976) considered the lineations and co-linear fold hinges to represent the second phase of folding (F2) in the area, which predated doming, accompanied Acadian peak metamorphism, and produced west-plunging (present orientation) tight to isoclinal folds, and accompanying north- to northwest-striking, west-dipping axial-planar schistosity that is the dominant foliation in the region. These second-generation folds strongly deform an older, larger, nappe-stage F1 fold of uncertain original orientation and possibly of Taconian age, producing a very complex "jelly roll" structure (Figure 10 in Stanley and Caldwell, 1976). The F2 movement direction may locally have been south over north, based on the rotation sense of minor folds in a calc-silicate layer just northwest of South Britain (Stanley and Caldwell, 1976). However, their cross-section A-A', drawn normal to the west-plunging fold axes, does not suggest a dominant overall rotation sense (Plate 2 in Stanley and Caldwell, 1976). They also recognized a post-peak-metamorphic Acadian F3 fold generation, and a late, F4 generation associated with retrograde, chlorite-bearing mineral assemblages that may represent effects of the Alleghanian orogeny (Stanley and Caldwell, 1976).

Based on the recent mapping around the Pomperaug basin, we agree with the Stanley and Caldwell (1976) model that the dominant structures in outcrop, including minor folds, lineations and schistosity, are second generation (S2/F2) in age, formed under peak metamorphic conditions, and are indicative of tight to isoclinal, minor to map-scale, west-plunging folds. It is possible, however, that the west to northwest-plunging lineations and fold hinges indicate transport direction, either up from the west, or up from the east and overturned by later folding. A first-generation foliation (S1) is also thought to be represented by locally preserved, fine-grained compositional layering—although F1 folds have not yet been recognized—and F3 folds locally form crenulations and kink folds in F2 schistosity. Many questions remain, particularly concerning the respective roles of the Taconian and Alleghanian orogenies, and more work is planned in the less-well-mapped Roxbury and Woodbury quadrangles, to the north of the quadrangles mapped by Scott (1974) and Stanley and Caldwell (1976), and just east of the New Milford quadrangle recently mapped by Walsh (2004).

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### ROAD LOG

8:30 AM Assemble at park and ride lot in Southbury, directions to which are as follows: Take I-84 West to Exit 14 for South Britain and State Route 172, turn right at bottom of ramp onto Rte. 172, take right again at light onto Main Street South, and right again into park and ride parking lot. Here we will consolidate cars.

#### Mileage

- 0.0 Park and ride lot on Main Street South and State Route 172, Southbury, CT.
- 0.1 Take left from parking lot onto Main Street South and take right at light on Route 172, heading northwest.
- 0.8 cross bridge over Pomperaug River, into village of South Britain.
- 0.9 turn right on Library Road.
- 1.0 turn right on Flood Bridge Road. Platt Farm Nature Preserve is on your left—hills are underlain by Orenaug Basalt. Note the large glacial boulder train of Orenaug Basalt in the field to your left.
- 1.2 turn left onto road passing between the buildings of Platt Farm 1.4 parking lot for Platt Farm Nature Preserve. Here we will park and head out on foot.

**STOP 1. Platt Farm Nature Preserve.** (2 HOURS) Park in the nature preserve parking lot. Refer to Figure 13. Small pavements of arkose in the slopes on both sides of the parking lot are probably outcrop. From the parking area, take the grassy trail west between bushes. Take the right fork, up hill, following the blue trail blazes on trees. Note large outcrop of Orenaug Basalt on hill on right with well-developed columnar jointing. The columns here are not normal to regional dip of basalt flow; elsewhere they are and can be used (with caution!) as a proxy for strike and dip of stratigraphy. Follow blue blazes to pond, at pond dam take left and then left again on yellow and red-blazed trail, heading up hill. At fork take yellow-blazed trail which curves back north along small ridge. Note moss-covered float and small outcrop ledges of South Britain Formation arkose, underlying ridge. The arkose dips

east about 25 degrees. Across ravine to west is massive Orenaug Basalt. Since this basalt stratigraphically overlies the arkose, which is here dipping away from it, a N-S-trending, west-side-down normal fault is required in the ravine just east of the basalt to explain the geometry.

Turn around and go back down hill to the pond. Before the dam, take a right on white and purple-blazed trail, on west side of and parallel to a creek flowing south. After about 70 meters, note small outcrop on right of vesicular East Hill Basalt, which stratigraphically overlies South Britain Fm. arkose and appears here to have subvertical columnar jointing. As trail curves to right going downhill, look on the hill slope to right and notice float chips of brown, fine-grained sandstone and siltstone of Cass Formation, which overlies the East Hill Basalt and South Britain arkose and underlies the Orenaug Basalt, as part of an east-dipping section. Large, glacially-transported blocks of Orenaug Basalt on left. Right at turn in trail are large slump blocks of Cass Fm. sandstone and siltstone.

Take trail back to pond and take right across dam on red-yellow-blazed trail. Low ridge of Orenaug Basalt on right. Fork right on red-circle-on-white-blazed trail. This trail climbs over low ridge of Orenaug Basalt and then descends down to junction with main, blue-and-white-blazed trail. Cass Brook is just to east, down in valley. Take right on this trail, heading south. Pass back through parking area, and head back down road to yellow-blazed trailhead, and take left. At bridge over stream, small outcrop of vesicular, East Hill Basalt is on right (downstream). Just downstream from here about 20 meters is small outcrop of arkose--contact between the two outcrops is interpreted as a fault. Directly under the bridge on the south bank of the stream, and in the stream bed, are exposures of the lower ~5 m of the Cass Formation, including fossiliferous lacustrine limestone, shale and siltstone. These strata produce a flora and fauna typical of the Shuttle Meadow Formation in the Hartford basin, including a Corollina-dominated palynoflora (Cornet, 1977), stem fragments of *Equisetites* sp., leafy shoots of the conifer *Brachyphyllum*, occasional fragments of the cycad *Otozamites*, fossil fishes pertaining to the redfieldiid *Redfieldius* sp. and *Semionotus* sp. The contact with the underlying East Hill Basalt is immediately to the west and northwest. Leave trail at bridge and descend into brook, heading upstream (north). First exposures up stream on right are East Hill Basalt, followed by exposures on left (west) of east-dipping, cross-bedded, pebbly South Britain arkose, in stratigraphic contact with the basalt. Where stream takes sharp bend to left (west), complete section below Orenaug Basalt is exposed. Pavement outcrops in the stream are of pebbly to locally conglomeratic South Britain arkose. Steep embankment to east is underlain mostly by East Hill Basalt capped by ledge of ripple cross laminated siltstone of Cass Formation.

At the upstream end of arkose pavement stream take sharp bend to right, heading north. Here is exposed highly fractured East Hill Basalt, overlain by west-dipping Cass Fm. strata. This west-dipping section must be separated from the east-dipping arkose just to the east by a NE-trending, west-side-down normal fault. Head upstream past exposures of pebbly arkose on right, on east side of fault. This is succeeded upstream by exposures of basalt, arkose, and mineralized fault breccia, as we follow the trace of the fault upstream. Chemistry indicates that most of the vesicular basalt outcrops are actually basal Orenaug Basalt. Proceed upstream through right bend and then left bend, to large exposure of vesicular Orenaug Basalt on west side of stream. In the stream itself is a submerged, steeply-east-dipping contact with beds of Cass Formation. Although this exposure suggests a dike intruding sediment, here it is interpreted as an east-dipping sedimentary contact, steeply tilted by faulting.

From this exposure, head uphill (east) to yellow-blazed trail, turn right and head south. In about 20 meters, trail crosses over ravine on small bridge—note exposures up hill in ravine. Here we see west-dipping, brown, fine-grained sandstone and siltstone of the Cass Formation, to east (uphill) of which is mineralized fault breccia (“reibungsbreccia” of Hobbs, 1901—but see discussion in text), which contains blocks of East Hill Basalt. Above that on hillside is ledge of east-dipping South Britain Fm. arkose. This mineralized fault zone, separating west-dipping Cass Fm. sediment from east-dipping South Britain arkose, is typical of intra-basinal faults in the Pomperaug basin.

Climb up the hill to the end of road, and head north about 30 meters to paved road heading off on right, and head up that road about 30 meters. Note crumbly exposures of East Hill Basalt on both sides of road. We are now in the normal, east-dipping section. Turn around and head back to south end of road. Veer off to left, down wooded slope, crossing old barbed-wire fence, and head towards clearing. Head east along north side of clearing, up to gap in trees. Here, note crumbly pavement exposure of vesicular East Hill Basalt. About 20 meters south along ridge is ledge of NE-striking, SE-dipping South Britain arkose, which is stratigraphically below East Hill Basalt but here dips away from it. A west-side-down normal fault is therefore mapped between the arkose and basalt exposures.

Wooded ridge from here south has nearly continuous ledges of South Britain arkose, possibly cut by NE-trending faults with minor offsets.

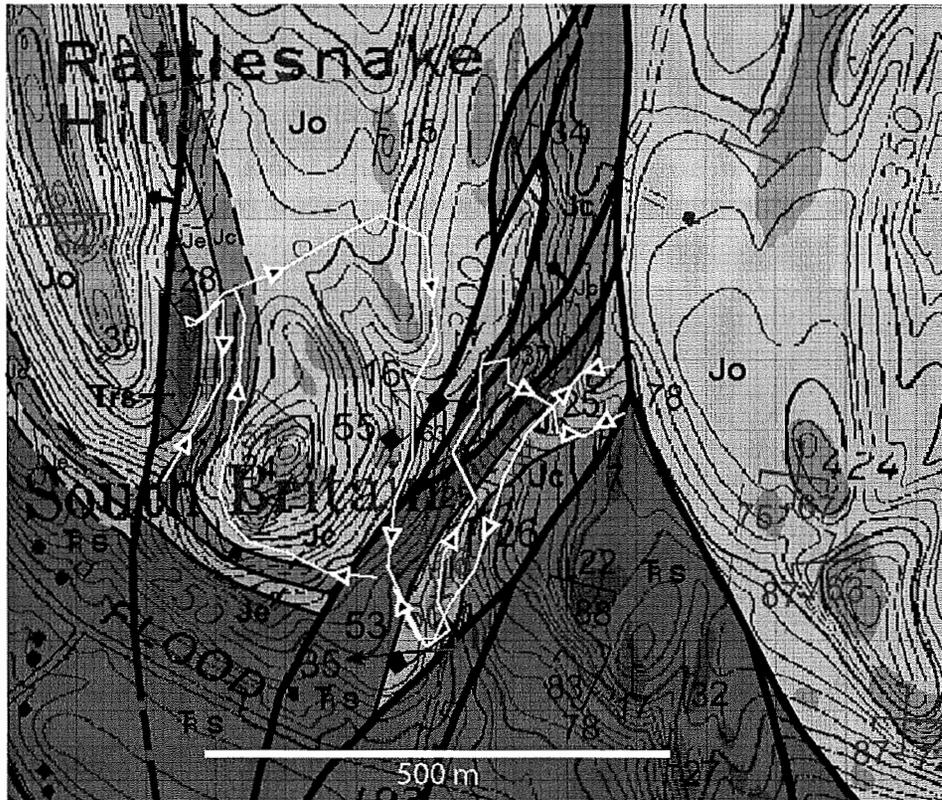


Figure 13. Close up of bedrock geologic map of Pomperaug basin showing walking route for Stop 1 in Platt Farm Nature Preserve. 1:24,000-scale topographic base from Southbury 7.5-minute quadrangle. Darker shaded areas denote outcrops. Symbols include strike and dip of bedding (standard symbol), joint (bracket), minor fault (bracket with black boxes), a plane normal to basalt columns (standard symbol with two ticks), and trend and plunge of slickenlines (arrow). Heavy lines denote normal faults, with ball and bar on downthrown side. Map units same as Figure 1.

Retrace steps to end of road, and take old dirt road down hill, heading south towards parking area. Note pavement of vesicular East Hill basalt underfoot, which is continuous with exposure just to west, above stream. Return to parking area and cars, retrace route to Route 172, and take right towards South Britain village center.

- 0.7 South Britain Country Store—if necessary, we will stop here for lunch supplies. On hillside to north, at north end of church parking lot, are ravine and hillside exposures of trough crossbedded, pebbly arkose of the South Britain Formation. Continuing north on 172, hills to south and west are underlain by crystalline basement (Rowe Formation schist) and hills to east underlain by South Britain arkose up through ridge-forming Orenaug basalt.
- 1.1 Turn right into development just before Southbury Training School. Bear right and drive up small hill, continuing to bear to the right, passing the traffic oval and past a locked gate that is posted. Continue straight onto cul-de-sac at far SE end of development. Park at the end of the cul-de-sac and follow footpath due south into woods a short distance until east-west flowing stream is encountered.

**STOP 2. South Britain Formation section.** (1 HOUR) The base of the section begins at or just above the intersection of the trail and the stream. This location, called the Southbury Training School (hereafter STS) section (Fig. 14), exposes about 90 m of strata representing the middle and upper South Britain Formation, which is the initial basin-fill unit of the Pomperaug basin and has an estimated thickness averaging 250 m. Our knowledge of the formation comes primarily from a ~190 m-thick composite section based on three roughly cross-strike outcrops that expose different overlapping portions of the unit. These outcrops include the Pomperaug River section at South

Britain Village, described by Schutz (1956), Hubert et al. (1978), Weddle and Hubert (1983), and Lorenz (1987), the O. Mitchell Brook section briefly described by Hobbs (1901), and this section (Fig. 14), which until our recent work, had never been studied, though Hobbs (1901) makes vague reference to its location. Other, limited exposures of the South Britain Formation are found in various areas of the basin, but typically represent several meters or more of the uppermost Rattlesnake Conglomerate Member.

The basal ~60 m of the South Britain Formation is not exposed, and the stratigraphically-lowest outcrops of the formation occur at the base of the Pomperaug River section. The middle and upper South Britain Formation is composed primarily of three facies associations that maintain their sedimentological character and stratigraphic position for at least 8 km along strike in the western area of the basin. The first facies consists of 0.5-3 m-thick beds of medium to coarse grained, well-sorted to poorly-sorted, trough cross-bedded arkosic sandstone that is often capped by relatively thin (0.25-2 m-thick) massive, burrowed and/or pedoturbated to ripple cross-laminated sandy siltstone. The lower 20 m of the STS section is composed of this facies, which we interpret as deposits of sandy, braided streams that flowed in an easterly direction from the western margin of the Pomperaug basin.

The second facies consists of thin beds (0.5-2 m-thick) of trough cross bedded, pebbly arkose that are overlain by as much as 18 m-thick sequences of massive, burrowed and/or pedoturbated to ripple cross laminated sandy siltstone and silty mudstone. This facies occupies a stratigraphic interval that encompasses approximately 110 m of the middle South Britain Formation, and represents the upper 70 m of the STS section (Fig. 14). The facies association is a classic fining-upward, fluvial point bar sequence deposited by meandering streams and rivers. Furthermore, the facies are conformable and gradational with underlying braided stream deposits, documenting maturation of the local drainage network as basin surface area increased over time, and possibly indicating the influence of a more humid regional climate. A remarkable aspect of the basal lag pebble conglomerates in this sequence is the abundance of carbonate pebbles. Some appear to be the products of reworked caliche, but others are metamorphosed limestone (marble). As paleocurrents at most exposures of the South Britain Formation are largely confined to a 40 degree sector (N290-330E,  $n = 250$ ), it is obvious that the extra-basinal carbonate clasts were derived from basement terrain located to the west and northwest of the basin. The only metamorphic carbonate unit located in this region of the western uplands of Connecticut is the Stockbridge Marble, according to the Bedrock Geological Map of Connecticut (Rodgers, 1985), the closest outcrops of which are located some 12 km west and northwest of the basin.

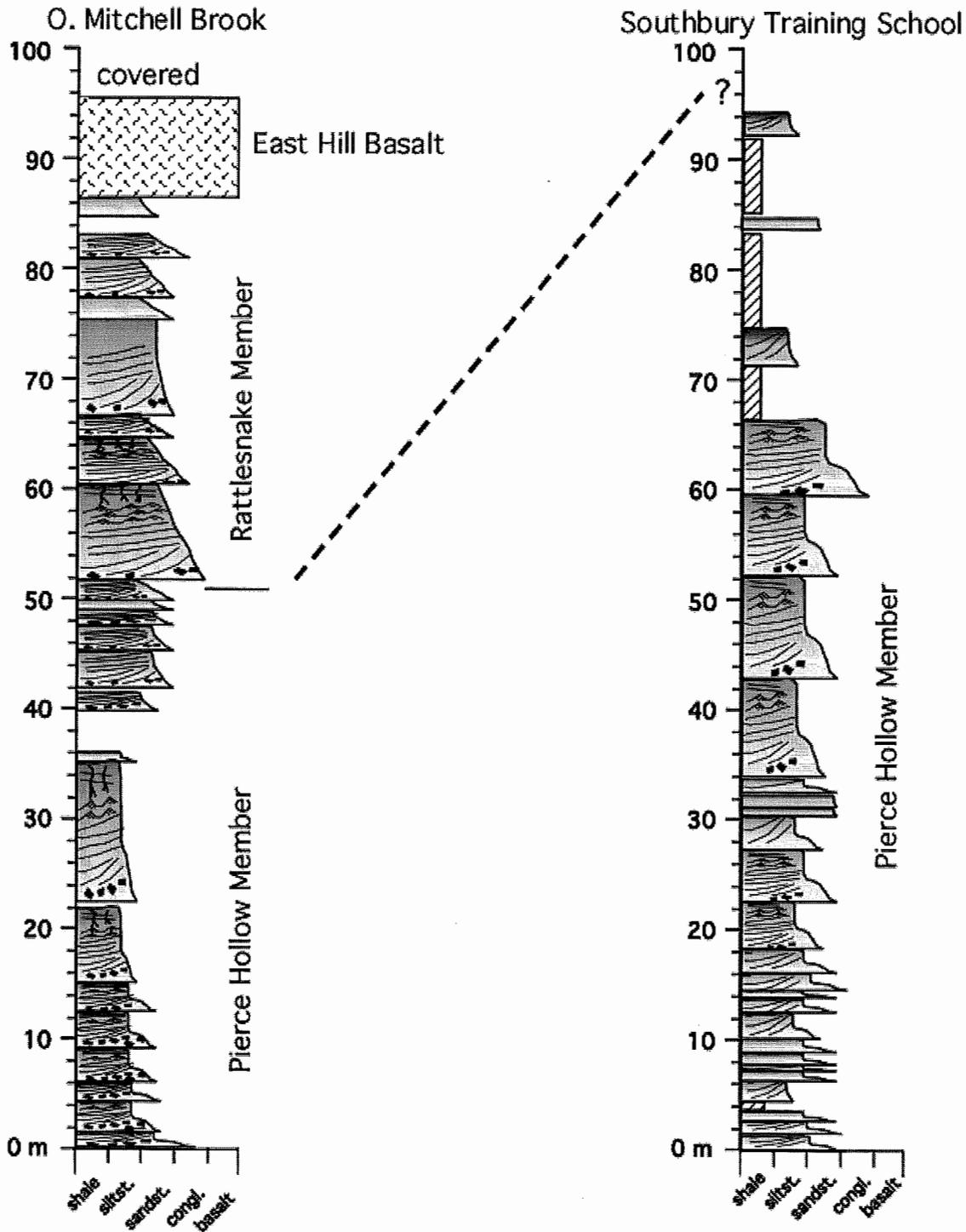


Fig. 14. Stratigraphic columns for the O. Mitchell Brook and Southbury Training School sections.

The third major facies association of the South Britain Formation, the Rattlesnake Member (Fig. 14), is composed of meter-scale beds of trough cross-bedded, pebbly arkose that dominate the upper ~50 m of the formation. This facies is not exposed at the STS section, the top of which is covered by thick glacial clays. However, further south at O. Mitchell Brook a complete section of the Rattlesnake Member is exposed, and excellent three-dimensional exposures are found along the bluff above the church parking lot opposite the South

Britain Country Store. Other outcrops also occur at the Platt Farm Nature Preserve, which we examined at our previous stop. This gravel-dominated facies of the South Britain Formation may reveal the progradation of alluvial aprons from the western margin toward the central basin floor. The western margin of the basin is located less than one km west of the localities described here.

Return to Route 172, and reset odometer.

- 1.1 Cross Transylvania Brook; outcrops of Rowe schist in brook to right.
- 2.0 4-way-stop intersection with State Route 67; continue straight on 172 (Transylvania Road), which involves a slight jog to the left.
- 2.8 New roadcut in schist of Ratlum Mountain Formation; pull off on right side of road.

**STOP 3. Schist of Ratlum Mountain Formation.** (10 MINUTES) Well-developed foliation defined by alternating quartz-rich and micaceous layers, believed to represent Paleozoic S2 foliation or possibly S1 reactivated by S2. This NW-striking, SW-dipping foliation is typical of the Ratlum Mountain Formation in this area and is responsible for the prominent ridges northwest of the Pomperaug basin. The *Bedrock Geological Map of Connecticut* (Rodgers, 1985) shows faults parallel to these ridges, but they are merely strike ridges that have been accentuated by glaciation.

- 0.1 North of Stop 2, turn left into driveway with Skinner sign out front, and then fork right immediately onto gravel road. This is the home of Brian and Cathy Skinner, and our lunch spot.

**STOP 4. Granite sill of Ratlum Mountain Formation.** (45 MINUTES--LUNCH STOP) Outcrop 10 meters south is of light gray, medium-grained, moderately well-foliated, porphyroblastic granite or granite gneiss. This granite typically forms NW-striking lenses and sills concordant with foliation in the surrounding schists and gneisses. The contact of this thin, NW-striking body with schistose country rock to the south runs right under the Skinners' house. The contact with quartz-rich gray granofels to the north is on the ridge immediately north of us. A two-kilometer-wide, NW-striking belt of foliated granite, not shown on the *Bedrock Geologic Map of Connecticut* (Rodgers, 1985), emerges from the Pomperaug basin not far to the north.

Head back south on Transylvania Brook Road.

- 0.8 Turn left (east) onto Route 67.
- 1.5 Turn left into gravel parking area in front of old cinder-block shed.

**STOP 5. South Britain Formation faulted against Orenaug Basalt.** (10 MINUTES) Cross road to north-facing hill slope at east end of parking area, proceed into woods and see small ledge of pebbly arkose, interpreted as outcrop. Cliffs 50 meters east, across small ravine, are of Orenaug Basalt and are known as Bates Rock. Ravine is interpreted as locus of NE-trending, intra basinal fault between arkose and basalt that cuts out intervening East Hill Basalt and Cass Brook Formation.

Continue east on Route 67.

- 0.2 Exposures of vesicular Orenaug Basalt on left, followed by disappearance of bedrock under glacial deposits.
- 0.9 Turn left into unmarked entrance of O&G trap rock quarry. After stopping at quarry office, proceed left around concrete plant, between large gravel piles. Watch for trucks!

**STOP 6. O&G Industries Southbury ("Silliman") Quarry.** (1 HOUR) The Southbury O & G Industries quarry has been in operation for many decades, and is famous for its spectacular assemblages of zeolite minerals, some of which were collected near this locale by Benjamin Silliman early in the 19<sup>th</sup> century. The quarry currently exposes 60+ m of the Orenaug Basalt, including the lower contacts of its two or three flow units. Over the past two decades the active quarry operations have intermittently uncovered and re-buried significant portions of the Cass Formation, including its upper contact with the overlying Orenaug Basalt. At one point, between 1990 and 1994, an 11 m-thick, sandstone-dominated sequence of the upper Cass Formation (Fig. 15) was exposed that included conglomerates and coarse, pebbly sandstones containing abundant molds and casts of dinosaur and other tetrapod bones (Huber and McDonald, 1992). This particular exposure has been re-graded and now lies beneath the grass covered slope along the northwest side of the quarry, opposite a small island of basalt that upholds quarry equipment.

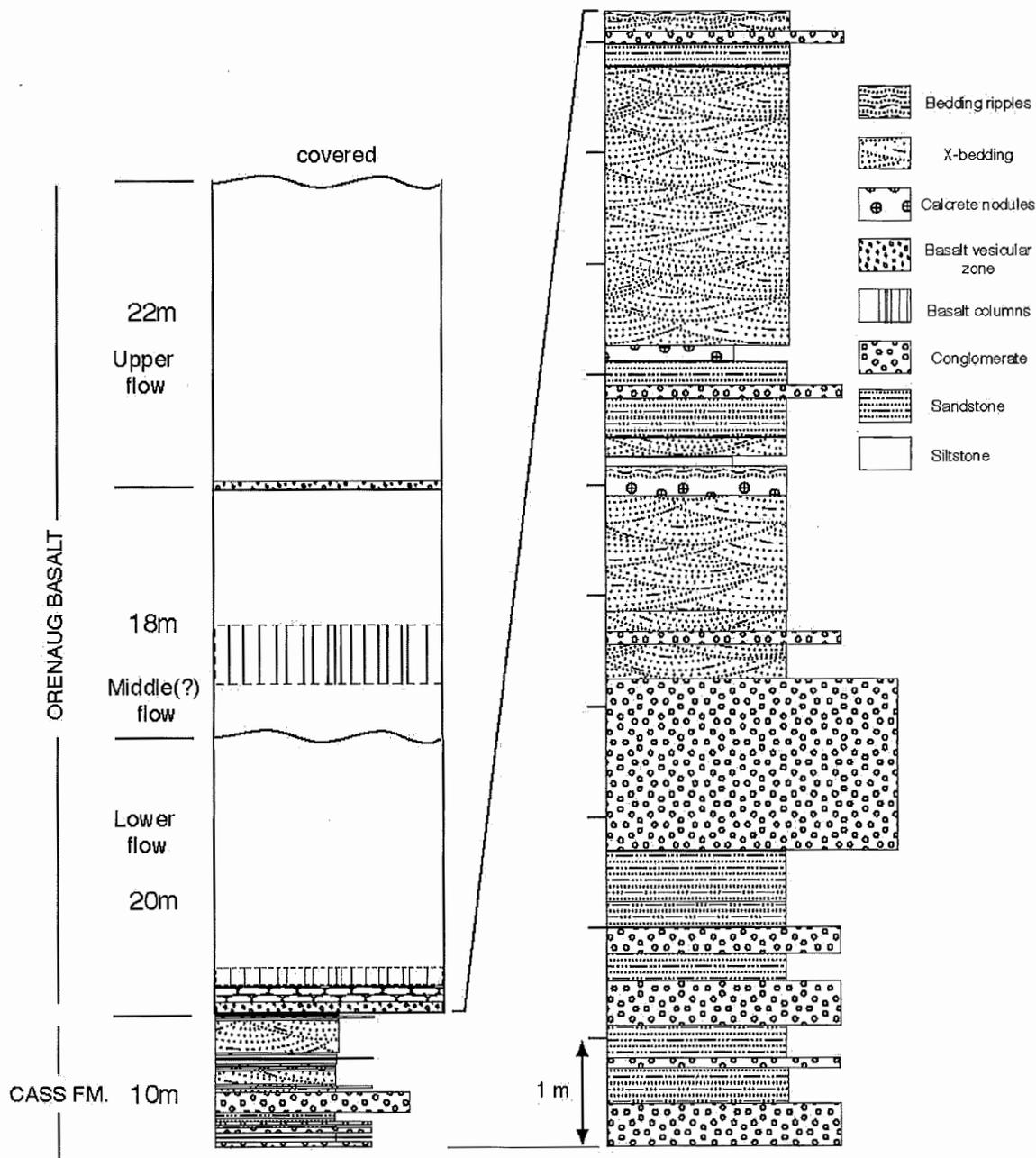


Fig. 15. Measured section for the O&G Woodbury quarry.

The upper 2.7 m of the Cass Formation observed in the quarry consists of fluvial, pebbly, quartz arenite and litharenite deposited by high flow regimes, as indicated by trough cross bedding and parting lamination. These beds are overlain by 1.5 to 2 m of coarse-grained, poorly sorted arkosic sandstone and conglomerate. The sandstone forms the basal 0.5 m of the unit (at the 3 m level on Figure 15), and contains abundant natural sandstone molds and casts of tetrapod bones, including the femur, tibia and vertebra of a probable theropod dinosaur. The encasing sediments are coarse grained and very porous, and all of the original bone material has been leached out, leaving semi-consolidated clayey sandstone casts of skeletal elements nested within three dimensional, undistorted molds. The bones are randomly distributed throughout the sandstone and include some that were buried in near vertical orientation. The sandstone is overlain by ~1 m of conglomerate (Fig. 16) that is crudely bedded and poorly sorted, and includes lentils that are matrix- and clast-supported. The dominant clast types are pebble to cobble-sized (up to 28 cm diameter) vein quartz (Figure 16D) and garnet-mica schist (Figure 16C) that are as much as 10 cm in length,

plus less-abundant pebbles of relatively fresh feldspar and rare granite. The top of the conglomerate unit is marked by a rippled siltstone parting surface. We interpret these sediments as an alluvial fan debris flow deposit.

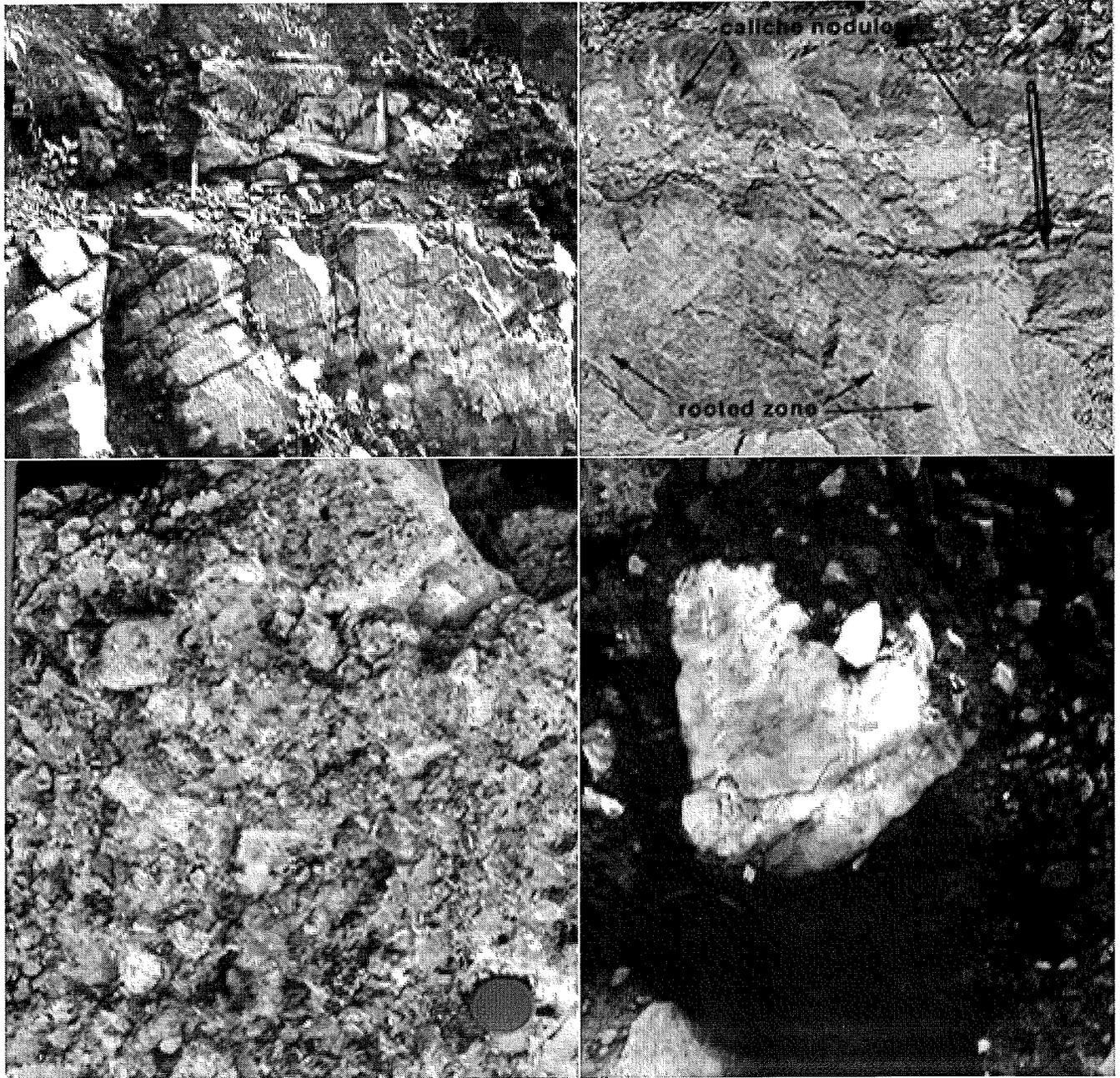


Figure 16. Upper Cass Formation at Southbury Quarry.

- A. (top left). Contact of Cass Formation eolian dunes and upper lacustrine re-worked eolian sandstone with overlying, vesicular Orenaug Basalt. Note well-developed pillow resting on contact at upper center of photo.
- B. (top right). Paleosol horizon below the eolian interval, showing well developed rooted zone and caliche development.
- C. (bottom left). Alluvial fan conglomerate about 6.5 m below contact with Orenaug Basalt. Note clast-supported framework. Dominant clast types are vein quartz and blades of garnet schist. Mottled appearance from secondary clay minerals derived from weathering schist clasts.
- D. (bottom right) Large clast of vein quartz, ~6 m below Orenaug Basalt, from formerly-exposed alluvial fan sequence.

The arkosic debris flow deposit is sharply overlain by 1.5 m of trough and planar cross bedded, fine-grained, subarkosic that contains a prominent, 4 cm-thick conglomerate zone near the base that is composed of flat, ellipsoid siltstone intraclasts (up to 5 cm in length) and a lesser percentage of extrabasinal pebbles. Above this unit is a 0.6 m-thick, trough cross bedded, subarkosic/sublithic arenite that contains abundant, grayish green (5G 5/2) nodules of pedogenic carbonate. The middle of the unit is punctuated by a prominent siltstone parting surface that displays asymmetric ripple marks spaced 25 cm apart, crest to crest (at the 6.2 m level on Figure 3).

The following 0.8 m of strata consist of siltstone and arkosic sandstone and conglomerate. The base of the unit is a mud cracked, siltstone parting surface that contains abundant, though poorly preserved, reptile trackways referable to the ichnogenera *Batrachopus* sp., *Grallator* sp., and *Eubrontes* sp., as well as meter length casts and impressions of equisetalian and small leafy shoots of conifer foliage (cf. *Brachyphyllum* sp.). Above the parting surface, sandstone coarsens upward into a 12 cm thick zone of pebble conglomerate that contains abundant, angular pebbles and small cobbles of metamorphic carbonate, which we believe were derived from the Stockbridge Marble (at 6.8 m level on Figure 3). The top 10 cm of the unit are composed of fine grained sandstone and sandy siltstone that are mottled grayish green, and contain spectacular rhizoconcretions and caliche nodules indicative of semi-arid depositional conditions on the former floodplain (Figure 16B). The caliche paleosol horizon is, in turn, overlain by a noteworthy 2 to 3 m thick layer of trough cross bedded, eolian quartz sandstone (Figure 16A) that is capped by a thin interval of ripple-laminated, pebbly, reworked eolian sandstone and occurs just below the undulating, pillowed contact with the Orenaug Basalt (LeTourneau and Huber, 1997 and in review).

**Significance of the eolian sandstone.** While common in the Fundy Basin, eolian sandstones have been regarded as notably scarce or absent in the Hartford-Deerfield and Newark basins. Smoot (1991) identified a thin, meter-scale eolian sandsheet in the upper New Haven Formation of the Hartford basin, and the ongoing work of Letourneau (2002) in the Hartford basin has revealed significant eolian sandstones in the lower Jurassic

Portland Formation at the Portland, Connecticut brownstone quarry and in the upper New Haven Formation /lower Shuttle Meadow Formation at Newgate Prison State Park in Granby, Connecticut. Identification of eolian sandstone is important for reconstructing early Mesozoic paleoclimate. Recognition of the eolian sandstone is based on comparison of sedimentary structures and bedding features (Figure 17A-C) with analogs from both modern environments and ancient dune deposits. Important criteria that indicate an eolian origin for this sandstone at Stop 7 include:

- 1) inverse grading of thin beds and laminae comparable to pinstripe lamination (Fryberger and Schenk, 1988), and subcritically climbing translent strata (Hunter, 1977; Kocurek and Dott, 1981);
- 2) sandflow cross strata and grainfall laminae (e. g. Hunter, 1977; Schenk, 1990);
- 3) high-index ripples with coarser grains near the ripple crests (e. g. Fryberger et al., 1979; Fryberger and Schenk, 1988);
- 4) tabular, wedge-shaped planar and trough cross-stratification with high- to low-angle laminae (e. g. McKee and Weir, 1953; Kocurek and Dott (1981); and
- 5) High sorting, high porosity and high permeability (Ahlbrandt, 1979).

A remarkable aspect of the Pomperaug basin eolian sandsheet is its small thickness relative to its wide aerial distribution. The eolian sandstone is traceable from Platt Farm Park to the Southbury Quarry north and east to the Woodbury Quarry (Stops 7 and 8), a distance of about 12 km, indicating that the eolian dune field was widespread across the basin, and not a local feature. Thin, but widespread dune fields are observed in many of the modern valleys of the Basin and Range Province of the western U.S. In particular, the Stovepipes Wells dune field of Death Valley and the dune fields of the Panamint Valley bear a striking resemblance to the Pomperaug basin dune field.

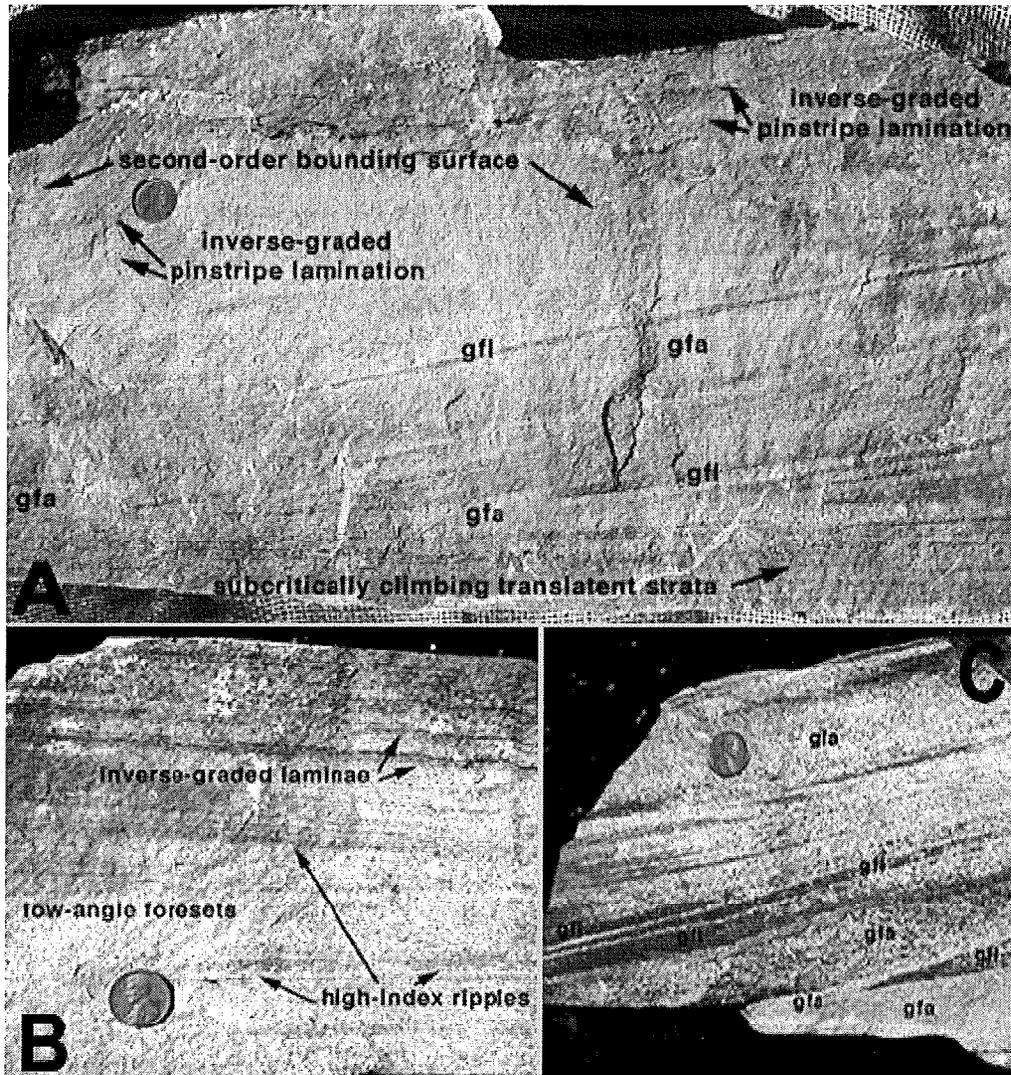


Figure 17. Eolian sandstone of the upper Cass Formation.

- A) Block of sandstone from the O&G Southbury quarry showing eolian sedimentary structures including toesets of small dunes. Coarse normal to reverse graded grain-flow (gfl) wedges are interbedded with massive to laminated grain-fall (gfa) wedges. Grain-flow (gfl) layers are formed by small avalanches of sand on the lee-side dune face; grain-fall (gfa) layers are formed by wind-blown sand streaming across the dune brink-point and settling on the lee-side foresets. Inverse-graded pinstripe laminations and subcritically climbing translent strata are formed by migrating ripples. Porosity of this unit is high....
- B) Block of sandstone from the O&G Southbury quarry showing inverse-graded wind-ripple laminae, low-angle foresets, and isolated high-index ripples typical of modern and ancient eolian deposits.
- C) Block of sandstone from the O&G Southbury quarry showing grain-flow (gfl) and grain-fall (gfa) dune toesets. The occurrence of these distinctive features is diagnostic of eolian deposits.

Laboratory measurements of samples of the Pomperaug eolian sandstone revealed porosities ranging from 15 to 20 %. Gas permeability reached values as high as 395 millidarcies (mD) and liquid (Klinkenberg) permeability up to 368 mD. The relatively high porosity and permeability values support the high sorting and grain support framework observed in hand samples. The porosity and permeabilities of the eolian beds indicate that the horizon may be a significant aquifer or pathway for contaminated groundwater. Furthermore, due to its location directly beneath the Orenaug Basalt it has a significant potential to be a semi-confined, artesian aquifer, similar to an artesian occurrence in the Hartford basin (observed in the field by LeTourneau).

Reset odometer at quarry entrance, and turn left onto Route 67, heading east.

- 0.1 Cross Pomperaug River.
- 0.3 Turn left onto Route 6 East.
- 2.6 Intersection with Route 64 East (Sherman Hill Road). Continue straight, entering village of Woodbury.
- 3.1 Take right onto Orenaug Avenue, at south end of village center. Pass over ridge of Orenaug Basalt, then pass Orenaug Park on left.
- 3.5 Unmarked entrance on left to O&G Woodbury trap rock quarry. Proceed slowly into quarry, keeping an eye out for trucks.

**STOP 7. Fault-offset Cass Formation in O&G Woodbury Trap Rock Quarry.** (45 MINUTES) This quarry exposes a contact between eolian sandstone of the Cass Formation and overlying Orenaug Basalt that has been offset along a N-S-trending, high-angle normal fault, which until recently was perfectly exposed along the west quarry wall. Our first stop in the quarry is at a small pavement of the sandstone atop the west quarry wall, which marks the western, upthrown side of the east-side-down fault. The color of the sandstone -- normally white, but here an orange hue--is due to thermal alteration produced during the flow of lava over this sediment. Walk carefully over to the linear pile of rubble that marks the quarry wall edge, and look towards the north end of the quarry. Visible at the bottom of the quarry near the north end is another, larger sandstone pavement, marking the same horizon but displaced downward some 30 meters and rotated into a more easterly strike. The fault, which runs north-south virtually under our feet at this spot, is marked by a one to two-meter-wide, planar zone of extensive brecciation and mineralization, and secondary growth of quartz, calcite, stilbite, and other minerals, as well as bitumen. Outcrop-scale slickenlines on the mineralized fault surface plunge about 25 degrees north (Fig. 15), indicating that there was a left-lateral strike-slip component to the fault movement, as discussed in the main text.



Figure 18. Mineralized fault zone exposed in west wall of O&G Woodbury quarry that offsets bedding pavement of Cass Formation. Note slickenlines dipping moderately to the right (north), indicating strike-slip component to down-to-the-east movement sense.

Driving down into the central part of the quarry, we pass the wall to our left that contains the fault exposure, now covered in rubble, and head north to the large sandstone pavement. This pavement was also better exposed previously, but most of the salient features are still visible underneath the gravel. These include ripple marks that

indicate a NE-SW current direction, bowl-shaped impressions left by lava pillows that formed as the lava flowed over wet sediment, and small offsets in the bedding surface caused by minor, mineralized normal faults. This sandstone likewise shows discoloration due to thermal alteration by the overlying lava, which is still preserved as small patches of a dark contact rind, particularly in the bottoms of the pillow impressions.

Proceed back out the quarry, and from quarry entrance turn left on Orenaug Avenue.

- 0.1 At stop sign, take left onto Bacon Pond Road.
- 0.2 Turn left into unmarked eastern entrance of quarry, proceed to end of drive marked by large boulders.

**STOP 8. Contact Between Orenaug Basalt and Cass Brook Formation.** (10 MINUTES) This ledge exposes Orenaug Basalt in contact with underlying fine-grained sandstone of Cass Brook Formation, the latter which is brownish-weathering, greenish-gray fresh, medium to fine-grained, locally pebbly sandstone, similar to sedimentary rock exposed in quarry. Here sandstone and basalt/sandstone contact are gently west-dipping and help to define east limb of Woodbury hanging-wall syncline (Fig. 1). Note N-S trending, steeply east-dipping curvilinear fault surface in sandstone with subhorizontal slickenlines. NNE-trending eastern border fault of basin passes along road or edge of woods just to the east. Hills beyond are underlain by schist of Taine Mountain and basal Taine Mountain formations (Ot and Otb, respectively, Fig. 1).

Turn around and turn right on Bacon Pond Road, heading south.

- 0.3 Continue straight through four-way intersection.
- 0.6 Continue straight through after stop sign, curving left.
- 0.7 Intersection with Old Sherman Hill Road (Route 64). Proceed through intersection and take right fork onto Middle Quarter Road.
- 1.0 Stop sign and one-lane bridge over South Brook. Cross bridge and pull over on the left into gravel parking area. Walk back over bridge and take right up gated gravel road on north side of South Brook. The road follows the bed of an old trolley that once ran between Woodbury and Waterbury. About 100 meters up road to the right (south) is flush-mounted USGS monitoring well CT-WY-87. Core extracted during the drilling of this well is of massive fine-grained basalt, which geochemical analysis shows to be Orenaug Basalt.

**STOP 9. Contact Between White Oaks Formation and South Brook Basalt.** (1 HOUR) Small exposures in E-W reach of stream are of blocky, fractured, gently E-dipping pebbly sandstone in contact with overlying vesicular basalt. Just upstream begins continuous exposure of this basalt. Because of its vesicular texture, the basalt superficially resembles East Hill Basalt and was originally mapped as such, since the exposed section here is roughly equivalent in thickness. However, geochemical analysis of a sample from this exposed contact in the stream, plus a more altered sample from upstream near the border fault, shows it to be chemically distinct from either Orenaug or East Hill basalts (Fig. 10). We therefore feel justified in calling this the "third basalt", in agreement with the interpretation on the Bedrock Geological Map of Connecticut (Rodgers, 1985), and have named it the South Brook Basalt (Fig. 1). The underlying sedimentary rock is named the White Oaks Formation, after a crossroads halfway between Woodbury and Southbury. The only other known exposure of this rock, now gone, was reported by Hobbs (1901) adjacent to an oil well drilled near the village of Southbury, about five kilometers to the south, although its existence is also suggested in well logs from the White Oaks area (Fig. 1), and blocks of black shale of probable White Oak affinity were recovered from a gravel quarry in Southbury (Fig. 19).

Proceeding farther upstream along road is another USGS monitoring well, CT-WY-86, on the left (northeast) side of the road where it bends to the south, following the brook. This well cored pegmatite and schist at about 20 feet, representative of the footwall of the eastern border fault. Fine-grained mica schist of the Taine Mountain Formation (Ot, Fig. 1) is also exposed in a small outcrop just south along the east side of the road. Scramble down hill towards brook, where along this N-S reach of the brook are more outcrops of vesicular basalt. This rock is brecciated and highly altered due to its very close proximity to the eastern border fault, which parallels and probably underlies the brook. The basalt contains abundant secondary interstitial silica, chlorite and calcite, as well as two copper prospects, one visible here near north end of ridge and one about 100 meters south. Exposed at one spot along this stretch of outcrops, between the two copper prospects, is a gently-west-dipping contact between South

Brook Basalt and underlying White Oaks conglomerate. This is the eastern limb of the South Brook hanging-wall syncline, comprising the two formations, which has been extrapolated southward based on well records (Fig. 1).



Fig. 19. Block of black shale (White Oak Fm.?) recovered from a quarry in Southbury

Clasts in the conglomerate include rounded pebbles of quartz, plus elongate chips of garnet mica schist and siltstone that suggest a proximal source. The elongate schist fragments likely represent material shed off the scarp of the eastern border fault, just to the east. The fault crosses underneath the brook and intersects the ridge towards the south, where fine-grained mica schist of the footwall is exposed.

#### **End of Trip**

To return to original meeting area, go south on Middle Quarter Rd. to intersection with Route 6, just ahead, and take left on Route 6. Proceed 3.1 miles south on Route 6 to Southbury village center, take right at second light on Main Street South, opposite Southbury Plaza, go 1.8 miles west to intersection with Route 172; park and ride is on the left just before intersection.

**APPENDIX 1. WHOLE-ROCK CHEMICAL ANALYSES OF POMPERAUG BASALTS**  
(see Appendix 2 for descriptions of samples and localities).

## Major Elements in weight percent

Sample	Type	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	P <sub>2</sub> O <sub>5</sub>	LOI
RS-1	1	47.4	1.09	13.5	10.4	0.15	9.12	8.34	1.41	0.14	0.14	6.99
RS-2	1	51.9	1.03	13.1	9.54	0.18	7.08	8.92	1.37	0.20	0.13	5.03
RID-1	1	42.2	1.07	13.7	10.7	0.14	9.30	6.79	3.48	0.21	0.17	11.4
WT-1	1	46.8	1.37	16.1	10.4	0.15	8.90	1.56	4.01	0.72	0.19	8.42
SQ-1	2	51.1	1.06	13.7	13.8	0.22	5.71	9.48	2.67	0.50	0.16	1.51
SQ-2	2	50.0	1.05	13.2	13.4	0.18	5.31	7.94	3.91	0.57	0.15	3.00
SQ-3	2	41.8	0.92	12.6	12.1	0.29	4.53	10.5	4.34	1.07	0.14	10.7
ORE-1	2	51.6	1.06	13.4	13.6	0.22	5.47	9.99	2.01	0.29	0.15	1.38
PF-2	2	51.9	1.03	13.4	13.4	0.23	5.77	6.64	4.23	0.65	0.15	1.76
SBH-1	2	50.6	1.03	13.5	13.8	0.23	5.94	9.85	2.07	0.56	0.15	1.52
ST-1	2	50.9	1.04	13.2	13.3	0.20	5.55	8.20	4.12	0.34	0.15	2.76
ORE-2	2	51.4	1.05	13.5	13.4	0.22	5.49	10.0	2.00	0.26	0.15	1.90
PB057	2	51.1	1.04	13.0	13.1	0.21	5.68	6.30	4.25	1.02	0.14	3.14
SB-2	2	51.6	1.02	13.5	13.5	0.22	5.69	10.2	1.93	0.28	0.15	2.05
WY-87	2	52.7	1.07	12.9	13.1	0.29	5.21	3.68	1.33	6.94	0.14	2.78
SB-1	3	43.8	1.53	14.8	14.6	0.22	7.88	11.4	0.28	6.64	0.19	6.54
PB060	3	42.6	1.47	14.7	14.9	0.34	6.88	3.69	0.79	5.73	0.18	8.50

## Trace Elements in Parts Per Million

Sample	Type	Ba	Ce	Cr	Cs	Cu	Ga	La	Mo	Nb	Nd	Ni	Pb	Rb	Sr	U	V	Y	Zn	Zr
RS-1	1	139	23	221	81	17	9			7		82			177		230	23	62	94
RS-2	1	148	24	195	89	15	7			7		75	3	3	219		213	22	51	86
RID-1	1	82	23	405	109	15	11			8	13	115	3	8	107	10	272	19	142	64
WT-1	1	439	21	387	85	19	9	2	9	13	88		23	160	5	259	19	113	101	
SQ-1	2	114	19	21	66	18	10	3	6		33	4	30	162		293	29	94	88	
SQ-2	2	92	22	20	72	20	10	3	6		34	5	23	94		305	29	95	84	
SQ-3	2	52	22	23	57	15	11	2	4		34	6	35	70	7	270	37	89	93	
ORE-1	2	175	20	21	13	72	18	9	3	5	30	6	31	188		273	29	94	89	
PF-2	2	47	21	21	61	16	9	4	6		38	5	29	536		288	28	96	87	
SBH-1	2	130	19	23	71	19	9		6		35	4	18	193		288	29	93	85	
ST-1	2	38	20	20	64	13	11	3	6		28	4	12	96		289	29	96	86	
ORE-2	2	119	22	21	16	72	17	9	6	3	53	2	8	182		271	29	91	87	
PB057	2	162	19	21	55	16	9	2	6		34	4	47	224		263	29	74	89	
SB-2	2	264	20	20	8	78	19	10	2	5	37	5	23	184		276	30	93	83	
WY87	2	206	24	15	48	14	10	3	6		20	5	457	603		279	28	70	93	
SB-1	3	293	19	95	208	19	6		5		69		415	337	4	416	33	70	110	
PB060	3	556	17	96	258	16	9	3	7	11	40	3	341	80		371	35	243	108	

Type 1 = East Hill Basalt ; Type 2 = Orenaug Basalt; Type 3 = South Brook Basalt

Major elements by USGS WDXRF, analyst T. Hannah, 20 September 2004. Trace elements by USGS EDXRF, analyst T. Hannah, 26 July 2004 and 9 September 2004.

## APPENDIX 2.. POMPERAUG BASIN BASALT SAMPLES FOR CHEMICAL ANALYSES

Sample	Date	Location Notes	Latitude deg min sec	Longitude deg min sec	Elev. feet	Strata
RS-1	7/17/2003	Red Spring, spheroidal ball, bank above the spring	41 28 49	073 14 32	416	East Hill
RS-2	7/17/2003	Red Spring, solid stream outcrop up north branch	41 28 52	073 14 33	436	East Hill
RID-1	7/10/2003	Ridge basalt, in bench below Rattlesnake Hill	41 28 13	073 14 55	252	East Hill
WT-1	6/6/2003	White Trail, near drill hole So. of Rattlesnake Hill	41 28 11	073 14 41	305	East Hill
SQ-1	5/3/2003	Southbury Quarry, upper member on NE face	41 30 51	073 12 54	400	Oreanaug
SQ-2	5/3/2003	Southbury Quarry, middle member on NE face	41 30 50	073 12 55	390	Oreanaug
SQ-3	5/3/2003	Southbury Quarry, lower member W. side of pit	41 30 46	073 12 57	354	Oreanaug
ORE-1	5/9/2003	Oreanaug Park, base of hill east of pond at south end	41 32 10	073 12 14	318	Oreanaug
PF-2	6/6/2003	Platt Farm, low wall, upper trail N of parking lot	41 28 24	073 14 38	364	Oreanaug
SBH-1	5/9/2003	South Britain Hill, 3 m below South end top of hill	41 28 13	073 14 44	367	Oreanaug
ST-1	6/6/2003	Sawtooth Road near Rte. 67 E. of Bates Rd	41 30 15	073 14 08	370	Oreanaug
ORE-2	5/9/2003	Oreanaug Park along so. trail near top of hill	41 32 26	073 12 13	462	Oreanaug
PB057		not described				Oreanaug
CT-WY-87		South Brook, 75.5 ft in WRD core from well site about 100-200 ft W of PB060				Oreanaug
SB-2	5/29/2003	South Brook, 150 m from gate (stream bank boulder?)	41 31 28	073 11 58	275	Oreanaug
SB-1	5/29/2003	South Brook, 300 m from gate, eastern stream bank	41 31 25	073 11 52	298	South Brook
PB060		South Brook, immediately above the conglomerate contact in the E-W segment of the brook				South Brook

# WESTERN END OF THE HONEY HILL FAULT ALONG THE EASTERN BANK OF THE CONNECTICUT RIVER

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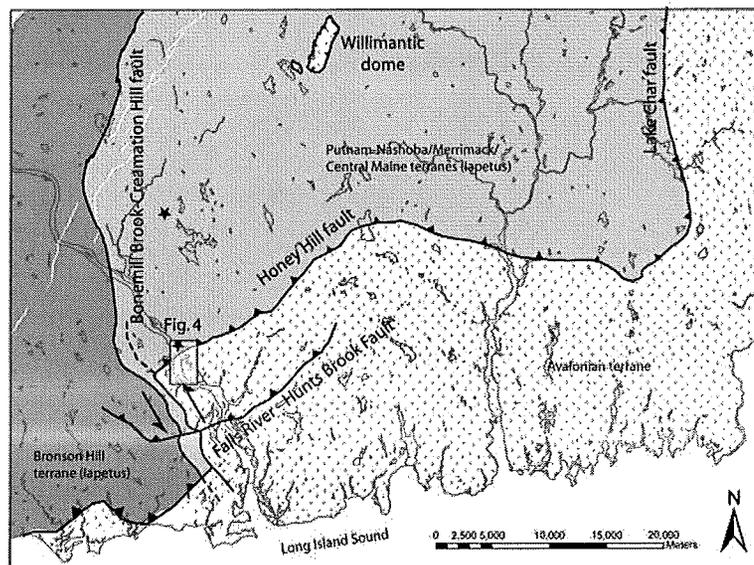
## BACKGROUND

### Introduction

Lundgren et al. (1958) first identified the Honey Hill fault in southern New England (Fig. 1) as a major thrust fault with significant offset. However, Lundgren (1963) also recognized that the structural development of the region was more complex, noting that the evidence for major thrusting came primarily from the Norwich quadrangle (Snyder, 1961) while in the Deep River quadrangle (the location of this field trip) the Honey Hill fault was dominated by retrograde top-to-the-northwest motion.

Lundgren's initial work predated the plate tectonic revolution, however, Wilson (1966) in his seminal paper on the closing and re-opening of the Atlantic proposed that the Honey Hill fault might be the trace of a former Lower Paleozoic ocean. Lundgren rejected the idea (Lundgren and Ebbelin, 1972; Lundgren and Ebbelin, 1973) and it remained until the late 1970's and early 1980's with the concept of suspect terranes before the Honey Hill fault was accepted as a major terrane boundary (e.g. Williams and Hatcher, 1983).

The complexity of this boundary as well as its tectonic significance has led to a debate over the exact nature of the Honey Hill fault that has lasted almost 50 years. In the following contribution we summarize the previous work on the zone highlighting some of the past and present controversies. We also briefly discuss the present day stress and strain of the region. The associated field trip visits outcrops that expose spectacular ductile and brittle fault features associated with the Honey Hill Fault zone that should stimulate a lively discussion regarding the nature of this fault.



**Figure 1.** Geologic terrane map of southeastern Connecticut. Teeth on faults illustrate hanging wall block, but do not necessarily indicate thrust motion. In many localities faults have been active as both thrusts and detachments. Box in lower left quarter shows location of field trip and figure 4. Stars are approximate borehole locations (upper – Moodus, lower – Gillette Castle) Based on state geologic map (Rogers, 1985) modified based on mapping by Wintsch (Wintsch et al., 1990; Wintsch et al., 2003) and London (London, 1988).

## Regional Geology

The area of southern Connecticut near the mouth of the Connecticut River was first mapped at a large scale by Lundgren (1963; 1964; 1966; 1967). The area is underlain by upper amphibolite facies schists, gneisses, and calc-silicates. Lundgren (1963) described the major structural pattern of the area as a series of gneiss-cored domes separated by schists and calc-silicate gneisses that he interpreted as recumbent folds. Within this general pattern he identified one major fault, the Honey Hill fault, a 1500-ft thick zone of retrograde "blastomylonitic" gneiss. He located the fault surface at the contact between the overlying Putnam Gneiss and the underlying Monson Gneiss and noted that the fault continued more than 25-mi to the east. To the west the fault continues to Chester where he interpreted it to bend toward the south due to folding. More recent studies have traced the Honey Hill fault system to the east where it connects to the northerly trending Lake Char fault (Fig. 1), into the subsurface where it has been drilled in the Moodus (at 809-m depth) and Gillette Castle (at 221-m depth) wells (Goldberg, 1992) and to the north where it surrounds the Willimantic Dome (Wintsch, 1979). Subsequent work has also led to reinterpretation of many of Lundgren's recumbent folds as ductile fault contacts between map units (London, 1988; Wintsch et al., 1990). Wintsch and coworkers (e.g. Webster and Wintsch, 1987; Wintsch and Aleinikoff, 1987; Wintsch et al., 2003) have used geochemistry and geochronology to identify litho-tectonic units with distinct protoliths and tectonic histories. This work has recently led to identification of Gander terrane rocks in the area (Wintsch, 2005) and will ultimately lead to new maps of the region (Wintsch, 2005 personal communication).

We interpret the following general structural chronology from field observations and published reports for the region. The earliest ductile fabrics which include upright north-south trending folds are cut by pegmatites that are inferred to be Permian in age (~260 Ma, e.g. de Boer and Brookings, 1972). The pegmatites are deformed by westward-vergent folding (London, 1988; Lundgren, 1963) suggesting east-west contraction. These folds are in turn overprinted by structures associated with largely north-south contraction including left-lateral slip on the Cremation Hill fault zone (London, 1988) and reverse-slip on the Honey Hill fault (Wintsch et al., 1992). These phases of deformation all occur under amphibolite facies conditions. Overprinting structures are associated with retrograde metamorphism and ultimately brittle deformation. These structures including widespread doming (Lundgren, 1963; Lundgren and Ebbin, 1972; Wintsch et al., 2003), ductile décollement along bedding planes associated with the Honey Hill fault zone which spans greenschist to brittle conditions (Goldstein, 1989, 1994; Lundgren, 1963), and subsequent left-lateral brittle-ductile to brittle strike slip faulting on steep northwest trending faults (London, 1988).

## Tectonic Significance

Although most workers appear to agree on the general structural chronology for the area it is the relative importance of these events, and thus the tectonic significance of the Honey Hill fault zone and the broader Alleghenian orogeny that has been the topic of heated debate since the fault was recognized as a major terrane boundary. The two general interpretations are that the Honey Hill fault zone is a major thrust fault associated with Alleghenian docking of the Avalon terrane with Laurentia (e.g. Wintsch and Sutter, 1986) or that the zone is primarily a décollement associated with either orogenic collapse (Getty and Gromet, 1992) or a releasing bend in an orogen parallel strike-slip system (Goldstein, 1994) with the Avalon terrane arriving during Acadian (Devonian) orogenesis. The ongoing arguments focus on the relative importance of thrusting vs. detachment faulting and the interpretation of the geochronological data.

Following Lundgren et al. (1958), Wintsch (e.g. Wintsch et al., 1990; Wintsch et al., 2003; Wintsch and Sutter, 1986) has interpreted the Honey Hill fault zone as a major thrust fault associated with the docking of the Avalon terrane during the Pennsylvanian/Permian Alleghenian orogeny. Wintsch and Sutter (1986) cited the juxtaposition of higher grade rocks over lower grade rocks (all amphibolite facies), imbrication of the Tatnic Hill Fm., NE-SW trending open folds, SE trending

mineral lineation, the rotation of boudins and porphyroblasts and the SE trend of tear faults as evidence for thrusting along many of the ductile faults in southeastern Connecticut.

Goldstein (1989; 1994) and Getty and Gromet (1992) interpreted many of the same faults as principally detachment faults with top-to-the northwest (normal-sense) motion. Goldstein (1989) points out that all of Wintsch and Sutter's (1986) lines of evidence for thrusting are ambiguous except for the rotation of boudins and porphyroblasts for which they provided no evidence in their paper. Goldstein (1989) demonstrates the abundance of normal-sense shear indicators within the Lake Char – Honey Hill fault system and states that no examples of thrust-sense indicators were found anywhere in the fault zone. Furthermore, Goldstein (1989) states that the shear zone clearly records top-to-the-northwest slip over a range of metamorphic conditions from lower amphibolite to greenschist to brittle-ductile conditions (nicely illustrated at stop 3). This sequence suggests unroofing associated with the décollement. Goldstein (1994) interprets the Alleghenian orogeny as two-phase with early northward directed shortening associated with the formation of synistral shear zones followed by a period of dextral faulting at which time he interprets the Honey Hill – Lake Char system to have formed as a normal fault within a releasing bend.

The two models for the development of the Honey Hill fault have very different implications for the timing of prograde metamorphism, deformation, and the arrival of Avalon. Geochronological studies have thus become a critical tool for evaluating models of the Alleghenian orogeny. Alleghenian mineral cooling ages have long been recognized in southern New England (e.g. Clark and Kulp, 1968). These ages could be interpreted as evidence of Alleghenian metamorphism or as delayed cooling from Acadian (Devonian) metamorphism. U-Pb sphene ages have the ability to more directly record deformation and metamorphism (e.g. Resor et al., 1996) since the mineral has a relatively high closure temperature (~700°C) and grows under a variety of metamorphic conditions, often below closure temperature (Cherniak, 1993; Frost et al., 2001). Getty and Gromet (1992) used multiple isotopic systems, including U-Pb sphene ages as well as microstructural observations to argue for complete recrystallization and isotopic resetting during Pennsylvanian-Permian (304±Ma) retrograde mylonitization in detachment-related fault rocks of the Willimantic dome. They also noted, however that there was no evidence for Acadian metamorphism in the Avalonian basement below the detachment so that juxtaposition of the terranes is likely Alleghenian. Wintsch et al. (2003) and Aleinikoff et al. (2002) recently reported Pennsylvanian and Permian U-Pb sphene (ca. 305-265 Ma) ages from the Bronson Hill terrane (north and west of the Honey Hill – Lake Char fault system). Wintsch et al. (2003) state that the sphene grew during prograde metamorphism (without presenting microstructural or metamorphic evidence) and thus interpret the sphene in combination with amphibole and white mica  $^{40}\text{Ar}/^{39}\text{Ar}$  ages as evidence for a major prograde metamorphism associated with thrust stacking from the south during the Alleghenian orogeny.

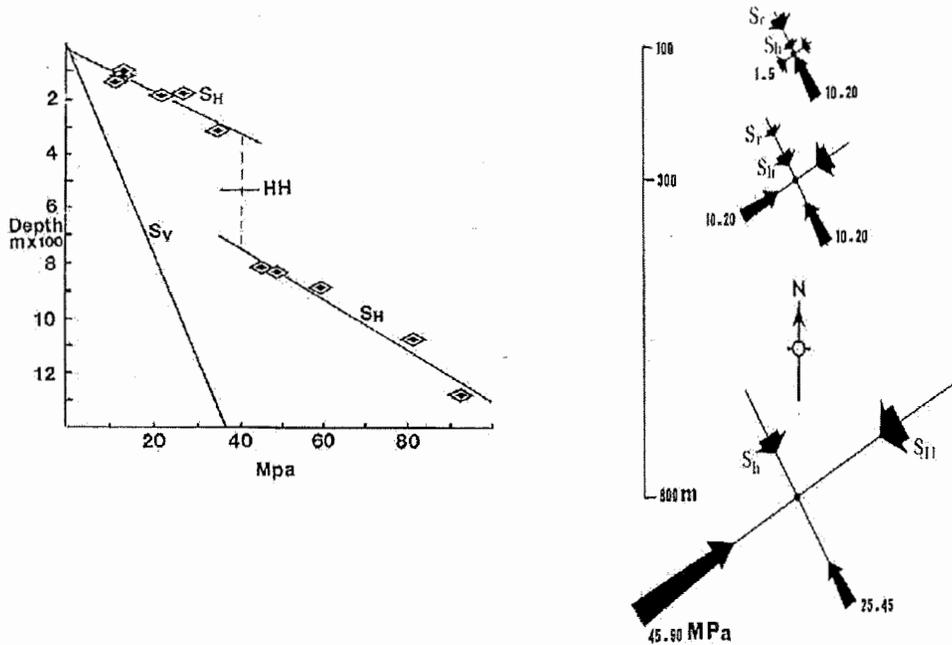
Future studies that generate hypotheses based on these two models and then test them may help to better define the tectonics of the Alleghenian orogeny. The most likely approach appears to be integrated structural, metamorphic, and geochronological studies as paleomagnetic studies provide only limited constraints due to primarily longitudinal motion of Avalonia during the time period in question (Tait et al., 1996).

### **Present-Day Stress and Strain**

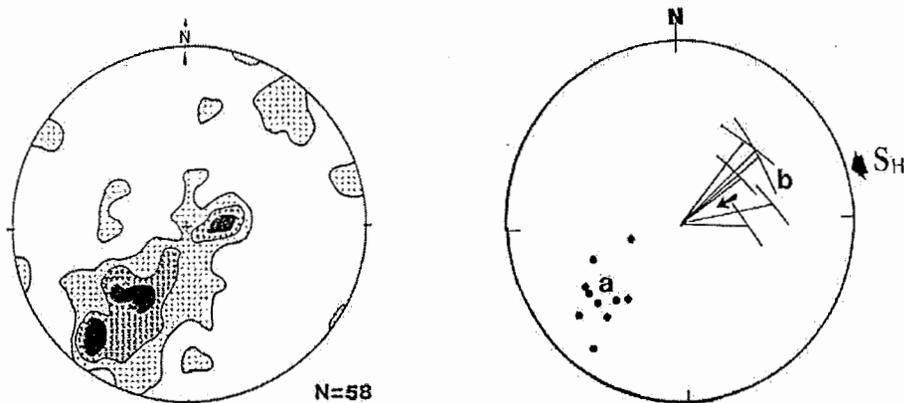
In addition to playing an important role in Paleozoic tectonics, the Honey Hill fault also appears to be of significance to the neotectonics of southern New England. South-central Connecticut is one of the more seismically active areas in the northeast (Ebel and Kafka, 1991) and the nearby Moodus area has had several earthquake swarms, most recently in the 1980's (Ebel, 1989).

Two research wells drilled in the area – at Gillette Castle to 500m total depth (TD) and near Moodus, CT to 1458-m TD - provide unique insights into the neotectonics (Fig. 1). In the Moodus hole it was found that the Honey Hill fault (~810-m) marked a distinct change in stress magnitude and orientation (Fig. 2) (Fang, 1989, and references therein). Above the Honey Hill fault zone maximum horizontal stress was oriented NNW, while below the fault the maximum principal stress

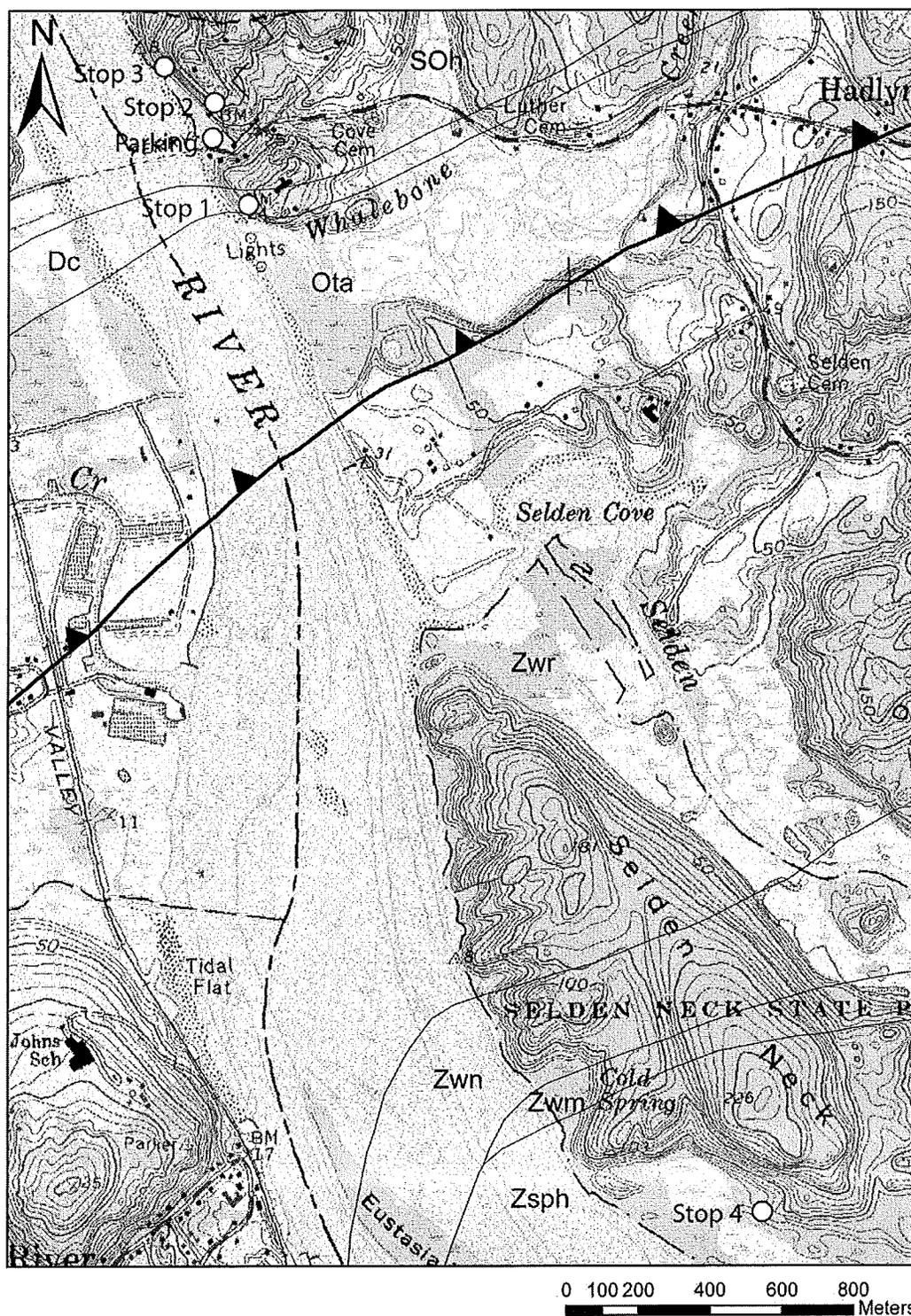
is oriented WSW. The deep stress is consistent with ridge push forces while the shallow stress may reflect a glacial or post-glacial effect (?). During construction of a nearby freeway (CT 11) shallow stresses were sufficient to cause reverse-sense slip along north-dipping foliation planes in the newly blasted median (Block et al., 1979). In the Gillette Castle borehole there were no breakouts, so stress orientations could not be determined, however there were abundant fractures (Fig. 3) and a family of NW-striking, westward dipping fractures with slickenlines indicating NE-SW slip. These features are consistent with slip in response to the deeper stress field (ridge push).



**Figure 2.** Left: Plot of stress magnitude with depth from the Moodus borehole. Magnitudes were determined through hydrofracture experiments. Modified from Woodward-Clyde (1988). Right: Illustration of horizontal principal stress orientations and magnitudes at three representative depths for southern New England. Magnitude data determined from hydrofracture experiments in Moodus area boreholes. Orientation data compiled from regional well and outcrop data. Modified from de Boer and Fang. Note different vertical scales.



**Figure 3.** Equal-area stereoplots of fractures from Gillette Castle borehole. Left: Contour plot of poles to all fractures observed using a borehole televiewer. Right: Fractures with slickenlines from a) Poles to fracture planes. b) slip vectors (lines from center) with their associated great circles (curved segments). Modified from Fang (1989).



**Figure 4.** Geologic map of field trip area. Field trip stops described in text. Nashoba/Merimack/Central Maine (Iapetus) formations: Ota – Tatnic Hill Fm., Dc – Canterbury Gneiss, Soh – Hebron Gneiss; Avalonian terrane formations: Zw – Waterford Group, Zwr – New London Gneiss, Zwn – Mamacoke Fm., Zsph – Potter Hill Granite Gneiss. Honey Hill fault is toothed bold line separating Ota from Zwr. Geology from Rogers (1985). USGS Topo base.

## TRIP LOG

The entire trip will be conducted on foot and boat from the parking lot at the eastern end of the Chester-Hadlyme ferry (Fig 4).

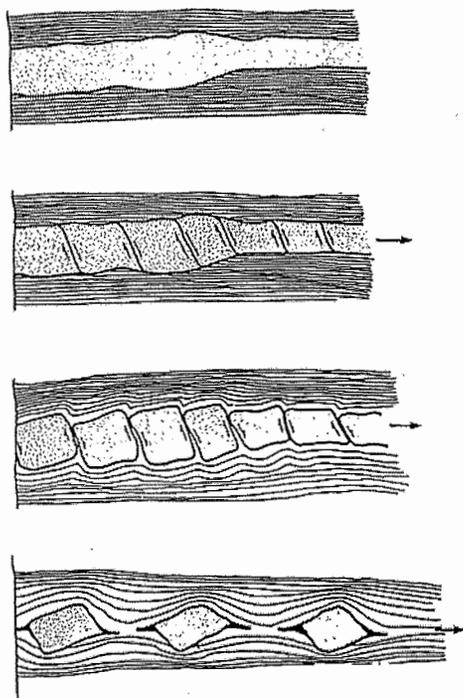
**Approximate Walking Distance (m)**

- 0 From ferry parking lot walk south along east bank of the Connecticut river approximately 100 m. Follow old road bed diagonally up hill to abandoned quarry.  
200 Arrive at old quarry location (end of road bed).

**Stop 1. Quarry in hanging wall of Honey Hill Fault (30 minutes)**

**NOTE: This stop is on private property. Permission must be granted by owner before trespassing. Please respect private property rights so that this stop remains accessible to future groups.**

The quarry exposes a spectacular example of a boundinaged pegmatite vein within gneisses mapped as the Devonian Canterbury Gneiss (Rogers, 1985) or Ordovician Hebron Gneiss (Lundgren, 1963). The two quarry walls are nearly orthogonal and approximately parallel to the axes of the finite strain ellipse within the foliation plane. A biotite lineation can be seen on the foliation planes trending  $\sim 285^\circ$ . This lineation is interpreted as the maximum stretching direction. The southwest facing wall of the quarry clearly shows a greater degree of stretching than the south east facing wall. The stretching on the southwest facing wall is accommodated by brittle-ductile boudinage with fracturing of the pegmatite into blocks and block rotation within the more micaceous matrix (Fig. 5). The block rotation suggests normal-sense (top to the NNW) motion. The southeast facing wall shows more classical ductile pinch and swell of the pegmatite with roughly symmetric tails. Principal stretches within the foliation plane are estimated to be 1.7 and 1.5, indicating a bulk general shear.



**Figure 5.** Sketch illustrating progressive development of faulted boudins. From top to bottom. 1) Pegmatite was intruded ( $\sim 260$  Ma?) parallel to foliation. 2) Initial layer-parallel extension leads to faulting in relatively stiff (and brittle) pegmatite. 3) With fault slip blocks begin to rotate. Surrounding gneiss deforms ductilely. 4) With continued extension and top to NW shear blocks are separated, rotate, and develop sigmoidal tails of quartz overgrowths. Modified from Chen (1990).

- 200 Return to Hadlyme ferry.
- 400 Turn right and walk east along Ferry road.
- 500 Head left up old roadway (now hiking path) toward Gillette Castle. The roadway is lined by stone walls.
- 600 Turn left onto hiking trail with wooden railing.
- 610 Veer left off trail (duck under railing) to low west-facing outcrop at top of steep slope.

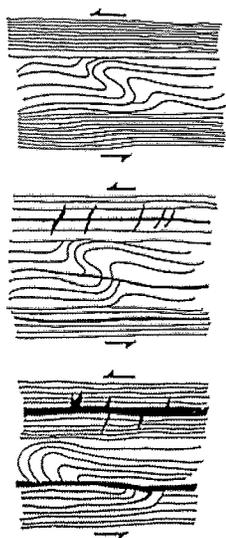
**Stop 2. Thrusting within the Honey Hill Fault zone. (10 minutes)**

This outcrop beautifully illustrates shortening in a roughly up dip direction (top to the SSE) in the hanging wall of the Honey hill fault. Pegmatitic layers within the Hebron Formation (Rogers, 1985) are folded with an updip vergence and are locally duplicated across brittle-ductile shears.

- 610 Return to path. Continue downhill (north) on path.
- 785 At prominent switchback near base of slope duck under fence and continue on less worn footpath to north.
- 800 Series of outcrops begins upslope of path.

**Stop 3. Multiphase deformation associated with the Honey Hill Fault zone. (60 minutes)**

This series of Hebron Gneiss outcrops extending > 10 m along the Connecticut River illustrates the complexity of the Honey Hill Fault zone. The stop is also in the hanging wall of the Honey Hill fault *sensu stricto*, however, it shows many of the deformation features described in previous studies of the fault (see discussion in background above). Multiple phases of deformation can be observed from amphibolite facies ductile thrusting to brittle normal faulting (Fig. 6). Pseudotachylites can be found along several foliation parallel surfaces with clearly developed generation planes and injection veins. This outcrop should fuel a vigorous discussion on the nature of the Honey Hill Fault.



**Figure 6.** Sketch illustrating progressive overprinting (top to bottom) of fault structures at stop 3. 1) Top to NW ductile shearing lead to layer bound flow folding. 2) With cooling and or increasing strain rate brittle faulting began. 3) Seismic slip on some fault surfaces lead to frictional melting and injection of melt into dilatational fractures. Modified from Chen (DATE).

- 800 Return to main path.
- 815 Duck under railing and continue south and down hill to field and parking lot.
- 1015 Arrive at parking lot. Boat will be waiting at landing for trip to Seldon Neck Island.

~ 3km boat ride to Seldon Neck Island.

- 1015 Land at beach on west side of island. Walk along old causeway (blue-blazed trail) toward east.  
 1415 Arrive at old quarry site.

**Stop 4. Quarry within footwall rocks of Honey Hill Fault.** (45 minutes)

This stop provides an opportunity to look firsthand at the footwall rocks of the Honey Hill fault. These rocks are granitic gneisses of the Potter Hill Fm. (Rogers, 1985). The gneiss is foliated (strike  $\sim 275^\circ$ , dip  $\sim 20^\circ$ ) parallel to compositional layering. There is also a well-developed biotite lineation trending  $\sim 140^\circ$  and plunging  $19^\circ$ . The gneiss is cut by a series of fractures striking  $160^\circ/340^\circ$  and dipping between  $\sim 80^\circ\text{NE}$ – $\sim 80^\circ\text{SW}$ . Some of the fracture surfaces have slickenlines with rakes of  $\sim 60^\circ$  from the north. Some fractures are also associated with retrograde metamorphism (greenschist) and ductile transposition of the foliation. We thus interpret these fractures as largely brittle-ductile faults.

- 1415 Return to boat along blue-blazed trail.  
 1815 Beach.

Return to Ferry parking lot by boat.

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# THE KILLINGWORTH COMPLEX: A MIDDLE AND LATE PALEOZOIC MAGMATIC AND STRUCTURAL DOME

by

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## INTRODUCTION

The Killingworth complex is exposed in the Killingworth dome, the southern-most of a sequence of structural domes in the Bronson Hill terrane of central New England (Fig.1). These Ordovician intrusive and extrusive rocks (Zartman and Leo, 1985; Tucker and Robinson, 1990; Moench, and Aleinikoff, 2003) were metamorphosed primarily during the Devonian Acadian orogeny. They have been considered to be related by folding and east-directed nappes to younger metasedimentary and volcanic rocks (Lundgren, 1962; Dixon and Lundgren, 1968). We present here geochronologic and isotopic results that complicate this interpretation significantly. We have found that the Middle Ordovician dome is cored by a Mississippian tonalite-trondhjemite. Furthermore we have not been able to recover any direct evidence supporting the contention that these rocks were metamorphosed by the Acadian orogeny. On the contrary, there is compelling evidence that they were strongly metamorphosed, partially melted, and assembled by ductile faulting in the latest Mississippian, Pennsylvanian, and Permian.

### Protoliths

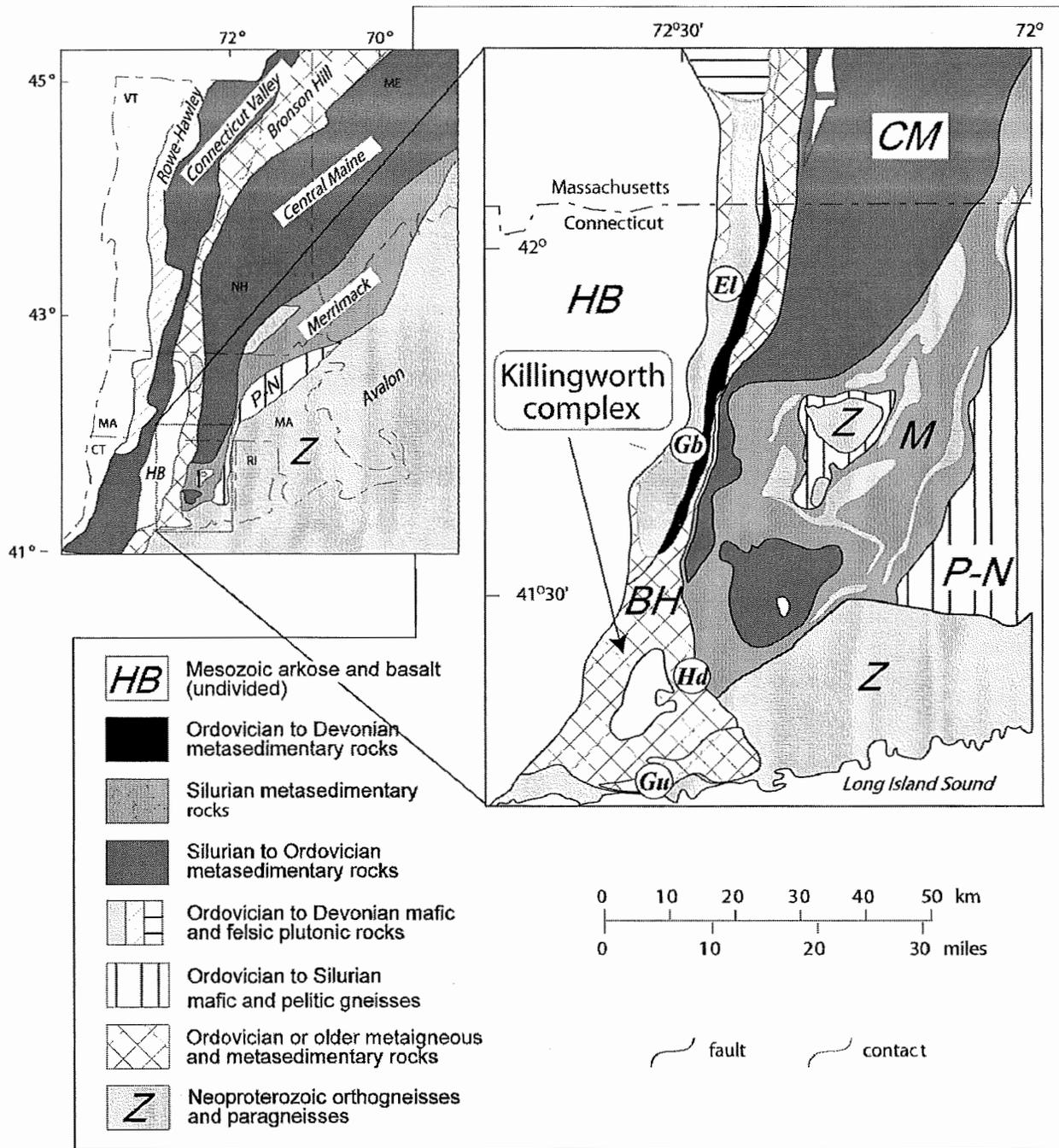
Ortho- and paragneisses of the Killingworth complex have a rather uniform modal mineralogy (primarily of plagioclase and quartz) that has stymied subdivision of these rocks based on field and petrographic relationships alone. Consequently they have been identified collectively as an undivided body of plagioclase gneiss in most previous geological investigations (Percival, 1842; Rice and Gregory, 1906; Foye, 1949; Rodgers, et al., 1959; Dixon and Lundgren, 1968; Fig.1). Varying facies have been recognized, but boundaries were never identified in the field. The differences in these facies were based on km-scale plagioclase compositional zoning in the core region (Mikami and Digman, 1957), and on the quality of foliation development, and on biotite and amphibole content in the Clinton quadrangle (Lundgren and Thurell, 1973). Rodgers (1985) thus had no option but to assign all varieties of these plagioclase gneisses to "Monson Gneiss," an assignment that we avoid here because of the unproven geologic continuity with type-Monson rocks in Massachusetts.

Webster and Wintsch (1987) were the first to recognize that rocks of the Killingworth dome could be unambiguously subdivided into four units. Because of the similarity in appearance of all units, they based their discrimination primarily on major and selected trace element geochemistry (e.g. Fig. 3A), together with cohesive map-scale distribution (Fig. 2). No units have been formally named, and Webster and Wintsch (1987) identified these domains by reference to their location around the dome (eastern, western, southern, and core gneisses). Our new results reinforce the interpretation that these bodies are fundamentally different, thus justifying formal names for these bodies. While all names used at this stage are informal, we maintain that all of these rocks as igneous lithodemic units rather than lithostratigraphic units. We follow the North American Stratigraphic Code (1983) in discriminating between groups and complexes, and formations and lithodemes.

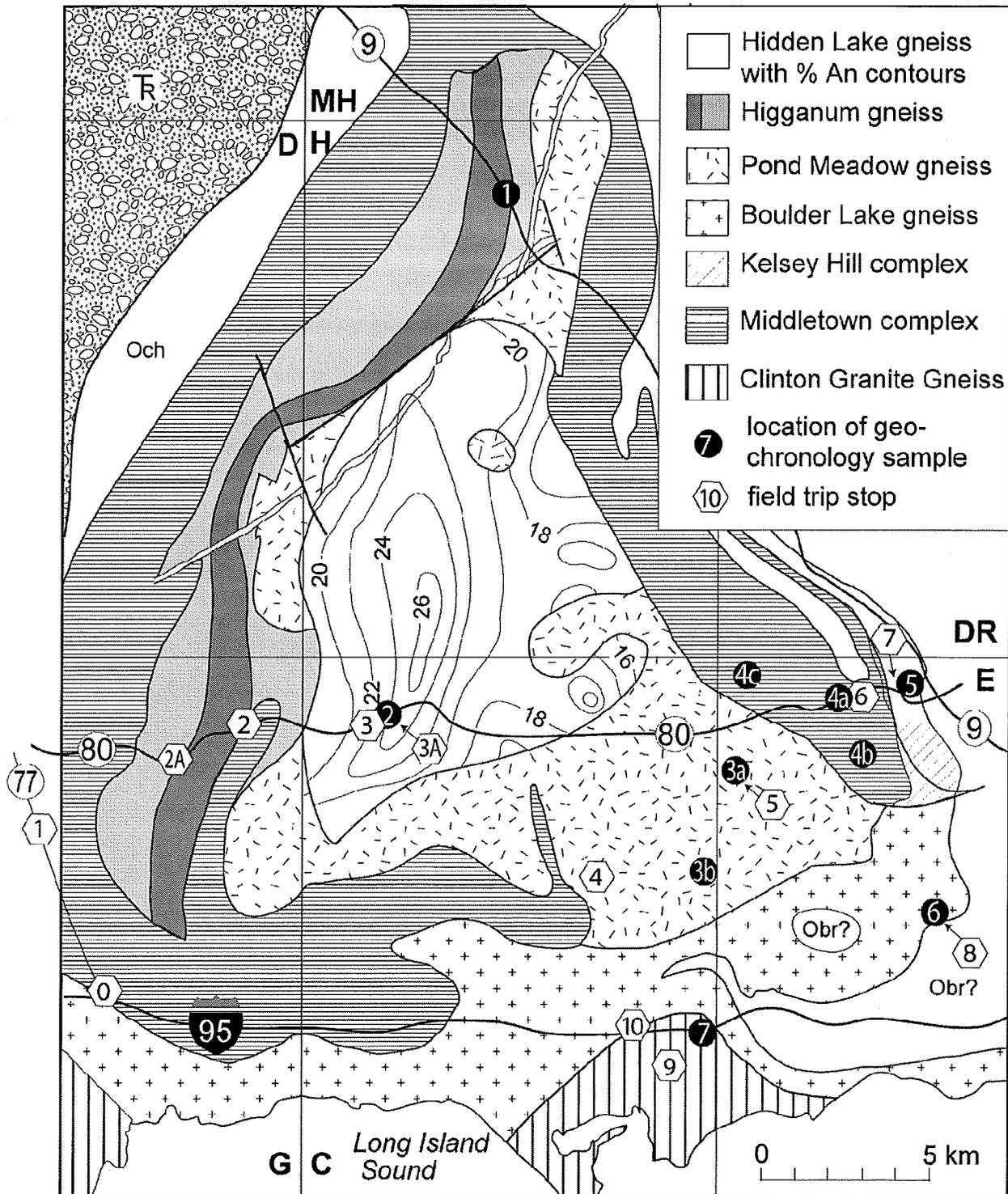
### Field Trip Strategies

The purpose of this trip is to visit typical outcrops of all of the units that comprise the Killingworth complex. Our goal is to show that although their mineralogy is similar (Fig. 3B), the rocks can nonetheless be separated. All units contain essential plagioclase (50-60 vol. %) and quartz, but the variable amounts of accessory biotite, hornblende, magnetite, or garnet (Fig. 3B) are useful discriminators. These accessory minerals are typical, but not always diagnostic of, the primary igneous and volcanic rocks. For example, the hornblende/biotite ratios of the Boulder Lake and Higganum gneisses are higher than in the Hidden Lake and Pond Meadow gneisses. Most magmatic features have been overprinted in the west by moderate to high-grade metamorphism and deformation, and in the east also by ductile deformation associated with Late Paleozoic anatexis. This metamorphic overprint also

leads to differences in the appearance of the rocks in the field, but reflects regional metamorphism rather than differences in protoliths. Mafic enclaves are preserved in some units, but whole rock compositions and zircon age data have been the only features of these rocks that have survived this metamorphism, and that allow objective subdivision of the rocks in this complex.



**Figure 1.** Map showing the Killingworth complex at the southern end of the Bronson Hill terrane. Abbr.: BH, Bronson Hill terrane; CM, Central Maine terrane; HB, Hartford Basin; M, Merrimack terrane; P-N, Putnam-Nashoba terrane; Z, Gander and Avalon terranes, undivided (From Aleinikoff et al., 2002). Letters in circles refer to locations where metamorphic history has been modeled by Wintsch et al., (2003); Gu, Guilford; Hd, Haddam; Gb, Glastonbury; El, Ellington.



**Figure 2.** Geologic map of the Killingworth complex (modified from Webster and Wintsch, 1987), showing subdivision of Rodgers' (1985) Monson Gneiss. The Mississippian Hidden Lake tonalitic gneiss intruded the Ordovician Higganum, Pond Meadow, and Boulder Lake gneisses. The lower contact of the Higganum gneiss and the Middletown complex are probably faults, and the upper contact of the Clinton Granite Gneiss (Stop 10) is probably a terrane boundary. Rodgers' (1985) Collins Hill and Brimfield formations (Och, Obr?) are

Many of the rocks of interest to us on this trip are moderately to well exposed along Conn. Rt. 80 between North Guilford and Deep River. However, fast and heavy traffic along Rt. 80, together with narrow to no shoulders, make most of these stops too dangerous to visit. Consequently, most localities selected for this trip are along smaller roads that are safer for viewing the rocks. Some of these outcrops are on private property, and permission is definitely required before the rocks can be examined. Finally, the authors have not done detailed mapping in any areas outside the Essex 7.5' quadrangle. After reconnaissance in the Guilford and Clinton quadrangles, localities were selected that illustrate critical features of these units. Much more detailed fieldwork is needed to resolve critical field relationships between Ordovician and Carboniferous intrusions, and in identifying Late Paleozoic fabrics and migmatites. Alternate stops will be visited if traffic is light, and if the number of cars in our caravan is small enough to allow us to park safely. Useful references for this trip include Bernold (1976), London (1988), Lundgren (1962, 1963; 1964; 1966; 1979), Lundgren and Thurrell, 1973; Mikami and Digman, 1957; Wintsch (1985; 1994) and Wintsch and Aleinikoff (1987), and Wintsch et al. (1990a; 1990b; 1992; 1993; 2001; 2003).

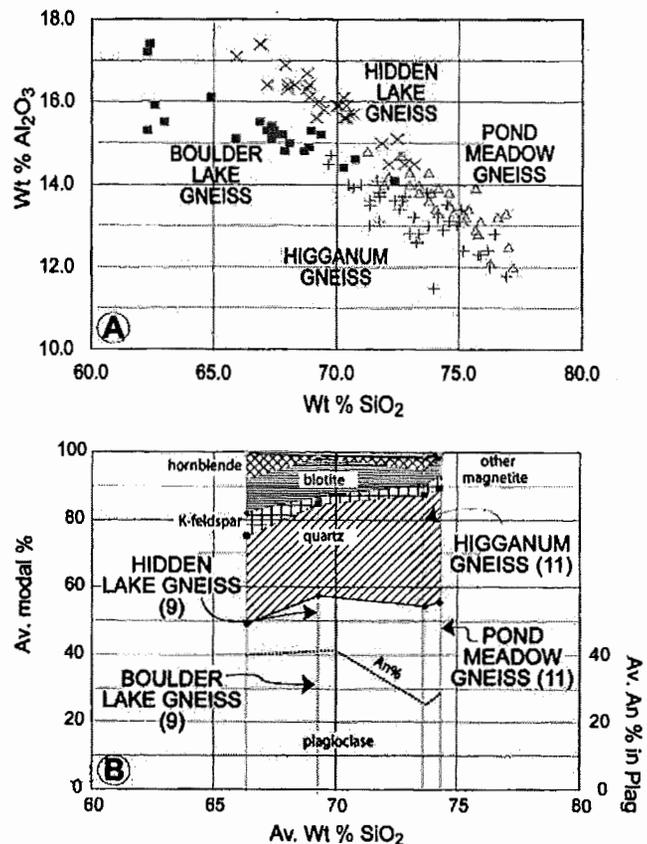
### METHODS

This study builds on the work of Webster and Wintsch (1987). We provide new age data obtained by analysis of zircon using the sensitive high resolution ion microprobe (SHRIMP), and new whole-rock analyses of Sr, Nd and Pb isotopic ratios. About 10 kg of rock were collected for each sample dated. Zircon (and sphene, where present) were extracted using standard mineral separations techniques, including crushing, pulverizing, Wilfley table, magnetic separator, and heavy liquids. Individual grains were hand-picked, mounted in epoxy, ground to half-thickness to expose internal zones, polished, imaged in transmitted and reflected light, and in cathodoluminescence (CL) or by back-scattered electrons (BSE). Selected pristine areas within crystals were dated by the U-Pb method using the USGS/Stanford SHRIMP-RG at Stanford University, Palo Alto, CA, following the methods of Williams (1998). The excavated area of a typical analysis is about 25-35  $\mu\text{m}$  in diameter and about 0.5-1  $\mu\text{m}$  deep. Crystallization ages are determined by calculating the weighted average of individual  $^{206}\text{Pb}/^{238}\text{U}$  ages.

### RESULTS

The units identified by Webster and Wintsch (1987) were differentiated on the basis of compositional clusters like those shown in Fig. 3A, together with cohesive map-scale distribution, as shown in Figure 2. These units take on new significance given their surprisingly different ages and origins. We integrate our new results with existing data below, and present them in the order that they will be visited on this field trip.

**Figure 3.** Summaries of the compositions of the four plagioclase gneisses dated in this study. Figure 3A is a variation diagram showing the variance of  $\text{Al}_2\text{O}_3$  with  $\text{SiO}_2$  (from Webster and Wintsch, 1987). Figure 3B compares the average compositions of the  $\text{SiO}_2$  contents of these four gneisses with average modal compositions of the same gneisses (9 or 11 samples, number in parentheses). The total range of  $\text{SiO}_2$  compositions of these gneisses can be evaluated in Figure 3A. Note that the primary differences in modal composition are in the quartz and mafic mineral contents and the ratios of biotite to hornblende.



**Middletown-Collins Hill complex (northwestern belt)**

The Middletown and Collins Hill formations of Rodgers (1985) lie on the west side of the Killingworth complex. They contain a variety of rock types including biotite schist, plagioclase gneiss, anthophyllite-bearing gneiss, amphibolite, and calc-silicate schist (Bernold, 1976) of mixed igneous and sedimentary protoliths. Distinction between the two formations is one of proportions, making the boundary gradational. With this mixture of igneous and sedimentary protoliths, and a minimum thickness of 1500 m, we refer to these rocks informally and together as a complex. The rocks of the western belt include plagioclase granofels and schist mixed locally interleaved with muscovite-biotite schist. Both of these features are well exposed at Stops 1A and 1B in rocks included by Rodgers' (1985) in the Ordovician Collins Hill Formation. Our geochemical reconnaissance in rocks identified by Rodgers (1985) as Middletown Formation shows that it is more primitive than other Bronson Hill igneous rocks, with positive  $\epsilon_{Nd}$  values, and relatively higher ratios of  $Pb^{207}/Pb^{204}$  (Fig. 4).

**Higganum (western) gneiss or complex.**

We informally name the northwestern body of plagioclase gneiss the Higganum gneiss (western gneiss of Webster and Wintsch, 1987), for the fresh exposures along Rt. 9 near the village of Higganum. This unit dominates the northwestern margin of the Killingworth complex (Fig. 1B). These rocks are distinguished from other plagioclase gneisses by being truly tholeiitic tonalites and trondhjemites, while the other rocks are calc-alkaline (Webster and Wintsch, 1987). In addition, rocks of the Higganum gneiss have positive  $\epsilon_{Nd}$  values overlapping with rocks of the Middletown Formation, suggesting a genetic relationship between the Higganum and Middletown lithodemes.

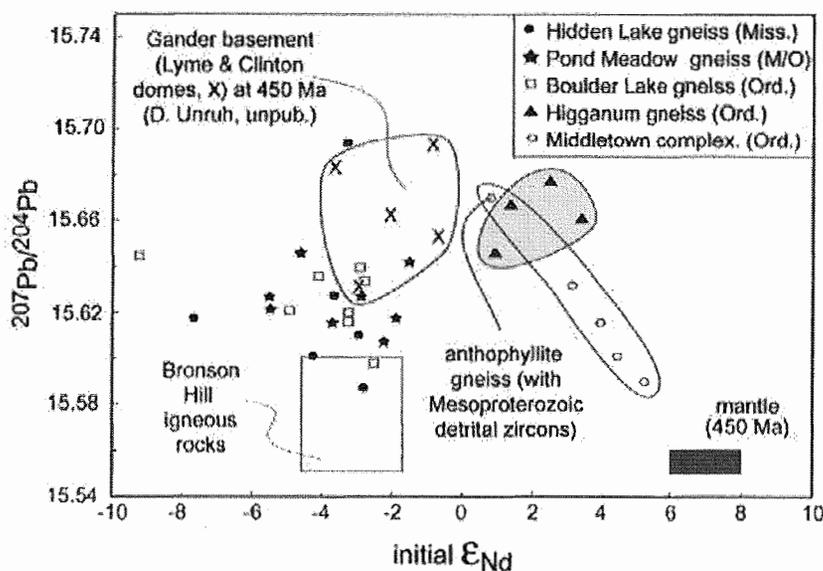
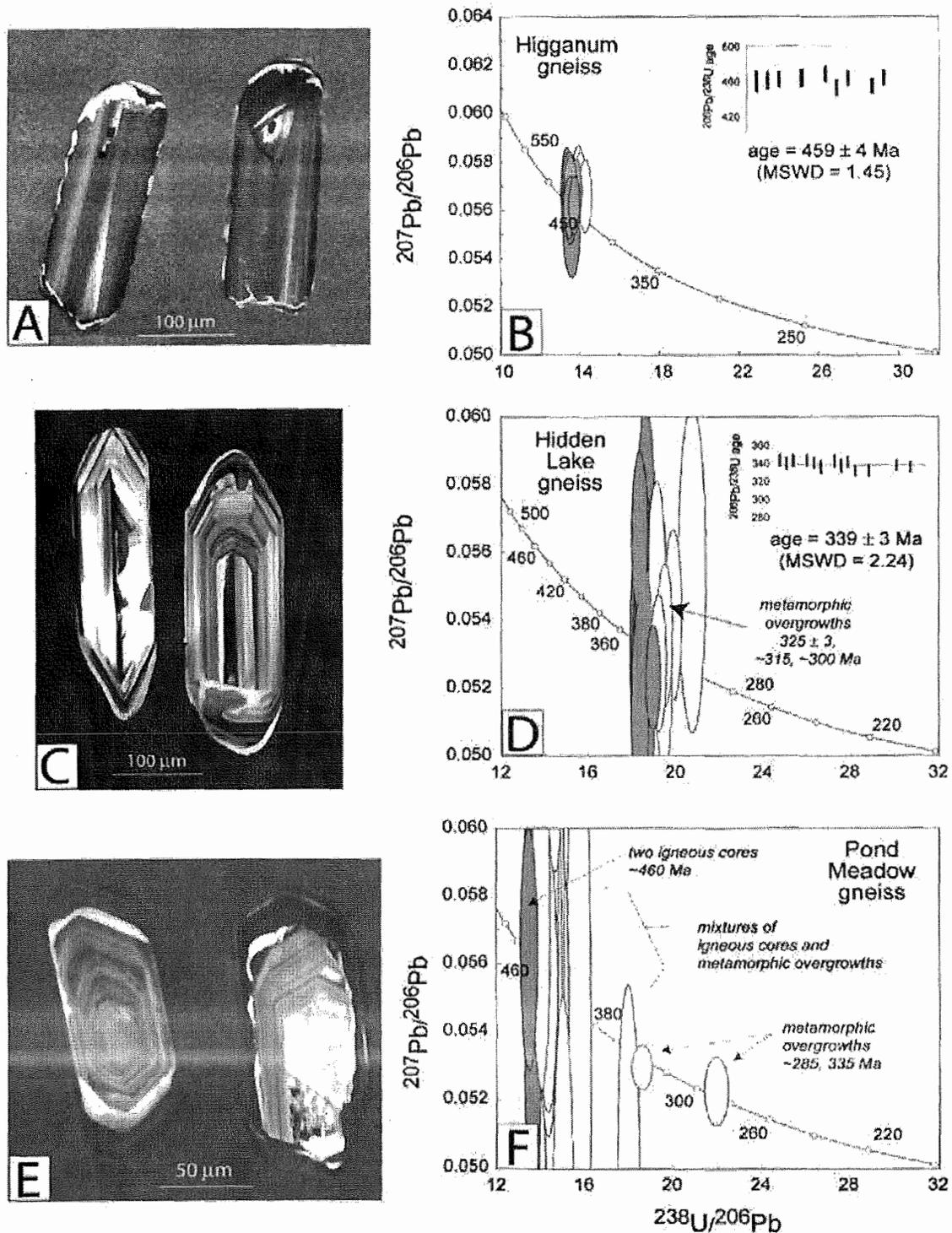


Figure 4. Diagram comparing the range of  $^{207}Pb/^{204}Pb$  and  $\epsilon_{Nd}$  values for several of the igneous bodies that we will see on this trip (D. Unruh, unpub.) with other related rocks (data from Lathrop, et al., 1996; Moench and Aleinikoff, 2003; Tomascak et al., 2005; Zartman and Haines, 1988). Note that the compositions of the rocks in the interior of the Killingworth complex lie between the fields defined by Gander zone rocks and rocks of the Bronson Hill terrane from central New England. However, rocks of the Higganum gneiss and the Middletown complex have more primitive  $\epsilon_{Nd}$  values.

Zircons from the Higganum gneiss collected from a locality along Rt. 9 near Higganum (loc. 1, Fig. 2) are prismatic, subhedral to euhedral, light brown to colorless, and contain numerous opaque inclusions. In CL, these grains show oscillatory zoning that is parallel to the c-axis of the grains but is not concentric (Fig. 5A). Also present are very small overgrowths that are black and unzoned in CL. The zircons contain relatively high concentrations of uranium (mostly 600-800 ppm) and have Th/U of 0.2-0.3, indicative of igneous origin. U-Pb SHRIMP data (9 of 12 analyses) indicate an age of  $459 \pm 4$  Ma (Fig. 5B), interpreted as the time of emplacement of the protolith of the Higganum gneiss. The three slightly younger analyses probably are due to minor Pb-loss.



**Figure 5.** Zircon morphology and age data from the Higganum, Hidden Lake, and Pond Meadow gneisses (localities 1, 2, and 3, Fig. 2). The subhedral, oscillatory zoning in the cores of each grain suggests crystallization from a magma. The narrow weakly zoned rims suggest growth during metamorphism, as confirmed by the relatively young ages of the rims wide enough to analyze.

### Hidden Lake gneiss

The Hidden Lake pluton ("core gneiss" of Webster and Wintsch, 1987) is located in the central portion of the Killingworth complex. Rocks of this region have already been identified as containing relatively calcic plagioclase (Mikami and Dignan, 1957), and as having a low mafic mineral content and flat foliations (Lundgren and Thurrell, 1973). In fact, these rocks constitute a zoned calc-alkaline meta-tonalite, with a relatively calcic core and sodic rim (Webster and Wintsch, 1987) and a low amphibole content (Fig. 3B). It is also quite distinctive from other plagioclase gneisses in its relatively high Sr and Ba contents (Webster and Wintsch, 1987).

The big surprise is the age of the Hidden Lake gneiss. Zircons from the core gneiss (loc. 2, Fig. 2) are euhedral, colorless, and contain few opaque inclusions. All grains show fine concentric, oscillatory zoning in CL; many grains have dark, unzoned tips (Fig. 5C). The oscillatory-zoned portions contain moderate amounts of U (about 200-400 ppm) and relatively high Th/U (0.5-0.8) typical of magmatic zircons, whereas the tips have higher U-contents (mostly 400-700 ppm) and much lower Th/U (0.09-0.14) typical of metamorphic zircons. SHRIMP U-Pb data (13 analyses) for the oscillatory-zoned interiors of these grains yield an age of  $339 \pm 3$  Ma (Fig. 5D), interpreted as the time of crystallization of the igneous protolith of this gneiss. Overgrowths yield a range of ages, including  $325 \pm 3$  Ma and  $\sim 300$  Ma, that date the times of high-grade metamorphic overprint. This metamorphism produced a moderate foliation and banding that dips gently NNW.

### Pond Meadow Gneiss

The most sodic and least mafic gneiss identified by Webster and Wintsch (1987) is the Pond Meadow (their 'eastern') gneiss. It is tonalitic and trondhjemitic in composition, though always gneissic and migmatitic in texture and structure (Stops 4 and 5; Lundgren and Thurrell, 1973). Zircons from the eastern gneiss were obtained from two samples. The first sample (Stop 5; loc. 3, Fig. 2) yielded grains that are prismatic to nearly equant, and colorless. Some grains contain fractured cores. CL images show that these zircons are composed of broad sector-zoned mantles surrounding small, partially resorbed, oscillatory-zoned cores (Fig. 6A). The extensive mantles have high U (800-1400 ppm) and very low Th/U (0.02-0.03). The combination of CL zoning pattern and low Th/U indicates that these mantles are metamorphic in origin. Only one pristine oscillatory-zoned core (of presumed igneous origin) was analyzed in this sample. It has about 370 ppm U and Th/U of 0.21. Ten of 13 analyses of the sector-zoned mantles yield an age of  $335 \pm 2$  Ma (Fig. 6B). Three other analyses are slightly older ( $\sim 345$  Ma) or younger ( $\sim 325$  Ma). The oscillatory-zoned core is  $461 \pm 4$  Ma.

The second sample was collected in an effort to obtain zircons that preserve original igneous zoning. These grains are prismatic, medium to dark brown, and contain numerous fractures. Many grains are composed of large, partially resorbed cores showing concentric, oscillatory zoning and medium to small dark, unzoned overgrowths (Fig. 6A). We attempted to date the oscillatory-zoned portions of these grains to obtain additional information about the time of emplacement of the protolith of the Pond Meadow gneiss. Unfortunately, most of the oscillatory-zoned areas are intimately invaded by overgrowth material, yielding a wide range of ages spanning about 110 m.y. (Fig. 6B). The two oldest ages, about 460 Ma, probably are analyses solely of cores. Two analyses of dark overgrowths are about 335 and 285 Ma, in agreement with data from the other sample of Pond Meadow gneiss. All intermediate ages from oscillatory-zoned areas are mixtures of the two age components, and therefore are geologically meaningless.

We interpret these data to indicate that the protolith of the Pond Meadow gneiss, now a migmatitic rock, was emplaced as a tonalite in the Ordovician. Migmatization (especially conspicuous at Stop 4) was apparently responsible for resorption of most of the Ordovician cores and formation of the pervasive broad metamorphic mantles. The ages of these mantles show that most of migmatization occurred in the Mississippian, with minor growth in the Permian.

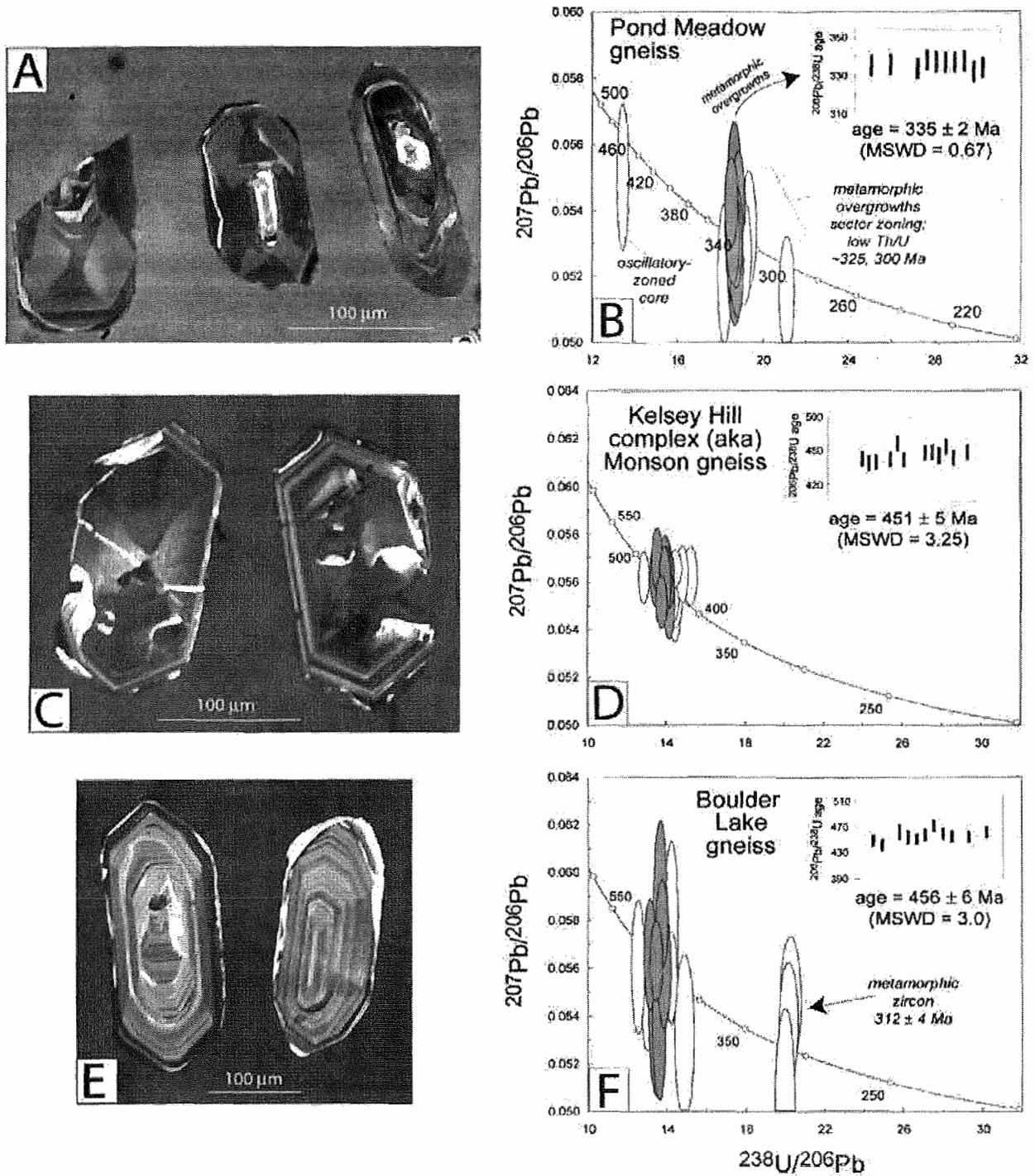
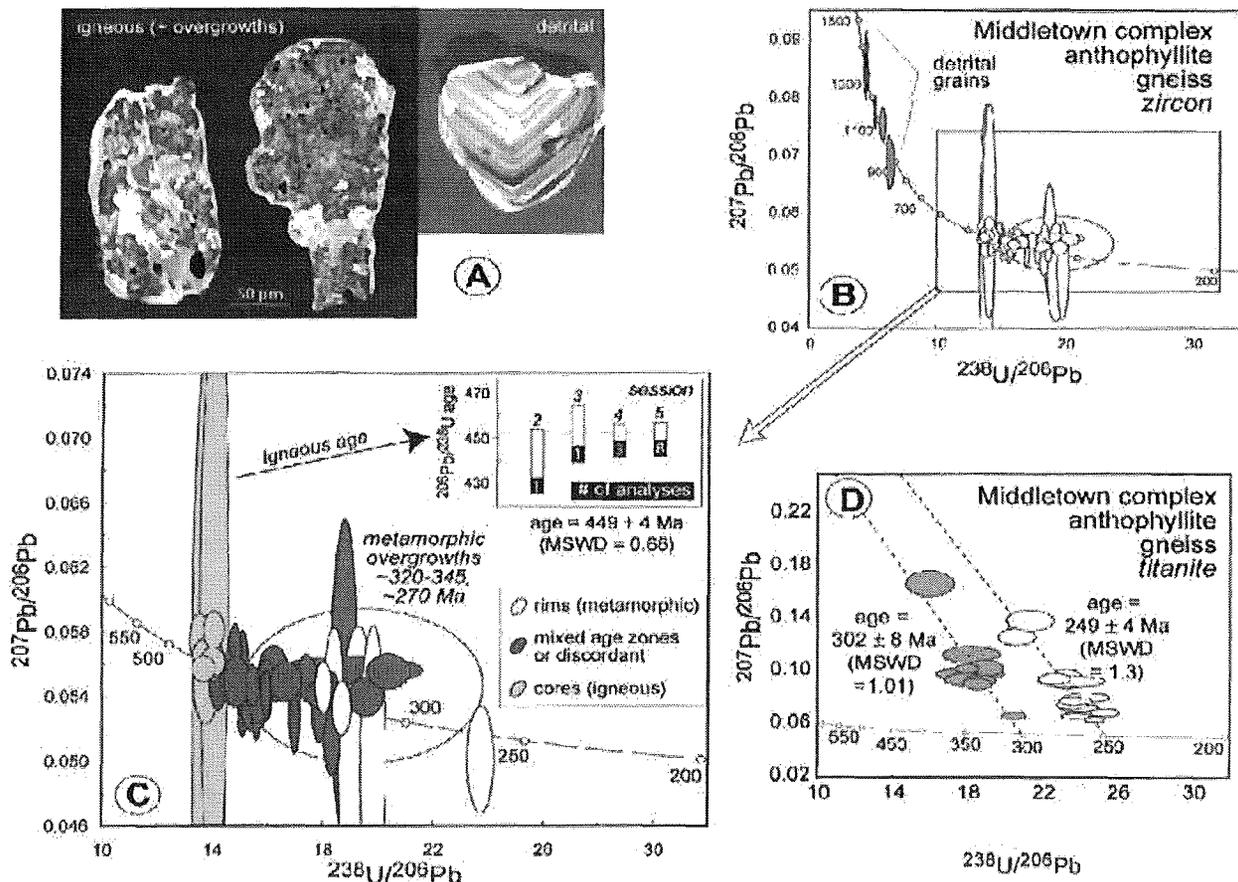


Figure 6. Zircon morphology and age data from the Pond Meadow, Kelsey Hill, and Boulder Lake gneisses. The oscillatory zoning in cores of zircons from the Kelsey Hill and Boulder Lake gneisses indicates crystallization in a magma, whereas sector zoning in broad mantles of zircons from the Pond Meadow gneiss suggests growth during metamorphism.

## Middletown complex (eastern belt)

The rocks in the eastern belt of the Middletown complex are historically recognized for anthophyllite-bearing plagioclase gneiss, which we will see at Stop 6. These gneisses are mixed with biotite schist, sillimanite schist, amphibolite, and cotecule, suggesting a volcanic-sedimentary package of rocks. Zircons from three localities (4a, 4b, 4c, Fig. 2) the anthophyllite gneiss within the Middletown complex are very unusual, but support this interpretation. Almost all grains, from all three samples, are anhedral (amoeboid shapes) and dark brown. In CL, they show mottled textures with sparse very dark spots, irregularly overgrown by light rims (Fig. 7A). In BSE, most grains contain small, randomly distributed, very bright spots that are thorite (determined by the SEM-energy dispersive analysis). The rim material commonly invades the grains, so that there are very few pristine, homogeneous areas large enough to accommodate a SHRIMP analysis without contamination by an adjacent area. We think the unusual anhedral, mottled grains are primarily igneous in origin, with discontinuous metamorphic overgrowths. A second population of zircon (less than 1% of the total) occurs in the anthophyllite gneiss. These grains are roughly equant, rounded, light brown to colorless, and have a variety of types of oscillatory zoning, all of which are truncated at grain boundaries. We interpret these rounded grains to be detrital in origin, as supported by the U-Pb age data (Fig. 7B).



**Figure 7.** Zircon morphology and age data from the anthophyllite gneiss of the Middletown complex. Many of these grains show very complicated intergrowths (Fig. 7A). After excluding data from mixed analyses (dark gray, Fig 7C), age data from the largest and most homogeneous cores yield an age of ~450 Ma. In addition, metamorphic rims show early Carboniferous and Permian overgrowths. Two populations of titanites exist in these rocks (Fig. 7D): dark brown grains that formed at ~300 Ma during Alleghanian heating, and pale grains that crystallized at ~250 Ma (during decompression following the Alleghanian orogeny).

By imaging using both CL and BSE, and examining thousands of grains during five separate SHRIMP sessions, we were able to obtain sufficient U-Pb data to have some confidence in assigning an age to this rock. The anhedral, mottled grains have high U contents (800-3600 ppm) and igneous-like Th/U ratios of 0.2-0.6. The light colored overgrowths have lower U (mostly 800-1200 ppm) and the very low Th/U of 0-0.04 typical of zircons of metamorphic origin. Rounded, oscillatory-zoned grains interpreted as detrital in origin have modest U contents (100-300 ppm) and igneous Th/U of 0.2-0.6. Thirteen analyses from four sessions resulted in a coherent group of data with an age of  $449 \pm 4$  Ma (Fig. 7C). Many more analyses are younger, probably the result of mixing of Ordovician core regions with Late Paleozoic overgrowths. Six of these overgrowths were dated, yielding ages of about 350-320 Ma; one other overgrowth is about 270 Ma. Finally, zircons interpreted as detrital in origin have much older ages, between about 0.9-1.3 Ga, indicating derivation from Laurentian (i.e. Grenville) sources (Fig. 7B).

Titanite from the anthophyllite gneiss occurs as light brown and yellowish to colorless, anhedral varieties. As is typical in titanite of metamorphic origin, these grains contain fairly high contents of common Pb. U-Pb ages are determined by calculating regressions through the data to determine concordia intercept ages of  $302 \pm 8$  and  $249 \pm 4$  Ma for the brown and colorless grains, respectively (Fig. 7D). The ages record the times of titanite growth due to Alleghanian metamorphism and decompression.

The presence of detrital zircons within an otherwise igneous population of zircon suggests that the protolith of the anthophyllite gneiss was a felsic, probably dacitic, volcanic rock. This interpretation is supported by the intercalation of amphibolite and biotite schists at other localities within the Middletown complex.

#### **Kelsey Hill Complex (aka Monson Gneiss)**

The name 'Monson Gneiss' has been used to refer to a large variety of rocks in the Killingworth complex, and in particular the rocks in the vicinity of Stop 6. However, this region includes granitic gneiss, tonalitic gneiss, amphibolite, and dunite (Wintsch, 1994). Such a collection of orthogneisses suggests a lithodemic complex, and not the Monson Formation of Rodgers (1985). We selected a granitic gneiss located 300 m east of our Stop 7 for dating (loc. 5, Fig. 2). Zircons from this gneiss are prismatic, stubby, subhedral to euhedral, and medium to dark brown. In CL, the grains appear mostly sector-zoned and surrounded by a thin oscillatory-zoned rim (Fig. 6C). Some grains contain very thin, dark, unzoned outermost rims, and most of the grains are extensively fractured. Zircons from this granitic gneiss have fairly high U contents (600-2100 ppm) and Th/U of 0.3-0.6. Eleven of 20 analyses form a coherent group with an age of  $451 \pm 5$  Ma (Fig. 6D), interpreted as the time of emplacement of the igneous protolith. Two analyses are older (~464 and 486 Ma) and may reflect derivation of this gneiss from slightly older Ordovician (Bronson Hill?) source rocks. Younger ages (~410-435 Ma) probably are due to Pb-loss from damaged, fractured grains, and thus, are geologically meaningless.

#### **Boulder Lake Gneiss**

The Boulder Lake gneiss was named informally by Lundgren and Thurell (1973) for rocks in the Clinton quadrangle, and has been carried east into the Essex quadrangle as 'southern gneiss' by Webster and Wintsch (1987) and Wintsch (1994). Bulk compositions show the protolith to be a zoned granodiorite-tonalite pluton with a more sodic interior, and a more calcic margin (Webster and Wintsch, 1987; Wintsch et al. 1990b). It is distinguished in the field by a relatively high amphibole content, a general lack of well-developed layering, and the common occurrence of mafic and calc-silicate inclusions (Fig. 3; Lundgren and Thurell, 1973; e.g. Stop 8) where strain is low. However, where strain is high, as it is along the southern (terrane) boundary with the Clinton dome, primary igneous structures are not apparent (Stop 10). The truncation of the chemical zoning pattern in this pluton against this fault along the southern margin of the body (Wintsch et al. 1990b) is consistent with this interpretation.

Zircons from the Boulder Lake gneiss (loc. 6, Stop 8, Fig. 2) are prismatic, euhedral, and light brown. Most grains show fine concentric, oscillatory zoning in CL (Fig. 6E), and some grains have very narrow dark, unzoned overgrowths. U contents of Boulder Lake zircons are moderate (150-400 ppm); Th/U is 0.2-0.4, typical of an igneous origin. Eleven of 16 analyses of the oscillatory-zoned portions of the grains yield an age of  $456 \pm 6$  Ma, interpreted as the time of crystallization of the tonalitic protolith. Two analyses are older (478 and 495 Ma). Another core, with an irregular rounded boundary and oscillatory zoning truncated by the igneous mantle, is of Grenville origin, as shown by the age of 1.33 Ga. Several ages younger than 450 Ma probably are due to Pb-loss, as these were obtained from oscillatory-zoned mantles. Metamorphic overgrowths yield an age of ~317 Ma.

### Clinton Granite Gneiss

The Clinton Granite Gneiss of Lundgren (1964) and Lundgren and Thurell (1973) lies at the structural base of the Killingworth complex along Long Island Sound. It is granitic in composition, and most outcrops show evidence of migmatization and variable texture (Lundgren and Thurell, 1973). Zircons in a sample of the Clinton Granite Gneiss from a road cut along Interstate 95 (too dangerous to visit) are subhedral to euhedral, and medium brown. Some grains are prismatic whereas others are stubbier. Most grains are fractured and contain numerous opaque inclusions. All zircons show fine concentric, oscillatory zoning in CL (Fig. 8A). Only a few grains have very small, dark, unzoned overgrowths. Eleven of 13 analyses yield an age of  $605 \pm 4$  Ma (Fig. 8B), interpreted as the time of crystallization of the granitic protolith. Two analyses are slightly younger, probably due to Pb-loss. Thin overgrowths are present, but were not dated.

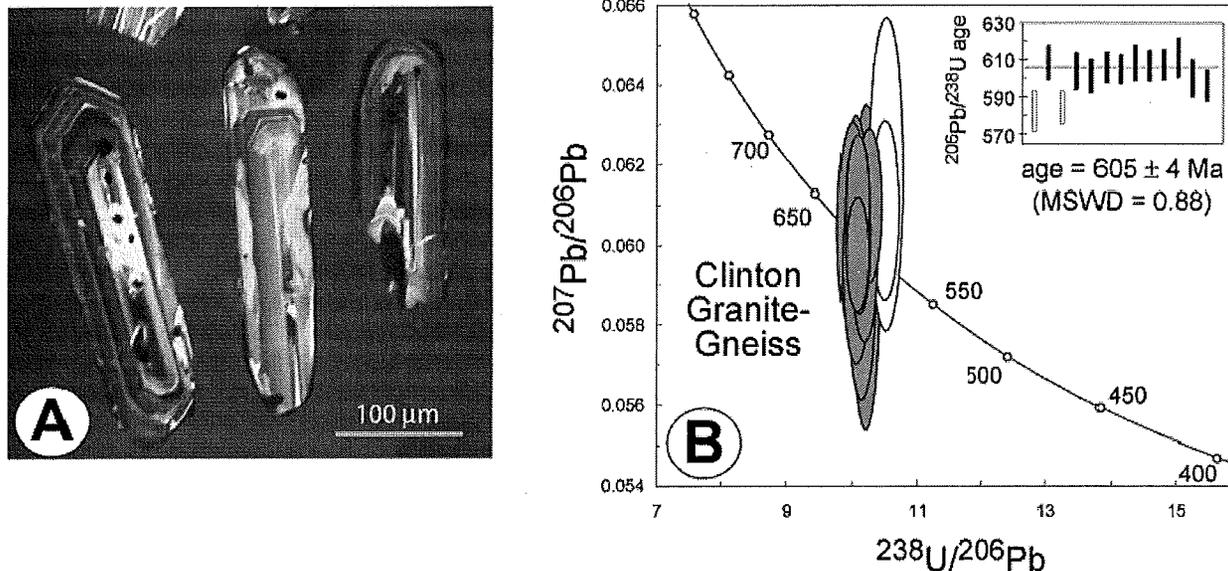


Figure 8. Zircon morphology and age data from the Clinton Granite Gneiss. The concentric, oscillatory zoning in the cores of these grains suggests crystallization for a magma. The very narrow dark rims on some grains that are too small for SHRIMP analysis suggest growth during metamorphism.

## REGIONAL IMPLICATIONS

### Ages of Intrusion

Our geochronological work shows that several of the units in the Killingworth complex are Late Ordovician, similar to other Bronson Hill rocks (Leo et al., 1984; Tucker and Robinson, 1990; Aleinikoff et al., 2002; Moench and Aleinikoff, 2003). These include the Higganum, Kelsey Hill, and Boulder Lake gneisses, all between 450 and 460 Ma. Their temporal, spatial, and isotopic similarities reported here indicate that they were emplaced together in a volcanic arc from a common source. However, their distinct chemical compositions (Webster and Wintsch, 1987; e.g. Fig. 3) show that they evolved as discrete magmas. The Higganum gneiss, in particular, has a more primitive  $\epsilon_{\text{Nd}}$  values (Fig. 4), suggesting that it has a closer link to the mantle than the other gneisses.

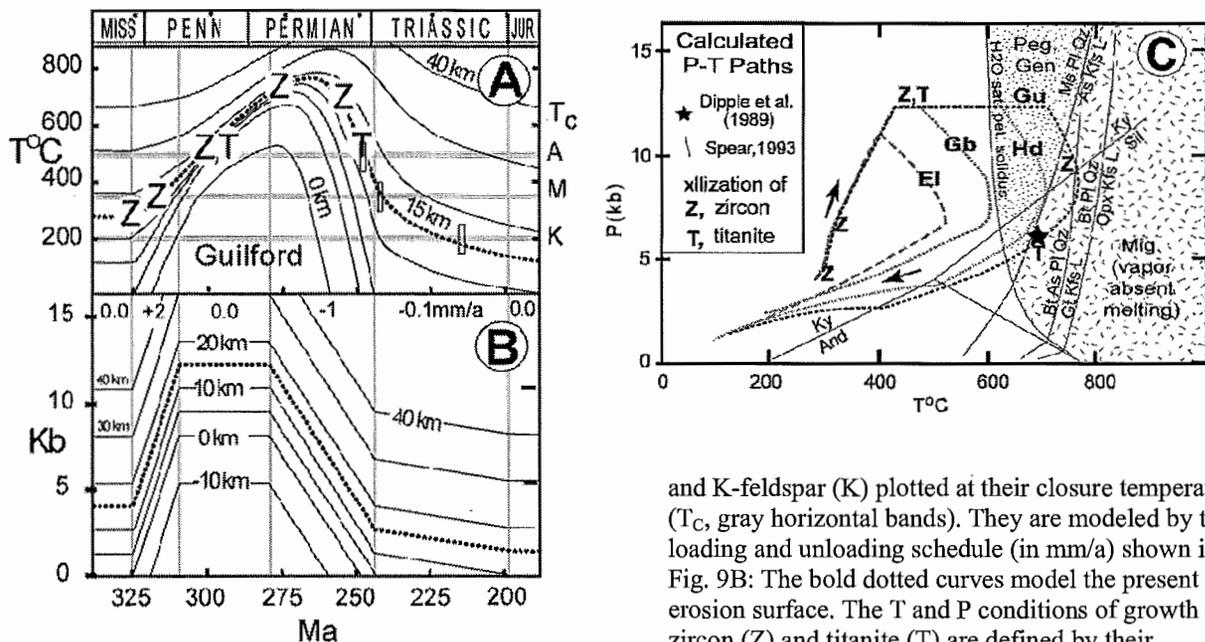
The interior body, the Hidden Lake gneiss, is a discrete pluton, with distinctly higher Sr and Ba contents (Webster and Wintsch, 1987), and with smooth chemical zoning shown by km-scale differences in normative anorthite content (Fig. 2). Our new results now show this body to be Mississippian (Fig. 5D). However, the Nd and Pb isotopic composition of this body overlaps with the similar but Ordovician Boulder Lake and Higganum gneisses (Fig. 4), suggesting that the Hidden Lake pluton was derived from the root zone of the Bronson Hill terrane, or from the same basement that gave rise to the Bronson Hill arc.

### Status of the Higganum and Middletown gneisses as a complex

More work is necessary to determine whether the Higganum gneiss is intrusive or extrusive, an issue made difficult by the intense metamorphic and structural overprint. The local intercalation of schistose rocks, especially in Guilford (note the finger of schist invading N into the Higganum gneiss in the Guilford quadrangle, Fig. 2) suggests that the Higganum gneiss may be of extrusive origin. However, the large massive outcrops of coarse-grained rock along Conn. Rt. 9 (geochronology sample 1, Fig. 2) suggest that part of it may be intrusive. A rough chemical zonation within the body (Webster and Wintsch, 1987; dark gray band, Fig. 2) parallels the structurally overlying contact with Rodgers' (1985) Middletown Formation, suggesting that the Higganum-Middletown-Collins Hill block may constitute a single volcanic-sedimentary complex. If so, then the upper Ordovician age of the dated rock may place a maximum age on the overlying Middletown and Collins Hill complex.

### Ages of Metamorphism

In northern and central New England, rocks of the Bronson Hill terrane are well known to have been metamorphosed by the Middle Devonian Acadian orogeny (e.g. Tucker and Robinson, 1990). In the Killingworth complex, however, we find no direct evidence of a Devonian metamorphism. On the contrary, much evidence is accumulating that a high grade to anatectic event metamorphosed these rocks during the Carboniferous and Permian (e.g. Lundgren, 1966; Dipple et al., 1990; Wintsch et al., 2003). Metamorphic overgrowths on zircons from continuous range of ages, including approximately 345, 325, 320, 315, 300, and 270 Ma. In addition, titanites from the Middletown gneiss yield ages of ~300 and 250 Ma. These data are time curves of Fig. 9A are constrained by data presented by Wintsch et al. (2003) including the cooling ages of amphibole (A), muscovite (M),



**Figure 9.** The metamorphic history of the southern Killingworth complex (Guilford area, Fig. 1). The temperature-time curves of Fig. 9A are constrained by data presented by Wintsch et al. (2003) including the cooling ages of amphibole (A), muscovite (M),

and K-feldspar (K) plotted at their closure temperatures ( $T_c$ , gray horizontal bands). They are modeled by the loading and unloading schedule (in mm/a) shown in Fig. 9B: The bold dotted curves model the present erosion surface. The T and P conditions of growth of zircon (Z) and titanite (T) are defined by their placement on the T-t (and P-t) curves. Fig. 9C: the model P-T path derived from Figs. 9A and B showing the inferred prograde and retrograde metamorphic conditions of zircon and titanite growth (see text). Abbreviations on nested P-T paths show model metamorphic histories for localities given in Fig. 1

We suggest that following the Ordovician emplacement of the Higganum, Pond Meadow, and Boulder Lake gneisses, the intensity of Acadian metamorphism was too low to form zircon overgrowths. An alternative possibility is that Alleghanian metamorphism was so high-grade that it destroyed earlier metamorphic overgrowths. However, selective destruction of metamorphic zircon is considered unlikely. The latter generally contains lower concentrations of minor and trace elements than does magmatic zircon (Hoskin and Schaltegger, 2003), and so

should be less reactive. The occurrence of inherited magmatic zircon as well as Mississippian metamorphic zircon thus makes doubtful that a more stable Acadian metamorphic overgrowth ever existed in these rocks.

In the Mississippian, about 110 m.y. after initial Ordovician magma emplacement, a second magma generated in the deep core of the Killingworth dome intruded the dome as the protolith of the Hidden Lake gneiss. At the same time, the Pond Meadow gneiss was extensively migmatized. We cannot discriminate among several causes of this Mississippian partial melting event: (1) The heating of the Pond Meadow gneiss may have been due to advection *in situ* by the intrusion of the Hidden Lake pluton, (2) the Ordovician protolith may have been heated at depth by the same event that formed the Hidden Lake magma and both the Hidden Lake and the Pond Meadow bodies were emplaced together, or (3) the Pond Meadow gneiss may have risen with the Hidden Lake magma, and migmatization of the tonalitic composition was achieved by decompression melting. The lack of migmatization in the adjacent Higganum gneiss to the west suggests that the latter may have been placed against the Hidden Lake gneiss later in the Carboniferous, or in the Permian, after metamorphic temperatures had fallen from their peak.

### Tectonic Significance

The ~340 Ma date of the Hidden Lake gneiss is an uncommon age in the Appalachians. It is older than typical Alleghanian events, and younger than is probably too young to be associated with the "Neo-Acadian" of Robinson et al. (1998). However, deposition of the large Mississippian clastic wedge in Pennsylvania and Virginia suggests that the early stages of the approach of Gondwana toward North America and the assembly of Pangea were operating at this time. It is thus possible that the orogenic event that caused the Hidden Lake pluton was a manifestation of an early volcanic arc associated with this Mississippian collision.

New isotopic and geochronologic evidence support the interpretation that the rocks of the Clinton dome are part of the Gander zone (Wintsch et al., 2005). If this interpretation can be confirmed, then the boundary between the Clinton granite gneiss and the overlying Killingworth complex should be strongly tectonic, and probably a major suture. The fact that the Hidden Lake pluton has isotopic compositions that are consistent with derivation, in part, from this Neoproterozoic crust allows the possibility that this leading edge of Gondwana wedged these Gander zone rocks further under the Bronson Hill terrane, causing the heating of both Gander and Bronson Hill tonalites. Magma mixing might then account for the isotopic compositions midway between the more primitive Bronson Hill rocks and the more enriched Gander zone rocks (Fig. 4) in the early Mississippian.

## ROAD LOG

**STOP 0.** Trip begins at the Park and Ride on the NW corner of the intersection of I95 and SR77. Assemble here at 7:45 and consolidate cars. Trip departs at 8 AM.

0.0 miles. Leave parking lot, turning N on Rt. 77.

3.7 miles. Turn right (east) into Andrews Ridge development. Follow Alexander Drive through cuts of Stop 1B, and bear right at the fork in the road.

3.9 miles. Turn around at the cuts on both sides of the road. Park on the east side of the road.

**STOP 1. Middletown-Collins Hill Complex.** (20 minutes) **Stop 1A.** Here we see (through rust stained west-dipping joint surfaces) a plagioclase-quartz-biotite granofels. It is only weakly foliated, and layered. Magnetite is absent. Chemical weathering is apparent; rocks closer to the surface are pale tan, while deeper at the base of higher cuts the rocks are a moderate gray color, reflecting the dominance of plagioclase. Retrace path toward Rt. 77.

4.1 miles. Park on right side of the road adjacent to a large road cut.

**STOP 1B. Middletown-Collins Hill Complex.** (20 minutes). This outcrop exposes primarily biotite-muscovite schists interlayered with plagioclase granular schists and concordant pegmatites, all dipping steeply west. The high density of thin quartz veins suggests that the biotite-muscovite-rich schists are the result of metamorphic differentiation of a more typical greywacke-like protolith. The dominant planar fabric, S<sub>n</sub>, is pervasively overprinted by a later crenulation (S<sub>n+1</sub>) with an axial plane dipping gently NW. This S<sub>n+1</sub> fabric even penetrates a coarse

grained pegmatite. The gray plagioclase-rich granular schists suggest that dacitic sills or volcanic layers were intercalated within the metasediments.

This set of two outcrops demonstrates two of the critical questions of this trip. First, how many different plagioclase gneisses exist in the southern Bronson Hill terrane? The plagioclase granofels at Stop 1A probably now exists as a lens within the schistose rocks. Presumably it is an Ordovician igneous rock that escaped fabric development because it was stronger than the surrounding schists, and the strain that produced the  $S_n$  fabric was partitioned into the schists. But is it possible that the granofels is younger than the  $S_n$  foliation? Given the Carboniferous ages of both intrusive and anatectic events to the east, this possibility cannot be ruled out. The second point raised is one of the age of the metamorphism. These rocks are not anatectic, and pervasive veining is largely restricted to quartz veins. Could the metamorphism during the development of  $S_n$  be Acadian? The  $S_{n+1}$  fabric is almost certainly Late Paleozoic. More work is needed to establish the time(s) of metamorphism in these rocks. Return to Rt. 77.

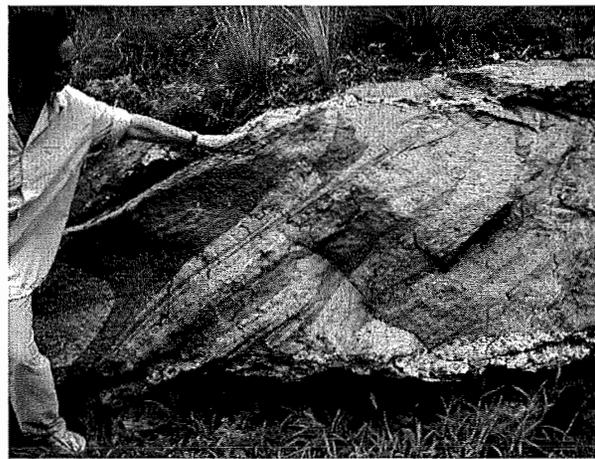
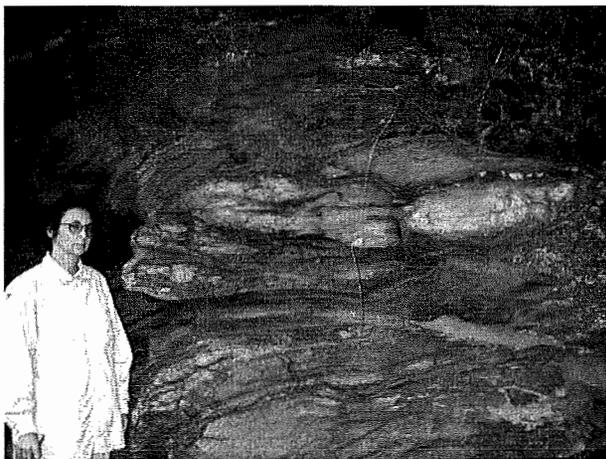
4.2 miles. Turn right (N) on Rt. 77.

4.8 miles. Turn R (E) on to Rt. 80.

7.8 miles. Passing alternate **STOP 2A. Higganum Gneiss.**

9.4 miles. Continue E through the round about. After only about 100 m turn R (S) into the North Madison shopping center.

9.6 miles. Park in the shopping center. **CAREFULLY** cross Rt. 80 to the north side of the road to cuts of Stop



**STOP 2. Higganum Gneiss.** (20 min) The purpose of this stop is to show the well-developed gneissosity of this orthogneiss. Evidence that it is an orthogneiss is best preserved in the zircon population. In outcrop, on the other hand, the gentle west-dipping dominant foliation deforms and isoclinally folds crosscutting pegmatites as shown above, and this foliation in turn contains gentle infolds of intercalated schists.

At **ALTERNATE STOP 2** layers of medium gray and dark gray plagioclase-quartz-biotite granofels are isoclinally folded, and some fold noses float in the pale gray tonalitic granofels. Here also relatively late granitic and pegmatitic sheets intrude the axial planes, and are themselves foliated. Together these relationships show that a very high grade, and probably anatectic metamorphic event has overprinted and destroyed any primary evidence of the Ordovician igneous protolith

Return to cars Turn right, continuing E on Rt. 80.

10.9 miles. Turn right (S) on to Owl Hollow Lane, and park in the cul-de-sac.

**STOP 3. Hidden Lake Pluton** (20 minutes) This stop is included to illustrate the high grade of the metamorphism of the core rocks of the Killingworth complex. Here, and at alternate **STOP 3A** the rock displays a moderately well developed foliation and a biotite streak lineation that plunges gently north. Biotite preferred orientation in this foliation is only moderately, but the biotite is segregated into bands giving the rock the appearance of gneissic

banding. Locally the dominant foliation includes asymmetrically boudined veins and “macrolithons” of gneiss in which an earlier gneissosity dips gently south. The inclusions containing an earlier gneissosity reveal that this dominant foliation is not the first foliation, and the tails that taper to the north suggest a top to the N sense of shear later in the metamorphic history. Note the moderate dip to the west of the prominent joint surfaces with local slickensides in this outcrop. This is probably a Mesozoic structure, which may suggest deformation from the Hartford basin working its way west into the footwall.

11.0 miles. Turn R (E) on Rt. 80.

11.7 miles. Passing through cuts of **ALTERNATE STOP 3A**.

12.5 miles. Passing Chatfield Hollow State Park. Large exposures of Hidden Lake Pluton here are covered with lichen and very difficult to study.

13.7 miles. Pass through the roundabout, staying on Rt. 80.

14.6 miles. Turn R (S) on Roast Meat Hill Rd.

15.0 miles. Turn left on Ironworks Road.

17.4 miles. Turn R on Ben Merrill Rd. Continue to #28 Ben Merrill Rd. passing Elsie Lane.

18.1 miles. Park along Ben Merrill Rd. at #28.

**STOP 4. Pond Meadow Gneiss** (20 minutes) This stop is included to illustrate the high degree of migmatization of these rocks, and the multiple schistositys present. The primary schistosity  $S_n$  dips gently west. It is defined more by the concentration of plagioclase, quartz, and the small amount of biotite into discrete wispy bands than by the preferred orientation of the biotite. This banding is interrupted by and deformed by moderately to steeply N-dipping  $S_{n+1}$  fabric defined by a dextral crenulation, small scale shear zones, and migmatitic zones. Both magnetite and biotite ‘eyes’ or flecks locally punctuate the texture. Both of these schistositys locally ‘dissolve’ into a hypidiomorphic (granitic) granular texture (figure below, around the pencil), suggesting very high-grade metamorphic conditions sufficient to partially melt a tonalitic rock. These rocks were interpreted by Webster and Wintsch (1987) to be Higganum (their western) gneiss, but their isotopic compositions suggest these rocks belong to be the Pond Meadow gneiss.



- 18.2 miles. Turn around and return to Ironworks Rd.
- 18.7 miles. Turn R on Ironworks Rd.
- 19.8 miles. Turn L on Kelseytown Rd.
- 21.0 miles. Turn R on the unsigned(!) Chittenden Hill Rd. Caution! This is a windy country road. Stay to the right and drive slowly!
- 22.5 miles. Turn L (on the unsigned Old Horse Hill Road) at the stop sign at the end of the road. Follow this road around to the next stop sign.
- 22.7 miles. Turn L at the next stop sign on to Rt. 145 (aka Stevenstown Rd.).
- 24.0 miles. Turn R (E) into the 'Emergency Access Only' chained road. Walk N for 100 m to road cuts on the left (W) side of the road.



**STOP 5. Pond Meadow Gneiss (20 minutes)** This stop is included in the trip to show the locality of one of the two dated samples. It shows features similar to Stop 4, including two schistositys, an earlier W-dipping, and a later N-dipping cross cutting one. Both fabrics are marked by metamorphic differentiation, and apparent partial melting. Some joint faces show apparent pinch and swell structures, probably caused by the interference of these two fabrics. At least one amphibolite boudin seems to 'float' in the Sn schistosity, suggesting partial melting. Together, these structures support the interpretation that very high grade metamorphism led to both the fabrics and zircon recrystallization at about 335 Ma.

- 24.0 miles. Continue N on Rt. 145.
- 24.8 miles. Turn R (N) on Rt. 80.
- 26.5 miles. Passing cuts of the Middletown complex too dangerous to visit on this trip.
- 26.7 miles. Turn R (S) on Woodbury Rd.
- 26.8 miles. Turn R on Industrial Park Rd. Low outcrops are at the immediate right.

**STOP 6. Middletown Complex. (20 minutes)** These small outcrops and a large local block (probably from the quarry in front of us) give us a flavor of these gneisses. More and fresher rock is exposed at alternate Stop 6, but is too dangerous to include on this trip. Here we see a strong foliation dipping moderately to the W, and a lineation both defined by 2-3 cm long anthophyllite needles. Layering in this rock seems in part due to metamorphic differentiation, but included detrital zircons (see results) suggest that the protoliths were volcanics.

- 26.9 miles. Turn around and return to Rt. 80.
- 27.4 miles. Continue straight E toward Ivoryton (do not following Rt. 80).
- 27.8 miles. Turn L onto Kelsey Hill Rd.
- 28.0 miles. Turn L into the Valley Regional High School grounds. Park the cars. Follow the hill slope along the west side of the hill to the new bleachers, and to a pavement outcrop of granitic gneiss.

**STOP 7. Kelsey Hill Complex. (20 minutes)** This outcrop was selected to show several things. First, the rocks in this belt are a mixture of granite-gneiss, tonalite gneiss, amphibolite, and dunite, suggesting a lithodemic complex, and not the Monson Formation of Rodgers (1985). This outcrop shows a granitic gneiss. Second, the dominant fabric in this rock now strikes ~320, and dips moderately to the NE. This trend follows the structure of 'Lundgren's Appendix,' and is outside the domain of the Killingworth complex. The rocks also show a long history of high-grade metamorphism, with older pegmatites strongly attenuated into the foliation, and younger ones only slightly affected by this deformation.

- 28.1 miles Return to cars. Bear left at the stop sign, and left again on Kelsey Hill Rd.  
 28.9 miles Crossing the terrane boundary between the Bronson Hill and Merrimack terranes. 29.0 miles  
 Outcrops of the polymetamorphic Hebron Formation.  
 29.3 miles Turn R (S) on Rt. 154 at the east end of Kelsey Hill Road. You are now driving along the strike of  
 "Lundgren's appendix".  
 30.8 miles Turn R (W) at the T intersection with the traffic light. Then take an immediate L (S) at the Cumberland  
 Farms, following the road to Clinton.  
 31.6 miles Turn R (S) on Rt. 154.  
 33.6 miles Pull over onto the soft shoulder opposite large road cuts on the L (E).

### STOP 8. Boulder Lake Gneiss.

(20 minutes) This stop is included to show a typical example of Boulder Lake gneiss and the locality from which the sample for dating was collected. This outcrop shows a plagioclase gneiss with minor hornblende and magnetite. Mafic inclusions are present, but not conspicuous, and calc-silicate xenoliths that are characteristic of this unit (Lundgren and Thurrell, 1973) are also present locally. Magnetite and biotite 'eyes' are present locally, probably reflecting porphyroblast-like growth at high and perhaps anatectic metamorphic conditions. The dominant foliation strikes ~210 and dips ~60W. It is cut by a pervasive  $S_2$  striking ~280 and dipping ~75 N and axial planar to  $F_2$  folds. This late fabric can be found in most outcrops at the west end of the Killingworth complex, and locally shows meters and even tens of meters of left lateral displacement. The rocks are cut by syenitic pegmatites and a few quartz veins. The pegmatites and quartz veins cannot have been magmatic; they are relatively undeformed and late, and probably Permian.

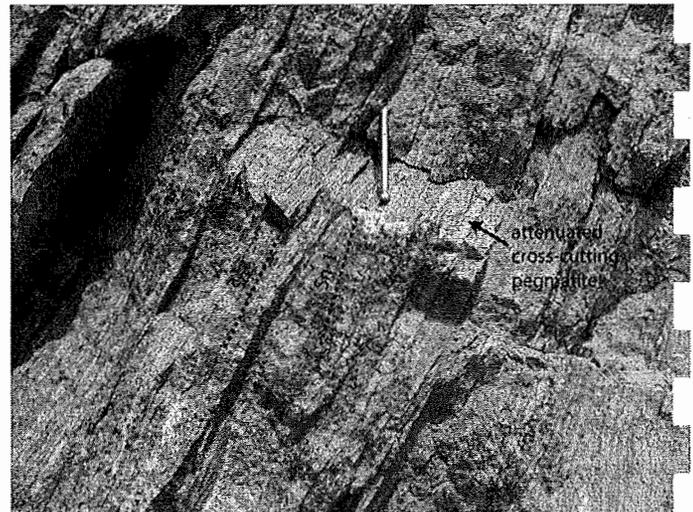


- 33.6 miles Return to cars. Continue south on Rt. 153.  
 36.0 miles Turn R (W) at the fork with Rt. 1, following signs for Clinton.  
 39.6 miles Passing Rt. 145, but continuing on Rt. 1.  
 40.8 miles Turn R (N) on North High St. (at the Loft Antiques, and Friendly's Restaurant). Follow North High St.  
 bearing R as it angles up hill to  
 41.1 miles #31 N. High St. Turn R following driveway to the factory. Bear L at the 'Y' and pass the plant keeping  
 it on your right.  
 41.4 miles Outcrop is under the wooded island in the middle of the parking lot.

**STOP 9. Clinton Gneiss.** (20 minutes) This stop is included in this trip to show an example of the generally poorly exposed Neoproterozoic granitic gneiss. We interpret these rocks to lie in the foot wall of a major suture between the Gander and Bronson Hill terranes. This outcrop shows a typical migmatitic and differentiated schistosity dipping steeply E and SE but cut by a later schistosity that strikes more EW. We infer from this differentiation that these fabrics developed with the help of an anatectic liquid.

- 41.4 miles Return to cars, and back track to North High St.  
 41.6 miles Turn L and return to Rt. 1 (West Main St.).  
 42.5 miles Turn R (N) on to Highland Drive and park in the lot of Snitzels foreign car parts store.

**STOP 10. Boulder Lake Gneiss and Late Paleozoic Suture.** (20 minutes) This outcrop is included in this trip to highlight a possible cause for the heterogeneity of the Killingworth complex. Regional geochemical analytical work suggests that this outcrop exposes the Boulder Lake gneiss of our earlier studies. This outcrop is in the immediate hanging wall of the boundary between the Clinton Granite Gneiss and the Boulder Lake gneiss, and its structures suggest that it lies in a ductile fault zone. The rock is unusually strongly layered, foliated, and differentiated. In these rocks biotite flakes have a strong preferred orientation, and tend to be segregated into discrete folia. Banding is parallel to this foliation, and is defined by the segregation of quartz and feldspar into discrete layers. Sn also envelops mono- and poly-crystalline inclusions of feldspars (upper figure) interpreted as tectonic inclusions derived from earlier pegmatites. This Sn fabric is in fact overprinting an earlier one Sn-1 locally preserved in "macrolithons" (lower figure). Earlier migmatitic and pegmatitic veins are strongly attenuated in both Sn and Sn-1. Finally, the intercalation of migmatitic layers (best viewed on the top of the eastern outcrop) suggests that much of this deformation was melt assisted. All of these features indicate that the strain in these rocks is very high, consistent with hypothesis that this contact is a major terrane boundary between Gander zone and Bronson Hill rocks.



- 42.5 miles Return to cars, and back to Rt. 1. Turn R (W) on Rt. 1.
- 43.5 miles Turn R at the light on the Hammonasset connector.
- 43.7 miles Passing cuts of Boulder Lake gneiss.
- 44.8 miles Enter I95 and follow this road to exit 58 (Rt. 77, Guilford, North Guilford) exit.
- 51.5 miles Exit 58. Your parking lot is at the bottom of the off ramp.

End of trip.

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