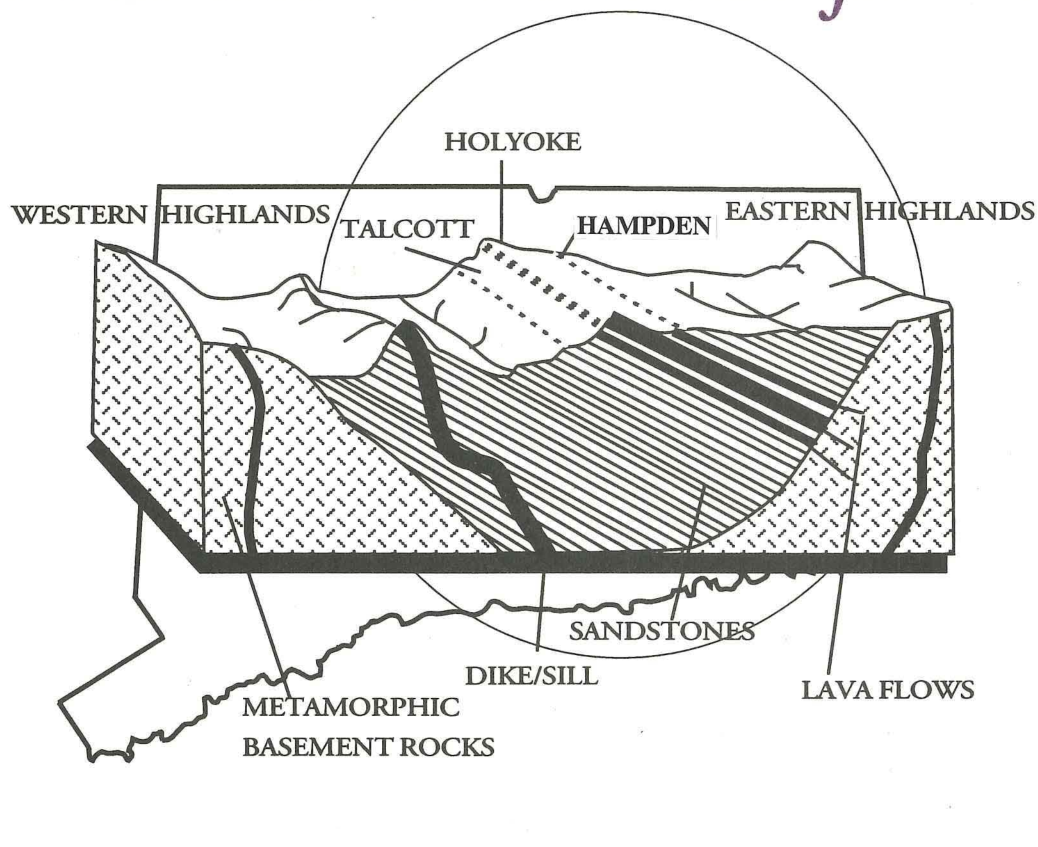
NORTHEAST SECTION
GEOLOGICAL SOCIETY OF AMERICA
30TH ANNUAL MEETING
CONNECTICUT

Guidebook for Fieldtrips in Eastern Connecticut and the Hartford Basin



STATE GEOLOGICAL AND NATURAL HISTORY SURVEY OF CONNECTICUT
THE NATURAL RESOURCES CENTER
DEPARTMENT OF ENVIRONMENTAL PROTECTION

MARCH 19, 20, 21, AND 22, 1995

Guidebook Number 7

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Guidebook for Fieldtrips in Eastern Connecticut and the Hartford Basin

Editor
Nancy W. McHone
State Geological and Natural History Survey of Connecticut
Guidebook Number 7
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State Geological and Natural History Survey of Connecticut

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Editor's Preface

It has been twenty-four years since the last Northeast Section of the Geological Society of America meeting in Connecticut. Since that time our understanding of the geological history of northeastern USA and southeastern Canada has greatly increased. The fieldtrips described in this guide incorporate, and add to, our understanding of that history.

Trip A examines metamorphic rocks, using mineral cooling ages to constrain the boundaries of terranes and the timing of terrane assembly. The sedimentary and basalt units of the Hartford Basin are the subjects of trips B and D. Variations within the sedimentary units are related both to climatic changes and to tectonic activity. Igneous rocks are the subject of Trip C, specifically the Holyoke Basalt. This unit is spectacurally exposed in a quarry in North Branford, Connecticut. Trip E is an in-depth examination of the Lantern Hill quartz lode, which has been well exposed by many years of quarrying.

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Trip A

Thermochronologic Evidence for Alleghanian Assembly of "Thermotectonic" Terranes, South Central New England

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Purpose of Field Trip

This field trip is designed to contrast the rock types, metamorphic grades, and especially metamorphic histories of the rocks of southeastern New England that constitute its lithotectonic terranes: the Avalon, Putnam-Nashoba, Merrimack, Central Maine, and Bronson Hill (Figure 1). We will focus on the lithologic, metamorphic, and geochronologic evidence for the definition of lithotectonic terranes, and show how contrasting Pressure-Temperature-time (PTt) paths reflected by the mineral cooling ages of the rocks in the various terranes require that significant assembly of these terranes post-dates middle to late Paleozoic metamorphism. We will show that much metamorphic history can be extracted from a metamorphic rock by dating its minerals, in spite of a low variance mineral assemblage that does not narrowly constrain the metamorphic conditions. We use the term "terrane" to refer to a body of rock with internal lithologic and metamorphic continuity, isolated by brittle and/or ductile faults. Thermochronologic data are critical to establishing a uniform cooling history within a terrane and discontinuities in mineral cooling ages define the boundaries of terranes. The term "thermotectonic" emphasizes this requirement that terranes share a continuity of metamorphic history.

The most contentious issue regarding the timing of terrane assembly has centered on the arrival of the Avalon terrane. Some argue for a Middle Ordovician arrival, causing the Taconic orogeny; some favor an early Devonian arrival causing the Acadian Orogeny; and some hold that at least the last stages of assembly occurred during the late Paleozoic Alleghanian orogeny. Stockmal *et al.* (1990) and van der Pluijm *et al.* (1990) have proposed that initial collision occurred in the Silurian, and continued in a protracted way into the late Paleozoic. Wintsch *et al.* (1992) go on to suggest that the arrival of western parts of Avalon (or its outboard terranes) underplated the eastern terranes, uplifting them, and causing the end of the Acadian orogeny. On this field trip we will examine some of the evidence for the latter proposal by contrasting middle and late Paleozoic PTt paths of the various terranes, and show how they provide evidence for the Late Paleozoic assembly of these terranes.

METAMORPHIC PRESSURE-TEMPERATURE-TIME PATHS

Identification of lithotectonic terranes has been based on both lithologic assemblages and on metamorphic history. Thermochronologic data in the form of mineral ages strongly compliment P-T data derived from mineral thermobarometry by defining the *time* interval of metamorphism. Mineral ages are well known to be reset by heating events, and in fact the ages of many minerals in igneous or high grade metamorphic rocks reflect the time of cooling from the last metamorphism, rather than the time of crystallization. Closure temperatures, or the temperatures below which minerals effectively retain radiogenic daughter isotopes, are for lead: monazite 720 ±20°C, and sphene 575 ± 50°C (Cliff and Cohen, 1980; Copeland *et al.*, 1988); and for argon: hornblende, 500 ±50°; muscovite, 350 ± 25°C; biotite, 300 ± 25°C; and K-feldspar, 200 ± 50°C (McDougall and Harrison, 1988).

Mineral ages are useful in several ways. For instance, in rocks east of the Willimantic window (Figure 2), hornblende ages reflect cooling from pre-Alleghanian metamorphism. However, in and southwest of the window, their younger Permian ages represent middle amphibolite facies Alleghanian overprinting. U-Pb ages of sphene behave in the same way except that sphene crystals that grow below their closure temperatures reflect the time of *crystallization*, because temperatures were not high enough to allow the diffusive loss of radiogenic lead. In the Avalon terrane, the Pennsylvanian ages of sphenes reflect prograde metamorphism to lower amphibolite facies conditions. Integration of these and other geologic and petrologic data provide very strong evidence for the Alleghanian assembly of these terranes. Some of the data upon which the trip is based is presented in Wintsch *et al.* (1991; 1992), Wintsch (1992) and Moecher and Wintsch (1994), and most of the text is taken from Wintsch *et al.* (1993). Prior reference to these is suggested.

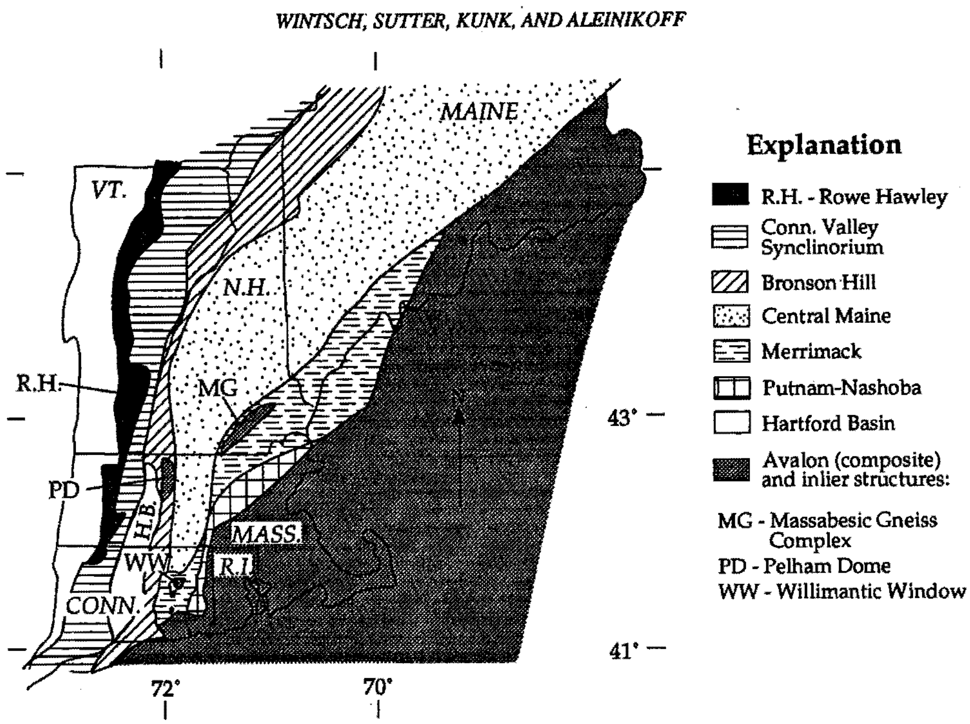

Figure 1. Generalized map of New England, showing the distribution of the lithotectonic terranes (modified from Zartman, 1988; Hutchinson et al., 1988; Robinson and Goldsmith, 1991). The locations of the Avalonian inliers in domal structures is indicated.

METAMORPHIC AND GEOLOGIC SETTING

The distribution of the high grade metamorphic rocks that comprise the terranes in the area of this field trip is shown in Fig 2. The overall NNE strike of the terranes reflects the gentle dips to the west of all the rocks, except in the vicinity of the Willimantic window. The Avalon terrane underlies all terranes to the west, and is exposed again in the Willimantic window, as are rocks of the Putnam-Nashoba terrane. All terranes are separated from one another by ductile (mylonitic) faults, which commonly mark discontinuities in metamorphic grade and/or regional cooling histories. Field trip stops included on this trip (Figure 2) are selected to show the maximum information about the terrane, primarily because of excellent exposure, but also because of the availability of petrologic and thermochronologic information.

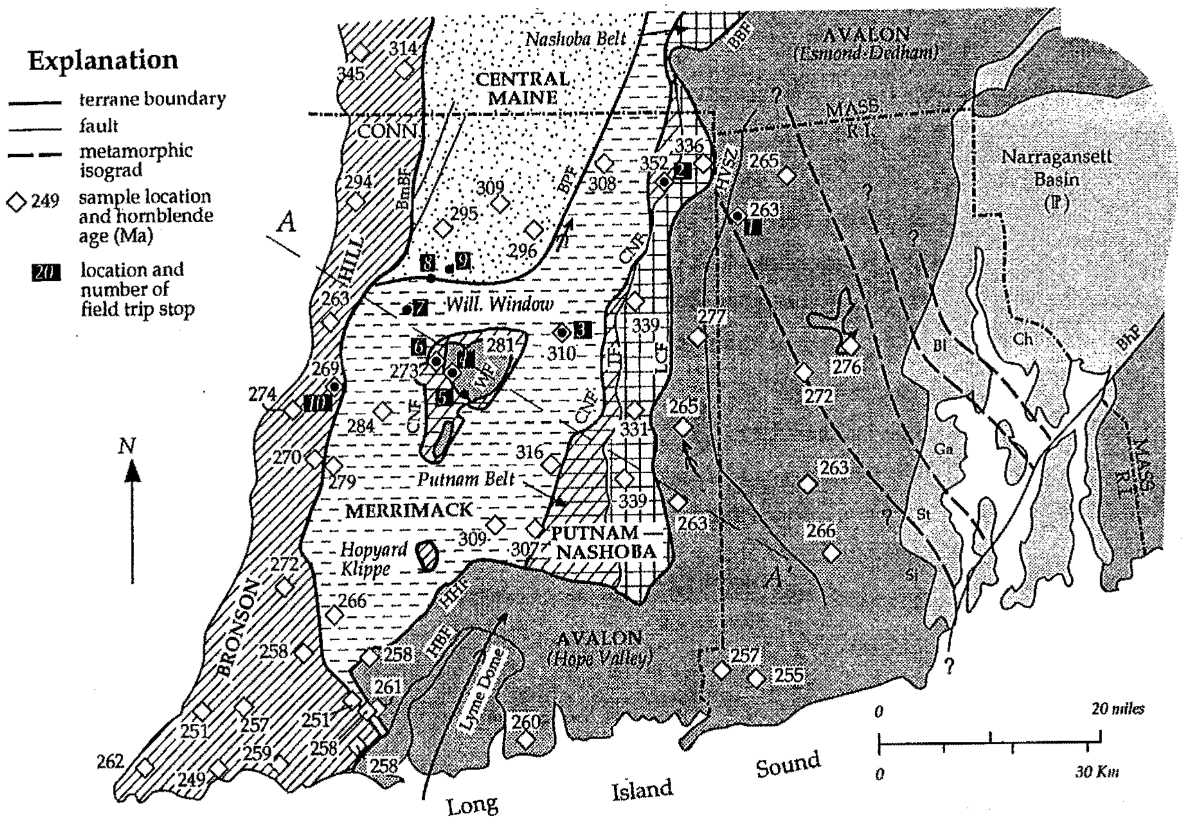

Figure 2. Map of southeastern New England, showing the distribution of terranes (after Wintsch, et al., 1992). Faults at terrane boundaries are: HHF, Honey Hill fault; LCF, Lake Char fault; BBF, Bloody Bluff fault; CNF, Clinton-Newbury fault; BPF, Black Pond fault; BmBF, Bonemill Brook fault. Faults within terranes: HBF, Hunts Brook fault; HVSZ, Hope Valley Shear zone; BhF, Beaverhead fault; TF, Tatnic fault. Alleghanian isograds in Rhode Island separate the Chlorite (Ch), Biotite (Bi), Garnet (Ga), Staurolite (St), and Sillimanite (Sl) zones.

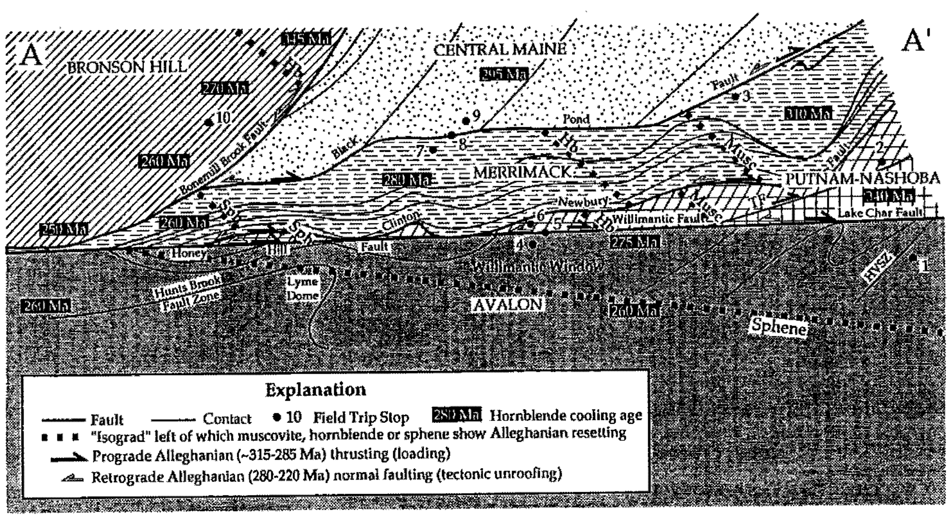

Figure 3. A partially schematic section across lithotectonic terranes in SE New England reconstructed from regional, quadrangle, and detailed (Wintsch, unpublished) mapping. The unfolded Honey Hill-Lake Char-Willimantic fault system is the plane of reference. The geometry of the terranes is projected onto plane A-A' (Fig. 2) with considerable vertical exaggeration. Alleghanian metamorphic history of each terrane is indicated by the sphene, hornblende, and muscovite 'isograds', that separate rocks containing mineral ages reset in the Alleghanian on the west from those to the east with unreset, pre-Alleghanian ages. A net thrust motion (→) is required by the structural position of terranes with younger cooling ages overlying terranes with older ages; significant thrust motion must postdate peak metamorphic conditions. Later (Permian-Triassic) reactivation of these faults in a normal sense (←), contributed to tectonic exhumation.

Avalon Terrane

The Avalon terrane consists of Late Proterozoic rocks of coastal New England, and three inliers of rocks in windows or domes in central New England west of the main coastal outcrop area (Figure 1). These three structures (the Willimantic window, Pelham dome, and Massabesic Gneiss complex) share with the main body of the Avalon terrane a Late Proterozoic crystallization age of igneous rocks and only a late Paleozoic metamorphism, and are thus correlated with it (Wintsch *et al.*, 1992). The presence of Avalonian rocks in these structures indicates that Avalonian rocks structurally underlie all terranes to the west at least as far as the Hartford basin. This common history further suggests that faults that cut this terrane (e.g. Hunts Brook, Hope Valley, Beaverhead, Figure 2) are less regionally significant than those that separate Avalon from structurally higher terranes.

The Avalon terrane contains primarily Late Proterozoic felsic plutonic (Stop 1), volcanic (Stop 4), and minor metasedimentary rocks. These are cut by Ordovician, Devonian, and late Paleozoic intrusive rocks, and several pre-, syn-, and post-metamorphic faults. The structural top of the Avalon terrane is defined by the Honey Hill-Lake Char -Willimantic (Stop 5) (and Bloody Bluff, Massachusetts) fault system. In the west, the fault was last active under middle to upper amphibolite facies conditions, but in the east it was active (or reactivated) at higher structural levels and greenschist facies conditions. Metavolcanic and metasedimentary rocks are cut out at a low angle and folded along the foot wall of the Honey Hill fault system (Figure 3).

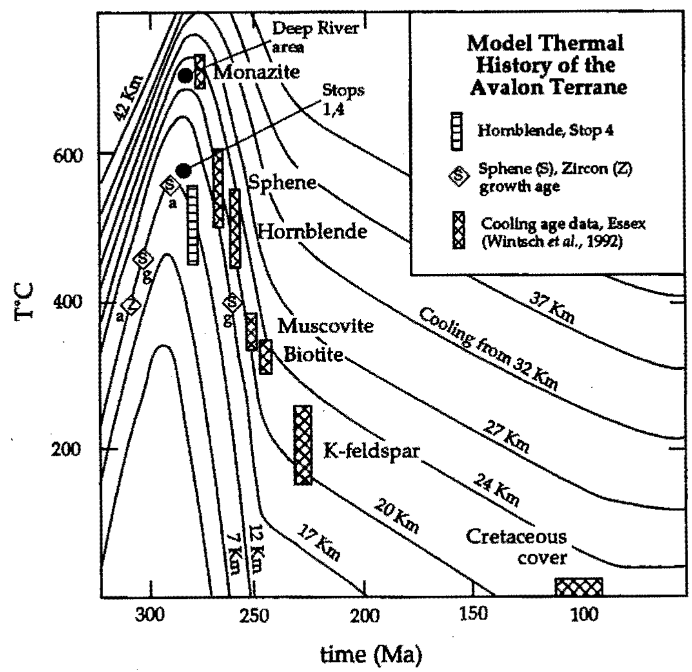

Figure 4. Cooling history of the southwestern part of the Avalon terrane, and model thermal history (from Wintsch, et al., 1992). Hornblende cooling ages from the structurally higher parts of the Avalon terrane are superimposed on the model cooling curves. Temperature coordinates for sphene and zircon ages interpreted as crystallization ages from the Willimantic Window (Getty and Gromet, 1992b; g) and near Chester (unlabeled, Aleinikoff, unpublished; a) are plotted to be consistent with the calculated Temperature-time curves.

The grade of metamorphism varies from unmetamorphosed Late Proterozoic igneous rocks and Cambrian and Pennsylvanian sedimentary rocks in eastern Massachusetts to upper amphibolite facies in southwestern Rhode Island, and southern Connecticut. Permian isograds (Figure 2) have a general northwest strike, with the Ponaganset Gneiss in northwestern Rhode Island lying approximately between the staurolite and sillimanite isograds (see Stop 1). In the Late Proterozoic rocks of eastern Connecticut this metamorphic gradient is manifest in muscovite + biotite ± garnet-bearing rocks conspicuously lacking migmatites and feldspar-bearing veins (Dixon, 1974; Moore, 1983; Stop 1), whereas to the south and west, abundant pegmatites and migmatites reflect anatectic conditions (Lundgren, 1966; Wintsch and Aleinikoff, 1987; Dipple *et al.*, 1990). Amphiboles in

TABLE 1. SUMMARY OF MINERAL APPARENT AGES

Terrane	*NW/ SW*	*Center*	*East*
Bronson Hill			
Sphene	350 / 284		
Hornblende	345 / 250		
Muscovite	250		
Biotite	245		
K-feldspar	220		
Central Maine			
Monazite	365(345)		
Sphene			
Hornblende	295		
Muscovite	250		
Biotite	248		
K-feldspar	>230		
Merrimack			
Sphene	305 / 281		315
Hornblende	266 / 251	285	310
Muscovite			250
Biotite	243		
K-feldspar			230
Putnam- Noshiba			
Monazite		400	
Sphene	-/285	335	350
Hornblende	-/264	280	340
Muscovite			275
K-feldspar			247
Avalon			
Zircon	270 L.I. (310)		
Monazite	278		
Sphene	270(295)	(305)	600 (315?)
Hornblende	260	280	277
Muscovite	250	247	250
Biotite	243	243	245
K-feldspar	230	228	>232

Table 1. Monazite, sphene data: U-Pb system. Hornblende, muscovite, biotite, K-feldspar data: $^{40}Ar/^{39}Ar$ system. All ages are cooling ages except those in ( ), which are interpreted as crystallization ages. L.I. refers to lower intercept of discordia. Columns refer to positions on section A-A', Fig. 2, where NW, Center and East refer to the northwestern, Willimantic, and eastern parts of the section. SW refers to the Deep River area (Wintsch, 1994). Data is from Wintsch et al (1992; 1993), Tucker and Robinson (1990), Getty and Gromet (1992b), Thomson et al. (1992), and Boyd et al.

structurally higher parts of the Avalon terrane are fine grained and acicular, but to the south and west they are coarse grained, nearly equant, show straight extinction, and generally share smooth, straight boundaries with all adjacent grains.

The time of metamorphism of these rocks is quite uniform; hornblende cooling ages are all late Paleozoic (Table 1; Figure 2), which precludes any km-scale tilting since the late Permian. However, the apparent metamorphic gradient south of the Honey Hill fault strongly suggests 10 km or more tilting of the Avalon terrane toward the north. A similar tilting to the west under the Lake Char fault is also likely. The history of metamorphism of the Avalon terrane rocks reflects relatively rapid loading and unloading during the Alleghanian orogeny. Rocks close to the structural top of this terrane, modeled by the 12 km isopleth of Fig 4, were heated and grew metamorphic sphene in the Pennsylvanian, reached peak metamorphic temperatures of < 600°C at about 280 Ma, and have hornblende cooling ages of slightly less than 280 Ma. Rocks deeper in the terrane in the southwest have monazite ages of ~280 Ma (Stop 18), and younger Permian sphene and hornblende cooling ages (Wintsch *et al.* 1993). The exposure of these deeper rocks reflects folding or tilting of the Avalon terrane between 280-265 Ma.

Putnam-Nashoba Terrane

Rocks of the Putnam-Nashoba terrane are exposed in eastern Massachusetts (Nashoba belt) and in eastern Connecticut (Putnam belt), and immediately overlie the Avalon terrane. These rocks are Late Proterozoic (?) pelitic metasedimentary and mafic metavolcanic rocks (Goldsmith, 1991). They were intruded by several post-metamorphic Silurian mafic rocks (Zartman and Naylor, 1984) and cut by several ductile faults (Dixon and Lundgren, 1968; Zen *et al.*, 1983), including the Tatnic fault (Figure 2). The base of this terrane is strongly cut out by the Honey Hill fault system, such that the lower mafic volcanic unit (Quinebaug Formation) is cut completely out, and farther west the entire terrane is missing (Wintsch *et al.* 1993; Wintsch 1994). The top of the terrane is cut out by the Clinton-Newbury fault.

Upper amphibolite facies metamorphic conditions of ~700°C and 5.7 kb (Moecher and Wintsch, 1994) predate Silurian (~425 Ma, Zartman and Naylor, 1984) intrusions in the Putnam belt and monazite mineral ages in the Willimantic window (~400 Ma, Getty and Gromet, 1992b). Available thermochronology suggests very slow cooling (3°C/m.y.) from a Silurian or older metamorphism to the Triassic (Table 1; Figure 5). Data from both the Putnam belt and the Willimantic window define parallel cooling curves of nearly identical age that support the lithotectonic correlation of the two belts of rock. In eastern exposures resetting by Alleghanian metamorphism was minimal; even muscovite apparent ages are consistent with slow cooling (Figure 5). However, in the Willimantic window (Stop 6) Alleghanian metamorphism (~600°C, 8 kb) has reset hornblende but not sphene, suggesting first sillimanite-grade Alleghanian metamorphic overprint (Moecher and Wintsch, 1994).

Merrimack Terrane

Rocks of the Merrimack terrane structurally overlie Putnam-Nashoba rocks (Figs. 2, 3) and are now correlated with the Oakdale Formation in Massachusetts (Pease, 1989; Robinson and Goldsmith, 1991) in the Merrimack Trough zone of Zartman (1988). East of the Willimantic window the rocks are dominated by calcareous metasiltstones and argillaceous metasediments. The base of the terrane is marked by the Clinton-Newbury fault, except in the Deep River area where the base is cut by the Honey Hill fault. North of the Willimantic Window, the top is cut by the Black Pond fault, which locally cuts out parts of the upper section, and is locally associated with low amplitude asymmetric folds. In the Deep River area, the top is cut out by the Bonemill Brook fault (Figure 3; Wintsch 1994).

These rocks are metamorphosed to staurolite-kyanite zone of the epidote-amphibolite facies. Preliminary thermochronology in these rocks shows that hornblendes in the east have not been affected by Alleghanian metamorphism; they show plateau or isochron ages of about 310 Ma whereas south of the Willimantic window they yield Permian ages (Figure 2) suggesting Alleghanian overprinting. Muscovite from the Scotland Schist member shows a ~250 Ma age, which is consistent with slow cooling from pre-Alleghanian metamorphism (Figure 5), and cannot be unambiguously assigned to cooling from Alleghanian metamorphic overprinting.

Central Maine Terrane

Rocks in the Central Maine terrane are dominated by calcareous and pelitic metasediments, metamorphosed to upper amphibolite and lower granulite facies conditions (Thomson *et al.*, 1992). Summaries of rock descriptions are given in Peper *et al.* (1975) and Robinson and Goldsmith (1991). The basal Black Pond fault (Figs 2, 3) cuts out the base of most units. The upper part of the terrane is cut by the Bonemill Brook fault.

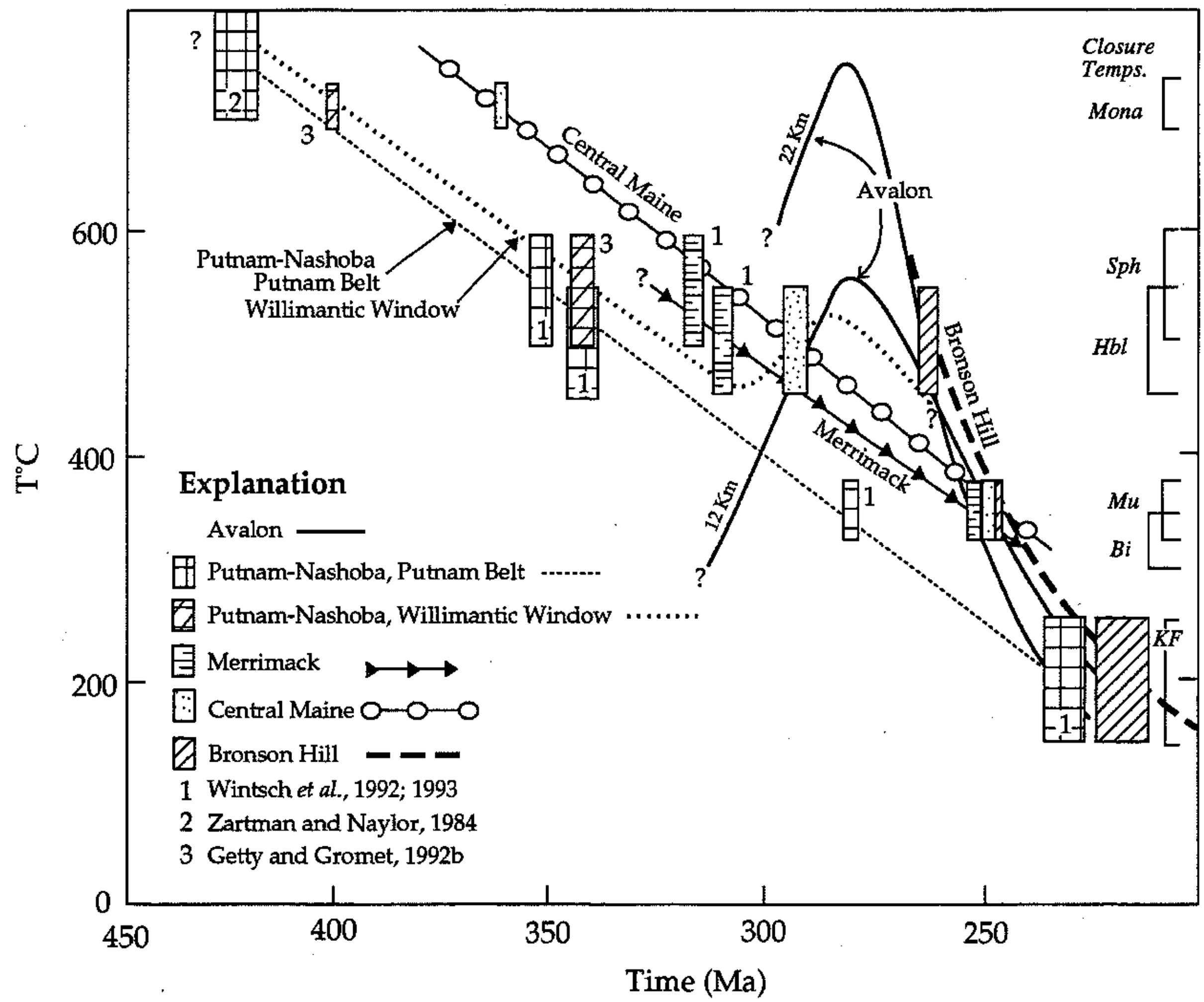

Figure 5. Comparison of the thermochronologic data from the Avalon, Putnam-Nashoba, Merrimack, Central Maine, and Bronson Hill terranes. Although some data are preliminary, the trend of decreasing cooling age from the eastern Putnam-Nashoba to the western Bronson Hill terrane is evident.

Available geochronology on monazites (Thompson *et al.*, 1992) suggests peak metamorphic conditions ceased by the late Devonian (~365 Ma) with younger monazite ages probably recording crystallization events. Our preliminary hornblende and muscovite ages of ~295 Ma and 250 Ma, respectively, suggest the end of amphibolite facies metamorphism by the Permian. Again, late Permian muscovite ages are consistent both with slow post-Acadian cooling, and with cooling from Alleghanian overprinting (Table 1).

Bronson Hill Terrane

Rocks of the Bronson Hill terrane are dominated by granodioritic orthogneisses (Stop 10), and dacitic, dioritic and basaltic metavolcanics; (Webster and Wintsch, 1987; Leo *et al.*, 1984), but metapelitic, calcsilicate and manganiferous (coticule) rocks are also present. Igneous activity is late Ordovician (Zartman and Leo, 1985; Tucker and Robinson, 1990). The base of the Bronson Hill terrane is marked by the Bonemill Brook fault. It deforms an amphibolite along much of its length, and locally cuts it out completely. A 100-million-year north-to-south age gradient in hornblende ages (Figure 2) shows cooling from middle amphibolite facies Acadian metamorphism in the north, and from upper amphibolite facies Alleghanian metamorphism in the south. The latter is probably an Alleghanian overprint on an earlier metamorphism (Boyd *et al.* 1993).

DUCTILE FAULTS AND TERRANE BOUNDARIES

All terranes are separated by ductile faults. All are pre- or syn-metamorphic, but some have been reactivated under post-peak metamorphic greenschist facies and even brittle conditions. Consequently, a wide variety of fault rock types are present among terrane boundaries, from mylonitic schists and gneisses (and straight gneisses [Stops 2, 6, 8]), to blastomylonites, phyllonites, and cataclasites (Stops 5, 10, 11, 12, 14). The spectrum of faults and fault rocks (Figure 6) is remarkable because some of the most spectacular mylonitic fault zones occur within terranes, and some of the most regionally significant faults are the least impressive in outcrop as high strain zones. Lithologic boundaries are quite sharp, with a minimum of imbrication (Stop 5).

Tectonic blocks are common in many ductilely deformed rocks; they probably formed during extension under simple shear, and can be thought of as huge boudins in various stages of rotation (Wintsch, 1979; Fig 7). They are characteristic of the pelitic unit in the Putnam-Nashoba terrane, especially near the base of the Tatnic Hill Formation and can be seen in various stages of development (compare Stops 2 and 6) along the Tatnic and Willimantic faults, and at the Black Pond fault. They are also exposed along the Honey Hill fault (Dixson *et al.*, 1968, Stop 3; Wintsch, 1985, Stop 7).

Most ductile faults have been intruded by pegmatite bodies, commonly at several stages of strain in the fault zone. Consequently, a wide spectrum of textures and structures are exposed in some fault zones, from strongly foliated, layered, and boudinaged gneisses to weakly layered and foliated tabular bodies, to undeformed dikes. We argue that at least some of these bodies were deposited from hydrothermal fluids rather than crystallization from silicate liquids. This opens the tantalizing question of what the compositions and pressures of aqueous fluids were during ductile faulting in the lower crust.

TECTONIC SIGNIFICANCE OF METAMORPHIC HISTORY

Putnam-Nashoba Terrane, and History of Motion of the Honey Hill Fault System

Several lines of evidence reflect important motion on the Honey Hill fault system. The regional truncation and folding of structurally higher rocks below the plane of the Honey Hill fault reflect thrust motion of the hanging wall. The gentle east dip of the boundary between reset and unreset sphene, and the decrease in hornblende ages to the west reflect the deeper level and higher grade of Alleghanian metamorphism in the west. Rocks of the Putnam-Nashoba terrane are also progressively cut out to the west. In the east, the Lake Char fault separates the Avalon terrane from the basal, mafic unit (Quinebaug formation) of the Putnam-Nashoba terrane. In the Willimantic window, the Willimantic fault places the upper, pelitic unit (Tatnic Hill formation) directly on Avalon terrane rocks (Figure 7). Thus the Avalon terrane boundary apparently cuts out the lower portion of the Putnam-Nashoba terrane. In the Deep River area, the entire Putnam-Nashoba terrane is missing (Wintsch *et al.* 1993; Wintsch 1994).

The cooling history of the rocks from the Putnam-Nashoba terrane contrasts strongly with that of the Avalon terrane. Peak metamorphic conditions of about 700°C and 6 kb (Hudson, 1982) in these rocks occurred in the Silurian or older, while in the underlying Avalonian rocks they were reached in the late Paleozoic (Figure 4). The 80-million-year difference between the hornblende cooling ages (Figure 5) thus precludes the possibility of thermal equilibrium between Putnam-Nashoba and Avalon terrane rocks during the Alleghanian heating of the latter (Wintsch *et al.*, 1992). On the contrary, it requires that Avalonian rocks were undergoing high-grade metamorphism, while Putnam-Nashoba rocks were at lower greenschist facies conditions. Given normal geothermal gradients, this approximately 300°C temperature difference requires the presence of 8-10 km of rock between the Putnam-Nashoba and Avalonian rocks at the Late Pennsylvanian time of peak metamorphism. This thickness of rock must have been removed in the Permian, because by Triassic time Avalon and cover rocks were adjacent, and cooling together (Wintsch *et al.*, 1992). The truncation of isograds at the western edge of the Avalon zone by the Lake Char and Bloody Bluff faults (Figure 2) further indicates important motion on this fault zone that post-dates peak metamorphic conditions in both the hanging wall and foot wall rocks. The data require that: (1) Alleghanian metamorphism in the Avalon composite terrane occurred elsewhere, remote from rocks now exposed in the Putnam-Nashoba zone, and (2) peak metamorphism in the Avalon terrane predates significant late movement on the Honey Hill fault system (contrary to O'Hara, 1986).

The same argument can be made for the Willimantic window. Avalonian core rocks have exclusively an Alleghanian metamorphic history (Wintsch *et al.*, 1992) where Late Proterozoic and Pennsylvanian sphenes are not fully reset (Getty and Gromet, 1992b), showing that temperatures did not exceed ~600°C (Stop 4). In the overlying Putnam-Nashoba rocks, hornblende (~280 Ma) is reset (Wintsch *et al.*, 1992) but sphene and monazite (Getty and Gromet, 1992b) still record cooling from a pre-Acadian metamorphic event. The resulting event is well preserved in local syntectonic metamorphic mineral assemblages and fabrics (Stop 6), recording prograde Alleghanian P-T conditions of nearly 600°C and from 8 to 5 kb (Moecher and Wintsch, 1994), superimposed on the earlier event. Only in the actual terrane boundary fault (Stop 5) are the Putnam-Nashoba rocks so fully reconstituted.

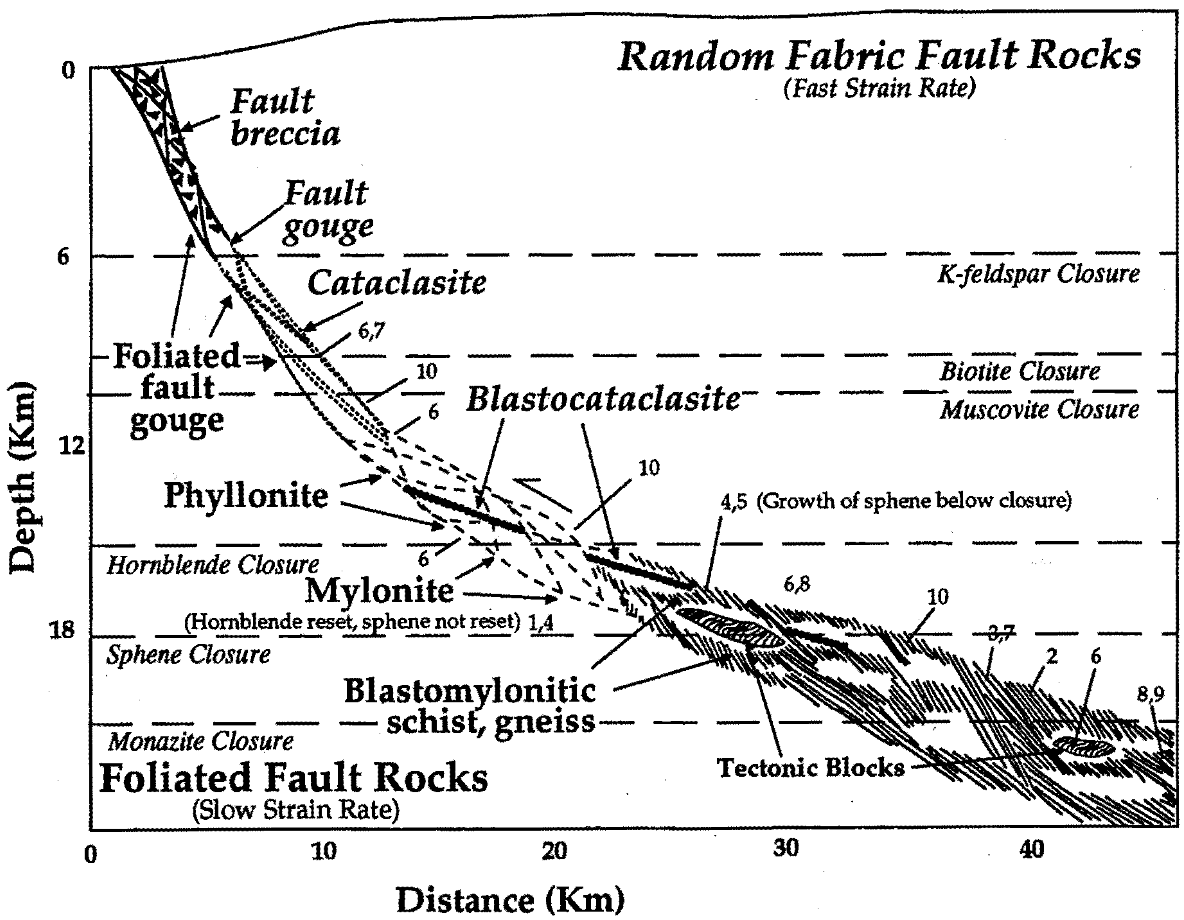

Figure 6. Conceptual model of a fault zone, showing possible P-T conditions of formation of various fault rocks: foliated fault rocks are in bold type; random fabric fault rocks are shown in italics. Numbers refer to field trip stops. The reactivation of ductile fault rocks under lower P-T and even brittle (Stops 6, 10) conditions indicates that exhumation of the fault block was partially tectonic, and that the position of the hanging wall (or footwall) relative to the active fault did not change during exhumation.

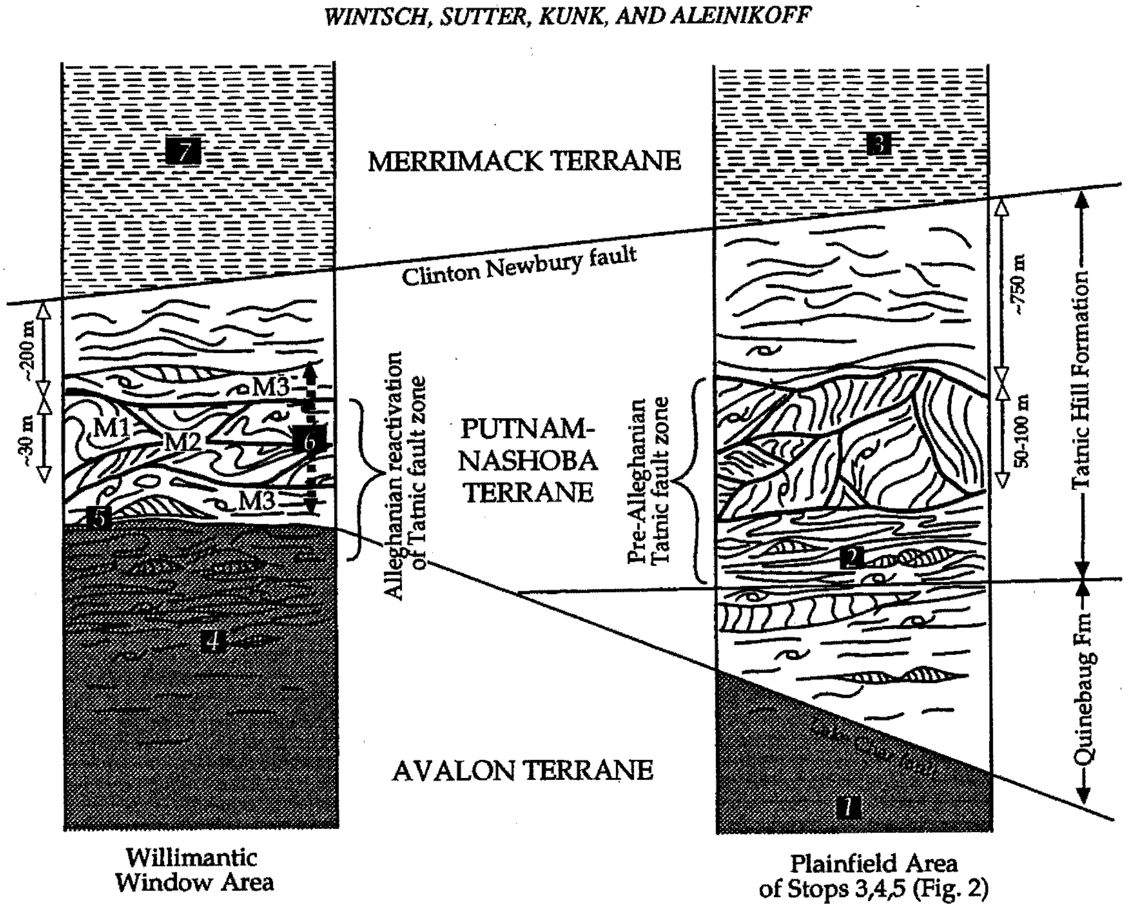

Figure 7. Partially schematic summary of the ductile structures in the Putnam-Nashoba terrane (after Wintsch, 1976), showing the relative position of pre-Alleghanian tectonic blocks in the Putnam belt (Fig.2) with the mafic Quinebaug Formation below. In the Willimantic window the Quinebaug Formation is cut out, and the zone of tectonic blocks is reactivated in the Alleghanian. All Putnam-Nashoba rocks were metamorphosed in pre-Devonian times (Dixon, 1982b), but in the Willimantic window this early metamorphism (M1) has been overprinted by Alleghanian high pressure staurolite + kyanite grade (M2) and lower pressure sillimanite-(andalusite) (M3) metamorphisms (Moecher and Wintsch, 1994). Numbers in sections refer to relative position of field trip stops.

Merrimack Terrane and History of Motion on the Clinton-Newbury Fault

The Clinton-Newbury fault separates the Putnam-Nashoba and Merrimack terranes, but its location in Connecticut is not known in detail. Its position on Figure 2 is put at the base of the lowest calc-silicate-bearing-Merrimack-like rock (locally named the Fly Pond member of the Tatnic Hill formation, previously assigned to the Putnam-Nashoba terrane). Rocks in this zone are blastomylonitic suggesting high strain, and this position allows correlation of the Pennsylvanian cooling ages of hornblende samples just above this boundary (307 and 316 Ma) with similar Merrimack terrane cooling ages in contrast to older cooling ages of Putnam-Nashoba rocks. Thus these hornblende cooling ages contribute to defining the terrane boundary.

The geometry and metamorphic history of these rocks suggest net thrust displacement along this boundary. Rocks of the Putnam-Nashoba terrane (north of Stop 3) in the footwall of the Clinton-Newbury fault are locally cut out, as are some Merrimack rocks in the hanging wall in the west, where the Honey Hill and Clinton-Newbury faults touch (Wintsch *et al.* 1993). Hornblende and sphene ages reflect Pennsylvanian cooling, but are younger than the same minerals in the underlying Putnam-Nashoba rocks. Because rocks structurally overlying and in thermal equilibrium with Putnam-Nashoba rocks would have cooled *before* the rocks below them, hornblende cooling ages from Merrimack rocks would be expected to predate hornblende cooling ages from Putnam-Nashoba rocks by approximately 5-10 million years. Merrimack rocks could not have cooled after the rocks of the Putnam-

Nashoba zone if they had been overlying them during the cooling of both. This is another line of evidence for thrust motion in the Alleghanian prograde metamorphism of Putnam-Nashoba rocks in the Willimantic window. In the window Putnam-Nashoba rocks show Alleghanian loading of ~ 3 kb and have hornblendes that are reset. Thus, the rocks of the Merrimack terrane must be allochthonous, and must have overridden the Putnam-Nashoba rocks since ~310 Ma, probably in the early Permian.

Central Maine Terrane and the Black Pond Fault

Rocks of the Central Maine terrane are separated from those of the Merrimack terrane by the Black Pond Fault. The upper part of the Merrimack terrane is cut out and map scale folds are present in the foot wall (north of Stop 3). Rocks at the base of the Central Maine terrane in the hanging wall are also strongly cut out. Motion sense here is readily argued to be thrusting.

Parallel to the arguments above, the thermal disequilibrium between rocks of the Central Maine terrane and the underlying Merrimack terrane suggested by the higher grade and younger age of hornblendes in the former require a net thrust motion. Moreover, Merrimack hornblendes in the west, deeper in the orogen, have been reset, and yield Permian cooling ages, which also require Alleghanian loading, presumably from rocks in the hanging wall. Work on these rocks is in progress.

Bronson Hill Terrane and the Bonemill Brook fault

The Bronson Hill terrane is separated from rocks to the east everywhere by the ductile Bonemill Brook fault system. Rocks and faults at the top of the Central Maine terrane are cut out by the Bonemill Brook fault. The rocks in the Bronson Hill terrane hanging wall are apparently less deformed. However, there is an important drop in metamorphic grade in rocks of the Bronson Hill terrane relative to rocks to the east. North and west of the Willimantic window the boundary dips west, where rocks of the Bronson Hill terrane cooled to hornblende closure temperature 10-15 million years later than the rocks of the Central Maine terrane to the east. Here, again, reverse motion on the Bonemill Brook fault system is implied. Stronger evidence for thrust motion is the presence of the small Hopyard Klippe of Bronson Hill rocks south of the Willimantic window (Figure 2). In the Deep River area the Bonemill Brook fault cuts out all of the Central Maine terrane, and much of the top of the Merrimack terrane. There the last motion was strike-slip.

Assembly of Lithotectonic Terranes in Southeastern New England

In summary, there is a progressive decrease in metamorphic age in terranes from east to west, from structurally lower to higher level rocks. Net thrust motion is required by the presence of metamorphic rocks with younger cooling ages overlying older, and in some cases by higher grade metamorphism overlying lower grade. A sequence of hinterland propagating thrust nappes is required by these results. This stack of nappes was thicker in the west, as shown by the reset ages and younger, higher pressure metamorphism in the west (Moecher and Wintsch 1994), and occurred at a time just older than the oldest reset ages, or ~280 Ma. By late Permian, however, rocks of all terranes were cooling (Figure 5). The cooling reflects exhumation, in part tectonic exhumation, with west dipping faults and terrane boundaries reactivated in a normal sense (i.e. hanging wall moves NW). This superposition of NW normal motion upon SE verging structures has led to spirited debate over the significance of kinematic indicators that reflect the various stages of motion. Our results show that the older loading history must have been achieved by thrusting. Kinematic indicators suggesting normal motion found in the younger Permo-Triassic (Goldstein, 1989; Getty and Gromet, 1992a) retrograde fault zones reflect the overprinting and destruction of earlier, ductile fabrics. It does not, however, provide evidence for a lack of loading during Alleghanian orogenesis in southeast New England; rather, it provides evidence that post-Alleghanian exhumation denudation was in part tectonic.

CONCLUSIONS

The distribution of lithotectonic terranes in southeastern New England reflects post-Acadian, Alleghanian, uplift and stacking in a westward (hinterland) propagating sequence of thrust nappes, followed by reactivation of these boundaries during NW verging tectonic exhumation by normal motion. Thus all terrane boundaries were active as thrust faults with regionally significant motion, but many were reactivated as normal faults, along which ductile fault fabrics and mineral assemblages were overprinted by lower grade fabrics and mineral assemblages.

ACKNOWLEDGEMENTS

This study has benefited from conversations with H.R. Dixon, R. Goldsmith, L. Lundgren, S. Quarrier, M.H. Pease, and the Connecticut Geological and Natural History Survey. Jean Reese is thanked for preparing drafts of figures as well as text. Appreciation goes to B. Hellickson, P. Ganguli and C. Jones for help in preparing the road log and to the McHones for support in the field. This research was partially supported by the U.S. Geological Survey, the Connecticut Geological and Natural History Survey, and NSF Grants EAR-8618305, 8803540, and 9104495.

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ROAD LOG

Mileage

Road log begins at Ma's Frosty Restaurant at the intersection of US 44 and SR 21. This can be reached by following US 44 west from its interchange with I 395 near Putnam, in northeast Connecticut.

0.0 Ma's Frosty Restaurant. Follow US 44 east.

2.5 Approximate trace of the Lake Char fault, separating Avalon and Putnam-Nashoba rocks.

3.6 Rhode Island state line.

5.6 Low road cuts on the left (N) side of the highway. Carefully make U-turn arc on north side of road.

STOP 1. AVALON TERRANE (30 Minutes)

Thompson Quad (Dixon, 1974; Wintsch, 1992). The purpose of this stop is to show a typical Avalonian granodioritic orthogneiss, and to show how much history can be extracted from it in spite of relatively poor exposure, and lack of constraining metamorphic mineral assemblages. The rock is a grayish-tan, massive weathering well foliated, plagioclase, quartz, microcline, biotite, epidote, hornblende, sphene gneiss, probably orthogneiss: the Late Proterozoic Ponaganset Gneiss (Dixon, 1974). Foliation and a conspicuous lineation (30°N15E) is defined by the parallel alignment of disseminated biotite flakes, but biotite-rich folia also give the rock a locally banded appearance. In contrast to K-feldspar porphyroclasts up to 1 cm in diameter, equant to elongate matrix grains of K-feldspar with weak cross-hatch twinning are 200 to 500 µm long. Hornblende (100 x 200 µm) is only ~2% of the rock, occurring as anhedral equant to elongate grains, always surrounded by, and locally embayed by biotite and epidote. These textures suggest the retrograde replacement reaction: hornblende + K-feldspar + Quartz + H_2O = biotite + epidote.

This outcrop contains a strong record of cooling from Alleghanian metamorphism. Hornblende from this outcrop plateaus at 263 Ma, biotite total fusion age is 249 Ma, and K-feldspar gives early Triassic ages (Table 1). However, sphene from this rock gives a concordant age of 602 Ma. Thus the temperature of Alleghanian metamorphism in this rock exceeded the closure temperature of hornblende, but not of sphene, and must have been ~550°C. In fact the age of 602 Ma is very easily interpreted to be a cooling age following late Paleozoic intrusion and cooling. This rock, then, was heated to about first sillimanite isograd conditions in the Permian, but last experienced second sillimanite grade temperatures in the late Precambrian. Given this cooling history, and a temperature estimate for hornblende replacement (above) of ~450°C, the lineation is constrained to be latest Permian, and a conspicuous flat quartz vein crystallizing in the lowest greenschist facies must be early Triassic.

5.6 Follow US 44 west from STOP 1.

7.3 Turn R (N) on SR21 at Ma Frosty's.

9.1 Turn R (N) on SR 193.

9.5 Jct. of SR 193 and 200 in the village of Thompson, and the approximate trace of the Tatnic fault. Continue NW on SR 200.

10.9 Intersection of Ct. Rts. 200 and 12. Turn left (S) on Ct. 12.

11.9 Bear left, leaving Ct. 12 (to the right), and enter the ramp to I395.

12.0 Park off the right shoulder of the ramp, before the RR overpass. Walk 100m up the ramp to Stop 2.

STOP 2. PUTNAM-NASHOBA TERRANE (45 Minutes)

Putnam Quad (Dixon, 1982a; Wintsch, 1992, Wintsch *et al.* 1993). The purpose of this stop is to show several characteristic rock types and structures of the upper pelitic unit of the Putnam-Nashoba terrane, here close to the base of the Tatnic Hill formation near the Tatnic fault. Foliations dip gently west, and steep ductile shear zones begin to define tectonic blocks (Figure Stop 2.1). The rocks of this zone are significant because they are clearly very high grade, above the second sillimanite isograd of the upper amphibolite facies. Garnets from this and nearby outcrops show only retrograde zoning (Figure Stop 2.2), and garnet-biotite geothermometry suggests temperatures of at least 650°C (Hudson, 1982). These rocks are higher in grade than rocks in the footwall Avalon terrane. The limited data available from the hanging wall Merrimack terrane (F, Y; Figure Stop 2.2) suggests lower grade, and even prograde zoning, again suggesting lower grade metamorphic conditions in these rocks.

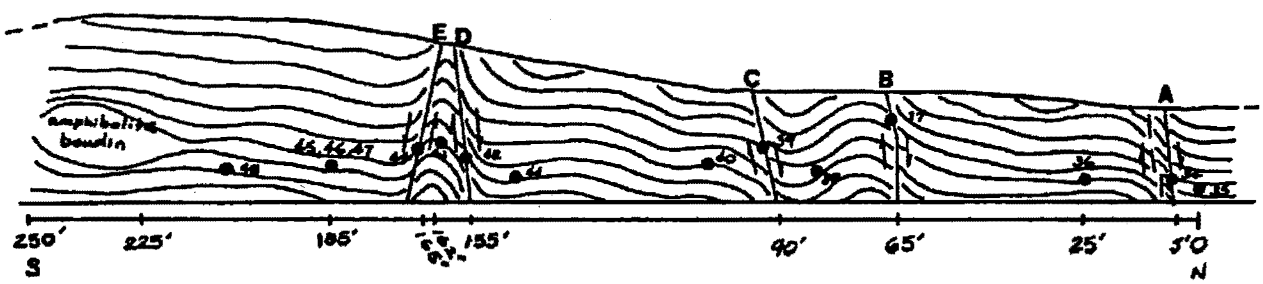

Figure Stop 2.1. Field sketch (from Hudson, 1982) of the western-most exposure at Stop 2, showing pre-Alleghanian foliation (strike parallel) categorized by dominantly north dipping, normal faults.

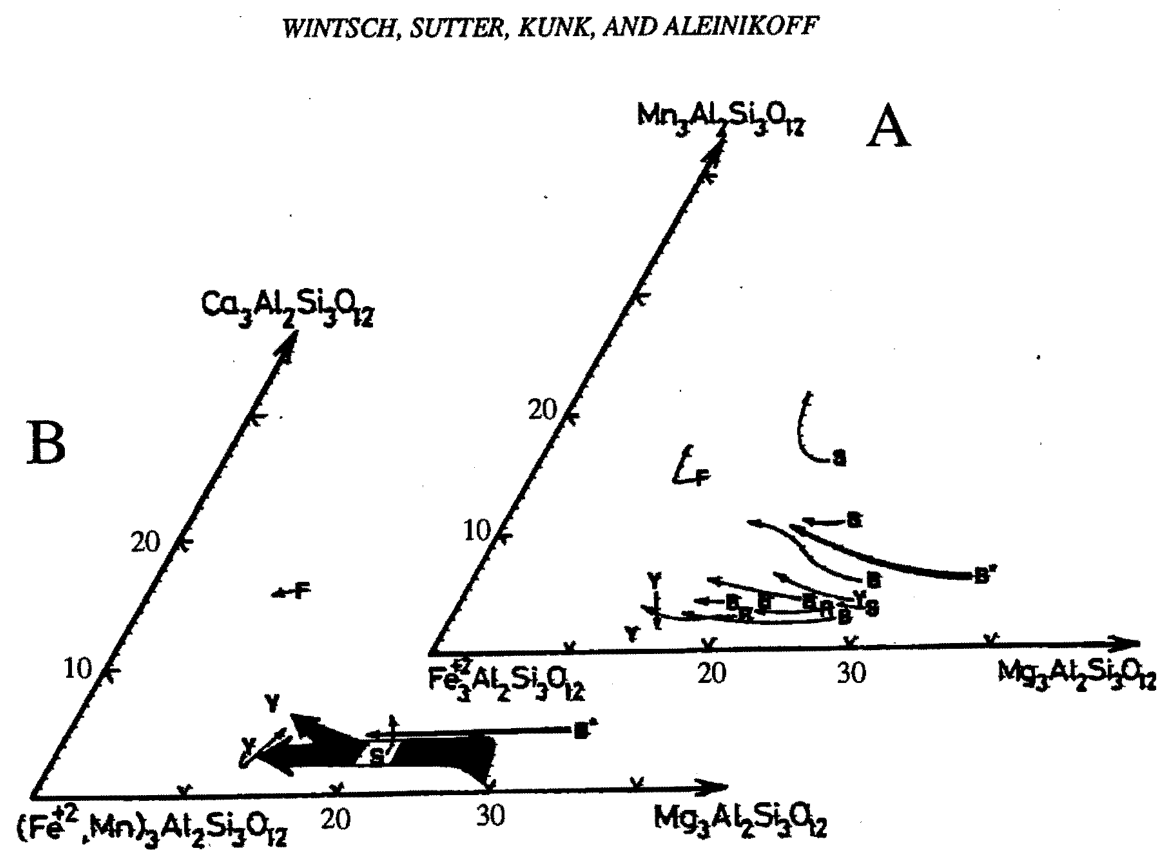

Figure Stop 2.2. Summary of compositional zoning in garnets (Hudson, 1982) from the Putnam-Nashoba terrane (R, B, S) and the Merrimack terrane (F, Y), from the Putnam area (near Stop 2). Arrowheads define rim compositions.

Part of the argument for a net thrust motion on the Lake Char fault comes from the higher grade Putnam-Nashoba rocks occurring over lower grade Avalonian rocks across this fault. In sharp contrast to Stop 1, amphiboles and sphenes from nearby rocks yield late Devonian cooling ages (Wintsch *et al.*, 1992; Wintsch, 1992), and indicate that peak metamorphism and most of the deformation in these rocks is Devonian or older.

12.0	Return to cars. Proceed south on I395.
13.8	Exit at interchange 97, following US 44 W through Putnam.
15.1	US44 joins SR12. Follow US 44.
17.1	Pomfret town line.
23.7	Jct. SR97. Turn L (S) on SR97 in Abington.
30.4	Jct. US 6. Turn L (E) on US 6.
30.6	Road cuts on N side of US 6, Blasted in May, 1993.

STOP 3. MERRIMACK TERRANE (30 Minutes)

Hampton Quad (Dixon and Pessl, 1966; Dixon *et al.*, 1968; Wintsch, 1992). The purpose of this stop is to show the ductile deformation of the well layered biotite schist and calc-silicate-granofels characteristic of the Merrimack terrane. This rock is dark greenish gray and purplish gray, medium grained, layered (mm-cm scale) hornblende granofels and biotite schist. The granofels layers, 1/2-2 cm thick, contain 100-300 mm grains of essential scapolite (porphyroblastic), hornblende, plagioclase and quartz, all in random orientation, with accessory zoisite, clinozoisite, calcite, sphene and zircon. Subhedral porphyroblastic hornblende grains are between 100-1000 mm long, and 50-300 mm wide. They very commonly contain rounded inclusions of scapolite, quartz, and plagioclase, but show straight extinction, and uniform pale green birefringence. The biotite schist layers 0.5-1.5 cm thick contain essential quartz, plagioclase, biotite and scapolite, with accessory zoisite, hornblende, sphene and zircon. All grains are anhedral except biotite, which is subhedral and evenly disseminated; it does not define biotite-rich folia. Several isoclinal Z-folds are typically overturned to the east (Dixon *et al.*, 1968), and here plunge ~15° N60W. They probably reflect eastward vergence of thrust nappes over and under these rocks.

The age spectrum from hornblende in this outcrop plateaus at 310 ± 2 Ma, and sphene yields an age of 315 ± 5. In contrast, muscovite and biotite are Permian. The Pennsylvanian age of this sample is distinctly different from hornblende ages in both the Avalon and Putnam-Nashoba terranes of Stops 1 and 2. In fact the *younger* age of this hornblende requires that it cooled about 30 m.y. after the hornblende in the structurally lower Putnam-Nashoba terrane rocks.. Normal cooling in a stack of rocks in thermal equilibrium would require that structurally higher rocks cool before those below. Thus the younger age of these hornblendes and sphenes requires that Merrimack rocks cooled to below about 500°C elsewhere (to the west?), and have been thrusted over Putnam-Nashoba rocks since amphibolite facies conditions, or in post-Pennsylvanian times.

30.6	Return to cars, continue west on US Rt. 6.
35.1	Crossing approximate trace of Clinton-Newbury fault zone.
36.5	Road cuts on left (S) are basal Putnam-Nashoba zone (Wintsch and Fout, 1982, Stop 3).
37.0	Crossing the Willimantic fault (Putnam-Nashoba - Avalon zone boundary.
38.6	Bear right on to U.S. Rt. 6 (limited access).
41.6	Exit at ramp to Ct. Rt. 32.
41.9	Park at bottom of ramp, well on to shoulder.

STOP 4. AVALON ZONE, WILLIMANTIC WINDOW (45 Minutes)

Willimantic Quad. (Snyder, 1964b; Wintsch and Fout, 1982; Wintsch, 1992). The purpose of this stop is to show the metavolcanic rocks at the structural top of the Avalon terrane, to document middle amphibolite facies metamorphism and ductile deformation, and to show that feldspar-rich sill- and dike- like structures can not be magmatic. These rocks contain Late Proterozoic metavolcanic Mansfield Hollow Lithofacies of the Hadlyme Formation (Wintsch *et al.*, 1990), and are correlated to part of the Waterford complex in the New London area. The most abundant rock type is a pale gray, massive weathering, well foliated and layered plagioclase, quartz, biotite K-feldspar, sphere, (muscovite or hornblende) gneiss and granofels. A second conspicuous rock type is a dark gray to black, massive, unlayered, well foliated hornblende, plagioclase, biotite, magnetite, sphene amphibolite. The outcrop is cut by several 0.5-1.0 m thick zoned granitic pegmatites, and by felsic sill-like layers parallel to the layering in the gneiss. The pegmatites are cored by quartz, have K-feldspar intermediate zones, and muscovite bearing margins. The sill-like structures are dominated by and are thickest where microcline-rich porphyroblasts bow out the surrounding foliation. They are joined together in necklace fashion by thinner layers of quartz and plagioclase.

Hornblende from the amphibolite produces an isochron age of 281 Ma on an isotope correlation diagram (Wintsch, 1992). Muscovite from a cross-cutting pegmatite, and biotite, and K-feldspar from the plagioclase gneiss yield cooling ages of 247, 243, and 228 Ma respectively. The cooling curve produced by these data (Figure 4) is very similar to that for the Hope Valley zone, except for a 15 m.y. older hornblende. Incompletely reset Late Proterozoic and Pennsylvanian sphene (Getty and Gromet, 1992b) in these rocks shows that the rocks did not reach the upper amphibolite facies temperatures of sphene closure (550-600°C) in the Alleghanian.

Knowledge that these rocks never exceeded 600°C is critical to interpreting the pegmatites and feldspar-rich 'sills,' in that they could not have been produced by the crystallization of silicate liquids. Partial melting is not a possibility, because even the most volatile-rich liquids do not form below 600°C. Thus all of these structures must have crystallized from an H_2O-rich liquid. The rocks also show moderate ductile deformation, but the necks between amphibolite boudins are rather straight, and are filled primarily with quartz. This is consistent with lower to middle amphibolite facies deformation, and with fracturing of the relatively strong amphibolite during ductile deformation of the relatively weak quartz-feldspar host gneiss. A locally strong lineation defined by biotite streaks and quartz and feldspar rods apparently formed at about 400°C (Wintsch and Fout, 1982), and by use of the T-t curve of Figure 4, they formed during latest Permian.

41.9	Return to cars. Drive south on SR 32.
42.7	Rejoin SR66, follow SR32 south.
43.5	Turn south at light, cross railroad tracks, and Willimantic River.
43.7	Proceed straight through intersection onto SR289.
44.3	Park under power line that crosses the highway, traverse uphill (W) across several road cuts on the east side of Hosmer Mountain to STOP 5.

STOP 5. WILLIMANTIC FAULT (30 Minutes)

Willimantic Quad. (Snyder, 1964; Wintsch, 1981; 1992; Wintsch and Fout, 1982; Getty and Gromet, 1992b). The purpose of this stop is to show the profile across the terrane boundary (Willimantic fault) between Avalon terrane plagioclase gneisses and Putnam-Nashoba terrane pelitic schists (Figure 7). The Avalonian plagioclase gneisses are similar to (but more mafic than) those at Stop 8, and to many south of the Honey Hill fault, but here are more highly strained. Here, an apparent strain gradient across the contact is suggested by an increase in the development of boudinage, of small scale folding, and of plagioclase and hornblende porphyroblasts, porphyroclasts, and tectonic inclusions, and a reduction in grain size of the matrix (Wintsch, 1979). The Tatnic Hill Formation here is a mylonite schist, and relative to the precursor rock has undergone a 10-50 X grain size reduction (Wintsch, 1979). Across the contact into the Putnam-Nashoba terrane, strain is very high, and the rocks contain totally reconstituted blastomylonites, as evident by kyanite and andalusite-bearing mineral assemblages, chemical composition, and mineral and isotopic composition (Wintsch, 1981; Moecher and Wintsch, 1994; Getty and Gromet, 1992b).

Strain is penetrative below the zone of tectonic blocks, where strain becomes very heterogenous. The lowest several meters of Putnam-Nashoba terrane rocks contain both kyanite and andalusite as well as sillimanite. The first two aluminosilicates embay, and andalusite also includes biotite, and both are associated with magnetite, suggesting the oxidation reaction (Wintsch, 1981): Biotite = Al_2SiO_5 + magnetite + quartz + ions. Hematite is also present in these rocks as thin blades intergrown with biotite. As grain size is reduced in these rocks, there is a decrease in SiO_2, Na_2O and CaO relative to Al_2O_3 and an increase in the Fe_2O_3/FeO ratio. This correlation of oxidized assemblage with small grain size supports the proposal of Wintsch (1981) that a pH increase caused by surface exchange could have been responsible for this oxidation. Thus the mineralogy and even the composition of these rocks do not reflect the upper amphibolite facies conditions which the rocks once experienced. Rather, they reflect a complex set of metasomatic reactions which by some path were probably strain induced at conditions that crossed the stability fields of all three Al_2SiO_5 polymorphs. The apparent discordance of this schist with the rocks both above and below suggests that some of the later strain in these rocks cut across both units. Further south along the road rocks higher in the structural section are exposed. They contain upper amphibolite facies assemblages and structures, including intrafolial folds, boudinage, tectonic blocks, and feldspathization — evidence of the earlier (Devonian or older), completely ductile deformation.

44.3	Return to cars, retrace path to Stop 4.
46.7	Enter SR6, westbound
48.1	Road cuts on SR6, 300 m northeast of its intersection with Ct. Rt. 66, 3 km west of Willimantic. STOP 6.

STOP 6. PUTNAM-NASHOBA TERRANE, WILLIMANTIC WINDOW (60 Minutes)

Columbia Quad. (Snyder, 1967; Wintsch and Fout, 1982; Wintsch, 1987; 1992; Getty and Gromet, 1992b; Wintsch *et al.* 1993). The purpose of this stop is to examine the pre-Devonian upper amphibolite facies rocks and structures of the Putnam-Nashoba terrane, and their overprint by prograde Alleghanian metamorphism. The most stunning features exposed are the large and very large tectonic blocks containing upper amphibolite facies metamorphic assemblages separated by anastomosing shear zones containing middle amphibolite facies assemblages (Figure 7). The dome shape of these blocks leads to quaquaversal foliation patterns in natural exposures in the Willimantic and Columbia quadrangles (Snyder, 1964; 1967) that is diagnostic of the basal Putnam-Nashoba rocks in the Willimantic window. Because they are usually 100 ft (30 m) or more long in the E-W direction, they are best viewed from a distance of 100 ft (30 m) or more. Face I (Figure Stop 6.1) is particularly well suited for viewing from a distance because the highway is not finished; but for those especially interested in these structures, a walk through all the cuts is imperative. The lack of continuity of the blocks or the shear zones on either side of the road along the north side of the interchange (at II, Figure Stop 6.1) indicates that these structures are not longer than 100 ft (30 m) in the N-S direction, and thus the blocks must be lens-shaped. Some of these blocks may have developed as large drag folds (e.g., 100m, Figure Stop 6.2) with axes striking N 30 E, evolving into these discrete blocks as strain was concentrated on the long limbs (Wintsch, 1979, Figure 5; see also Soula and Bessiere, 1980). However, many blocks do not appear to be rotated, and some degree of mega-boudinage (e.g. 200 m, Figure Stop 6..2) was probably also involved. Rotation of these blocks and many other small-scale structures were probably produced during southeast motion of the Merrimack and Putnam-Nashoba terranes over the Avalon terrane during the loading stage of the Alleghanian orogeny.

WINTSCH, SUTTER, KUNK, AND ALEINIKOFF

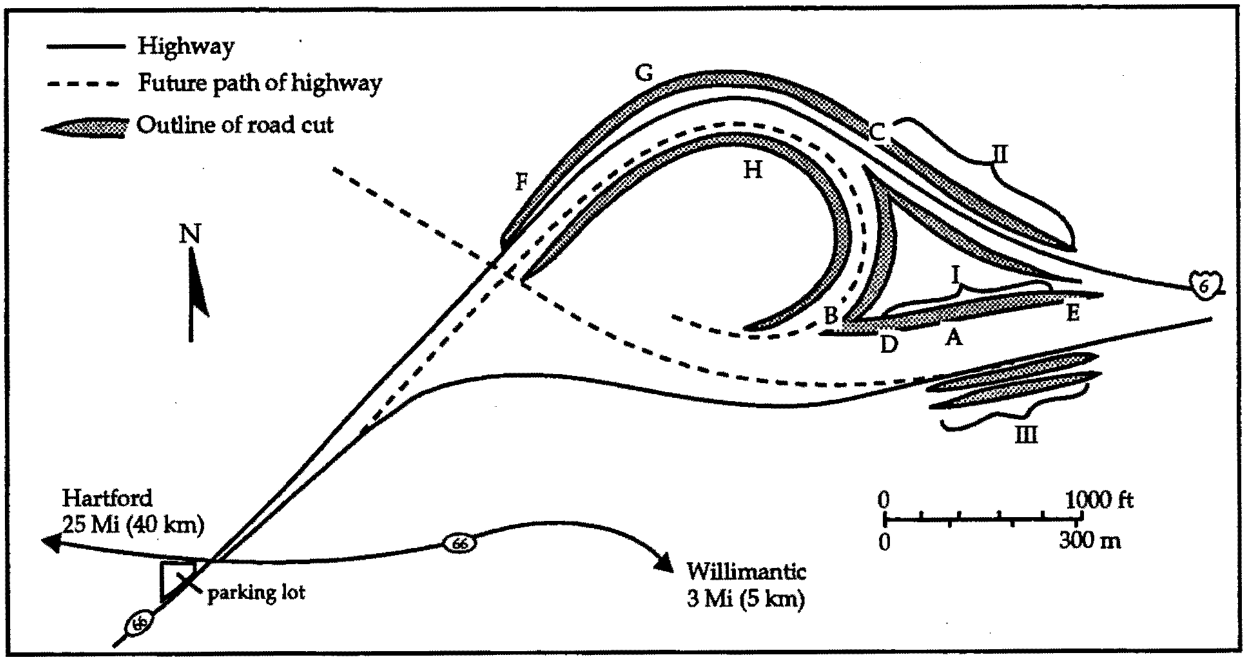

Figure Stop 6.1. Map showing the distribution of outcrop (shaded), road cuts and natural exposures along the unfinished interchange of US6 and SR66. Letter symbols refer to locations given in text.

Augen gneiss, blastomylonitic gneiss, mylonitic schist, and mylonite are present in order of decreasing abundance in these rocks. The highest grade assemblages are best preserved in the augen gneisses inside and away from tectonic blocks (e.g., east end, 150 to 250 m, Figure Stop 6.2). These quartz-plagioclase-biotite-K-feldspar-garnet-sillimanite bearing gneisses locally contain K-feldspar and plagioclase augen up to 5.5 in (14 cm) in diameter. These augen probably grew as porphyroblasts, and did not crystallize from a melt. Evidence for this comes from rotated biotite inclusion trails in some crystals. The incorporation of these rotated inclusions demands growth of a rigid crystal in a solid matrix. Moreover, the bulk composition of the augen gneisses is strongly syenitic or monzonitic, and does not reflect the minimum melt composition of these pelitic gneisses, which should be close to a granitic eutectic. Sillimanite is ubiquitous in these gneisses, and commonly occurs as randomly oriented needles in fibrolite mats. This metamorphism (M1, Figure 7) is early Devonian or older, and thus pre-Acadian (Getty and Gromet, 1992b), reaching greater than 700°C and 6kb (Moecher and Wintsch, 1994).

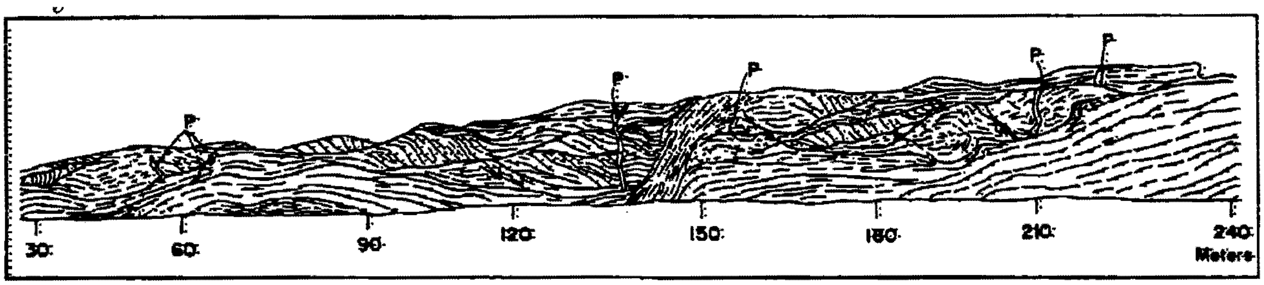

Figure Stop 6.2. Profile traced from photographs of road cut face I (Fig. Stop 6.1) showing anastomosing shear zones that define tectonic blocks. The clockwise rotation of foliation from gently west-dipping (e.g. 210-240 m) to horizontal, and east dipping within tectonic blocks, provide some of the evidence for top moves east (to the right) or thrust motion (Wintsch, 1979). Local reactivation of west dipping shear zones in a normal sense occurred in the late Permian (Getty and Gromet, 1992a).

The shear zones surrounding and cutting these blocks of high-grade gneiss contain blastomylonitic schists and gneisses with lower grade kyanite-bearing assemblages (e.g., localities A, B, C Figure Stop 6.1, if not collected out). This Alleghaninan metamorphism (M2, Figure 7) exceeded a pressure of 8kb (Moecher and Wintsch, 1994). These high pressure sheer zones are in turn cut by lower pressure (3-5 kb) shear zones, where sillimanite occurs with a strong preferred orientation trending approximately N 60° W, parallel to locally developed biotite streaks and quartz-feldspar rods. Still lower grade, slabby, strongly lineated and layered mylonitic schists and blastocataclastic rocks (Figure 6) are well exposed on the natural cliff face III (Figure Stop 6.1), south of the road. This schist projects under all rocks exposed in road cuts, and its strong N-S trending lineation parallel to tight isoclinal folds suggests a similar change in fault motion direction. Another less slabby, even textured, and finer grained mylonitic schist forms a gently folded 3 to 4 m thick layer at D (Figure Stop 6.1) between 60 and 120 m; (Figure Stop 6.2). This later foliation is itself foliated into small isoclinal folds, but they are difficult to find because the rock almost totally lacks compositional layering. True mylonites are rare, but late, fine-grained, middle to lower greenschist facies mylonite up to 2 cm thick cuts the other gneisses in the steeply west-dipping shear zone which cuts the entire exposure at A (Figure Stop 6.1; 140 m, Figure Stop 6.2). A brittle fracture zone occurs at E (Figure Stop 6.1) where the K-feldspar-chlorite-epidote bearing assemblage reflects very shallow alteration of this zone. Together these shear zones and their associated mineral assemblages demonstrate repeated deformation in this fault zone during shallower and lower metamorphic grade conditions (Figure 6).

Several other rock types are present in these cuts. Interlayered in the pelitic blastomylonitic schists and gneisses are at least 28 thin (30 cm) amphibolite layers, all boudinaged (between A and D, Figure Stop 6.1). Successive boudinage of a larger (1 m) amphibolite boudin at its tapering neck can be seen at 215 m (Figure Stop 6.2). Several 30 to 50 cm thick layers of diopside-bearing marble are present at F (Figure Stop 6.1). Sphene from one of these marbles yields a U+Pb age of 335 Ma. The margins of one of these contained hornblende porphyroblasts up to 10 cm in diameter (now apparently collected out). A 2 m diameter pod of ultramafic rock, now chlorite-talc schist is present at G (Figure Stop 6.1).

Hornblende from a massive weathering, well foliated hornblende, plagioclase, biotite, chlorite amphibolite, from the (unfinished) westbound entrance ramp (H, Figure Stop 6.1) was dated. Hornblende grains range from 0.5 to 3 mm in diameter, some with a few plagioclase inclusions. Most grains contain rectangular exsolution lamellae from 10-100 μm long. A few larger grains are composed of smaller rounded, mutually embayed, randomly oriented grains, suggestive of recrystallization and grain growth. Other evidence for a secondary event includes the partial replacement of biotite grains 1/2 to 3 mm in diameter and by equally coarse grained chlorite, and the strong development of deformation-twinning in plagioclase grains, 200-500 μm in diameter. This hornblende (Wintsch, 1992) produced a U-shaped age spectrum with an age minimum of 306 Ma. However, an isochron containing over 98% of the $^{39}Ar_K$ in the sample defines an age intercept of ~273 Ma, and is probably a better reflection of the time of cooling. This hornblende age is not consistent with monazite and sphene ages of 401 and 335 Ma respectively (Getty and Gromet 1992b), but is consistent with Alleghanian cooling in the Avalon zone. The monazite and sphene apparent ages are easily interpreted as a cooling age from pre-Silurian metamorphism, and define a cooling curve almost indistinguishable from that obtained from the rocks in the east (Figure 3). Moecher and Wintsch (1994) interpret the hornblende age of ~270 Ma as reflecting cooling following Alleghanian reheating. This is consistent with the occurrence of overprinting metamorphic assemblages containing andalusite and kyanite in the ductile, Willimantic fault zone within 100 m above the Avalon zone contact (Wintsch, 1980). Indeed, complex exsolution lamellae (not present in hornblendes in the east) may reflect such heating. This Permian resetting may have been generated either by conduction of heat across the fault or by frictional heating (Barr and Dahlen, 1989). In view of all the other data that support isotopic resetting in this fault zone (Getty and Gromet, 1992b), this hornblende apparent age is probably best interpreted as locally thermally reset to a temperature > 500°C, but less than ~600°C, because sphene is not reset.

Wintsch (1979) interpreted the overall motion sense in these rocks to reflect SE directed transport. Since then Goldstein (1989) and Getty and Gromet (1992a) have observed that some kinematic indicators suggest a NW normal sense of motion. The conflicting evidence for motion sense probably arises from overprinted fabrics. Earlier fabrics produced by SE directed motion on thrusts that loaded and rotated the blocks were overprinted by NW directed tectonic unloading during Permo-Triassic exhumation.

48.1	Return to cars, follow US6 W, make U-turn on US6, retrace path to SR32, following it north to US44a.
54.6	Turn left (west) on US44a.
55.9	Follow US44 west to STOP 7.

STOP 7. MERRIMACK TERRANE (30 Minutes)

This is a very dangerous highway cut! Stay off the pavement! Spread out along the cut! South Coventry Quadrangle (Fahey and Pease, 1977; Wintsch *et al.* 1993). The purpose of visiting this outcrop are to examine rocks of the Merrimack terrane on the NW side of the Willimantic window, and the evidence for ductile and brittle deformation in the foot wall of the Black Pond fault. These 2 m high road cuts expose quartz-plagioclase-biotite, and hornblende-diopside schists and granofels. Foliation and axial planes of intrafolial folds dip gently N, and fold and boudin axes plunge gently N. At least three generations of concordant pegmatites cut these rocks (e.g. east end of outcrop). Older pegmatitic layers are strongly foliated and boudinaged, younger pegmatitic sheets are weakly foliated, and the youngest pegmatites are undeformed and tabular. All these structures are cut by planes of microbreccia and rare breccias, some showing a sinistral sense of displacement. Hornblende ages in these rocks are late Pennsylvanian, and probably postdate most of these ductile structures, making the latter Acadian or older. Regional data on K-feldspar ages (total gas) suggest that these breccias are Triassic.

55.9	Return to cars. Follow US44 east.
56.4	Turn left (north) on Brigham Road.
57.4	Follow Brigham Road north to natural cliff exposures on the west side of the road, STOP 8.

STOP 8: CENTRAL MAINE-MERRIMACK TERRANE BOUNDARY BLASTOMYLONITES (30 Minutes)

South Coventry Quad, (Fahey and Pease, 1977). The purpose of this stop is to show the very highly strained rocks of the Black Pond fault, the boundary between the Merrimack and Central Maine terranes. These exposures show very well these north-dipping mylonitic and blastomylonitic schists and gneisses of the Black Pond fault. Amphibolite layers are strongly thinned, but not boudinaged in the surrounding biotite-plagioclase-garnet bearing schist, probably reflecting very high temperatures and/or slow strain rates. These rocks also contain rootless intrafolial folds, boudin-like, tectonic inclusions, and steeply dipping mylonites that cut the mylonitic schist, and begin to define tectonic blocks similar to those in the Tatnic fault zone (Figure 7).

57.4	Return to bus. Follow Brigham Road south.
58.4	US44. Turn left (east) on US44, cross SR32.
61.1	Turn left (north) into correctional facility.
61.3	Exit cars, at trail up the hill to water towers, and STOP 9.

STOP 9: CENTRAL MAINE TERRANE (30 MINUTES) (South Coventry Quad., Fahey and Pease, 1977). The purpose of this stop is to contrast the lithology, structure, and metamorphism of rocks assigned to the Central Maine terrane in the hanging

wall of the Black Pond fault with those of the Merrimack terrane. The rocks around this water tower are strongly layered quartz-plagioclase-K-feldspar-garnet-biotite granular schists. The rocks contain conspicuous porphyroblasts of feldspars (1-3 cm) and garnet (1/2 - 3 cm). Steeply dipping NNE trending foliations wrap to more easterly trending strikes with shallower dips, mimicking the larger scale structure of the Black Pond fault.

61.3 Return to cars. Return to US44.

61.5 Turn right (west) on US44. Follow US44 to SR84.

70.1 Turn left (south) on SR85. Follow SR85 south.

76.7 Turn right (west) on SR94.

79.9 Turn left (south) onto Mimichang Farm to STOP 10.

STOP 10: GLASTONBURY GNEISS, BRONSON HILL TERRANE

(45 MINUTES) Marlborough Quad. (Snyder, 1969; Leo *et al.*, 1984; Ambers, 1988). The purpose of this stop is to compare the structure, lithology, intrusive relationships, and metamorphism of rocks of the Bronson Hill terrane to those of the Central Maine terrane. The rocks in this exposure, studied by Ambers (1988), include a xenolith-rich strongly lineated granodioritic gneiss and mylonitic derivatives, an intrusive granitic gneiss, and cross cutting pegmatitic veins, all occurring in a sub-horizontal, right-lateral zone of deformation. Strain can be monitored using aspect ratios of xenoliths in the deformed granodiorite, that increase from a low of 12-15 in xenolith measurement domains 7-12 (Figure Stop 10.1) to over 135 in the mylonite domain 3. Regional studies show that hornblende + K-feldspar in the precursor assemblage has reacted to biotite-epidote-bearing assemblages. Contact metamorphism of this assemblage around granitic intrusions in the southern part of the outcrop

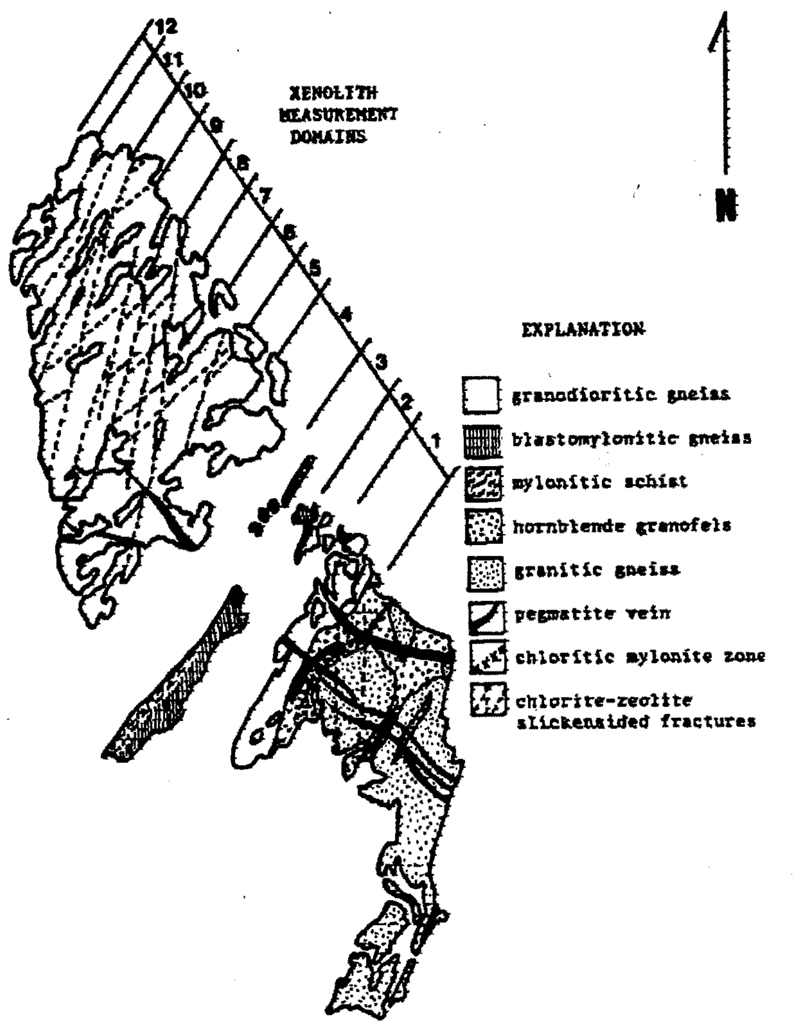

Figure Stop 10.1 Pace and compass map of Stop 14 (from Ambers, 1988) showing the distribution and spectrum of rock types present, and xenolith measurement domains used for semiquantitive strain analysis.

remetamorphoses the biotite and epidote schist to a hornblende granofels. Thus, some of the ductile deformation of the granodiorite predates the granite, but it too, is deformed with the strong horizontal N30°E lineation of the gneiss. Many interesting small scale structures are present in the mylonite and around the pegmatites.

The mylonite in domain 3 is derived from the granodiorite by syntectonic metasomatism, where microcline originally in the mylonite was replaced by plagioclase and biotite, and plagioclase in the shoulders of the mylonite zone was replaced by microcline as porphyroblasts. Mass balance calculations among mylonite, quartz-K-feldspar veins, and metasomatized gneiss show nearly isochemical and isovolumetic deformation on the scale of the outcrop (Ambers, 1988).

End of trip. Return to cars.

Trip B

Lake Sequences of the Central Hartford Basin, Newark Supergroup

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INTRODUCTION

The Hartford Basin of central Connecticut and Massachusetts contains sedimentary rocks and interbedded tholiietic basalts and dikes that belong to the Newark Supergroup (Froelich and Olsen, 1984). Strata of the Newark Supergroup occur in a linear series of NW-SE trending half-graben basins, both exposed and subsurface, along the eastern coast of the United States. They developed in response to crustal extension and incipient rifting of the supercontinent Pangaea during the Early Mesozoic (Manspiezer, 1988). This zone of rifting approximates the present circum-Atlantic margin, and other rift basins that contain broadly coeval basin-fill sequences are found in the Arctic Archipelago, Greenland, Spitzbergen, central and southwestern Europe, and northwestern Africa. Most of these basins appear to be reactivated late Paleozoic structures (Schlische, 1993), and the precise age of renewed subsidence is diachronous, but largely restricted to the early and middle Triassic. However, most synrift basins along the Atlantic seaboard that contain the Newark Supergroup initially formed during the early and late Carnian (Julian and Tuvalian substages), as indicated by palynostratigraphy, megafloral biostratigraphy, vertebrate biochronology, and magnetostratigraphy (Ash, 1980,1987; Litwin et al, 1991; Olsen et al., 1989; Lucas and Huber, 1993; Molina-Garza et al., 1991). The age of the oldest strata of the Hartford Group remains poorly constrained, but appears to coincide with the Carnian-Norian boundary (Table 1) (Cornet and Traverse, 1975; Cornet, 1977, Lucas and Huber, 1993).

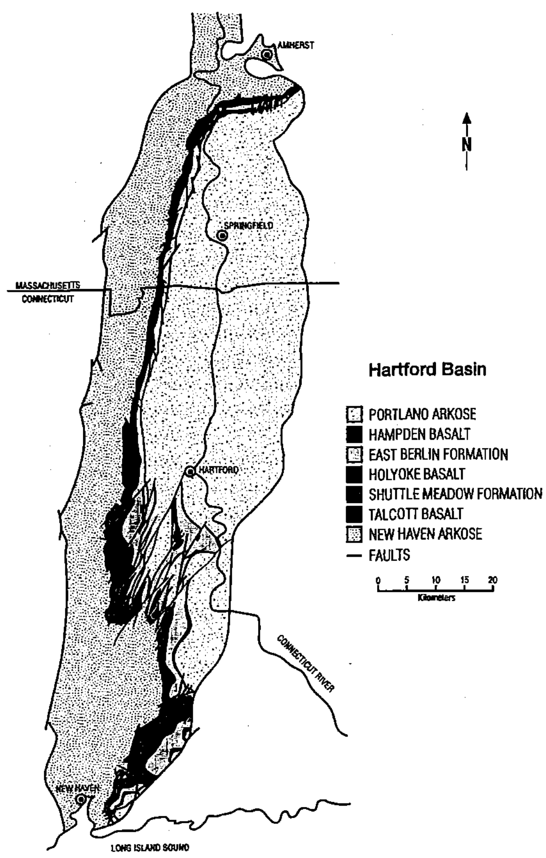

Figure 1 - Simplified geologic map of the Hartford Basin (after Lorenz, 1987). Graphics by Samuel Girton.

TABLE 1 - General Stratigraphy of the Hartford Basin

Series	Stage	Thickness (m)	Stratigraphic Unit	Lithology	Environmental Interpretation	Hydrologic Drainage	Tectonics
Lower Jurassic	Sinemurian-Hettangian	~2000 +	Portland Fm.	black shales mud/siltstones sandstones conglomerates minor limestones	lacustrine & alluvial	open ↑ closed	Subsidence slows
	Hettangian	30-65	Hampden Fm.	tholeiitic basalt	volcanic flow	-----	
		150-300	East Berlin Fm.	black shales mud/silt/sandstones conglomerates minor limestones	lacustrine & alluvial	closed	
		80-150	Holyoke Fm.	tholeiitic basalt	volcanic flow	-----	
		17-140	Shuttle Meadow Fm.	black shales mud/siltstones sandstones minor limestones	lacustrine & alluvial	closed	
		0-80	Talcott Fm.	tholeiitic basalt	volcanic flow	-----	
Upper Triassic	? Early Norian -basal Hettangian	~2400	New Haven Arkose	mud/siltstones sandstones conglomerates	alluvial	open	201 Ma - Extension & Igneous Activity ↑ 220 Ma - Extension

The Hartford Basin half-graben (Figure 1) is 140km long and has a maximum preserved width of 30km near the Connecticut-Massachusetts border. The multi-segmented, west-dipping eastern border fault zone defines the eastern boundary of the basin, and parallels the N-S trending system of mantled gneiss domes of mid-Paleozoic age. The western basin edge is delimited by either sedimentary rocks in fault contact with, or that conformably rest on, crystalline rocks of Paleozoic age. The basin-fill sequence comprises as much as 5km of sedimentary rocks and basalts named the Hartford Group. In the southern and northern portions of the basin, a number of transverse folds are developed adjacent to the eastern border fault system. As noted by Schlische and Olsen (1990), these folds die out basinward away from the hanging wall, and do not fold or cross the segmented border fault system, implying a syndepositional origin. Basinal strata dip easterly on average 7-15 degrees, with local variation most pronounced perpendicular to the axes of the transverse folds. Syndepositional border faulting in the basin is evidenced by a decrease in dip upsection in the younger synrift rocks (Wise, 1992; Schlische, 1993). The basin-fill of the Hartford Group is exposed in a series of fault blocks that are bounded by normal faults striking on average N 30° E. These fault blocks, mapped by Davis (1898) and Emerson (1898), are developed subjacent to the border fault zone in the southern portion of the basin, and are best-exposed progressively to the north and west toward the basin center. The southern terminus is not well-defined and the northern border with the Deerfield Basin is composed of a narrow 5km-wide "neck" where the eastern border fault system is continuous, but a transverse fold exposes unroofed, crystalline basement ("Amherst inlier").

At the onset of rifting during the late Triassic (late Carnian-Norian), the Hartford Basin was located around four degrees north of the paleoequator, and had drifted to eight degrees north paleolatitude by the early Jurassic (Hettangian) (cf. Witte et al., 1991). According to current paleoclimatic and paleogeographic models for the Early Mesozoic (e.g. Manspeizer, 1988; Dubiel et al., 1991; Parrish, 1993), the Hartford Basin was located with a tropical to sub-tropical belt subject to pronounced monsoonal seasons. Then during the 18myr span of the Norian, it became increasingly arid before entering a pronounced "wet" phase very close to, or at the Rhaetian-Hettangian boundary. Because the upper age limit of the Hartford Group, as well as other Newark Supergroup, rocks is poorly constrained, it is uncertain whether the early Jurassic "wet" phase was entirely restricted to the Hettangian or persisted into the Sinemurian and perhaps Pliensbachian.

Lower Jurassic deposition in the Hartford was significantly influenced by coeval climatic cyclicity and tectonism that ultimately controlled facies types and distributions, as in the other basins of the Newark Supergroup. The most pronounced climatic events were recorded as cyclic episodes of lacustrine deposition that may have been controlled by Milankovitch climatic forcing at the 21 ky wavelength. During wet phases, perennial stratified lakes that supported a moderately diverse invertebrate and piscine fauna occupied much of the basin surface area. During phases of lake contraction, sedimentation over the largely featureless basin floor was dominated by sheetflood events and playa deposition. Through these "dry" phases, a diverse terrestrial biota that included a variety of plants, invertebrates, and archosaurs inhabited the central basin area.

Many authors have attributed lake cycles in all the continental basins of the Newark Supergroup to climatic influence (Van Houten, 1962; Wheeler and Textoris, 1978; LeTourneau, 1985a; Olsen, 1986; Smoot, 1991; Glenn and Kelts, 1991). In general, such small-scale sequence changes as water-level fluctuations in lake basins appear to be mostly climatically controlled, whereas large-scale sequence changes affecting drainage patterns (such as open or closed drainage) appear to be tectonically controlled (see conclusions by IGCP 219 Spanish Group on Tertiary Basins, 1990). Many observations by Olsen et al. (1989), Schlische and Olsen (1990), Smoot (1991), and Hubert et al. (1992) on sedimentation patterns from the Newark Supergroup basins support this theory.

Table 1 outlines the stratigraphic units and their characteristics in the Hartford Basin, as well as the paleoenvironmental interpretation of each unit and postulated paleodrainage. Border faults were active throughout the history of the basin (Krynine, 1950; Schlische, 1993) and would have profoundly influenced sedimentation patterns (LeTourneau, 1985b; Schlische and Olsen, 1990; Hubert et al., 1992). Arguments for the exact nature and extent of tectonic controls and their resulting signatures in the stratigraphic record (Olsen et al., 1989; Schlische and Olsen, 1990; Smoot, 1991; Hubert et al., 1992; Schlische, 1993) are sketchy since information concerning the character of facies changes across the basin at different stratigraphic levels is lacking. The exact thickness of some stratigraphic units in the Hartford Basin is still imprecisely known (Table 1). Outcrop exposures across the basin are incomplete and difficult to correlate to basin-edge sedimentary sequences. Thermal maturity indices suggest that the original size of the sedimentary basin was much larger (Smoot, 1991) and a thick sequence of upper deposits have been eroded away (~1500 m according to Pratt et al., 1988). However, changes in hydrologic drainage (open vs. closed) appear to be related to tectonic activity associated with rifting in the Hartford Basin (Table 1; Hubert et al., 1992).

In this field guide we present the sedimentology of the three formations of the Hartford Group in central Connecticut that contain lake deposits: the Shuttle Meadow, East Berlin, and Portland Formations. Please note that all the field trip stops are under private and/or state ownership, and access to them can be obtained by permission only.

ROADLOG

START

Mile	
0.0	Assemble in the parking lot of the Radisson Hotel at the intersection of Interstate 91 and Route 372 in Cromwell, CT.
	Turn right onto 372.
0.3	Proceed directly to the on-ramp of Interstate 91 south (New Haven). (0.3)
6.1	Exit to 691 west, direction of Meriden. (5.8)
10.7	Use Route 322- Southington (Exit 4). Note outcrops of New Haven Arkose along off-ramp. (4.6)
10.9	Turn left onto road in the opposite direction to Southington. (0.2)
11.8	Turn left into entrance of Hubbard Park. (0.9)
12.0	Follow one-way road through the first stop sign. (0.4)
12.1	At second stop sign turn left and drive through gateway. (0.1)
13.5	Turn left and cross bridge at north end of Merimere Reservoir.(1.4)
14.9	Turn left at fork in road. (1.4)
15.3	Parking area for observation tower (weather permitting). (0.4)

STOP 1 - East Peak of Meriden (Castle Craig Tower of Hubbard Park)

Castle Craig Tower overlooks the East Peak of the Hanging Hills of Meriden, Connecticut. The stone observation tower was built by local stone masons and dedicated in 1900; a gift of Walter Hubbard.

Built on the Holyoke Basalt, the elevation at the base of the tower is 976 feet. The tower itself is 32 feet in height and 58 feet in circumference. The overlook affords a view to New Haven and Long Island Sound to the south and the Berkshires in southern Massachusetts to the north. To the east, the Eastern Highlands composed of igneous and metamorphic rocks define the eastern edge of the Hartford Basin. To the southwest, the Western Highlands, composed of Lower Paleozoic metamorphics can just barely be seen. "The Sleeping Giant" (Mt. Carmel) is a basalt sill or stock in the New Haven Arkose located to the south-southwest.

Basalt layers and sills in the Connecticut Valley define the hills which stretch from Long Island Sound to the Holyoke Range in Massachusetts. The general north-south strike and eastward dip of the basalt and intervening sedimentary strata in the Hartford Basin is altered in many areas by zones of northeast-southwest trending faults (Figure 1). Eastward from Castle Craig Tower one such fault offsets the Holyoke by nearly 13m. This faulting pattern defines en echelon, intrabasinal fault blocks composing the geometry of the half graben and its basement (Schlische, 1993: Figure 7). The sinuous outcrop patterns are related to the synclinal nature of folding of the basin fill, implying differential subsidence during extension (Schlische and Olsen, 1990; Schlische, 1993).

Alternate STOP 1 - Hubbard Park - at the base of East Peak.

19.0 Retrace route down through Hubbard Park. Turn right onto street back to 691 interchange. (3.7)

20.0 Entrance ramp back onto 691 west, direction of Waterbury. (1.0)

23.0 Take Interstate 84 east toward Hartford (Exit 2). (3.0)

31.9 Exit at Route 372 (Crooked Street) (Exit 34) (8.9)

32.0 Turn right at stop sign. (0.1)

32.1 Turn right at traffic light onto White Oak Ave. (0.1)

32.2 Turn left onto Ledge Road (right before bridge). (0.1)

34.7 Take right fork in road on downhill slope (2.5)

35.0 Take left fork in road past orchard store (Long Bottom Rd.). (0.3)

35.4 Take right onto Shuttle Meadow Street. (0.4)

35.5 Outcrop on left side of road. (0.1)

STOP 2 - Shuttle Meadow Formation (Shuttle Meadow Reservoir, Southington)

The Shuttle Meadow Formation, the oldest sedimentary formation of the Hartford Group (Table 1) containing lake deposits, attains a maximum estimated thickness of 140m in southern Connecticut and wedges out to the west and north near the Connecticut-Massachusetts border. Across its outcrop area in southern and central Connecticut, the lower Shuttle Meadow Formation begins with a pebble-cobble conglomerate that rests directly on, and includes well-rounded clasts of, the Talcott Basalt. The basal 10-20m of the formation are mostly massive to stratified sandstone, red and gray massive to laminated micaceous siltstone, stratified mudrocks, microlaminated black shale, and massive to laminated micrite. The coarser-grained lithologies of this interval occur proximal to the eastern border fault zone, and are replaced by muddy siltstones toward the central part of the basin. The black shale forms a single bed that is thickest at outcrops within several km of the eastern border fault in southern Connecticut. This shale unit can be traced to the north and west where it thins and grades laterally into disrupted shales and mudrocks of the central basin area near Meriden. Farther north and west at Southington, the shale unit occurs as a single bed 0.1m thick that is encased by stratified mudrock and associated minor limestone.

The upper Shuttle Meadow Formation throughout the Connecticut portion of the Hartford Basin is similar to the section at Cook's Gap, Plainville (Stop 3), with a predominance of ripple cross-laminated muddy siltstones. In the central basin area, sandstone occurs only as small, channel lenses. In the southern basin area, outcrops directly below the Holyoke Basalt within 100m of the eastern border fault zone include 20m of matrix-supported cobble-boulder conglomerates (with coarse-grained arkose) containing lenticular, clast-supported zones. The clasts are angular to sub-angular and are derived from the underlying Talcott Basalt, which in the vicinity of the border fault, lies approximately 90m below the base of the Holyoke. The presence of Talcott clasts in the uppermost beds of the Shuttle Meadow Formation implies that the Talcott Basalt overflowed the border fault zone. The same stratigraphic level is exposed 1.2km to the north along U.S. Route 1 (described by Sanders, 1970). Here the conglomerates contain well-rounded to subangular, pebble-cobble clasts that fine upward into coarse- to medium-grained sandstone and ripple cross-laminated muddy siltstone.

In northern Connecticut near the border, the Shuttle Meadow Formation thins to 13.2m at Tariffville Gorge before it wedges out 2km farther north. The Tariffville section begins with .75m of basalt-cobble conglomerate set in a muddy siltstone matrix which rests directly on the Talcott Basalt. The remainder of the section consists of ripple cross-laminated red siltstone to muddy siltstone and minor gray disrupted shales - all of which contain abundant evaporite? pseudomorphs.

Thus, fourteen facies have been identified in the Shuttle Meadow Formation across its outcrop belt in the basin. They include: matrix-supported sandy conglomerate (Gms), matrix-supported muddy conglomerate (Gmf), clast-supported conglomerate (Gm), conglomeratic sandstone (Gt), trough cross-stratified sandstone (St), Horizontally stratified sandstone (Sh), ripple cross-laminated sandstone and siltstone (Sr, Fr), massive mudstone (Fm), disrupted calcareous shale (Fl_3), stratified mudrock (Fl_2), black shale (Fl_1), massive limestone (Pm), and bedded limestone (Ph) (Table 2). Facies Gms, Gm, Gt, Sh, St, Sr, Fm, and Fr represent an alluvial fan to alluvial plain/sandflat paleoenvironment. Sheetflooding was the dominant depositional process with paleosols developing on the muddy sediments. Facies Fm, Fl_3, Fl_2, Fl_1, Pm, and Ph are interpreted as playa mudflat to lake paleoenvironments, with paleosols developing to varying degrees on the playa mudflats as well.

SHUTTLE MEADOW FORMATION - TABLE 2

FACIES DESIGNATION	FACIES NAME AND DESCRIPTION	DEPOSITIONAL ENVIRONMENT
Gms	**Matrix-supported sandy conglomerate** Matrix is poorly to moderately sorted arkose and silty sandstone; clasts are angular to subangular, pebble- to boulder-sized, of basalt and extrabasinal origin. No size grading. Beds are 0.5-2.4m thick, separated by erosional surfaces and/or overlain by wavy to laminated coarse to fine arkose and silty arkose in beds up to 12cm thick.	Debris flow on alluvial fan
Gmf	**Matrix-supported muddy conglomerate** Matrix is muddy siltstone: clasts are subangular to well-rounded pebbles and cobbles of basalt; no size sorting. Single locality in Tarriffville, CT, where it forms a bed up to 1m thick on scoured surface of Talcott Basalt. Top of this bed is gradational with overlying ripple cross-laminated muddy siltstone.	Reworked debris flow associated with sheetflood material

Gm	**Clast-supported conglomerate** Clasts: angular to subangular, pebbles to boulders. Origin: clasts are intrabasinal (?Talcott Basalt) and extrabasinal. Crude size sorting: fining upward. Morphology: lenticular pods up to 1.7m along strike with maximum observed thickness of 0.6m. Occurring at the top of Gms beds. Matrix: arkose to silty sandstone.	Sieve deposits
Gt	**Conglomeratic sandstone** Matrix is moderately to well-sorted, medium to fine arkose; clasts are mostly pebbles (some cobbles). Stratification: trough cross-bedding. Units up to 2.3m thick fine upward, in clast and matrix size. Set thickness: decimeter scale. Clast imbrication above scoured base of units. Normally topped by Fr or Fm.	Lower alluvial fan
St	**Trough cross-stratified sandstone** Fine- to coarse-grained sandstone, moderately to well-sorted. Pebbles present at bases of units (lag deposits). Units fine upwards; set thickness: decimeter scale. Commonly interbedded with Fr and Fm.	Sheetflood deposit
Sh	**Horizontally stratified sandstone** Medium- to fine-grained sandstone, moderately to well-sorted. Cm-scale beds associated with Gt and Sr/Fr. Unit contacts are continuous and planar for tens of meters along strike.	Sheetflood deposit
Sr,Fr	**Ripple cross-laminated sandstone, siltstone** Fine sandstones to muddy siltstones, commonly micaceous. Depositional units are decimeter scale and muddier units are commonly topped by mudcracks. Stacks of units can be over 20m thick in mid-basin. Associated with St, Sh, and Fm. Rare gypsum pseudomorphs.	Sheetflood deposit

Fm	**Massive mudstone** Sandy to silty mudstone containing mudcracks (filled with both sandstone and mudstone), floating intraclasts of original bedding, and random mm-sized pores which are empty or filled with mudstone or carbonate. Units can be up to 1.5-2.0m thick. Commonly associated with Sh, St, Sr, and Fr.	Paleosol
Fl_3	**Disrupted calcareous shale** Red to green/gray to black shale (clay/silt lamina) intercalated with carbonate lamina to micrite beds (up to 4cm thick) and rare sandstone lamina. Locally disrupted by mudcracks. Unit thickness on a decimeter scale. Transitional to Fl_2.	Playa mudflat
Fl_2	**Stratified mudrock** Mudrocks containing planar thick lamina of sandy siltstone, muddy siltstone, organic-rich claystone, and carbonate (dolomite-rich). Units up to 2m thick. Macrophytic plant debris and fish scales common. Associated with Fl_3 and commonly envelopes Fl_1.	Shallow lake
Fl_1	**Black shale** Shale containing discontinuous to continuous lamina of fine siltstone, organic-rich claystone, and micrite. Pyrite, bitumen common. Abundant fully-articulated fish fossils. Facies traced for tens of kilometers. Associated with Fl_3 and Fl_2.	Offshore lake
Pm	**Limestone** Micrite with a clastic texture containing siliciclastics and microbial structures. Rare ostracode shells and fish bone fragments. Gypsum crystals and rhizoliths also locally present.	Carbonate pond
Ph	**Coe Quarry Limestone** Bedded to laminated carbonate described by Steinen et al. (1987). Restricted to southern CT with a lateral extent of 1km or less (Krynine, 1950).	Hot spring deposit?

East Berlin Formation - Table 3

Facies	Designation	Description	Interpretation
Trough Cross-Stratified Sandstones	St	Fine- to medium-grained sandstones, medium- to well-sorted. Sets: 10-20cm thick within cosets 10 to 100cm thick. Small-scale cross-stratification, which commonly fines upwards. Cosets commonly with erosional bases and gradational upper contacts into overlying mudrocks. Color: mostly light red (some units gray-green or mottled). Grains: quartz, feldspar, sand-sized mud aggregates, and mudstone intraclasts up to 10 cm in length. Commonly interbedded with Sh, as sets up to 20cm thick within cosets up to 50cm thick. Invertebrate trace fossils are common in the mud drapes capping sedimentation units containing both St and Sh sequences. St and Sh sand bodies: sheet morphology. Associated with facies Fr and SF as part of fining upward sequences. Paleocurrents: unidirectional eastward trend.	Sheetflood deposits on an ephemeral alluvial plain/ sandflat
Horizontally Stratified Sandstones	Sh	Color, sorting, and composition similar to St. Parting lineations. Commonly interbedded with St, as sets up to 20 cm thick within cosets up to 50cm thick. Many sets have erosional bases, fine upward, and are overlain by a cracked mud drape. Trace fossils are common in the mud drapes capping sedimentation units containing both St and Sh sequences. St and Sh sand bodies: sheet morphology. Associated with facies Fr and SF as part of fining upward sequences.	Sheetflood deposits on an ephemeral alluvial plain/ sandflat
Interbedded Sandstones and Mudrocks	SF	Fine- to medium-grained sandstone laminae or thin beds up to 5cm thick intercalating with continuous to discontinuous laminae of mudstone up to 1mm thick. SF units: .25 to 1.5m thick with erosional bases. Colors: red, pink, and gray. Sandstone laminae/beds: asymmetric and symmetric ripple cross-lamination or horizontal lamination. Mudrocks: flasers or continuous laminae that are cracked. Flaser to wavy bedding represented. Invertebrate race fossils mostly restricted to upper portions of depositional units.	Sheetflood deposits on an ephemeral alluvial plain/ sandflat
Ripple Cross-Laminated Siltstones	Fr	Muddy to fine siltstones and very fine-grained sandstones. Ripple cross lamination: mostly asymmetric current ripples, both climbing and erosional. Uniformiry of paleoflow (unidirectional to the east) is notable. Colors: red and yellow/gray. Cosets: .25 to 1.5m thick with multiple sets separated by erosional surfaces. Upper few cm of cosets commonly with irregular patches of Fm or complex system of mudcracks. Yellow, carbonate-cemented, well-sorted fine siltstone beds (up to 5cm thick) exhibit primary ripple structures within associated muddier, more compacted siltstone units. Invertebrate trace fossils mostly restricted to upper portions of depositional units. Gypsum molds, calcite pseudomorphs after gypsum, and star-shaped calcite "balls" (which exhibit syndepositional features).	Sheetflood deposits on an ephemeral alluvial plain/sandflat
Disrupted Mudstones	Fm	Massive mudstones, composed of units up to several meters thick. Interbedded with Fr and Fl_3. Colors: red with rare green or red/ green mottled units. Textures: patchy with numerous mudcracks in polygonal networks, with curved clay slickensides, and small fenestrae or sub-spherical cavities (empty, clay-lined, or filled with dolomite or calcite). Large mudcracks: up to 1m deep filled with sandstone, mudstone, or both. Small mudcracks: 10-15cm long also with three types of fill. Many sandstone-filled cracks with no over-lying source bed. Dispersed calcitic/dolomitic nodules (up to 8cm in diameter locally). Gypsum molds and calcite pseudomorphs after gypsum. Petrography: skeleton grains of quartz, feldspar, and mica float in a groundmass of clay and iron oxides. Heterogeneous, agglomeratic to porphyritic texture indicating poor sorting. Corroded quartz grains. Distinct clay and carbonate coatings and concentrations.	Paleosols (vertisols) developed in various stages on alluvial plain/ sandflat as well as playa mudflats

Disrupted Shales	Fl_3	Red and green/yellow shales occurring as discontinuous thin beds (2-15cm thick) within massive mudstone (Fm) units. Complex mudcracks: sandstone- and mudstone-fill; from mm-scale through one lamina to cm-scale through entire beds. Displaced blocks of shale intraclasts: in-situ, displaced, to floating in massive mudstone matrix. Two styles of lamination: very thin "microbial" lamination and thicker fine alternations of siltstone and claystone (2-5mm thick). Some carbonate crusts. Mud curls. Roll-up structures. Rare ripple structures: transitional to facies Fr. Gypsum molds and calcite pseudomorphs after gypsum. Fossil plant fragments, very rare fish scales, and invertebrate trace fossils.	Playa mudflats
Stratified Mudrocks	Fl_2	Mudrocks with continuous, planar, thick laminations (up to 1.5cm thick). This facies conformably envelopes the black shale facies, but is also present as isolated units. Colors: black, green, red (siliciclastic mud), yellow, and white (micrite). Arkosic, fine-grained sandstone to siltstone intercalations (with ripple cross-lamination) in upper portions of isolated units and those above black shales. Mudcracks: simple to multiple with sandstone and/or mudstone fills: density increases upward in isolated units and those above black shales. Units above black shales: up to 2m thick. Units below black shales: only a few cm thick with rare mudcracks. Fossils: plant fragments, dinosaur trackways, rare small burrows, and rare conchostracans. Traces of analcime and gypsum. Very rare molds of halite, glauberite?, and gypsum.	Ephemeral shallow lake deposits
Black Shales	Fl_1	Three main types of shale (exclusively associated with Fl_2): (1) Irregularly laminated black shale - most common. Discontinuous, irregular laminae (mm to submm scale) of arkosic to carbonate, muddy siltstone to claystone (up to 20mm thick). True laminae: parallel lamination. Other laminae: laminated intraclasts with internal stratification parallel or oblique to general stratification. Some very rare chaotic bedding. Rare radiating crystal impressions on bedding surfaces. Rare conchostracans and poorly preserved fish fossils. (2) Magnesite-bearing shale with magnesite microspar nodules 2-3mm in diameter and larger magnesite crystal masses to lenses, disseminated in varying densities. Single example is 60cm thick (lake cycle 2). (3) Finely laminated black shale with continuous, alternating claystone and carbonate siltstone lamination. Single example is 25cm thick (lake cycle 6). Fossils: well-preserved, articulated fish, conchostracans, ostracodes, and coprolites.	Offshore sediments of a perennial saline lake

Alluvial fan/alluvial plain paleoenvironment

The conglomerate facies (Gms, Gmf, Gm, Gt) of the Shuttle Meadow Formation represent debris flow, sieve, and lower fan deposits. Reworking of the conglomerates as they are slowly moved down fan is evidenced by the mixing with sheet flood sediments (Gmf) and the intercalation of finer sheetflood material with Gms and Gt. These coarser-grained facies grade into the alluvial plain facies (St, Sh, Sr, Fr, Fm). Trough cross-stratification, horizontal to inclined stratification, and ripple cross-lamination are common structures within sheet deposits (e.g. Karcz, 1972; Sneh, 1983). The sheet morphology of these facies can be best seen at the new type section of the Shuttle Meadow Formation with paleocurrents pointing generally northward (Stop 3). Fm represents the pedogenic modification of the muddy sediments on the Shuttle Meadow alluvial plain, similar to that in the East Berlin Formation (Gierlowski-Kordesch and Rust, 1994). Mudcracked drapes on sandstones, possible evaporitic pseudomorphs in the finer-grained rocks, and mudcracked to massive tops of units of Fr, all point to ephemeral, semi-arid paleoconditions.

Playa/lake paleoenvironment

Disrupted calcareous shales (Fl_3) topped by massive mudstone units in the Shuttle Meadow Formation are interpreted as aggrading, dry playa mudflats containing carbonate crusts. The disruption of lamination by mudcracks and the overlying massive,

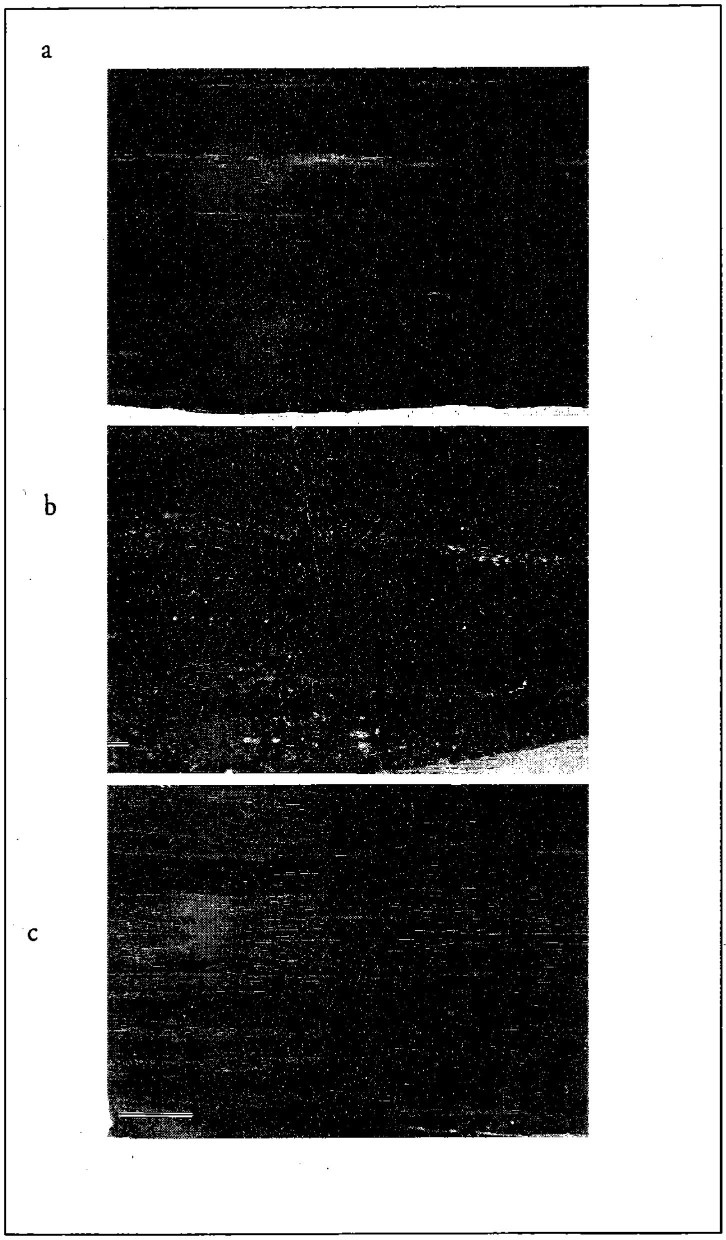

Figure 2 - Lamination styles of the black shales of the Shuttle Meadow Formation, Connecticut. (a) Black shale slab from Higby, central Connecticut. Continuous to discontinuous, low angle to horizontal lamination of claystone and carbonate lamina. Note erosive surfaces and deformed to brecciated carbonate lamina. Scale bar = 0.5cm. (b) Black shale slab from Durham, southern Connecticut. Continuous claystone, siltstone, and carbonate lamina are diffuse and undulatory, suggesting microbial origin. Vertical lines are fractures in slab. Scale bar = 0.5cm. (c) Black shale slab from North Guilford, southern Connecticut (Bluff Head locality in McDonald (1975)). Continuous claystone and carbonate microlamina are disturbed by microfaulting at some levels. Scale bar = 0.5cm.

pedogenically-altered mudstone units evidence aggradational processes common in playas (Hardie et al., 1978; Demicco and Gierlowski-Kordesch, 1986). Pm (massive micrite) layers are associated with the siliciclastic playa to alluvial plain deposits at the type section in Plainville. These are interpreted as carbonate ponds, after Platt and Wright (1992). Localized bedded carbonate deposits (Ph) in southern Connecticut (Steinen et al., 1987) are interpreted as hot spring deposits.

Stratified mudrock (Fl_2) and black shale (Fl_1) represent shallow, ephemeral to perennial, offshore lacustrine sediments. These lacustrine deposits are very calcareous and contain layers and lenses of micrite. The lamination style of the black shales is variable across the outcrop belt in the southern half of the Hartford Basin (Figure 2), denoting different paleoconditions and/or lakes. Continuous to discontinuous clay lamina alternate with carbonate lamina in an irregular fashion. Erosive lamina boundaries, micro-brecciated carbonate lamina, small-scale cross-stratification, microfaulting, and coarser possible "microbial" layering, all perhaps indicate changes from saline to fresh water phases in these sequences. Fish fossils are abundant though bioturbation is not pervasive. Research on these shales is in progress.

The section in Southington is within a small abandoned quarry on the east side of Shuttle Meadow Reservoir. This lake sequence, nearly 7.5m thick and composed of claystone, mudrock, and limestone, represents the lower portion of the Shuttle Meadow Formation. The floor of the orchard across the street is the dip surface of the Talcott Basalt. The lake deposits of the Shuttle Meadow Formation contain more carbonate than most other lacustrine sequences in the Hartford Basin. There is no intensive bioturbation in this sequence, interpreted as perennial lake sediments. Fossils include rare fish scales and carbonized *Equisetites* and conifer pinnules.

39.0 Retrace steps back to main road near bridge and turn right onto White Oak Ave. (3.5)

39.1 Turn left at traffic light. (0.1)

39.5 Turn right onto Route 372 east. (0.4)

39.6 Turn right into empty lot across from outcrop. (0.1)

STOP 3 - Shuttle Meadow Formation (Cook's Gap, Plainville)

The Shuttle Meadow Formation was named by Lehman (1959) for strata previously called "Anterior Shales" (Percival, 1842; Davis, 1898) or "Lower Sedimentary Member of the Meriden Formation" of Krynine (1950). Krynine defined this unit on the basis of two outcrops (his localities 29 and 30) of a near continuous stratigraphic succession separated by a short covered interval, and described as "typical facies" of the "lacustrine prototype". Section 29 was illustrated by a generalized graphic log, but section 30, which was to represent the upper portion of the intended type section, was mislabelled and actually described the type section of the Upper Sedimentary Member of the Meriden Formation (East Berlin Formation of Lehman (1959)). When Lehman raised the two sedimentary members of the old Meriden Formation to formational rank, he erected a formal type section for the East Berlin Formation, but merely referred to Krynine's original description of the "type" Lower Sedimentary Member of the Meriden Formation to describe the Shuttle Meadow Formation. This practice was also followed by Simpson (1966), who mapped the bedrock geology of the New Britain Quadrangle and provided the first accurate location of Krynine's intended stratotypes.

Since no subsequent workers have properly defined and described a stratotype for the Shuttle Meadow Formation, we designate the lectostratotype for the Shuttle Meadow Formation to be the outcrop on the north side of Cook's Gap in Plainville, Connecticut, as described in Appendix A. Our reasons for discarding the original intended type section follow. (1) The original type section is a composite stratigraphic section that consists of only 10m of strata for which no complete description was ever published. (2) Krynine (1950) described the original type section by a curious mixture of stratigraphic nomenclature and paleoenvironmental terminology (such as "lacustrine prototype"). (3) The original outcrops do not represent the bulk thickness of the Shuttle Meadow Formation, just a local facies. The newly designated type section at Cook's Gap represents approximately 80% of the formation across its outcrop belt. It is dominated by ripple cross-laminated siltstone to muddy siltstone (80%), massive silty mudstone (10%), fine- to medium-grained arkose and sublithic arenite (8%), and massive micrite (2%). And, (4) the Cook's Gap section displays much greater stratigraphic continuity (nearly 50m thick) and visible access.

This spectacular outcrop in Plainville of nearly 50m thickness represents over 80% of the Shuttle Meadow Formation, which is topped by the Holyoke Basalt. Permission must be obtained before viewing this abandoned quarry owned by Tilcon-Tomassa Inc. The lower half of this section is exposed behind the gas station and basket store to the west.

As shown in Figure 3, the lower half of this Shuttle Meadow sequence contains disrupted calcareous shale (Fl_3) (Figure 4), massive mudstone (Fm), ripple cross-laminated muddy siltstone (Fr) (Figure 4), and two beds of limestone (Pm). This is interpreted as a transition between playa mudflats and a muddy alluvial plain dominated by sheetflood deposition (meter 1-10; Figure 3). Both the playa mudflats

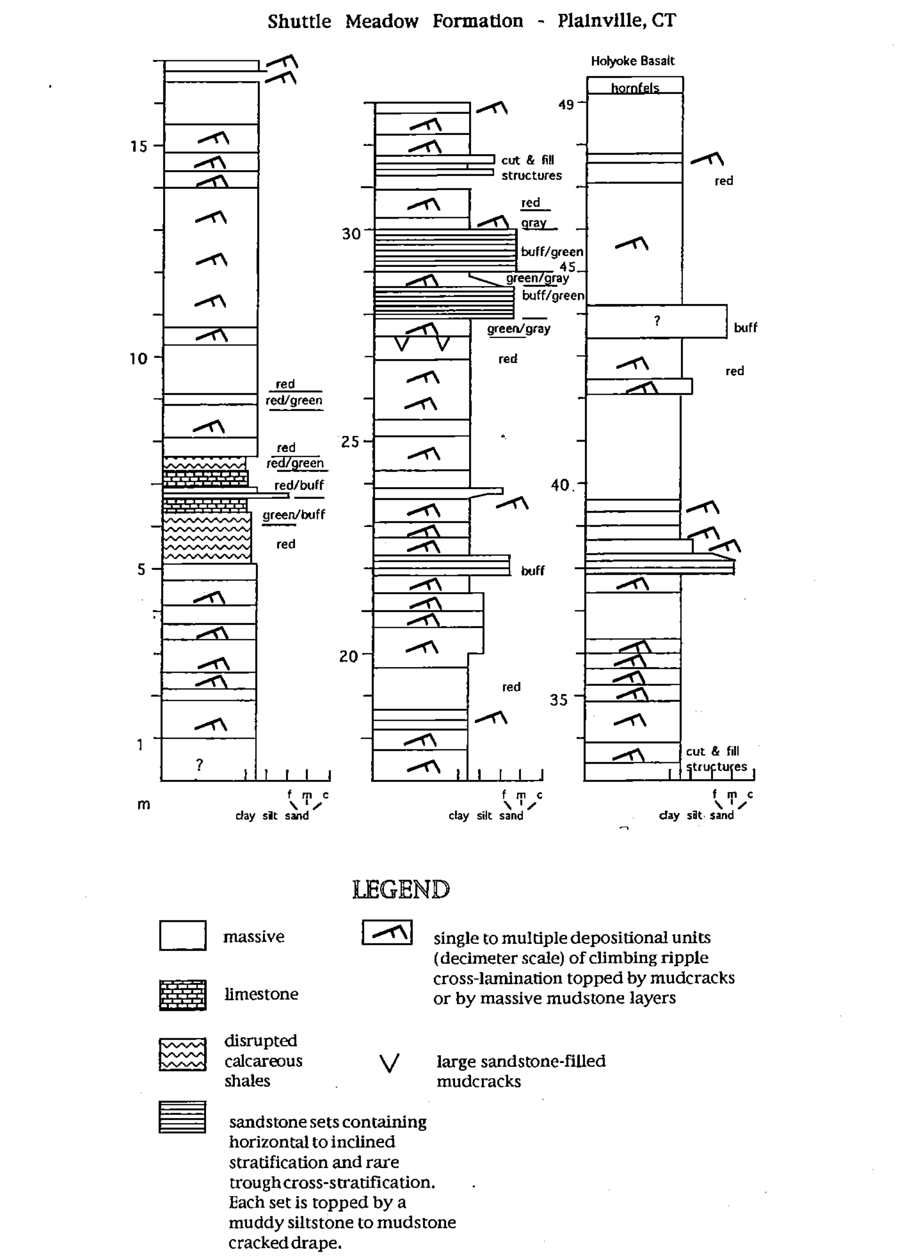

Figure 3 - Sedimentologic log (and its legend) of the Shuttle Meadow Formation exposed at Cook's Gap in Plainville, Connecticut. m=meters, grain size scales at bottom range from clay, silt, fine sand, medium sand, to coarse sand.

and the muddy alluvial plain were subsequently altered by pedogenesis to varying degrees. At meter 9, the green/red mottled mudstone exhibits deformed mudcracks and pedogenic structures evidencing a more saturated soil development, in comparison to other soil zones associated with the arid alluvial plain sequence. The limestone lenses (averaging 20cm in thickness) are

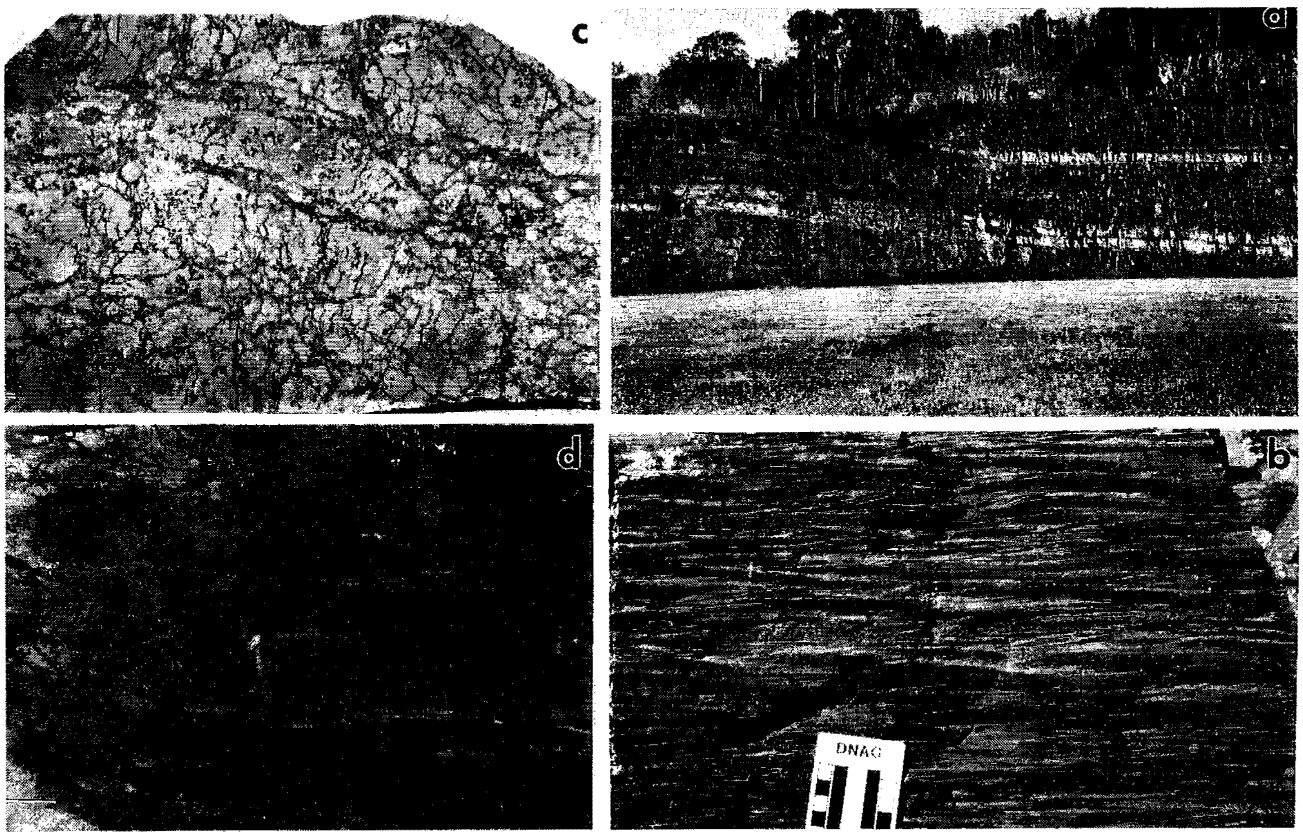

Figure 4 - Rock slabs and outcrop photos from the Shuttle Meadow Formation in Plainville, Connecticut. (a) Outcrop photo. Center of outcrop exhibits prominent, thick sheet sandstone channel at base of outcrop, surrounded by ripple cross-laminated muddy siltstone units.(b) Outcrop photo. Ripple cross-laminated muddy siltstone with climbing ripples. DNAG Scale: centimeters to left, inches to right. (c) Rock slab from upper limestone layer (meter 7). Features are variously oriented rhizoliths filled with clay and spar calcite. Scale bar = 0.5cm. (d) Rock slab of calcareous disrupted shale below carbonate layers. Note mudcracks to the left and well-defined lamination to the right. Scale bar = 0.5cm.

interpreted as carbonate ponds (after Platt and Wright, 1992) associated with the playa mudflats. A sheetflood deposit separates the two carbonate layers (Figure 3). Textures within the limestones include "clastic" carbonate grains, rare gypsum crystals (late diagenetic?), in-situ microbial growth, and root structures (Figure 4). Ostracode shells and fish bones have also been identified (Hubert et al., 1978). A succession of pedogenic alteration of the carbonate paleoponds is evidenced in thin section by the presence of different stages of rhizobrecciation and rhizolith development

The majority of the Shuttle Meadow in the abandoned quarry (Figure 3) contains large sheets of ripple cross-laminated muddy siltstones to fine sandstones (Fr, Sr) (Figure 4), stratified medium-grained sandstones (Sh), and massive mudstone (Fm). These deposits are interpreted as sheetflood deposits containing sand grains as well as pedogenic mud aggregates; sheet morphology is visible across the extent of the outcrop. Deep scour and sheetflood channeling are evident in the red siltstones as well as the yellow/green sandstones. A prominent sandstone bed halfway up the exposure (meter 28-30; Figure 3) illustrates lateral accretion processes associated with sheetflood channeling as successive flood events erode and deposit material along different tracts (Figure 4). Erosional features within channels containing red muddy siltstones, such as cut and fill structures, are less obvious (inclined surfaces and elongate scour of Hubert et al., 1978). The overprint of pedogenic processes on the muddy facies is variable vertically and laterally. The best developed paleosol (vertisol) sequence is visible at meter 27 (Figure 3), just below the prominent sandstone channel. The green-colored silty mudstones interbedded with these channel sandstones, previously interpreted as lakebeds, contain relict ripple cross-lamination preserved by early diagenetic cements associated with soil formation (as in the East Berlin Formation (Gierlowski-Kordesch and Rust, 1994)). These silty mudstones were deposited during the powering-down of the sheet paleoflow; rare macrophytic remains also settled out. Subsequently, pedogenic processes altered its texture, then a rise in water table levels (next flood event?) may have caused reduction and produced the green color.

Fossils in this sequence include ostracodes, dinosaur (*Grallator*) and reptile (*Batrachopus*) footprints, conchostracans, fish bones and scales, and invertebrate trace fossils. A nearly 20-cm long escape burrow near meter 40 (Figure 3) illustrates the episodic but quick depositional processes on the alluvial plain. Fossils were found within the limestone beds and on the upper surfaces and/or mud drapes of siliciclastic depositional units.

40.8 Turn left onto 372 west and turn left onto street leading to Route 72 east/Interstate 84 east. (1.2)

41.4 Enter highway (0.6).

42.8 Take Route 72 east (Exit 35). (1.4)

45.7 Take Route 9 south to Middletown, Berlin (2.9)

49.4 Use exit for Route 372 (Exit 22) (3.7)

49.7 Go straight through traffic light and cross straight over Routes 5/15 (0.3)

49.8 Park on right berm of entrance ramp onto Route 9 south. (0.1)

STOP 4 - East Berlin Formation (Route 5/15 & Route 9, East Berlin)

The East Berlin Formation was named by Lehman (1959) for strata between the Holyoke and Hampden Basalts (Table 1). The East Berlin Formation ranges from nearly 300m in thickness in the southern part of the basin, and like the underlying Shuttle Meadow Formation, thins to the west and north across its outcrop belt to 170m in central Connecticut and 100m near the Massachusetts border. In central Massachusetts, 120m of East Berlin strata are present in the central basin area which thicken and progressively coarsen eastward along the strike valley on the back side of the Holyoke Range, toward the eastern border fault system. The predominate lithologies of the East Berlin Formation include black shales, red mudstones, variegated stratified mudrocks, red silty mudstones and siltstones, red shales, red to yellow arkosic sandstone and minor litharenites, pebble to cobble conglomerates, clast-supported breccias and minor carbonates. The coarser deposits are restricted to exposures in the southern and northern portions of the Hartford Basin, while fine-grained facies comprise the bulk of exposures in the central basin area.

Mid-basin facies encompassing the upper 107m of the East Berlin Formation are well exposed in a series of roadcuts and stream exposures located in central Connecticut (Stops 4 and 5). Nine facies (Table 3) are recognized: trough cross-stratified sandstone (St), horizontally stratified sandstone (Sh), interbedded sandstone and mudrock (SF), ripple cross-laminated siltstone (Fr), black shale (Fl_1), stratified mudrock (Fl_2), disrupted shale (Fl_3) and disrupted mudstone (Fm). These facies form assemblages that represent three continental depositional systems: (1) alluvial plain to sandflat; (2) mudflat-playa and (3) fresh to saline perennial lake. Red/buff sandstone/siltstone facies St, Sh, SF and Fr are attributed to sheetflooding on a distal braidplain to sandflat. These deposits are interstratified with disrupted mudstones (Fm), which are interpreted as vertisols developed on alluvial and playa muds. The lacustrine facies assemblage includes Facies Fl_1, Fl_2, Fl_3, and Fm. Playa mudflat (Fl_3) to ephemeral, shallow lake facies (Fl_2) are distinguished by their lamination style and color. Fl_3 and Fm are interstratified with alluvial to sandflat facies and can contain scattered to bedded carbonate nodules. The offshore perennial lake facies (Fl_1) is mostly gradational with encasing shallow lake deposits (Fl_2), but can rest unconformably on sandsheets of facies Sh. Playa conditions alternated with perennial fresh and saline lake environments throughout East Berlin deposition. Supporting evidence for these interpretations are discussed below.

Alluvial plain to sandflat paleoenvironment

Units of the disrupted mudstones (Fm) are interpreted as alluvial, sandflat and playa muds (Fr, SF, Fl_3) that are pedogenically modified. The paleosols of this facies are interpreted as vertisols in various stages of development. Vertisols are characteristic of semi-arid to arid climates and form as a result of alternations between saturation and desiccation of mostly swelling clays (churning) (Yaalon and Kalmer, 1978; Wilding and Tessier, 1988). Characteristics include lack of horizon development, wide and deep cracks, curved and intersecting slickensides, granular ped development and common carbonate concretions (Retallack, 1988; Martins and Pfefferkorn (1988); Buol et al., 1989). Vertisol development in the East Berlin Formation is demonstrated by sandstone to mudstone-filled cracks up to 1 m in depth; pedogenic textures and clay (argillans) cutans in thin-section

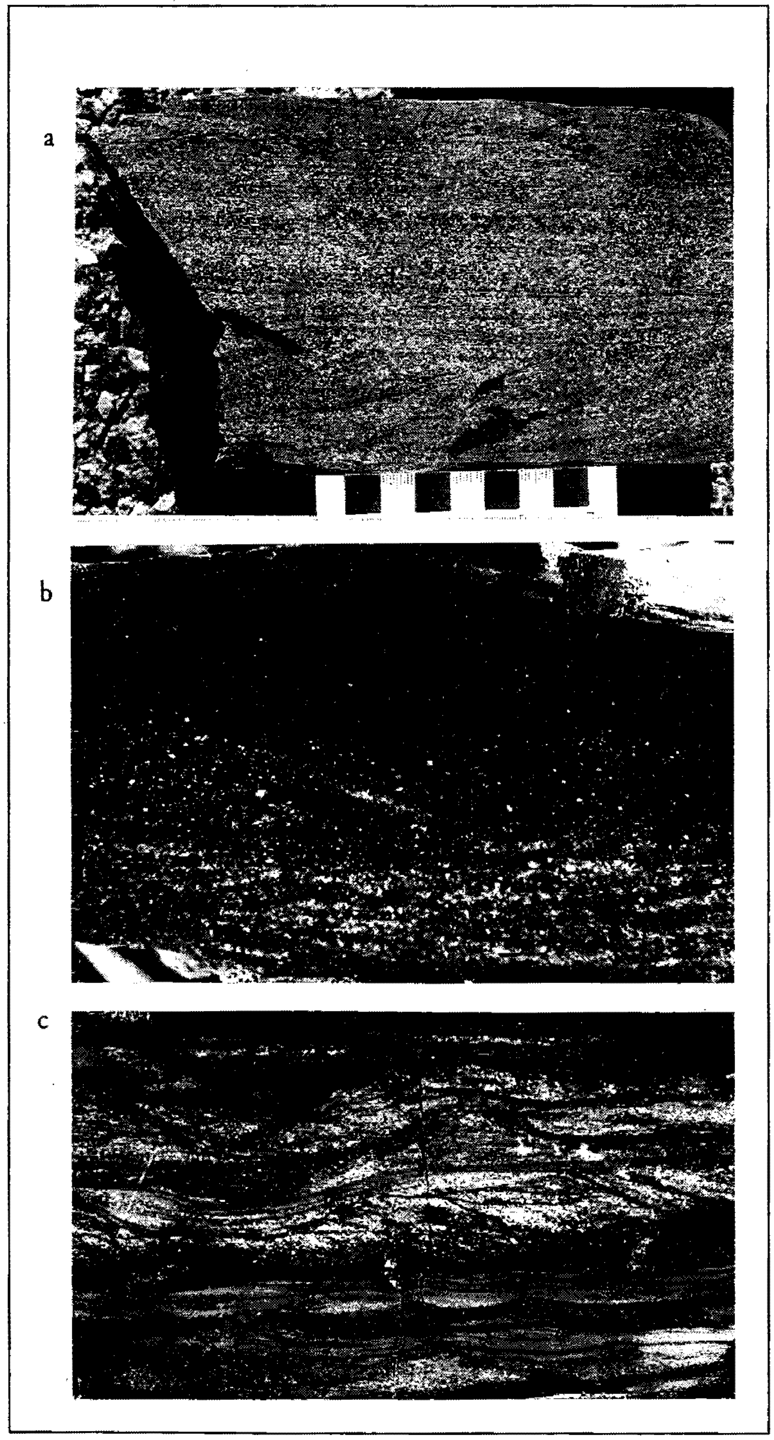

Figure 5 - Sheetflood deposits of the East Berlin Formation in central Connecticut. (a) Rock slab. Trough cross-stratified to horizontally stratified, fine- to medium-grained arkose. Note large mud intraclasts and mud lamina incorporated into cross-sets of trough bedding. Scale bar in centimeters. (b) Rock slab. Fining-upward sequence of ripple cross-laminated muddy siltstone into fine sandstone. Black square at top is 1cm wide. (c) Rock slab. Interbedded sandstone and mudrock exhibiting ripple cross-lamination, cracked mud lamina, and curled mud lamina. Black square at upper left is 2cm long.

(Demicco and Gierlowski-Kordesch, 1986); carbonate nodules; and the presence of swelling clays (Gottfried and Kotra, 1988).

The sandstone and rippled siltstone facies (St, Sh, SF, Fr) (Figure 5) are interpreted as deposits of an ephemeral alluvial plain to sandflat paleoenvironment in which most of the mud was transported as bedload in sand-sized mud aggregates. Mud aggregates act as sand and can develop bedforms in lab experiments (Nanson et al., 1986) and in nature (Rust and Nanson, 1989), as documented from the ephemeral alluvial plain of Cooper Creek in the Lake Eyre Basin of Australia. Evidence for bedload aggregates in the East Berlin Formation includes their occurrence within sandstone-filled mudcracks and as microclasts in sandstone facies (cf. Rust and Nanson, 1989). The sandstone facies, though in minority when compared to the volume of mudrock-dominated facies, exhibit sedimentary structures, sheet morphology and consistent unidirectional paleocurrents (Gierlowski-Kordesch and Rust, 1994), which are characteristic of ephemeral sheetflood deposits (e.g. Karcz, 1972; Sneh, 1983). The ephemeral nature of the depositional system is also shown by the abundance of mudcracks on the upper surface of a depositional unit; i.e. cracked mud drapes on sandstones, mudcracked tops of units of ripple cross-laminated muddy siltstone (Fr), or cracked mud layers or flasers within the SF facies. Also, bioturbation traces are limited to the upper portions of depositional packages (Gierlowski-Kordesch, 1991).

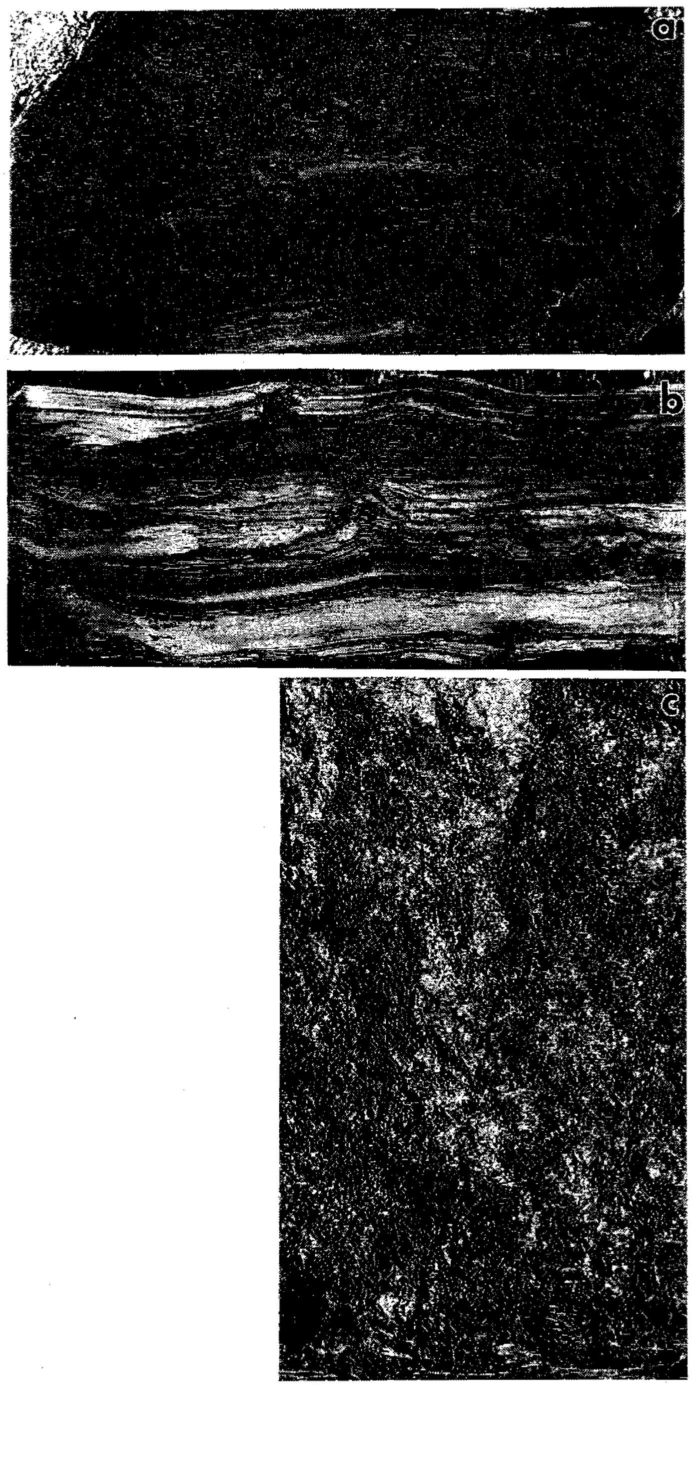

Figure 6 - Playa deposits of the East Berlin Formation in central Connecticut. (a) Rock slab. Disrupted shale block surrounded by mudcracked shale to massive mudstone and sandstone-filled mudcracks and layers. Lamination style of disrupted shale: thicker, fine alternations of siltstone and claystone with rare carbonate lamina (crusts). Mound feature at bottom implies microbial origin. Black square at bottom left is 2cm long and 1 cm wide. (b) Rock slab. Disrupted shale with mudcrack features and fine clay laminae intercalated with disrupted to massive muddy siltstone lamina - thin "microbial" lamination contains tiny bumps and ridges. Black square at bottom left is 2cm long. (c) Outcrop photo. Laminated to massive sequence of aggrading playa deposits. Note sandier texture of disrupted to massive mudstone above cracked horizontal lamination. Tip of pen in lower right for scale.

Sheetflood sandstones occur mainly within red alluvial plain to sandflat sequences containing paleosols. Some sandstones, however, are present within gray-green sequences (complete lake sequences) where they commonly are mottled gray/red and exhibit soft-sediment deformation. These sandstones are attributed to sheetfloods which spread across the playa surface and were subsequently inundated by a rise in lake level, or which entered marginal parts of the lake. They are synonymous with facies LM3 of Smoot (1985, Figure 2.3) and Smoot and Lowenstein (1991; Figure 3.21) for small basins where alluvial deposits intersect lake level. The main difference between our interpretation of these East Berlin rippled siltstones and that of Smoot (1991) lies in the exact distribution of this facies. The rippled siltstones in green-gray sequences can be defined as deltaic sheets associated with lake margins, whereas those in red sequences are sheetflood deposits within an alluvial-sandflat setting (similar to Cooper Creek). Both types of rippled siltstones are deposits of sheetflood deposition resulting from flow deceleration due to high water levels, but deposition on a flooded lake margin (which is changing position perhaps with each flood) versus on a flooded alluvial plain are differentiated.

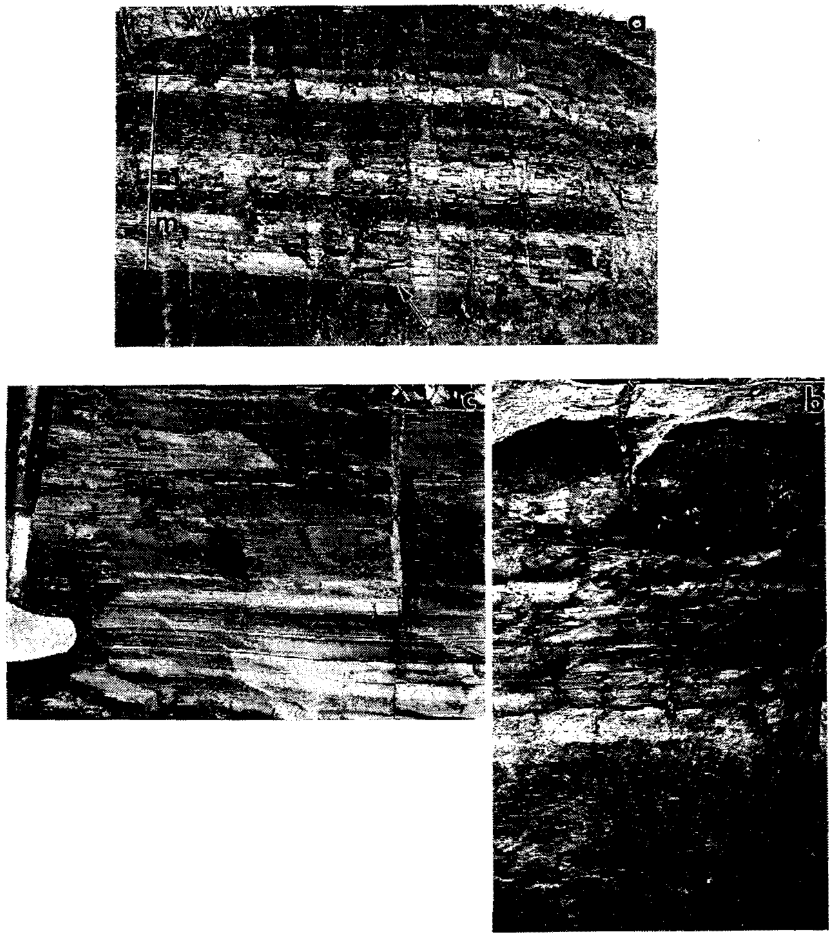

Figure 7 - Complete and incomplete lake cycles in the East Berlin Formation in central Connecticut. (a) Outcrop photo. 7th lake cycle (4th complete lake cycle) at Section A, East Berlin. The green, yellow, and black rocks (extent of line) comprise one complete lake cycle containing a black shale core (b) encased by stratified mudrocks (m) and green disrupted shales and mudstones and buff sandstones. Rock hammer (arrow) for scale. (b) Outcrop photo. Incomplete lake cycle (5th lake cycle). Massive muddy sandstone at bottom grades upward into red/yellow stratified mudrock containing many levels of mudcracking. Uppermost layer of massive mudstone contains broken intraclasts from mudrock below. Lens cap at right for scale. (c) Outcrop photo. Close-up of greenish/grey stratified mudrock with continuous planar lamination below the shale of the 4th complete lake cycle. Hammer for scale.

Lacustrine paleoenvironment

The green-yellow to red disrupted shales (Fl_3) of the East Berlin Formation are interpreted as dry playa mudflats. These shales are termed disrupted because bedding is rarely preserved across a distance of meters. Disrupted shales are normally topped by and laterally equivalent to units of facies Fm; this transition from laminated to massive in vertical sequence (Figure 6) is characteristic of aggradational processes in playas (Demicco and Gierlowski-Kordesch, 1986). Lamination style includes two main types: (1) thicker, fine alternations of siltstone and claystone and (2) possible very thin microbial lamination (Figure 6). Similar modern features in playas are described by Hardie et al. (1978); Bauld (1986) and Renaut (1993). Supporting evidence for lacustrine microbes in these East Berlin mudrocks includes geochemical analysis of kerogen by Spiker et al. (1988), Kotra et al. (1988) and Kruge et al. (1990;1991).

The disrupted shales contain rare ripple cross-lamination and appear to be transitional in form to the associated Fr facies. This gradation in sedimentary structures from Fl_3 to Fr indicate that the playa to alluvial plain/sandflat transition was gradual on a uniformly sloping surface. Other evidence includes the observation that no well-defined shoreline deposits have been identified in the central basin outcrop area of the East Berlin Formation. Changes in water level caused rapid shoreline migration, but lake level does not appear to have been at equilibrium for the length of time required to develop beach deposits - a typical feature of playa lakes (cf. Smoot, 1985). In addition, the nearly constant pedoturbation of the muds in playa, sandflat, and alluvial plain paleoenvironments contribute to the gradual nature of the alluvial/lacustrine transition (cf. Bowler, 1986; Smoot and Lowenstein, 1991). The East Berlin Formation is unusual because its sandflat/alluvial plain contained mostly muddy sediment. Vertisol features can develop in 10 to 1000 years (Allen and Wright, 1989) and eradicate primary depositional structures in both playa mudflat and alluvial plain/sandflat environments.

Stratified mudrocks (Fl_2) with continuous and planar lamination are interpreted as ephemeral shallow lake deposits which record the farthest extent of lacustrine influence. Bioturbation only occurred on a lamina scale; shallow conditions were neither oxic nor fresh. These mudrocks typically envelope black shales (Fl_1), but also occur without any associated black shales (Figure 7). The upward increase in mudcrack density in isolated units and in regressive subfacies above black shales is attributed to lake desiccation, as suggested by rare evaporites (Hubert et al., 1976;1992; Kruge et al., 1990;1991). The rarity of mudcracks within mudrocks in transgressive subfacies directly below and gradational with black shales in contrast implies rapid rise of base level at the onset of the ensuing 'wet' cycle.

Black shales (Fl_1) represent offshore sediments in perennial lakes. The three types of black shales (Figure 8) represent different paleoconditions. The irregularly-laminated shale is similar to "offshore" deposits that may be only a few meters deep, such as in modern shallow saline lakes cored in Australia and Canada (Last, 1990; Last and De Deckker, 1990; 1992). Discontinuous carbonate lamina undergo syndepositional to very early cementation and can easily break and shift due to compaction and gravity sliding (cf. Jackson and Galloway, 1984). The magnesite-bearing shale cycle is interpreted to represent saline lake conditions. Modern saline lakes precipitating magnesite include Lake Beeac in Australia (Teller and Last, 1990; Last and De Deckker, 1992), Milk Lake, British Columbia (Renaut and Stead, 1991), and Freefight Lake, Saskatchewan (Last, 1993). The finely microlaminated shale type resembles laminites (cf. Platt and Wright, 1991) deposited in a freshwater lake. Additional evidence for salinity in East Berlin Formation mid-basin areas include evaporite pseudomorphs in facies Fl_2, Fl_3 and Fr and traces of evaporites in Fl_2. Reduced bioturbation patterns are attributed to substrate toxicity and dysoxia common in shallow saline lakes and playas (Smoot and Lowenstein, 1991; Gierlowski-Kordesch, 1991).

The relatively new road cut (Section A) in East Berlin exposes the upper 107m of the East Berlin Formation as well the Hampden Basalt above. The remaining portion of the Type Section is just to the north on Route 372. Figure 9 gives a schematic overview of Section A (legend in Figure 10). Eight facies are recognized: trough cross-bedded sandstones (St), horizontally stratified sandstones (Sh), interbedded sandstones and mudrocks (SF), ripple cross-laminated siltstones (Fr), black shales (Fl_1), stratified mudrocks (Fl_2), disrupted shales (Fl_3), and disrupted mudstones (Fm) (Table 3). Interpreted as a continental depositional system, the facies are divided into two assemblages. The sandflat/alluvial plain facies assemblage (St, Sh, SF, Fr, Fm) is composed of sheetflood deposits from a sandflat/alluvial plain complex with the muddier sediments altered into vertisolic paleosols. The lacustrine facies assemblage (Fl_1, Fl_2, Fl_3, Fm) represents a saline lake-playa system also affected by vertisol development (Gierlowski-Kordesch and Rust, 1994).

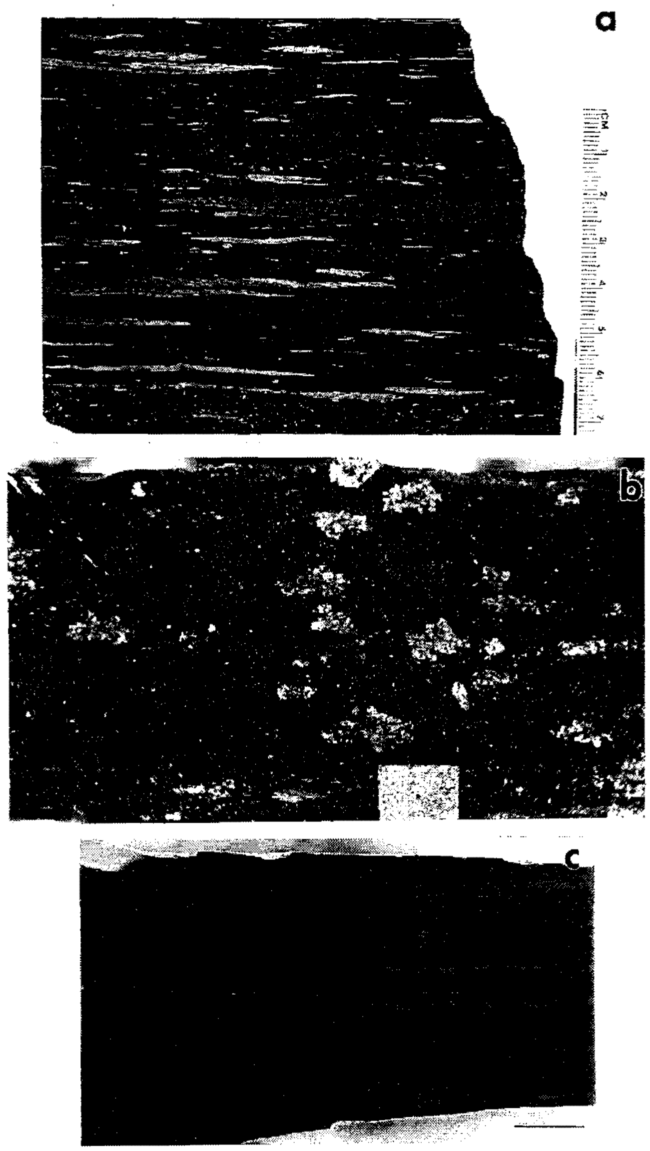

Figure 8 - Black shale types of the East Berlin Formation of central Connecticut (see Table 3). (a) Rock slab. Irregularly laminated shale with discontinuous carbonate and claystone lamination. Note internal stratification of some carbonate "lamina" are oblique, not horizontal. Ruler at right for scale. (b) Rock slab. Magnesite-bearing shale with magnesite microspar nodule. White block at bottom is 1cm wide. (c) Finely laminated shale with continuous to discontinuous lamina of claystone and carbonate. Black lenses at bottom half of slab are fish or coprolite remains. Scale bar = 0.5m.

Nine of the ten lake cycles identified in the upper East Berlin Formation are exposed here as incomplete and complete lake sequences. Incomplete lake cycles contain red/yellow stratified mudrocks with disrupted shales or sandstones. Complete lake sequences or cycles are composed of a core of black shale enveloped by green/gray stratified mudrock followed by green disrupted shales; sandstones commonly compose the outer layers above and/or below (Figure 7). Please note that the first complete lake cycle (lake cycle 2) contains the magnesite-bearing black shale and the third complete lake cycle (lake cycle 6) contains the finely micro-laminated black shale (examples in Figure 8). The yellow/green sandstones associated with lake cycles are interpreted as deltaic sheets, formed as sheetfloods entered the paleolake. Red/yellow strata between lake sequences contain playa mudflat facies, sheetflood facies, and massive paleosol mudstones. Prominent paleosol sequences, interpreted as vertisols, can contain meter-long sand-filled mudcracks and yellowish carbonate nodules within massive red mudstones.

Small-scale and large-scale mudcracks pervade the entire East Berlin sequence. Mudcracks commonly define the tops of depositional units of ripple cross-laminated muddy siltstones and occur on mud drapes topping sandstone sets. Mudcracks pervade stratified mudrock and disrupted shale sequences (shallow lake to playa mudflat deposits). Invertebrate trace fossils are commonly found only at the top of depositional units (Gierlowski-Kordesch, 1991). All these features indicate ephemeral depositional conditions. Playa sequences are rarely continuous, but grade laterally into massive mudstones, illustrating the processes of pedogenic alteration. Evaporite pseudomorphs are rare but can be found in most of the fine-grained facies. Fossils in this sequence and those of Stop 5 include fish and macrophytic plant remains, fish scales, coprolites, conchostracans, ostracodes, invertebrate trace fossils, and dinosaur footprints.

52.1 Drive on Route 9 south and stop on berm after the exit sign for Interstate 91 south (Exit 20S) (2.3)

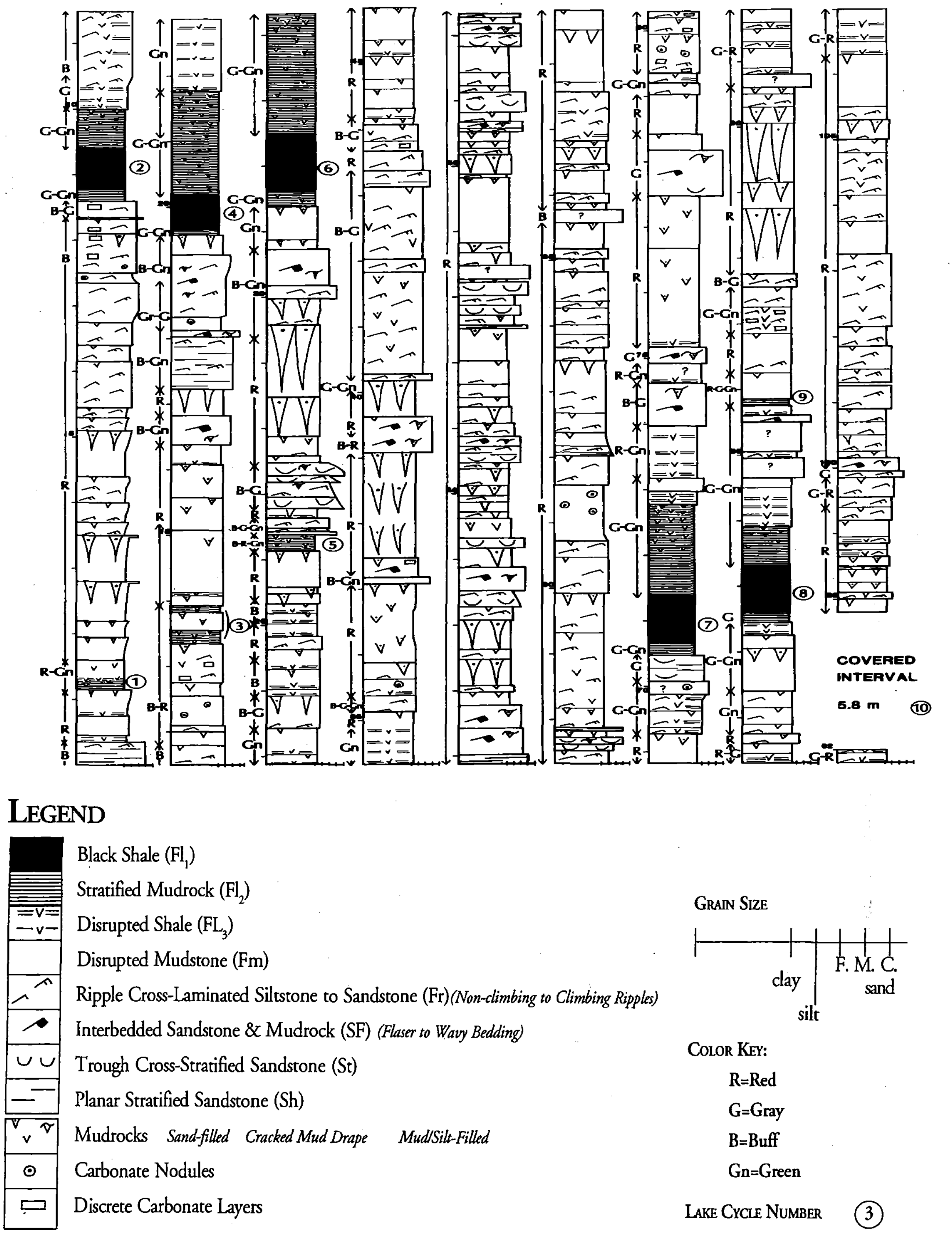

Figure 9 - 10 Sedimentologic log of Section A of the East Berlin Formation located near East Berlin, Connecticut. Tick marks and numbers on vertical scale indicate thickness in meters. Horizontal scale at bottom shows grain size scale. From Gierlowski-Kordesch and Rust (1994) - copyright permission.

STOP 5 - East Berlin Formation
(Route 9 & Interstate 91, Cromwell)

The upper 44m of the Q, R, and X sections in Stop 5 are compared to the Type and A Section of the East Berlin Formation in Figure 11. Correlation shows the tenth lake cycle is exposed here. Of interest is the upper portion of the East Berlin with its contact with the Hampden Basalt. Fine red siltstones (with evidence of possible paleosol gleying) exhibit two different types of evaporite pseudomorphs - star-shaped balls and probable gypsum pseudomorph crystals (Gierlowski-Kordesch and Rust, 1994). The star-shaped balls exhibit evidence for contemporaneous transport and deposition. Nearer to the contact with basalt, probable diagenetic pseudomorph crystals as well as zeolites occur, related to baking by the Hampden lava paleoflow. Perhaps the evaporite pseudomorphs in the red siltstones of the upper East Berlin Formation are especially well preserved here because the adjacent "baking effect" enhanced preservation potential and occurred before removal of the evaporites by normal groundwater processes. More research is needed on differentiating between early and late diagenetic processes.

53.7 Continue on Route 9 south and exit at Route 372, Cromwell (Exit 19) (1.6)

54.0 Turn right onto Route 372 west. (0.3)

55.3 Continue straight on Route 372 and turn right into the parking lot of the Radisson Hotel.

LUNCH

55.4 From hotel parking lot, turn right onto 372 and immediately enter ramp leading to Interstate 91 north to Hartford. (0.1)

58.5 Use Route 3 exit for Rocky Hill (Exit 23). (3.1)

59.1 Turn right (east) onto Route 3 (West Street). (0.6)

59.8 Turn right into the entrance for Dinosaur State Park. (0.7)

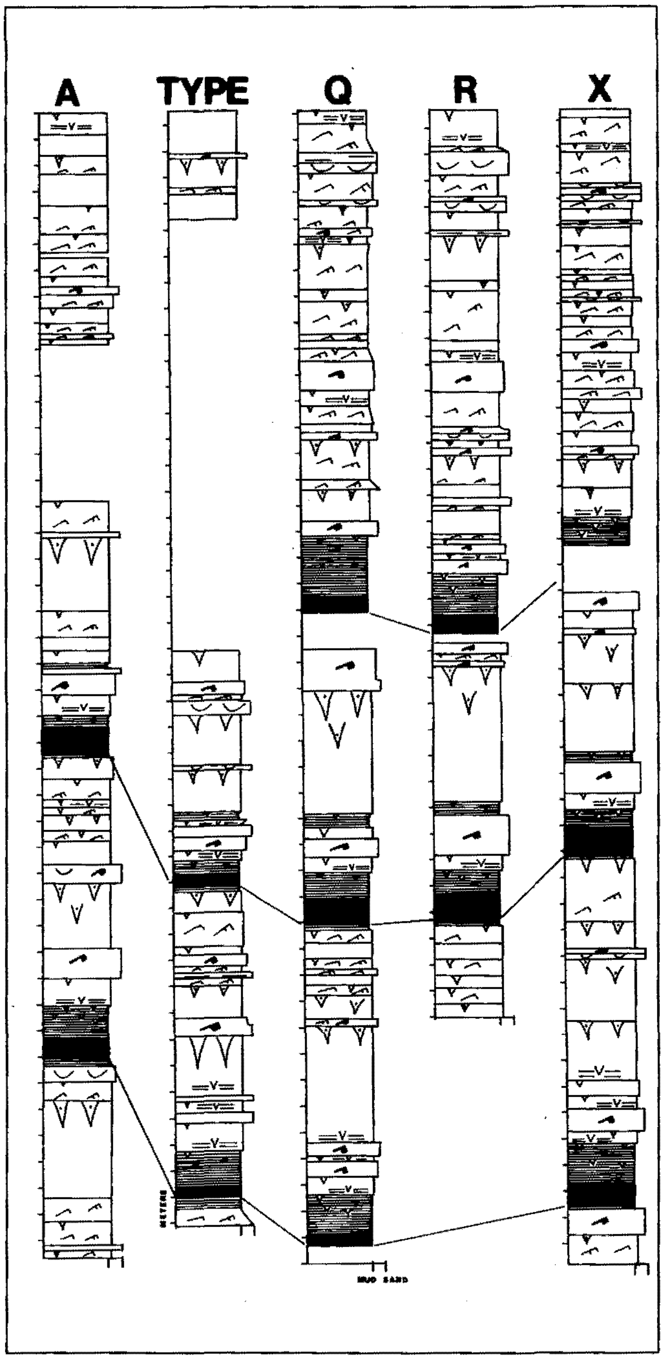

Figure 11 - Measured sections of the upper 44m of Sections A, Type, Q, R, and X of the East Berlin Formation in central Connecticut. Correlation stretches from East Berlin (A, Type) (Stop 4) to Cromwell (Q,R,X) (Stop 5). The top of each section is defined by the Hampden Basalt. Legend in Figure 10. From Gierlowski-Kordesch and Rust (1994) - copyright permission.

STOP 6 - East Berlin Formation (Dinosaur State Park, Rocky Hill)

During construction at a site in Rocky Hill in 1966, dinosaur trackways were uncovered. More than 35,000 sq. ft. of bedding plane was unearthed with abundant fossil footprints. Soon afterward, the area was designated Dinosaur State Park. A portion of the trackways are exhibited within the main building; the rest of the exposure is buried. Examples of the typical strata and various dinosaur tracks are also displayed outside.

These trackways are found in an East Berlin Formation lake sequence which is tentatively correlated to the 10th lake cycle in the upper part of the formation (Ostrom, 1968). Two vertebrate ichnotaxa of footprints have been found at Rocky Hill : *Grallator* (*Eubrontes*) and *Grallator* (*Anchisauripus*). Walking patterns across the trackway appear to be random in orientation. Preservation of the footprints depended on the grain size and paleoconsistency of the sediment. Wave-ripple marks, soft sediment deformation, raindrop impressions, invertebrate trace fossils, and mudcracks also occur on the trackway within the shallow lake/playa and sheet delta deposits.

60.5	Turn left onto Route 3 and return to highway interchange.
	Turn left into entrance ramp for Interstate 91 north to Hartford. (0.7)
64.9	Take Route 3 north to Glastonbury. (4.4)
67.3	Cross over Connecticut River and take Route 2 east to Norwich. (2.4)
68.0	Take Route 17 south to Portland/ S. Glastonbury (0.7)
72.0	Pass through the junction with Route 160. (4.0)
73.7	Turn right onto Old Maids Lane. (1.7)
74.1	Turn left on road before third barn; it turns into a gravel road. (0.4)
74.4	Entrance to quarry. (0.3)

STOP 7 - Portland Formation (Quarry in South Glastonbury)

The Portland Formation is distributed across the eastern one-third of the Hartford Basin (Figure 1) and is the stratigraphically youngest formation of the Hartford Group. This formation was named by Krynine (1950, p. 69) for rocks previously called "eastern sandstone" by Percival (1842), but no formal stratotype was ever designated nor was a detailed stratigraphic section described. Krynine (p. 69-70) did state that "the best exposures are found near Middletown in the brownstone quarries" and that these exposures "provide a good series of outcrops illustrating all the lithologic variations from micaceous siltstone to medium-sized conglomerates".

The Portland Formation is approximately 450m thick in the Totoket syncline (Sanders, 1968; 1970), and increases northward to 1200m near Middletown (Lehman, 1959), attaining a maximum estimated thickness of 2600m near the Connecticut-Massachusetts border (McInerney, 1993). Because of the discontinuous nature of the Portland outcrops and the lack of encasing stratigraphic marker beds (i.e. basalts), it is not possible to precisely correlate any given exposures. However, detailed mapping by

LeTourneau (1985a) and field observations by the authors indicate the presence of at least nine gray siltstone-shale units, some of which may extend as far north as Hartford. Because of the easterly dip of Portland strata, a complete series of depositional facies can be examined that range from mid-basin sheetflood/playa/lake to river/beachfront/lake to eastern border alluvial fan.

Detailed sedimentologic investigations by Gilchrist (1979), Hubert et al. (1982), LeTourneau (1985a,b), LeTourneau and McDonald (1985), McDonald and LeTourneau (1988), and McInerney (1993) demonstrated the Portland Formation to be composed of nine major facies associations containing a total of 11 facies. These are listed from LeTourneau (1985a,b) with revised component facies designations below:

(1) Poorly-sorted, matrix-supported boulder and cobble conglomerates (Gms)

(2) Clast-supported lenticular conglomerate associated with pebbly coarse-grained sandstone beds (Gt, Gp, Sh)

(3) Trough and planar cross-stratified conglomerates associated with pebbly sandstones (Gt, Gp, St, Sp)

(4) Planar and ripple cross-laminated silty sandstone (Sh, Sr, Fr)

(5) Cross-stratified coarse and pebbly sandstone with thin-bedded to laminated, mudcracked siltstone (St, Sp, Fr, Fm)

(6) Massive to thin-bedded siltstone with lenses and thin beds of coarse, cross-stratified sandstone (Fm, Fr, St, Sp)

(7) Ripple cross-laminated to planar laminated medium to coarse sandstone and ripple cross-laminated to wavy- bedded siltstone (Sr, Fr)

(8) Black shale and gray mudrock which can be associated with limestone (Fl_1, Fl_2).

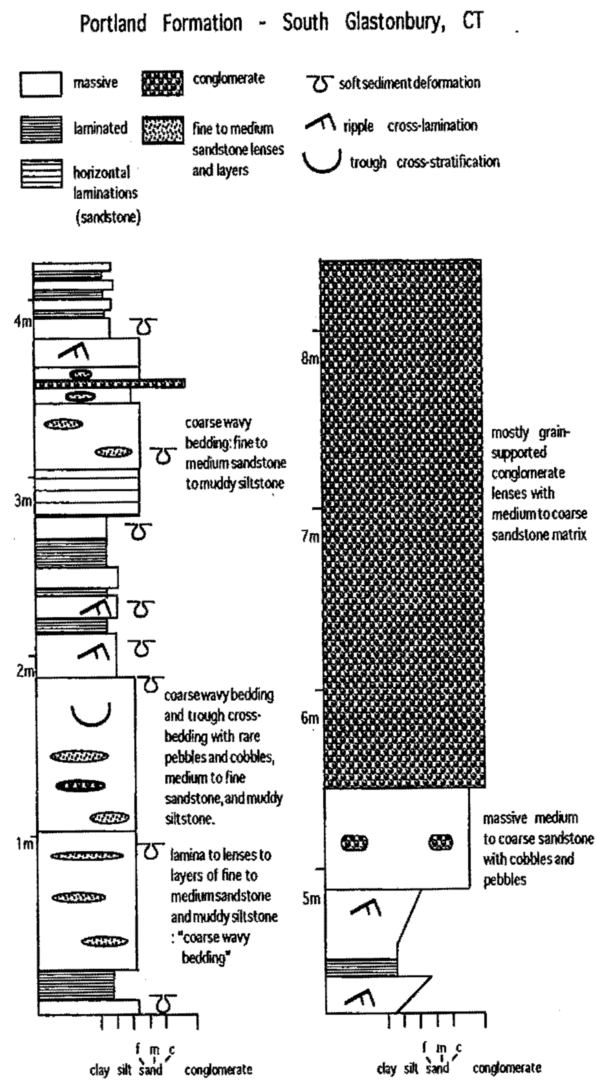

Figure 12 - Sedimentologic log of the Portland Formation outcrop in South Glastonbury, Connecticut.

Conglomeratic facies form sedimentary prisms along the eastern margin of the basin. Two types of sequences characterize the Portland Formation in central Connecticut. Facies associations 3-7-8-7 (phase 1) typically occur in 5-10m thick packages and facies associations 6-5-4-3-2-1 (phase 2) are found as up to 12m thick coarsening-upward sequences. Stacked sequences of these phases alternate and are each approximately 175m thick. Together with paleocurrent dispersal patterns, LeTourneau (1985a,b) proposed a "wet-dry" sedimentation model. Phase 1 represents a lake transgression over stream-dominated distal alluvial fan deposits during humid conditions. Phase 2 is interpreted as a more ephemeral, dry period characterized by prograding alluvial fans over sheetflood deposits. In addition, the influence of syndepositional tectonic activity is evidenced by changes in radial patterns of sediment dispersal with time. Coarse deposits extend only one to three kilometers from the eastern border fault, representing only mid- to lower-fan deposits. The upper fan deposits were probably eroded by post-depositional uplift. Eastward tectonic tilting of the basin floor may have prevented prograding alluvial fans from extending far into the basin by continually burying the coarser sediments under fine distal deposits. Stops 7 and 8 illustrate the wet Phase 1 cycle within the Portland Formation.

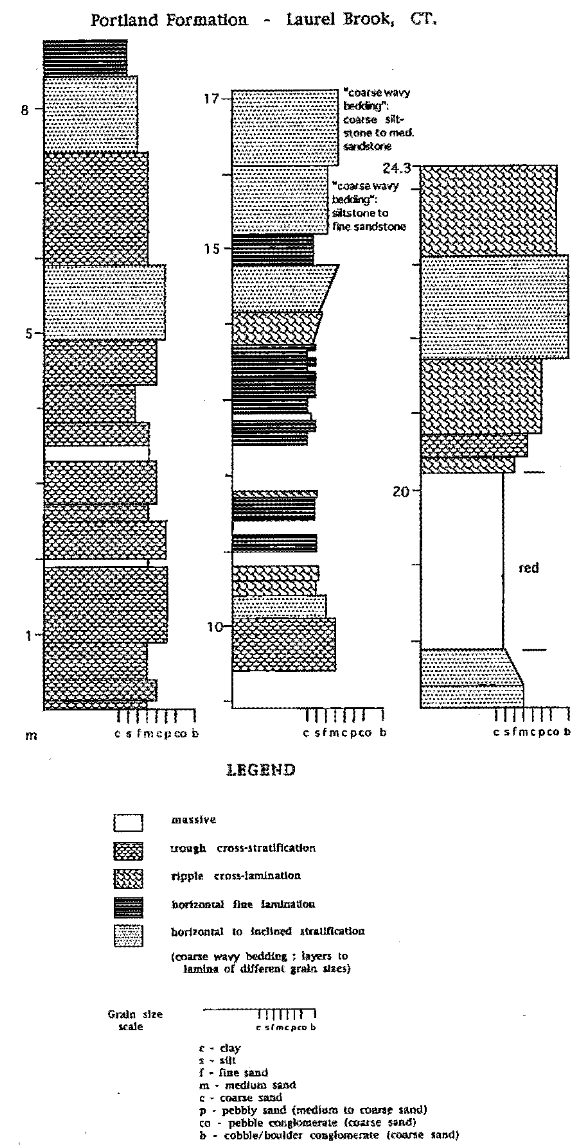

Figure 13 -14 Sedimentologic log of the Portland Formation stream exposure near Middlefield, Connecticut. Adapted from Letourneau (1985a).

The Glastonbury outcrop near Route 17, which is less than 1km from the eastern boundary of the Hartford Basin (Horne et al., this volume), exposes over 8m of the Portland Formation (Figure 12) . It contains a lake sequence associated with alluvial fan facies along the eastern border fault. The black muddy siltstones to mudrocks (Fl_1, Fl_2) are coarser-grained than more distal lake deposits in the basin; they contain well-preserved articulated fish fossils as well as macrophytic plant debris. The interbedded sandstone laminations, lenses, and layers throughout the Portland sequence, which also can contain macrophytic plant debris, indicate the episodic influx of siliciclastics into the lake (Fr, Sr, Sh (coarse wavy bedding), St). Soft sediment deformation features are common in the finer sandstones of the lower portion of this sequence. The coarser upper deposits of this sequence (pebbly to cobbly sandstones to cobble and boulder conglomerates - Gt, Gp, Sh, St, Sp) are interpreted as fan delta to alluvial fan sediments. The dip surface of the uppermost layer of conglomerate is beautifully displayed on the south side of the outcrop, polished by glacial action. The various source rocks from the Eastern Highlands are present within the clast population of the conglomerate.

75.1	Retrace steps back Route 17. Turn left (south). (0.7)
77.8	Turn right onto Route 17A. (2.7)
78.9	Enter the town of Portland. (1.1)
80.8	Go straight through junction with Routes 66/17. (1.9)
81.8	Go back over Connecticut River and enter the town of Middletown.
	Turn left at light following signs for Route 17. (1.0)
81.9	Turn right onto Routes 66/17. (0.1)
82.7	Take Route 17 south to New Haven (Exit 13). (0.8)
83.2	Turn left at traffic light following Route 17 south. (0.5)
84.8	Pass straight through junction with Route 155. (1.6)
87.0	Position of old weigh station, now under construction. (2.2)
87.3	Turn right into parking lot of industrial building. Outcrop is less than 1km behind the large building.

STOP 8 - Portland Formation (Laurel Brook, Middlefield)

This locality is 500m south of Laurel Brook Reservoir, 750m west of Route 17, south of Middletown, behind a transport company. Studied by Gilchrist (1979) and LeTourneau (1985a), this stream exposure (Figure 13, 14) contains up to 30m of conglomerates, sandstones, and finer-grained rocks of the Portland Formation. Facies exhibited here include: Cobble/boulder conglomerates (Gms), pebble/cobble conglomerates (Gm), pebbly sandstones (Gt), cross-stratified and inclined planar to horizontally stratified, fine- to coarse-grained sandstones (Sh, St), ripple cross-laminated siltstones to fine sandstones (Sr, Fr), silty mudstones (Fm), and mudrocks to shales (Fl_1, Fl_2) (LeTourneau, 1985a). These facies are interpreted as alluvial fan to alluvial plain/ mudflats to marginal to deeper lacustrine sediments. A complete transgressive to regressive lacustrine cycle is marked by a succession of fluvial to littoral sandstones, followed by siltstones and shales, and topped by littoral sandstones and alluvial fan conglomerates. The littoral to shallow lacustrine sequence below the shales are thinner than those above the shales. This is interpreted as a "quick" transgression followed by a regression of long duration. This asymmetry is also present in the lake sequences of the East Berlin Formation in the central part of the basin (Gierlowski-Kordesch and Rust, 1994).

The red silty mudstones, pebbly sandstones, and conglomerates above the fine lacustrine rocks can be represented as alluvial plain/mudflat sediments associated with shoreline deposits (well-sorted bars and wave-generated beach ridges) leading into alluvial fan deposits. The fossils found at this locality include well-preserved fishes, macrophytic remains, coprolites, and rare reptile tracks (McDonald, 1982).

91.2	Turn left back onto Route 17 north, back to Middletown. Turn right at traffic light toward Route 9. (3.9)
91.6	Take exit to Route 9 north, direction of New Britain. (0.4)
95.6	Exit for Route 372, Cromwell (Exit 19). (4.0)
95.9	Turn left onto Route 372 west. (0.3)
97.4	Drive straight on. Turn right into the parking lot for the Radisson Hotel. (1.5)

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Appendix A

Cook's Gap, Plainville, Connecticut
Shuttle Meadow Formation Lectostratotype
(UTM coordinates 575500, 310500)

Hartford Group - Shuttle Meadow Formation

Unit 30 - Basalt, dark greenish gray (5G 4/1), massive	not measured
Unit 29 - Siltstone, grayish black (N 2), massive, well- indurated, hornfels	0.3
Unit 28 - Siltstone, dark reddish brown (10R 3/4), massive	1.4
Unit 27 - Siltstone, grayish red (10R 4/2), ripple cross-lamination	0.1
Unit 26 - Siltstone, grayish red (10R 4/2), clayey, massive	3.4
Unit 25 - Sandstone, pale reddish brown (10R 5/4), fine-grained, well sorted, ripple cross-lamination, calcareous	0.8
Unit 24 - Siltstone, moderate reddish brown (10R 4/6), massive; interstratified with ripple cross-lamination zones up to 6 cm thick	4.4
Unit 23 - Muddy siltstone, grayish red (10R 4/2), ripple cross-lamination, more clayey zones contain vertebrate tracks	0.8
Unit 22 - Sandstone, grayish red (10R 4/2), medium- to coarse-grained, moderately to well sorted, massive, lenticular, inclined bedding unit, coarse fraction contains volcanic glass shards, calcareous	0.2
Unit 21 - Muddy siltstone, moderate reddish brown (10R 4/6), ripple cross-lamination to massive, more clayey zones contain invertebrate trails, vertebrate tracks, and conchostracans	7.2
Unit 20 - Sandy siltstone and muddy siltstone; sandy siltstone is moderately red (5R 5/4), forms two tabular inclined bedded units that cut and fill underlying siltstone units; interbedded muddy siltstone is pale reddish brown (10R 5/4), massive	0.7

Unit 19 - Sandy siltstone and muddy siltstone; sandy siltstone is grayish red (10R 4/2), ripple cross-lamination; muddy siltstone is pale reddish brown (10R 5/4), forms clayey zones on sandy siltstone units, calcareous	0.7
Unit 18 - Sandstone, siltstone, and muddy siltstone; sandstone is pale yellowish brown (10YR 6/2), fine-grained, well sorted, planar cross-stratification; siltstone is grayish red (10R 4/2), ripple cross-lamination to massive, forms thin zones interstratified with sandstone units; muddy siltstone is dark gray (N 3), pedogenically modified, massive, micaceous with relict ripple cross-lamination	1-3.5
Unit 17 - Sandstone, siltstone, and muddy siltstone; sandstone is moderate yellowish brown (10YR 5/4), fine-grained, well sorted, massive; siltstone is grayish red (10R 4/2), ripple cross-lamination; capped by clayey zones of muddy siltstone that are grayish red (10R 4/2). Unit is penetrated by mudcracks up to 0.3m deep	6.8
Unit 16 - Sandstone and siltstone; sandstone is pale reddish brown (10R 5/4), fine- to medium-grained; well sorted, lenticular with intraclasts of muddy siltstone, trough cross-stratification, calcareous; siltstone is dark reddish brown (10R 3/4), forms thin ripple cross-lamination zones intercalated with sandstones, calcareous, unit cut and fills top of underlying siltstone	0.1-0.4
Unit 15 - Siltstone, moderate reddish brown (10R 4/6), ripple cross-lamination, calcareous	2.3
Unit 14 - Siltstone, moderate reddish brown (10R 4/6), massive, mudcracked	1.0

Unit 13 - Siltstone, dark reddish brown (10R 4/6), ripple cross-lamination, soft sediment deformation structures, occur in units up to 25cm thick that fine up into massive, clayey units which contain relict ripple cross-lamination	1.8
Unit 12 - Siltstone, dark reddish brown (10R 4/6), ripple cross-lamination, abundant invertebrate and vertebrate trace fossils	3.8
Unit 11 - Siltstone, dark reddish brown (10R 4/6), massive, mudcracked	1.9
Unit 10 - Siltstone and muddy siltstone; siltstone is grayish red (10R 4/2), ripple cross-lamination; muddy siltstone is grayish red (10R 4/2), massive, mudcracked	1.6
Unit 9 - Muddy siltstone, moderate reddish brown (10R 4/6), ripple cross-lamination to massive; massive zones are mottled, grayish red (10R 4/2) and pale yellowish green (10GY 7/2)	0.2
Unit 8 - Sandy siltstone and muddy siltstone; sandy siltstone is grayish red (10R 4/2), massive; muddy siltstone is moderately reddish brown (10R 4/6), ripple cross-lamination	1.3
Unit 7 - Muddy limestone, pale greenish yellow (10Y 8/2) to moderately red (5R 5/4), massive, mudcracked micrite	.30
Unit 6 - Siltstone - moderately reddish brown (10R 4/6) and pale yellowish brown (10YR 6/2), massive, calcareous	.25
Unit 5 - Muddy and sandy siltstone, moderately reddish brown (10R 4/6), ripple cross-lamination, calcareous	.25

Unit 4 - Sandstone, grayish red (10R 4/2), fine-grained, well sorted, massive	.10
Unit 3 - Muddy limestone, pale yellowish green (10Y 8/2) to pale red (10R 6/2), massive, mudcracked micrite	.30
Unit 2 - Siltstone and muddy siltstone; siltstone is pale reddish brown (10R 5/4), ripple cross-lamination to massive; muddy siltstone is pale reddish brown (10R 5/4), massive, mud-cracked, calcareous	1.2
Unit 1 - Siltstone, moderately reddish brown (10R 4/6), ripple cross-lamination to massive, mudcracked	3.1
Base of section is covered, but begins approximately 40 meters above the Talcott Basalt	
	49.5 meters

Trip C

The Holyoke Basalt in Southern Connecticut

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INTRODUCTION

Geological travelers through the Mesozoic Hartford Basin of central Connecticut are certain to be impressed by its great longitudinal ridges, or hanging hills. In the western part of the Hartford Basin, several ridges are caused by differential erosion around basaltic dikes and sills, while steep cliffs in the central and southeastern Basin are produced by the up-tilted edges of great basaltic lava flows. The basalts are very similar to (and may represent the same magmas as) flows and sills in the Newark Basin, not far to the west in New York and New Jersey. In both basins, three flows or groups of flows are distinct; in the Hartford Basin, they are known as the Talcott, Holyoke, and Hampden basalts, from oldest to youngest. If these lavas flowed across and among northeastern North American basins, our hanging hills must be only small remnants of former flood basalts.

This field guide briefly describes the Holyoke basalt of Connecticut, with a prime example of its very long exposure in Tilcon Corporation's trap rock quarry at North Branford (one the largest such quarries in the world). Although the quarry company understands the economic potential of the Holyoke, the geological importance of this basalt, and others like it, is not fully appreciated. We finish with a comment on the former extent of this great flood basalt.

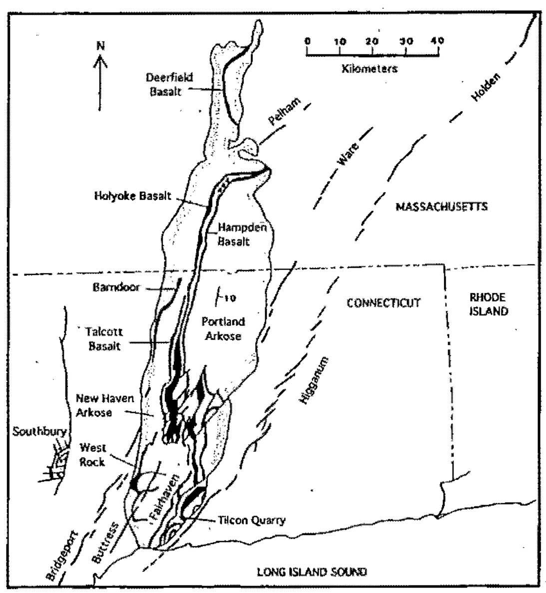
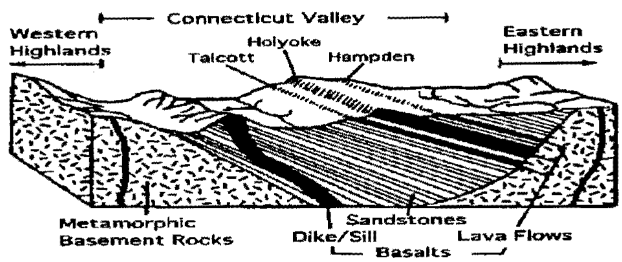


Figure 1. Sketch map of the Hartford, Deerfield and Southbury (Pomperaug) Mesozoic basins, Connecticut and Massachusetts. The Tilcon quarry of North Branford is cut into the arcuate outcrop of Holyoke Basalt in southeastern Connecticut. The extensive dikes probably mark former fissure sources for the basalts. Modified from Figure 1 of Philpotts and Martello, 1986.

STRATIGRAPHY AND AGE

The Mesozoic basins of Eastern North America (ENA) each contain portions of a sequence of sedimentary and igneous rocks that make up the "Newark Supergroup" (see other Hartford Basin field guides in this volume for references and details). Sandwiched between thick clastic formations (Figs. 1 and 2), three well-defined basalt units are prominent in the Hartford Basin. The oldest is the Talcott basalt (approx. 65 m thick), which poured onto New Haven arkose that is up to 2000 m thick in parts of the valley. The Talcott is known for its pillows and mineralized breccias, and at least four separate flows have been recognized. The Talcott is separated from the Holyoke basalt by 100 m or more of Shuttle Meadow siltstones and sandstones. The Holyoke is thick (100 to 150 m) and is characterized by prominent joints and columns within two or more massive flows. Still more red and gray siltstones and mudstones (the East Berlin formation, about 170 m thick) separate the Holyoke from the 60- to 100 m-thick Hampden basalt. The Hampden appears massive but also contains two or more flows. Up to 2000 m of clastic sediments (including fanglomerates) of the Portland formation overlie the Hampden. Except where folded and rotated by intrabasinal faults and fault blocks, the basalts and sediments dip 10 to 20 degrees towards the eastern fault margin (Figure 2).

Figure 2. Block diagram of the Hartford Basin, dimensions exaggerated. Note that all strata dip eastward, indicating considerable post-depositional offset. The sub-surface position of the basinal dike/sill is conjectural. Modified from Anonymous (no date).



The basalts are major stratigraphic markers within sediments that were formerly assumed to be completely Triassic in age ("Triassic basins" is still commonly used as a general name). New studies in the 1970's (Armstrong and Stump, 1970; Cornet and Traverse, 1975) led to the realization that both Late Triassic and Early Jurassic sediments are present in the Hartford Basin.

As part of Mesozoic correlation efforts and tectonic ("ocean opening") studies, many K-Ar and Ar-Ar analyses have been made of flows and dikes in this and other Mesozoic basins of Eastern North America (ENA). Unfortunately, the radiometric data show a great deal of scatter between 170 Ma and 230 Ma, although many of the fresher samples provide ages between 190 and 200 Ma. For example, Armstrong and Stump (1970) measured three whole-rock K-Ar ages for the Holyoke basalt at Tilcon's North Branford quarry, of 161 +/- 9 Ma, 197 +/- 12 Ma, and 201 +/- 12 Ma. Newer decay constants add 4 to 5 Ma to these older age calculations. Seidemann and others (1984) report a plagioclase K-Ar date of 189 +/- 6 Ma for Holyoke basalt from this area.

Careful measurement and analysis of stratigraphic sections by Paul Olsen and colleagues (Olsen and Fedosh, 1988) show that all of the lava flows are close together in age, probably spanning no more than 600,000 years during the Early Jurassic. They also argued that basalts in the Newark and some other ENA basins can be exactly correlated with the Hartford flows. This model received support from a paleomagnetic study (Prevot and Williams, 1989) that resolves some of the confusing scatter of paleomagnetic directions shown by the basalts. With more work using Ar isotopes (Michael Kunk, pers. comm. 1994), it is starting to look as if 196 Ma is approximately our "best age" for the Holyoke and other basin basalts, including dikes and sills.

It is also clear that, not far beneath the Holyoke, the Talcott basalt and its correlatives (the Orange Mountain basalt in the Newark Basin and the North Mountain basalt in the Fundy Basin) are very close to marking the beginning of the Jurassic Period; only a small portion of underlying New Haven arkose is post-Triassic. Is this coincidence, or could these grand basalts be related to some major event that also closed the Triassic Period?

Petrology

Hand samples of the Hartford Basin basalts can look pretty much alike. Fresh surfaces are usually fine grained and greenish gray, with small scattered plagioclase crystals. The columnar basalt from the Holyoke is so abundant and fine-grained that the Holyoke is commonly characterized as being nearly aphyric. At the North Branford quarry, non-columnar (interior) samples of the Holyoke range into medium-grained gabbro, with dark plagioclase and greenish-black pyroxene visible, as well as clots or phenocrysts of altered pyroxene or olivine. The Holyoke is also distinguished by immiscible sulfide globules (Philpotts and Martello, 1986). Hartford Basin basalts are notoriously altered, possibly by post-magmatic hydrothermal activities or simple burial metamorphism, to zeolite or epidote facies. The basalt is fairly hard, but it shatters into angular pieces and rapidly weathers to a rusty brown surface.

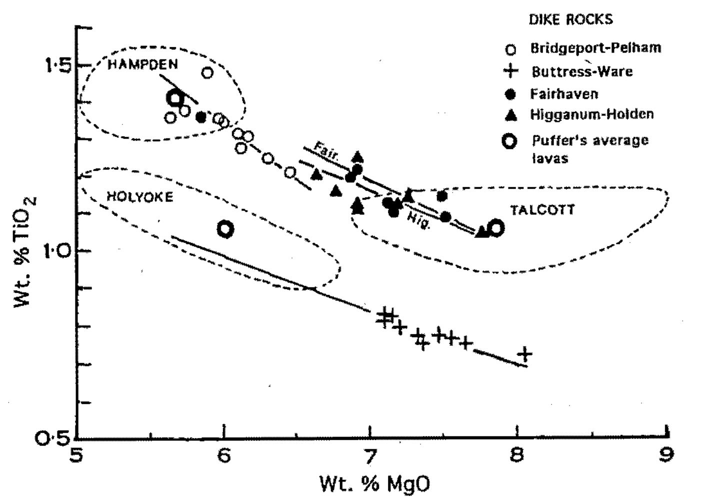


Figure 3. TiO_2 vs. MgO for Hartford Basin basalts and source dikes. The Holyoke chemistry is considerably fractionated from the proposed Buttress-Ware source magma. Modified from Figure 3 of Philpotts and Martello, 1986.

Like the other basalts, the Holyoke shows some variability in its mineralogy and chemistry. This is the result of both hydrothermal alteration (which adds volatiles and can move alkalies and some other elements around), and crystal fractionation. Most fractionation is probably from removal or concentration of olivine, pyroxene, and/or plagioclase phenocrysts, both within the lava itself and within earlier staging zones of the Holyoke magmas. Although all the basalts are classified as "high-titanium quartz tholeiites," there are "high iron" (low magnesium) and "low iron" (high magnesium) subgroups (Puffer and Philpotts, 1988) that result from Mg-rich mineral fractionation. Figure 3 illustrates the TiO_2-MgO relationships in the Hartford Basin basalts. The Fairhaven and Higganum dikes were conduits for Talcott-type magmas, which may be cousins to Bridgeport-Pelham and Hampden magmas. The Holyoke and Buttress-Ware magmas form their own range of compositions, probably related by fractionation of olivine and/or pyroxene (MgO-rich minerals) from a distinctive mantle melt. The more primitive Buttress magma is higher in Mg than the more evolved "high iron" Holyoke derivative.

FEATURES OF THE HOLYOKE AT THE TILCON QUARRY

Reactivation of border and intra-basinal faults have separated the southern Holyoke basalt and adjacent strata into a partial sub-graben (Sanders and others, 1963). A complex rim of normal fault movements produced a zone of inward downwarp for the strata in this section, forming the Gaillard graben or basin (Figure 4). As elsewhere, the curving ridge of basalt around the basin is caused by differential erosion of the softer sedimentary rocks above and below the Holyoke. The natural basin topography assisted engineers who dammed some fairly minor streams to create the large Lake Gaillard (some water is also brought to the lake by pipeline), which is an important water source for New Haven. This water is carried to the city by a large pipe that runs under the middle portion of the North Branford (or Totoket) quarry of the Tilcon Corporation. The lowest floor section of the quarry is close to the water level of Lake Gaillard, but the basalt is fairly water tight except in zones where it has been blasted.

Entrance into the quarry is made eastward off Rte. 22 just north of its intersection with Rte 80 (Figure 4). Access to this and other property of the Tilcon Corporation is strictly limited to groups that have a special arrangement such as ours; do not attempt to visit on your own. We will travel through much of this large quarry by bus, especially viewing its eastern wall that displays at least two major flows. Our stop will be near a large boulder field that is left in the quarry for processing, where samples show typical Holyoke features that we will examine. While outside the bus, DO NOT APPROACH THE QUARRY WALLS. Published descriptions of the quarry require the permission of Tilcon Corporation.

The North Branford quarry has been operating since the early part of the 20th century, cutting into the Holyoke basalt along the southern part of the ridge known as Totoket Mountain (Figure 4). Three levels are being developed, each about 60 feet (20 m) deep but all remaining completely

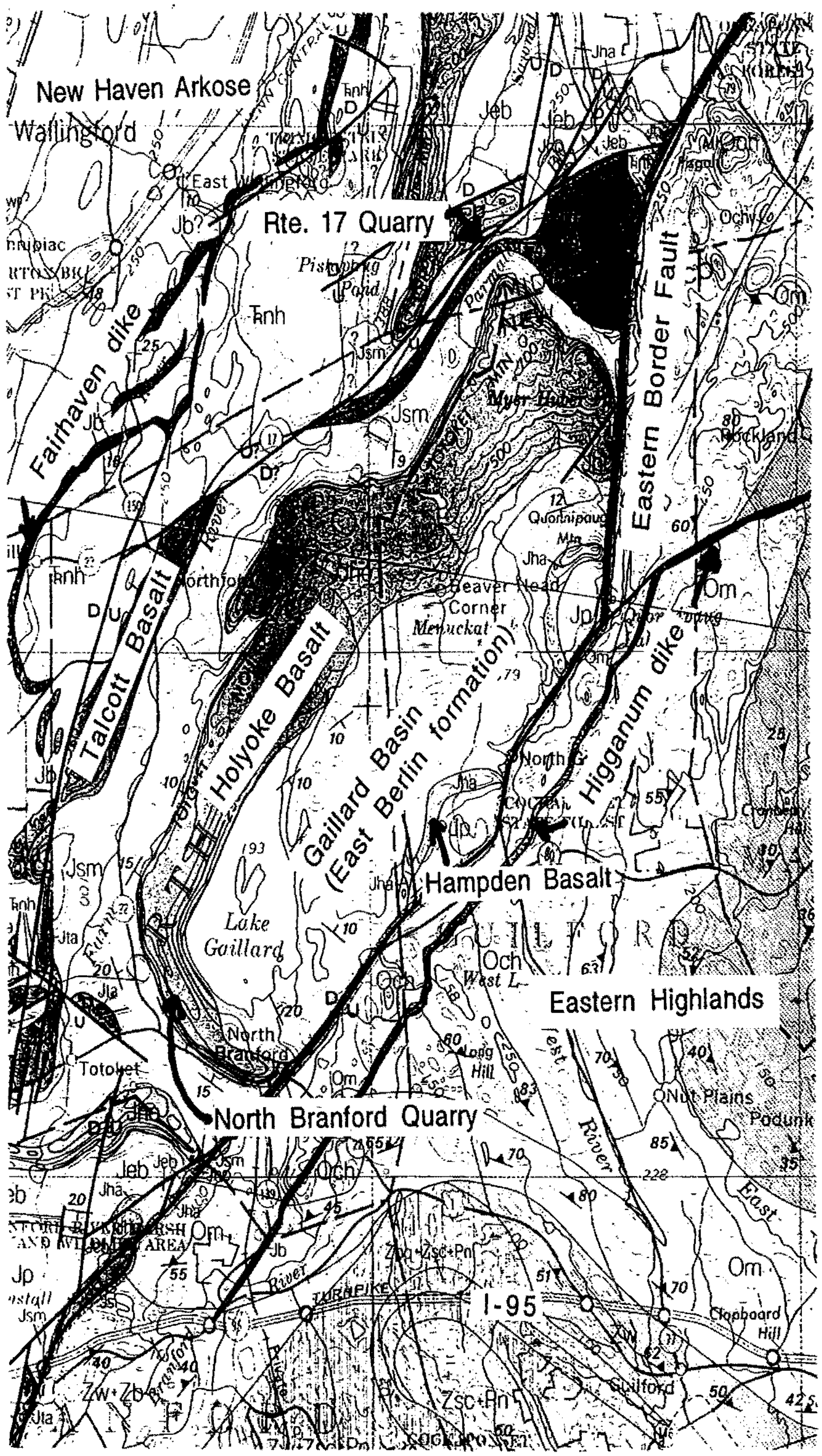


Figure 4. Geology of the area of the Gaillard Basin and the Holyoke Basalt of Totoket Mountain, North Branford. Slightly enlarged from the geological map of Connecticut (Rodgers, 1985), 1:250,000 scale.

within the Holyoke. The ridge to the northeast displays a continuous cut face nearly 3 miles (4 1/2 km) long, and present (1994) operations are extending this cut at the northwestern end of the quarry. This Tilcon Corporation quarry is well known for its good management, with a clean and tidy facility, settling ponds for run-off, a constant dust-abatement program, and a careful plan of production and uses for its trap rock. Large berms are maintained along the quarry to protect the Lake Gaillard watershed as well as other adjacent properties.

The lower contact of the Holyoke is passed on the entrance road immediately above the scale house. The reddish siltstones represent the Shuttle Meadow formation. Because of the 10 to 15 degree northeasterly dip of the basalt, the western face of the quarry is irregular and prone to mass wasting into the quarry. The eastern face holds its vertical cut better and the quarry is worked more deeply into that side. The northern end of the quarry is recently active and we may see the typical results of blasting to produce great piles of angular cobble to boulder size clasts. Our stop is near some large boulders that can be studied and sampled.

Even in its natural cliff faces on the western sides of hanging hills, the Holyoke shows a distinct break between two major flows. The contact is especially well marked in the eastern face of the North Branford quarry about 15 to 20 m below the top of the cut, where it shows columnar structure above and below the break. In places, the surface between the flows shows steep-sided convolutions or peaks and valleys of several meters relief (Figure 5). These amazing features are highlighted by radiating fans of columns that are perpendicular to the contact, both above it and below it. Both flows may have joined and cooled inward from this surface in some way that preserved its extreme relief. Columns gradually disappear into the coarser interior of the upper flow, but they remain large and continuous through much of the lower flow. A few minutes of discussion (or argument) about the origin(s) of these features are called for, if only to promote the progress of our science!

Along some sections, another flow appears to be preserved close to the top of the eastern cut. This unit occupies more gentle lows and highs, possibly with a zone of weathering (lighter colored materials) at its base. Basalt columns appear to be less developed in this unit than within lower flows.

Figure 5. Irregular surface between flows of the Holyoke in the eastern wall of the North Branford quarry. The quarry wall is approximately 40 m high. Note the short columns (2-4 m) of the contact zone of the upper flow, in contrast with the long columns (6-10 m) of the lower flow. Photo by permission of the Tilcon Corporation, 1994.

QUARRY ON ROUTE 17, DURHAM

If time allows, we may stop at a recently worked roadside cut on the west side of Rte. 17 south of Parmelee Brook, and 2.7 miles south of the intersection with Rte. 77 (Figure 4). The Holyoke at this cut lies just above its basal contact with the Shuttle Meadow, which shows pieces in the roadbed. Large sections of brittle, black glass and glassy basalt occur in the sides of the cut. Note the sooty black coatings on rock surfaces. The lower wall also contains breccias of mixed sandstones and basalt. Although this quarry is not posted, we assume that it is privately owned; please do not climb the hill or disturb the site any more than necessary.

THE HOLYOKE FLOOD, KNOWN AND INFERRED

In the model of Philpotts and Martello (1986), a local source for Holyoke magmas is the Buttress dike, which crosscuts the New Haven arkose and the West Rock sill in the southwestern basin (Figure 1). The Buttress dike cannot appear in any of the post-Holyoke basin strata, but it is mapped to the northeast in the eastern basement rocks into Massachusetts (the Ware dike, Figure 1). The Holyoke-Buttress correlation is based upon chemical and petrographic similarities as well as geographic relationships. The simultaneous (if not continuous) eruption of Holyoke type magmas across several basins must mean that several fissures, or dikes, were active to supply the huge volume of basalt that is still represented within the Hartford, Newark, and other basins. As proven by the same basalts in the Southbury (Pomperaug) basin, the Hartford Basin basalts also flowed across western Connecticut, although there the topography must have been less basin-like because flows are thinner. If we assume that the Hartford and Newark basins were connected according to the "Broad Terrane" model, it follows that the Holyoke-Preakness magmas covered at least both basins as well as the section between them; 90 km by 400 km (36000 square km) is a conservative estimate (perhaps less than the area of Sanders' 1963 model). A thickness of 0.1km indicates 3600 km^3 of basalt.

In addition, the Holyoke and equivalent Newark Basin Preakness basalt are clearly truncated by major border faults. In Connecticut, normal movements on the eastern border fault created the easterly dip of the lavas and other basin strata, with perhaps 8 km (or more) of total offset likely (Figure 2). Boulders of basalt are components of Portland arkose fanglomerates that washed down from and across the eastern fault, proving that basalts once flowed across an area now within the Eastern Highlands. Finally, the Buttress-Ware and other dikes essentially mark locations beneath former fissure eruptions that contributed to the present basalt remnants. This being true, the Holyoke has a high probability of formerly extending across portions of eastern Massachusetts as well as Connecticut (Figure 1). It is also likely that the Preakness equivalent extended westward into Pennsylvania as well as eastward across regions now under the Coastal Plain (including Long Island). Such reasoning can double our estimate of the former size of the Holyoke-Preakness basalt, not to mention its siblings the Talcott and Hampden.

ACKNOWLEDGMENTS

Frank T. Lane and the Tilcon Corporation provided guided access to their property and generous assistance for this summary and field trip to the North Branford quarry. Brian Skinner presented an introduction to the geology of this site during an earlier visit. We also thank Nancy McHone for her editorial assistance.

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Trip D

Paleoenvironmental Traverse Across the Early Mesozoic Hartford Rift Basin, Connecticut

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INTRODUCTION

This two-day field excursion focuses on reconstructing the sedimentary and tectonic paleoenvironments of the early Mesozoic volcanic and nonmarine sedimentary rocks exposed in the Hartford Basin, one of the earliest studied rift basins in North America. Following a general east-west traverse across central Connecticut, the ten field stops highlight both the vertical succession and the lateral diversity of facies and their interpreted paleoenvironments including: proximal alluvial fan and mid-fan facies, distal fan and lakeshore facies, alternating shallow-to-deep lacustrine facies, basalt flows, playa and shoreline facies, alluvial plain facies, and the unconformable and faulted western margin of the basin.

The Lower Mesozoic Newark Supergroup fills a series of block-faulted rift basins along the eastern side of the Appalachian orogen from the Maritimes to the Carolinas (Frodich and Olsen, 1984), formed during the incipient rifting of North America and Africa in the Late Triassic and Early Jurassic. One of these, the Hartford Basin, is an elongate north-trending asymmetric graben, approximately 140 km long and 30 km wide in central Connecticut (Figure 1). The Hartford Basin contains more than four km thickness of Upper Triassic and Lower Jurassic terrestrial sedimentary strata (Table 1), and several intercalated tholeiitic basalts (Cornet, 1977; Olsen el al., 1989). A transverse basement horst partially separates the Hartford Basin from the Deerfield Basin to the north (Wise, 1992); the basin terminates at some unknown distance to the south beneath Long Island Sound. West-dipping listric and en echelon faults define the eastern margin of the basin, toward which the strata generally dip and young (Wise and Robinson, 1982). Therefore, our paleoenvironmental traverse will descend from the youngest through to the oldest strata as we move west.

Structural Setting

The Hartford Basin displays the asymmetric half-graben morphology characteristic of most modern and ancient continental rifts (Leeder and Gawthorpe, 1987; Lambiase, 1990), though at depth, the structural geometry may be considerably more complex. Exposed strata typically dip 10-20° toward the eastern boundary of the basin, which is delineated by alternating segments of N-S trending and NE-SW trending normal faults (Figure 2). The eastern border faults are characterized by relatively steep westerly dips near the surface (45-65°), which shallow to about 20° at depths of more than 2 km (Wise, 1981; Zen, 1983). Gravity and seismic studies (Eaton and Rosenfeld, 1960; Chang, 1968; Wenk, 1984) reveal that the deepest part of the basin (~5 km) is near its axis, considerably west of the eastern basin margin.

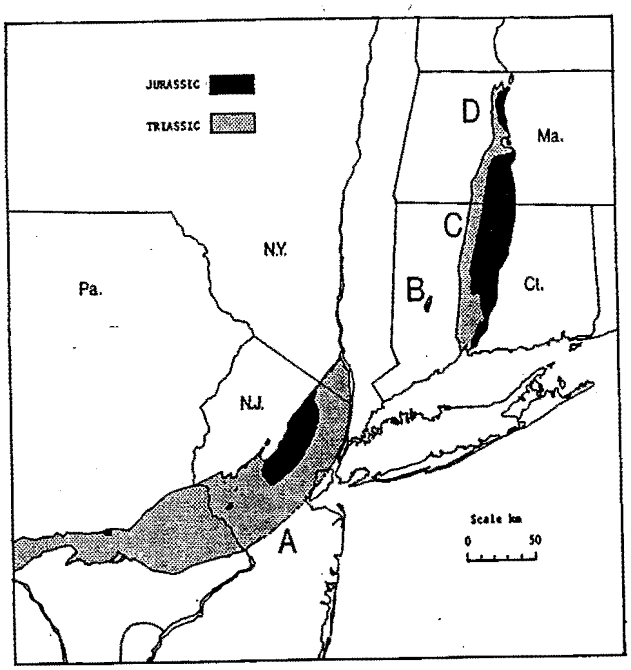

Figure 1. Index map of the (A) Newark, (B) Pomperaug, (C) Hartford, and (D) Deerfield basins (Olsen, 1980).

This presumably indicates the presence of step faults or rider blocks along the eastern margin of the basin. Eastward-thickening rock units, repeated coarsening-up vertical sequences, alluvial fan geometries and paleocurrent directions (Eaton and Rosenfeld, 1960; LeTourneau, 1985) confirm that the eastern border fault system controlled depositional processes in the eastern part of the basin throughout the Early Jurassic. However, its role during initial basin development and early sedimentation in the Triassic is unclear.

For much of its length the western margin of the Hartford Basin is marked by N-striking, E-dipping normal faults, but in some locations early Mesozoic strata unconformably overstep Paleozoic crystalline basement. Westward thickening of the New Haven Arkose (Wenk, 1989) indicates the presence of depocenters along the western margin, and suggests that the western fault system may have played an important role in the early phases of basin subsidence (de Boer and Clifton, 1988). In the earliest Jurassic, rising magma apparently utilized some western faults, emplacing large intrusive bodies adjacent to the western margin of the basin.

In the central part of the basin the structure may be very complicated. The presence of a major axial fault zone at depth is suggested by a swarm of normal faults that intersect and offset the ridges formed by the lava flows. Most of these faults trend northeasterly and dip steeply west. Cumulative displacement is estimated to be about 1.5 km down to the west. An accumulating body of evidence indicates that many of these oblique, intrabasinal, axial faults were syndepositionally active (Philpotts and Martello, 1986; Wenk, 1990a, 1990b). The presence of half-meter diameter crystalline clasts in the New Haven Arkose near the center of the basin (Davis, 1898; Rice and Gregory, 1906) argues for early activity of intrabasinal horsts. Lambiase (1990) has observed that in many continental rift basins, block faulting occurs prior to significant synrift sedimentation, and initially the basin floors are composed of a mosaic of tilted fault blocks at slightly lower elevations than the rift margins.

Seismic reflection, magnetic, and gravity data from the New York Bight Basin (Hutchinson el al., 1986), which may be an offshore extension of the Hartford Basin, indicate that its southern portion is underlain by chaotic block-faulted basement. Intrabasinal reflectors are tilted and discontinuous, suggesting that sediment was deposited between blocks contemporaneously with faulting. Early rifting in the Hartford Basin may have been analogous, resulting in a physiography similar to the Basin and Range region of the western U.S., where horsts and grabens, decollements, and complex listric faults are distributed over broad areas (Hutchinson el al., 1986).

Evolution of Paleogeographic Concepts

The Hartford Basin has been the subject of geologic investigation for more than two centuries. Noteworthy American geologists who conducted research in the basin include: B. Silliman, E. Hitchcock, J.G. Percival, J.D. Dana, J.S. Newberry, O.C. Marsh, I.C. Russell, W.M. Davis, B.K. Emerson, J. Barrell, C.R. Longwell, and P.D. Krynine. Only a few of these workers formulated depositional models or attempted paleogeographic interpretations for the basin, and a review of these concepts over the past century is informative.

William Morris Davis conducted structural and physiographic investigations in central Connecticut for nearly 20 years. In his 1898 monograph, *The Triassic Formation of Connecticut*, Davis hypothesized sedimentation taking place in an elongate, downwarped trough, bowl-shaped in cross section, with detritus being supplied from both margins and accumulating in near-horizontal layers. Recognizing the shallow-water nature of much of the strata, he realized that depression of the trough and deposition must have taken place simultaneously. Davis believed that basin margin conglomerates were laid down by streams, and suggested (1898, p. 35): "Shallow lakes may have now and then overflowed a *middle* [italics ours] strip or a greater part of the trough, and there finer sediments would gather. The lake floors were sometimes slowly shoaling, ...

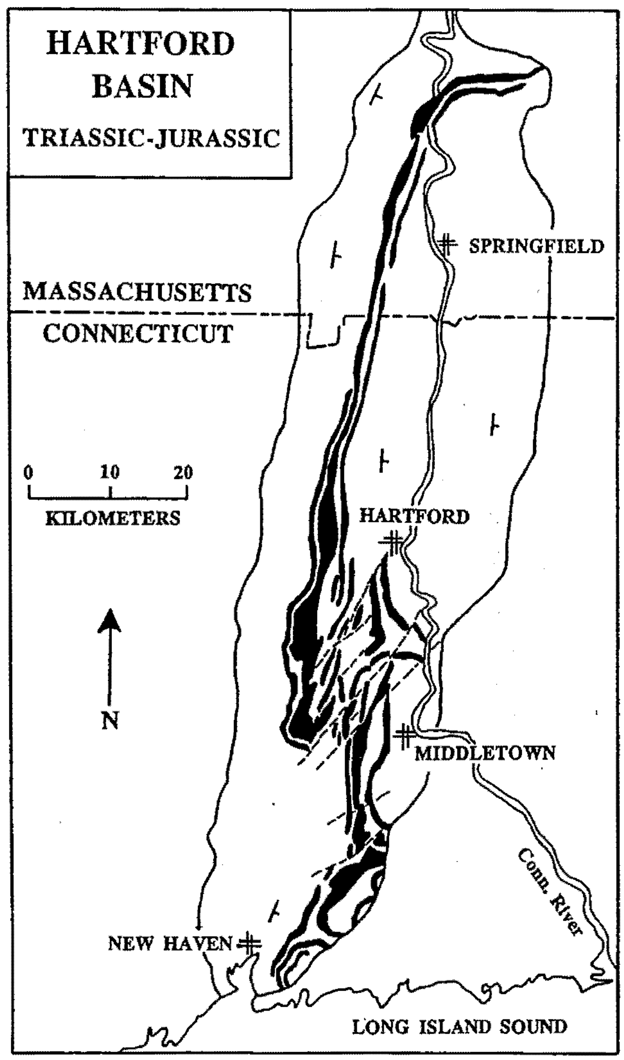

Figure 2. Outline map of the Hartford Basin, showing basalt flows in solid black.

sometimes sinking to greater depth." Later in his report, he postulated a master axial river draining the basin to the south (p. 155-156). Unfortunately, Davis did not recognize the subtle evidence for synsedimentary activity on the faults which he so carefully and accurately mapped, and he believed instead that most faults within and bordering the basin were postdepositional (Lorenz, 1988).

Joseph Barrell was the first to theorize that the faults which bound the eastern edge of the Hartford Basin were intermittently active during sedimentation, and exerted a primary influence on basin geometry (Barrell, 1915; Eaton and Rosenfeld, 1960). He hypothesized that deposition occurred in a flat-bottomed, wedge shaped trough, subsiding on the eastern edge, with most of the sediments originating from highlands adjacent to the eastern boundary faults. During sedimentation, according to Barrell, the surface of the basin remained almost horizontal, or perhaps slightly inclined to the west, but the bottom of the trough assumed an increasing eastward inclination due to the weight of the detritus entering the basin from the east. Barrell makes little mention of a western hinged margin of the basin (or west-derived sediment), except to suggest that the basin probably extended well beyond its present western limits perhaps as far as New Jersey (1915, p. 29). In this regard it is clear that he was profoundly influenced by the now discredited "broad-terrane" hypothesis of Russell (1879), which advocated that the Hartford Basin and the similar Newark Basin of New Jersey (Figure 1) are erosional remnants of a much larger full-graben, the basins being later uplifted, block-faulted and separated by a broad postdepositional longitudinal arch (Olsen el al., 1989). Nevertheless, Barrell originated the half-graben model to explain the structural development of the Hartford Basin, a model which still has many adherents.

Chester R. Longwell (1922) examined the apron of coarse alluvial fan deposits along the eastern margin, confirming Barrell's ideas of synsedimentary faulting along that boundary. In a later report with E.S. Dana (1932, p. 56-57), Longwell comments: "... the red sandstones and shales, with included sheets of trap, probably covered much of the area between central Connecticut and northern New Jersey." Further studies on the relationship between faulting and sedimentation in the basin led Longwell to proclaim that the entire body of sediments in southern Connecticut was furnished by the highlands to the east of the basin: "Thus, there is indicated a wide piedmont deposit, tapering in thickness westward ... all derived from a block that was lifted progressively by faulting as sedimentation proceeded." (Longwell, 1937, p. 437-438).

Paul D. Krynine carried out detailed mineralogical and petrographic investigations of sedimentary rocks from the Hartford Basin in the 1930s, but his comprehensive report remained unpublished until 1950. Although Krynine's petrographic work was of value in the areas of stratigraphy and structural geology, his quantitative and statistical data added little to the existing understanding of the origin or conditions of deposition of the rocks of the Hartford Basin (Lorenz, 1988). In the area of paleogeography, his findings were almost entirely in accord with those of Barrell and Longwell: 1) the eastern border faults were syndepositionally active, 2) the sediments show an exclusive eastern provenance, 3) the basin floor was gently inclined to the west during deposition, and 4) tilting and block-faulting of the basin occurred postdepositionally from the rise of a geanticline between the Hartford and Newark basins. Krynine (1950) agreed with most of the tenets of the broad-terrane hypothesis, but thought that the deposits in the Hartford Basin had originally wedged out in western Connecticut, some 3.5 km west of the Pomperaug Valley (Figure 1), a small basin of early Mesozoic strata often correlated with the Hartford Basin.

In the most thorough study of the western margin of the basin to date, Girard Wheeler (1937) found evidence for faulting along the western boundary, and concluded that the basin was in part a graben. However, he found no evidence for western sources of detritus in the basin, and did not believe the western faults were syndepositionally active. A staunch advocate of broad-terrane paleogeography, Wheeler argued for the interconnection of the Hartford and Newark basins.

Recent Paleogeographic Studies

Interest in the Hartford Basin resumed in the 1960s with the studies of John E. Sanders and George deVries Klein. Sanders (1968), a supporter of Barrellian ideas, interpreted some of the fine-grained, well-bedded strata of the Jurassic formations as lacustrine deposits with dominant east- and southeast-trending paleocurrent directions, but attributed the paleocurrent trend to irregularities of lake bottom topography, rather than to the influence of regional paleoslope. Sanders also recognized the interbedding of "deep lake" strata and conglomeratic units along the eastern margin, but maintained that in general, lacustrine deposits occupied positions near the center of the basin, passing laterally into coarser-grained strata at the basin margin (Sanders, 1968). Klein (1968) identified locally-derived clasts in the New Haven Arkose near New Haven, thus confirming a western provenance for at least some of the earliest basin deposits. Klein also recorded easterly directed cross-stratification in sandstones of the East Berlin Formation, and noted the varied lithologies and extreme cyclicity of that formation. In 1969, Klein took an anti-broad-terrane stance, and interpreted the provenance and paleocurrents in the Newark Basin to indicate that it had been filled from all directions of the compass, and thus there could be no persistent connection with the Hartford Basin during deposition. Klein's data also indicated that the Hartford Basin was filled from both sides, and he concluded: "... the dominant flow direction of depositional streams in Connecticut during [Early Mesozoic] sedimentation was to the west; data from the basal New Haven Arkose and the East Berlin Formation, however, also indicate deposition by some east-flowing streams" (Klein, 1969, p. 1827).

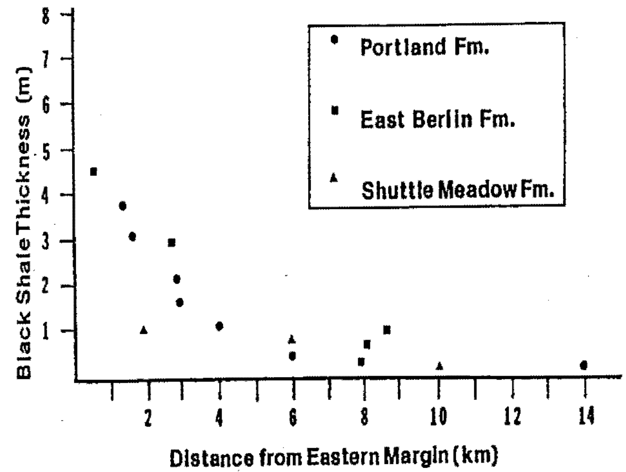

Figure 3. Relation between thickness of black shale beds and distance from the eastern faulted margin of the Hartford Basin.

A series of sedimentological studies by John F. Hubert and his students published over the last two decades are valuable contributions to understanding the geology of the Hartford Basin (Hubert el al., 1976, 1978, 1982). They described the diverse paleoenvironments found in the basin, assembled detailed paleocurrent data for the various sedimentary formations, produced the first paleogeographic maps of the region, and outlined the sedimentary history and distribution of alluvial, fluvial, lacustrine and playa deposits. In their earlier paleogeographic reconstructions of the basin, however, they did not consider the evidence for western provenance and easterly paleoslopes presented by Sanders (1968) and Klein (1969), and adopted a modified broad-terrane viewpoint essentially like that of Krynine (1950). Recently, Hubert el al. (1992) integrated their earlier studies with provenance and diagenetic data to analyze the structural and hydrocarbon history of the basin.

Recent studies (McDonald and LeTourneau, 1988, 1989, 1990; Smoot, 1991) have verified the presence of the easterly paleoslopes in the basin hypothesized by Klein (1969). In detailed investigations of alluvial fan deposits in the Middletown area, LeTourneau (1985) documented the interfingering of coarse conglomerates with perennial lacustrine strata along the eastern fault margin, and demonstrated that in most cases the lake deposits are thickest adjacent to that margin (Figure 3). Earlier investigators had envisioned lakes as occupying middle regions of the basin and shallowing toward both margins. The restriction of productive fossil fish localities (McDonald, 1975) to areas close to the eastern margin, and patterns of fossil preservation (McDonald and

LeTourneau, 1989) further confirm that perennial lakes were deepest and persisted longest adjacent to the eastern boundary of the basin. Syndepositional eastward tilting of the basin floor in the Jurassic also limited the western progradation of coarse detritus, producing a series of localized, discrete alluvial fans alongside the eastern margin (LeTourneau and McDonald, 1985). Recognition of western shoreline and deltaic deposits in the central portion of the basin (McDonald and LeTourneau, 1988) indicates that the lakes gradually shoaled to the west, where fine-grained detrital and allochemical sedimentation was predominant. In mid-basin areas the persistent trend of paleocurrents in the Shuttle Meadow, East Berlin and Portland formations is to the NE, E and SE (Hubert el al., 1978; McDonald and LeTourneau, 1988). The latest interpretations of Hubert el al. (1992), based on paleocurrent, provenance and petrographic studies, also support the hypothesis that the Hartford Basin received detritus from both eastern and western margins during the Late Triassic and Early Jurassic.

Stratigraphy and the Basin Model

Interpretations of the paleogeography and paleoclimate of the Hartford Basin have been presented by Krynine (1950), Hubert el al. (1978, 1992), and McDonald and LeTourneau (1988, 1989, 1990). The basal New Haven Arkose is mainly composed of coarse redbeds deposited on alluvial fans and braid plains by streams which flowed from Paleozoic crystalline highlands bordering both basin margins. Caliche paleosols suggest that the Late Triassic climate was tropical and semi-arid, with perhaps 100-500 mm of seasonal rain and a long dry season (Hubert, 1978).

An increase in lithospheric extension in the earliest Jurassic (Schlische and Olsen, 1990) led to greater subsidence along the eastern margin of the basin, periodically tapping underlying magma sources, and producing an internally-drained half graben with regional paleoslopes to the east. Following extrusion of the Talcott Basalt, mudstones and sandstones of the Shuttle Meadow Formation were deposited on floodplains and in ephemeral and perennial lakes. The western uplands were the primary sediment source area for the gently inclined hanging wall dip slope of the basin. Sheetfloods and small streams distributed sediment eastward across broad alluvial fans onto a wide braid plain punctuated by playas (Figure 4). During wet intervals stratified perennial lakes occupied much of the basin floor, perhaps attaining depths of 100 m adjacent to the eastern side of the basin. West flowing streams only supplied detritus locally along the eastern margin. The Shuttle Meadow is highly fossiliferous; vertebrate footprints are common in fine-grained redbeds, articulated fossil fishes and plants are present in laminated black shales, and plant fragments are locally found in gray mudstones (McDonald, 1992).

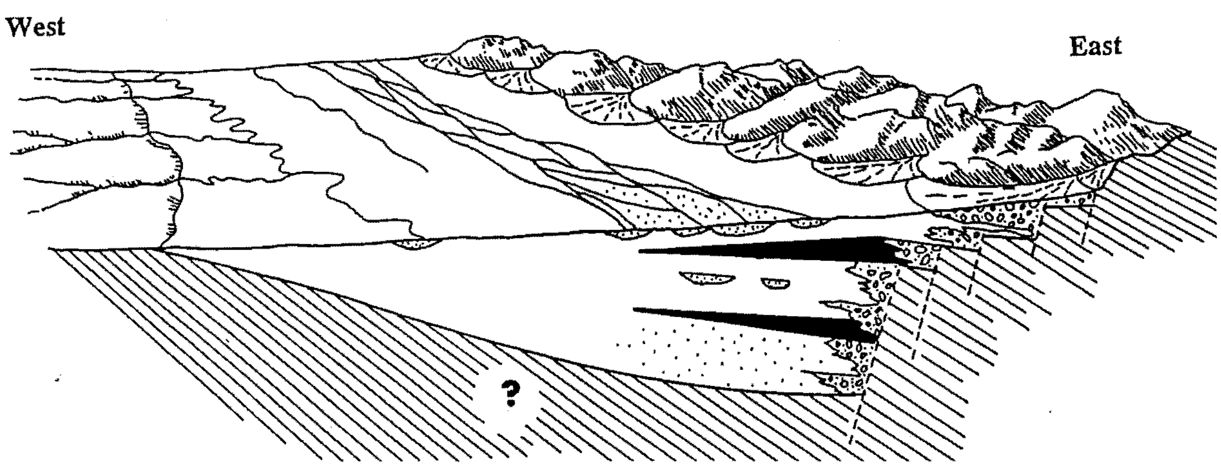

Figure 4. Paleogeographic model of the Hartford Basin during an arid interval in the Early Jurassic (modified from Steel, 1977). Solid black indicates perennial lake deposits; igneous rocks omitted.

After extrusion of the Holyoke Basalt, renewed alluvial fan, floodplain, playa and lacustrine environments are represented by the strata of the East Berlin Formation. Symmetrical cycles of gray mudstone-black shale-gray mudstone, which record the periodic expansion and contraction of large perennial lakes, are most conspicuous in East Berlin outcrops (Hubert el al., 1978; Olsen el al., 1989). Detailed stratigraphic analysis demonstrates that these cycles were climate controlled, perhaps in response to Milankovitch-type fluctuations in the earth's orbit (Olsen, 1986). The region was under the influence of a subtropical, probably monsoonal climate characterized by alternating seasons of high precipitation and aridity. The East Berlin is also rich in fossils.

A third episode of extension and renewed subsidence produced the flows of the Hampden Basalt and aided the development of extensive lacustrine conditions during the deposition of the lower portion of the Portland Formation (Figure 5). Some of the most productive fossil fish localities occur in perennial lake-bed strata of the basal Portland. The Portland Formation has locally produced numerous dinosaur tracks. The upper part of the Portland is an alluvial fan and alluvial plain facies consisting of red mudstone, sandstone, and conglomerate (LeTourneau and McDonald, 1985).

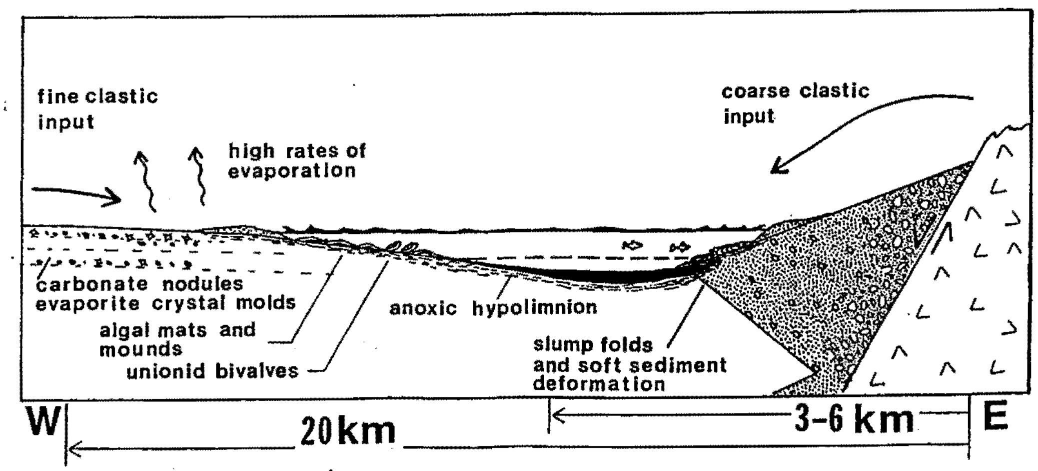

Figure 5. Paleogeographic model of stratified lacustrine conditions in the Hartford Basin in central Connecticut during a wet interval in the Early Jurassic.

HARTFORD BASIN FACIES

The deposits in the Hartford Basin are characterized by lateral and vertical variability, intertonguing and interfingering relationships, and a general lack of strata or fossils that can be confidently correlated across broad areas of the basin. Further complicating the stratigraphic relationships is the block faulting and pervasive eastward tilting of the strata. Strata become progressively younger from west to east across the basin, and also tend to become coarser grained toward the east. The stratigraphy of the basin is largely based on the position of strata relative to the basalt flows, although in some areas of structural complexity even that relationship is not entirely clear.

Early workers in the basin advanced only gross generalizations about the areal distribution of rock types. The lateral and vertical variability of the rocks prevented the development of a comprehensive paleoenvironmental model for the basin. Later workers (Hubert el al., 1978, 1982; LeTourneau and McDonald, 1985) recognized the repetitive association of certain rock types as facies assemblages, and this led to paleoenvironmental reconstructions and the development of a paleogeographic model for the basin. The definition of facies assemblages provided a sense of order to the distribution of rock types that early workers sought in individual units or beds.

The facies analysis presented herein is largely based on detailed studies in the Portland Formation (LeTourneau, 1985). However, work of others in the New Haven, Shuttle Meadow, and East Berlin Formations corroborates the Portland facies models (Hubert el al., 1976, 1978, 1982; Demicco and Gierlowski Kordesch, 1986; Olsen el al., 1989). Three distinct facies in the basin may be subdivided into nine subfacies (Table 2), each indicative of a discrete depositional environment based on grain size, color, diagnostic fossils, sedimentary structures, and associations with other subfacies.

Subfacies and Depositional Environments

Conglomerate-Sandstone Facies. The texture, fabric, sedimentary structures, clast-size distribution, bedding geometry, and sediment dispersal patterns of Subfacies 1 to 4 are characteristic of alluvial fan deposits. These are the coarsest deposits in the basin, and they occur in discrete lobes adjacent to the eastern border faults. These lithosomes are wedge and prism shaped bodies that thin and fine radially away from the coarsest and central parts of the lobes.

SUBFACIES 1: consists of poorly sorted boulder and cobble beds that compare favorably with descriptions of modern and ancient debris flow deposits. Diagnostic features include: random or chaotic orientation of major clasts in mud-rich matrix, concentration of larger clasts near the upper and outer contacts of the deposits, hummocky and irregular upper contacts, planar and distinct lower contacts, and indistinct or poor internal organization. Subfacies 1 lithosomes are interbedded within the deposits of Subfacies 2.

SUBFACIES 2: consists of thinly bedded and poorly stratified conglomerate and pebbly sandstone in normal-graded and laterally discontinuous beds. The depositional units fine upwards and are commonly capped with a silt drape, often with desiccation features. Cross-bedding is not common, although poorly developed inclined stratification or small scale trough cross-stratification is locally present. These beds are interpreted as ephemeral braided stream deposits on alluvial fans.

SUBFACIES 3: consists of trough and planar cross-stratified fine conglomerate and pebbly sandstone. Cross-bed set thickness is about 30 cm, but may be as much as 1.5 m. Cobbles and pebbles are dispersed in a sandy matrix or exist as small lenses within the beds; imbricate pebbles lie along the lower contacts and foresets. Upper and lower contacts are discontinuous and the beds are lenticular. These conglomerates and sandstones are interpreted as deposits of braided streams on alluvial fans. This interpretation is based on sedimentary features indicative of persistent stream flow: well developed large and small scale cross-stratification, moderate sorting, and absence of silt drapes and desiccation features. These features are common in modern deposits of seasonal streams with a substantial volume of flow.

SUBFACIES 4: consists of thin beds of planar and ripple cross-stratified, poorly sorted, silty fine sandstone with very thin granule interbeds. The sandstone typically contains carbonate cement and pore fillings. Subfacies 4 is interpreted as very shallow stream or sheetflow deposits at the distal portions of alluvial fans. It includes the finest-grained alluvial fan deposits, transitional between the lower fan and basin floor. These beds often form a distinctive horizon at the base of coarsening-upwards alluvial fan cycles.

TABLE 2 - ENVIRONMENTAL INTERPRETATION OF NEWARK SUPERGROUP LITHOFACIES, HARTFORD BASIN

FACIES	SUBFACIES	DESCRIPTION	INTERPRETATION		DEPOSITIONAL SETTING
Conglomerate-sandstone	1	Matrix-supported, poorly sorted cobble and boulder conglomerate	Debris flow	ALLUVIAL FAN	Proximal fan
	2	Clast-supported, poorly stratified pebble and cobble conglomerate	Ephemeral braided stream		Mid-fan
	3	Cross-stratified pebble conglomerate and sandstone	Perennial braided stream		Mid- to distal-fan
	4	Planar-stratified and ripple-cross-stratified silty fine sandstone	Sheetflood and very shallow stream		Distal fan
Sandstone-siltstone	5	Trough cross-stratified medium-to-coarse sandstone with interbedded siltstone	Ephemeral braided streams and desiccated floodplains	FLOODPLAIN	Basin floor
	6	Thin-bedded siltstone and medium-coarse sandstone lenses	Perennial meandering river		Basin floor
	7	Dark siltstone with interbedded well sorted ripple-cross-laminated sandstone	Lacustrine shoreline and sublittoral zone		Lake margin
Siltstone-shale	8	Dark siltstone and laminated black shale with fish remains	Lacustrine profundal zone		Perennial lake floor
	9	Massive mudstone with desiccation cracks and evaporite minerals	Playa		Ephemeral lake bed

Sandstone-Siltstone Facies. The sandstone and siltstone (and minor shale) of Subfacies 5 and 6 are typical of alluvial plain deposits and dominate the central portion of the Hartford Basin. Subfacies 5 and 6 are closely related, representing end members of varied sandstone/siltstone ratios. A large degree of variability exists within the fluvial deposits of Subfacies 5 and 6, depending on river sinuosity, channel deposits, depth and duration of flow, and floodplain character.

SUBFACIES 5: consists of fining-up beds of poorly sorted pebbly sandstone with subordinate pebble conglomerate, enclosed within surrounding siltstone. The sandstone beds typically are trough cross-stratified, often with irregular bases scoured into underlying siltstone. The siltstone commonly contains desiccation cracks and burrows; caliche horizons and carbonate intraclasts are locally present. In some areas angular clasts of siltstone from channel bank collapse and mud peloid ripples are abundant in the sandstone. Green-grey mottling from localized reduction of iron oxides is common. This subfacies is interpreted as deposits of ephemeral braided streams and desiccated floodplains.

SUBFACIES 6: consists of rhythmically stacked, fining-up sequences of sandstone, siltstone, and mudstone. Sandstone beds are often cross-bedded at their base, grading upwards to ripple cross-lamination, and with a bioturbated siltstone-mudstone cap. Internally, the sandstones often have large foresets with coarse grained toes, probably formed as lateral accretion surfaces on point bars. Soft sediment deformation features are common. Subfacies 6 is interpreted as deposits of perennial rivers meandering across broad floodplains on the basin floor.

Siltstone-Shale Facies. The sedimentary structures and fossils in the gray siltstone, minor gray sandstone, black shale, and massive mudstone of Subfacies 7, 8, and 9 are comparable to both modern and ancient lacustrine or playa deposits. These facies are most prevalent in the eastern half of the Hartford Basin, and also generally thicken towards the east.

SUBFACIES 7: consists of gray, ripple-laminated and wavy-bedded siltstone with intercalations of claystone and very fine sandstone, and gray sandstone beds with a wide variety of wave-generated primary structures, including oscillatory ripples and tabular, low-angle accretionary lenses. Both siltstone and sandstone are very well sorted and characteristically contain abundant plant remains. The sandstones become coarser grained and conglomeratic near the eastern margin of the basin where they interfinger with alluvial fan facies. Vertebrate tracks are locally abundant, and mollusk shell casts, carbonate peloids, and tufa crusts are occasionally present.

Subfacies 7 is interpreted as deposits of offshore-onshore lake margin environments. The gray, wavy bedded siltstone was deposited at or below wave base in a lake. The intercalations of laminated clay and fine sand is indicative of the variable energy conditions that existed between the nearshore and deeper water environments. The gray sandstone was deposited in the sublittoral and littoral zones. Large-scale tabular cross stratification in the coarse sandstone and minor conglomerate along the eastern margin may have been formed in beach ridges built by wave reworking of alluvial fan sediment.

SUBFACIES 8: consists of thinly bedded, laminated-microlaminated black shale with abundant and well preserved fossil fishes. The black shales were deposited under reducing conditions in periodically or permanently stratified lakes. The preservation of fine laminations indicates that burrowing organisms were not present. Anoxic bottom conditions which excluded a benthic fauna may have resulted from either thermal or chemical stratification of the lake. In the eastern portions of the basin within the Shuttle Meadow, East Berlin and Portland Formations the Siltstone-Shale Facies is interbedded with alluvial fan conglomerates and sandstones. The stratigraphic succession in these areas indicates that the lake margins expanded and contracted across the slopes of the alluvial fans. In central areas of the basin these subfacies are interbedded with fluvial sandstone and siltstone. At several locations within the Portland Formation deltaic sandstones are interbedded with lacustrine black shale.

SUBFACIES 9: consists of massive mudstone with extensive desiccation cracks, often from multiple generations of cracking, infilled with sand and silt. Euhedral-subhedral gypsum and halite molds or vugs are locally abundant. Carbonate deposits include thin limestone beds and locally abundant carbonate nodules and concretions. Thin, ripple cross-laminated sandstone lenses are widespread, and often contain sand-sized mud peloids defining ripple slip faces. This subfacies is interpreted as deposits of a dry lake bed or playa, with alternating inundation and desiccation, episodic aeolian influx of fine detritus, and evaporative reflux of interstitial brines.

Climate Hypothesis

Past workers have variously characterized the depositional environments of Hartford Basin deposits as estuarine, marine, fluvial, lacustrine, savannah, or playa, formed under climatic conditions ranging from arid to humid. Usually, one type of climatic regime was used to characterize all of the strata within the basin. The work of Hubert el al. (1978, 1982, 1992), LeTourneau and

McDonald (1985), Demicco and Gierlowski Kordesch (1986), and Olsen el al. (1989) has indicated that one type of climatic regime can not characterize all deposits in the basin. Mutually exclusive facies assemblages record the influence of contrasting climate regimes within specific stratigraphic intervals. For example, in the lower Portland Formation (LeTourneau, 1985) vertical alternations between distinct assemblages of subfacies probably represent repetitive wet and dry depositional cycles (Figure 6). Similar climatically controlled depositional cycles have been described from the Triassic Lockatong Formation in the Newark Basin (Olsen, 1986) and from the East Berlin Formation in the Hartford Basin (Hubert el al., 1978; Demicco and Gierlowski-Kordesch, 1986).

Dry cycle assemblages in the Hartford Basin are composed of Subfacies 1-2-4-5-9 (Table 2), which were deposited on alluvial fans and floodplains in the basin by ephemeral fluvial activity. Debris flows (Subfacies 1) are most common on fans in semiarid climates (Blissenbach, 1954; Hooke, 1967). The preservation potential of debris flow deposits is better in dry climates where they are not subjected to large scale reworking by streams. The presence of unreworked debris-flow deposits and desiccation features including abundant mudcracks, caliche horizons, and evaporite minerals suggests that this subfacies assemblage was deposited under arid to semi-arid conditions.

Wet cycle assemblages in the Hartford Basin are composed of Subfacies 3-6-7-8 (Table 2), which were deposited in fluvial and perennial lacustrine environments. These subfacies are indicative of meandering, sinuous rivers, lacustrine deltas, fan deltas, alluvial braided plains, and biologically productive floodplains and basin floors. The record of perennial, stratified deep lakes with well developed food webs (Olsen, 1980; McDonald, 1992) indicates humid climatic regimes. Coarse-grained subfacies of wet cycle assemblages are interstratified with more common dry cycle subfacies in several stratigraphic levels within the lower Portland Formation along the eastern margin of the basin (as close as one kilometer from the border faults). Interbedded alluvial fan and lacustrine deposits also are recognized near the basin margin in the Shuttle Meadow and East Berlin formations.

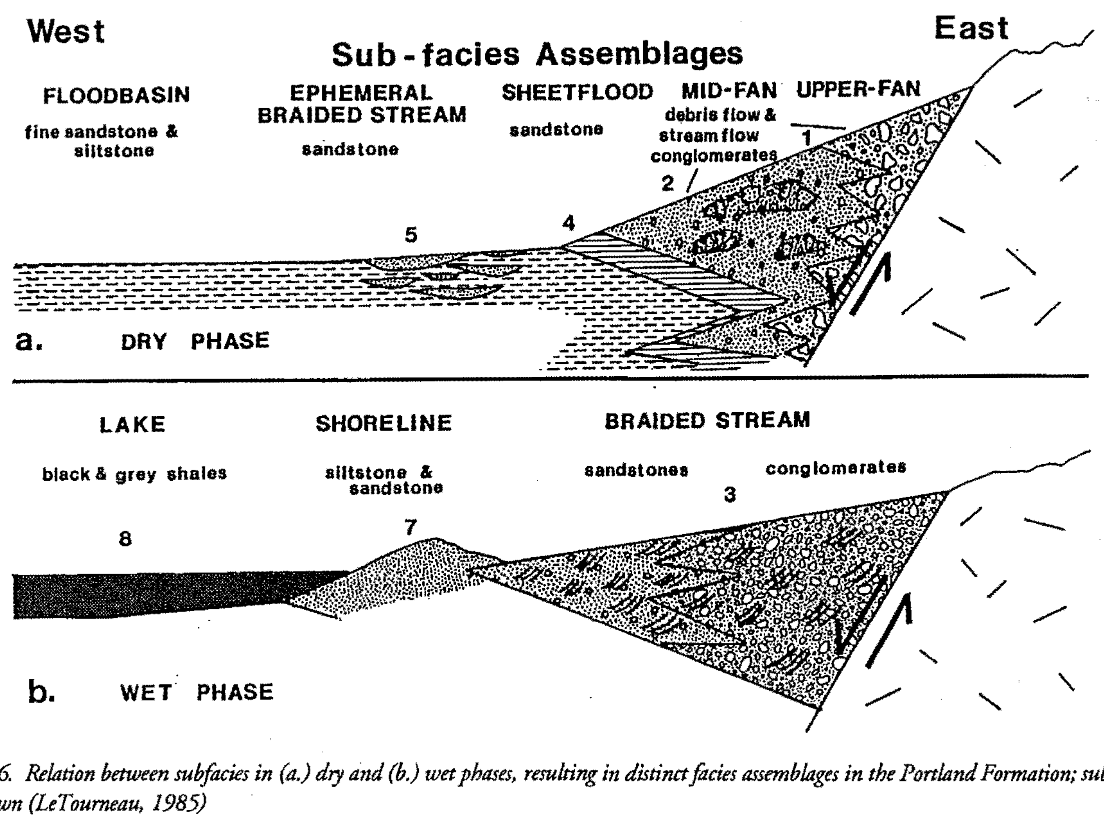

Figure 6. Relation between subfacies in (a.) dry and (b.) wet phases, resulting in distinct facies assemblages in the Portland Formation; subfacies 6 and 9 not shown (LeTourneau, 1985)

Tectonic Hypothesis

Various lines of evidence support the conclusion that syndepositional tectonic rejuvenation of the eastern border faults in the Early Jurassic also controlled the style of sedimentation and the distribution of lithofacies in the Hartford Basin. The grain-size distribution and the paleocurrent patterns in Subfacies 1 through 4 indicate that discrete, small radius alluvial fans were banked against the eastern border faults. This sort of fan distribution is typically associated with rapidly subsiding basin margins, as along the eastern side of Death Valley (Denny, 1965; Bull and McFadden, 1977). In contrast, the less active hinged margins of modern rift basins typically contain broad, low angle, coalescing fan complexes (Hooke, 1972; Steel, 1976; Leeder and Gawthorpe, 1987).

Syndepositional asymmetric basin subsidence is also suggested by the thickness of lacustrine strata. In the Shuttle Meadow, East Berlin, and Portland formations lacustrine strata consistently thicken eastward towards the border faults (Figure 3), where they become interbedded with alluvial fan conglomerate and sandstone (LeTourneau, 1985). The eastward thickening of individual black shales is also an indirect indication of increasing lake depth; quiet euxinic conditions persisted longer adjacent to the eastern margin (Figure 5).

Boulders and cobbles of tholeiitic basalt are common within several conglomeratic intervals in the Shuttle Meadow, East Berlin, and Portland formations near the eastern margin of the basin. The trace element geochemical signature of basalt clasts in the Portland Formation is very similar to that of the underlying Hampden Basalt (LeTourneau, 1985). If the Hampden were the source for the basalt clasts in the lower Portland, then tectonic activity is demanded. It is postulated that the relatively thick Hampden flow may have encroached into the drainage outlets of the eastern crystalline highlands during a pre-Portland interval of relatively low relief between the highlands and the basin. Renewed uplift of the highlands in early Portland time would result in erosion of portions of the flow located east of the basin margin and transport of basalt clasts onto the adjacent fans.

Implications of Cyclicity

Repeated alternations of depositional sequences indicate that cyclicity, both autocyclic and allocyclic in nature, exists on several scales. Small-scale (few meters or less) fining-up or coarsening-up sequences indicate individual depositional events or autocyclic channel entrenchment and avulsion. Syndepositional tectonic rejuvenation of relief at the basin margin produced progradational fan lobes consisting of coarsening-up sequences tens of meters thick. The alternations between wet and dry facies assemblages exist on a scale of hundreds of meters, and probably represent nonseasonal climatic fluctuations that may correlate with global cyclical events (Olsen, 1986).

Figure 7. Hypothetical eastward view across the central part of the Hartford Basin in the Early Jurassic.

The sedimentary succession within the Hartford Basin records the influence of both episodic (tectonic) and periodic (climatic) forces on depositional processes and environments. The interplay of tectonic and climatic controls has resulted in complex interfingering and intercalation of subfacies. In the central portion of the basin the climatic signature is more pronounced than the tectonic signature due to the widespread, fine-grained depositional regime. Small changes in climatically controlled lake depth produced wide lateral migrations of lake margins. Increased rainfall produced broad meandering rivers and extensive floodplains in areas adjacent to the lakes. In contrast, near the basin margin evidence of climatic influences on sedimentation are less obvious than the effect of tectonic subsidence, which controlled the geometry of both the alluvial fans and lake basins. Climatic cycles are still indicated by the intercalation of lake and fan deposits, but the marked asymmetry of the lake deposits is a result of the tectonic architecture of the basin.

STOP DESCRIPTIONS

Stops 1 and 2 are about 3 km west of the eastern margin of the Hartford Basin on Round Hill, a glacially scoured and heavily vegetated bedrock hill that typifies the quality of natural exposures in low relief areas of Southern New England such as the Connecticut Valley. Scattered outcrops of conglomerate and sandstone are found along numerous linear ridges, small ledges, and bedding plane exposures. No single outcrop in the Round Hill study area is adequate to characterize the range of local alluvial fan deposits. The alluvial fan depositional model was developed from observations of virtually all of the accessible exposures in the area.

STOP 1: UPPER FAN FACIES OF PORTLAND FM.

Stop 1 provides an excellent example of the range of depositional styles in the coarsest units to be observed in the Hartford Basin. The central areas of Round Hill are notable for extremely large boulders, up to 2 m in length. The vertical section at this stop consists of coarsening-up sandstone and conglomerate units interpreted from base to top as: sheetflow sandstone; shallow, ephemeral, braided stream deposits of cross stratified coarse and pebbly sandstone; and debris flow deposits of cobble and boulder conglomerate.

A boulder bed is the most prominent feature in the 10 m high outcrop. At the north end of the boulder bed a number of features indicative of debris flows can be observed, including: inverse grading, unsorted matrix supported major clasts with a random or chaotic fabric, depositional units with planar lower contacts and hummocky or irregular upper contacts, and abrupt textural contrasts with adjacent units. Gray-green weathered basalt clasts in the debris flow conglomerates are geochemically similar to the Hampden Basalt, which underlies the Portland Formation. Approximately 50 m north of the boulder bed several large boulders shelter underlying remnants of debris flow fabrics. Other large boulders occur as isolated clasts in a poorly sorted pebbly sand matrix. These isolated boulders are thought to have been freighted out on the alluvial fan surface by debris flows, which were subsequently eroded removing most or all of the finer-grained constituents, leaving a lag deposit of large, isolated clasts.

The debris flows are surrounded by discontinuous beds of poorly sorted and normal-graded conglomerate and pebbly sandstone that constitute the bulk of the alluvial fan deposits. These deposits were formed by deposition and reworking of alluvial sediment in ephemeral, shifting, braided or bifurcating streams on the fan surface. The beds internally contain variable planar to trough low-angle cross-stratification and scattered scour-and-fill structures. Draped lenses of silt are common along the upper surfaces of beds. The base of the exposure consists of planar to wavy, laminated, silty fine sandstone, interpreted as distal fan deposits. The vertical succession exposed here is thought to represent a progradational fan episode.

STOP 2: MID-FAN FACIES OF PORTLAND FM.

Stop 2 is laterally equivalent to the ridges seen at Stop 1, and is located only 1 km along strike to the north. Two parallel ridges of alluvial fan conglomerate and sandstone illustrate the lateral changes in facies from the previous stop. These exposures also comprise coarsening-up progradational sequences, but contain fining up units of finer texture and thinner succession. Debris flow deposits have not been recognized here. Maximum clast size is smaller than at Stop l, but cobbles and boulders do occur in clast-supported, imbricated layers and lenses.

The exposure along the western lower ridge consists of thin, fining-up depositional units representing decelerating flow events. Individual units are well stratified and most of the cross-stratification is small scale. At the south end of the ridge a shallow stream channel (~1 m thick and ~5 m wide) with a scoured lower contact, defined by a cobble-pebble lag deposit, and planar upper contact is present. The channel fill is noticeably finer grained and better sorted than the surrounding units, and internal cross-stratification is very well developed. Comparison of the depositional style within the channel with the surrounding units is indicative of the contrast between more frequent and persistent stream flow in an alluvial fan channel and less frequent short duration, high velocity flow over broad surfaces of a fan during high discharge events.

The exposure along the eastern upper ridge shows features similar to those described in the lower ridge. However, the eastern ridge provides a better example of the overall coarsening-up nature of the sequences, interpreted as successive alluvial fan progradational cycles. At the base of the exposure a small, cross-stratified channel deposit in fine-grained strata is present. Near the north end of the ridge a thick unit of well-stratified and well-sorted sandstone probably represents a distal fan deposit, but may be partly of aeolian origin.

STOP 3: EASTERN LAKE MARGIN FACIES OF PORTLAND FM.

Stop 3 is located less than one km from the eastern margin of the Hartford Basin, and illustrates several fundamental aspects of the paleogeographic model for the rift basin proposed herein. This stop is significant because of its proximity to the faulted eastern border, the extremely wide range of grain sizes in vertical association, the three dimensional aspect of the exposure, and the presence of paleoenvironmentally diagnostic fossils. In a very limited succession, this small exposure displays a wide range of rift basin depositional facies, including deep water lacustrine laminates, shallow lacustrine littoral sands, fan-delta conglomerates, and alluvial fan sandstones and conglomerates. Sediment grain size from base to top ranges from mud to very large boulders (~2 m long). The exposure illustrates the intimate association of alluvial fan and lacustrine environments along the eastern faulted margin of the rift basin.

The base of the outcrop consists of dark-gray to black laminated shale containing well preserved, articulated fish remains. The overlying fine-grained sandstone is ripple cross-laminated and contains abundant plant debris and evidence of soft sediment deformation; a large slump fold is present in the central portion of the outcrop, and asymmetric interstratal folds are present near the top of the unit. The top of the fine sandstone is marked by an abrupt contact with overlying pebbly and cobbly sandstone. The conglomeratic sandstone rapidly coarsens upward into cobble and boulder conglomerate.

The south side of the outcrop has an extraordinary bedding plane exposure of coarse conglomerate; clasts consist of both low-grade and high-grade metamorphic rocks. The western end of the outcrop displays several small sections of the internal fabric of the deposit along joint surfaces. The range of depositional fabrics through the outcrop indicate that the lower portion of the conglomerate was deposited subaqueously, presumably near the shore of a perennial stratified lake. The upper portion exhibits the crude stratification and internal laminations typical of alluvial fan or fan delta deposits.

The cyclic nature of the alluvial fan and lake deposits is indicated by the presence of a coarse, fining-up conglomerate located immediately north of the exposure viewed here. The conglomerate underlies the dark shale in this exposure, and is not as coarse grained as the upper conglomerate. However, it does contain a similar assemblage of clast types.

STOP 4: LACUSTRINE-FLUVIAL CYCLES IN EAST BERLIN FM.

The three extensive roadcuts at this stop expose the upper two thirds of the East Berlin Formation in a nearly continuous section 120 m thick. Overlying the East Berlin at the eastern end of the cuts, almost all of the Hampden Basalt is exposed. Although the variety of colors, fabrics, and textures in the East Berlin strata had been previously recognized (Davis, 1898; Krynine, 1950), the existence of lacustrine cycles in the formation was first clearly documented by Klein (1968) and confirmed by most recent workers (Hubert el al., 1976, 1978; Demicco and Gierlowski-Kordesch, 1986; Olsen el al., 1989).

The East Berlin Formation here consists of cyclical red, gray, and black lacustrine strata with subordinate interbedded fluvial units. Three cycles in the middle part of the East Berlin Formation are exposed at the west end of the southernmost roadcut as sequences of gray mudstone - black shale - gray mudstone, thought to record the expansion - high stand - contraction of large perennial lakes. The expansion and contraction of lakes was partly climate-controlled, perhaps in response to Milankovitch-type orbital forcing (Olsen, 1986). Similar cyclical deposits in the Triassic Lockatong Formation in New Jersey (Van Houten, 1964) have been called "Van Houten cycles" (Olsen, 1986).

The upper cycle in this triplet contains a distinctive black, laminated carbonate-rich shale bed correlative with the Westfield Fish Bed (McDonald, 1982; Olsen, 1988), a unit which is widely traceable the East Berlin Formation and contains a characteristic fossil assemblage. Here this bed has produced whole, but dephosphatized specimens of the fishes <u>Semionotus</u> and <u>Redfieldius</u> (McDonald and LeTourneau, 1989), as well as coprolites, conchostracans, and fern and conifer fragments. The associated gray mudstones in this cycle are palynologically productive and have also yielded carbonized leaf and twig fragments of the conifers <u>Brachyphyllum</u> and <u>Pagiophyllum</u> and the cycadophyte <u>Otozamites</u>. During the excavation of these roadcuts in 1988, numerous dinosaur tracks and invertebrate trails were collected from the red and gray units.

A 35 m-thick interval of red, gray, and purple strata separate the lower three Van Houten cycles here from three more lacustrine cycles in the upper East Berlin. The upper cycles are best exposed in the central (CT-9) and northernmost (CT-372) cuts, and are correlative with three dark shale cycles exposed at the I-91/CT-9 interchange, about 4 km to the east. The upper cycles have produced abundant dinosaur tracks (as at Dinosaur State Park), conchostracans, and fragmentary plants, but no articulated fish. Paleocurrents in the upper cycles trend strongly northeast (Hubert el al., 1978), indicating that the source for much of the East Berlin detritus lay to the west.

STOP 5: FAULTED AND MINERALIZED HAMPDEN BASALT

This stop provides an exposure of the Hampden Basalt (Fe-rich, high TiO_2, quartz normative tholeiite), the youngest of the three Early Jurassic volcanic flows in the Hartford Basin. The <100 m-thick basaltic unit presumably flowed east to northeast (Ellefsen and Rydel, 1985) away from a feeder dike and fissures in the metamorphic basement complex to the west (Philpotts and Martello, 1986). The Hampden consists of at least eight individual units separated by thin vesicular horizons which can be correlated over distances of more than 30 km (Chapman, 1965). The Hampden Basalt here is cut by numerous sub-vertical, NE-trending extensional fractures and normal faults. Hot brines apparently rose along these faults from underlying arkosic aquifers, leaching the basalt and emplacing quartz, sulfides, carbonates, barite, and small vitreous blobs of bitumen (Gray, 1988). Fluid inclusions in quartz crystals indicate that the brines were moderately saline, near CO_2 saturation, and precipitated most minerals at temperatures between 90° and 220°C.

Thermal maturity levels of the lacustrine black mudstones in the Shuttle Meadow and East Berlin formations indicate pyrolytic fields with temperatures between 435° and 460°C in the central portion of the Hartford Basin (Pratt el al., 1988). High heat flow along the axial zone of the rift valley may have been responsible for heating the brines, and seismic pumping may have aided in forcing the hot brines upward into a NE-trending set of fractures and faults. This thermal event postdated the three major volcanic events (Talcott, Holyoke and Hampden). Elsewhere in the basin it seems to have reset K-Ar ages (Seidemann el al., 1984; Sutter, 1988) and introduced a chemical remanent magnetization (de Boer and Snider, 1979; Witte and Kent, 1989). The relatively high radioactive content of the mineralized fractures allows them to be traced northeast into the lower section of the Portland Formation (Simpson, 1966). This indicates that the thermal event responsible for the mineralization occurred late in the evolutionary history of the Hartford Basin.

STOP 6: MID-BASIN FACIES OF SHUTTLE MEADOW FM.

DANGER: This is an unstable quarry face; DO NOT APPROACH THE EXPOSURE!

Approximately 60 m of mudstone, siltstone and sandstone from the upper half of the Shuttle Meadow Formation (in contact with the overlying Holyoke Basalt) are exposed in this former rock quarry. In general, the lower part of the exposure consists of red-brown, fine-grained floodplain deposits and the upper part consists of buff-brown to gray sandstone lenses and layers interbedded with similar fine-grained deposits. Two thin micritic limestone beds (rarely found elsewhere) are located near the base of the far west end of the exposure.

The fine-grained units dominate the section and contain a wide variety of sedimentary structures including: climbing ripple cross-lamination, mud peloid ripples, desiccation cracks, vertebrate tracks, soft sediment deformation features, burrowed horizons, planar lamination, parting lineations, low-angle planar cross stratification, and minor trough cross-stratification. The sandstones contain both planar and trough cross stratification.

A prominent channel located halfway up the exposure is a lateral accretion feature that grades into floodplain mudstones to the west and abruptly terminates against floodplain deposits to the east. The geometry of the sandstone channel fill and the type of sedimentary structures within the sandstone are typical of a high sinuosity meandering river. The thin gray mudstone associated with the channel sandstone may represent a shallow, organic-rich floodplain lake or an abandoned meander bend.

Hubert el al. (1978) interpreted the limestones here as deposits of shallow, alkaline lakes or playas with substantial dissolved magnesium, calcium, carbonate, and sulfate. The limestones are primarily a sandy and muddy dolomitic micrite. The lower limestone contains round fragments of micrite containing gypsum crystals and sparry calcite and pieces of algal tufa; sparse ostracod shells and fish bones have also been observed (Hubert el al., 1978). The sedimentary characteristics of the limestone units and surrounding strata are consistent with a playa depositional environment (Hubert el al., 1982). Massive mudstone surrounding the limestone contains abundant evidence of both desiccation and flashy, ephemeral flow.

The contrast between the lower and upper portion of the outcrop may represent a transition from dry to humid climatic conditions. The fine-grained deposits in the lower portion of the section contain more features indicative of dryer conditions, including desiccation cracks, paleosol features, and playa fabrics. The sandstones and mudstones in the upper pan of the section have fewer desiccation features, and appear to be meander plain deposits. The progression from thick, desiccation-cracked mudstones to fluvial mudstones and sandstones may indicate a long term change in climatic regime during the deposition of the section.

STOP 7: BRAID PLAIN FACIES OF NEW HAVEN ARKOSE

The middle part of the 2000 m-thick New Haven Arkose exposed in this large roadcut has been extensively studied by John Hubert and his students (Hubert, 1978; Hubert el al., 1978), from which the following description has been abstracted. This 72 m-thick section is interpreted as an alluvial plain sequence of sandstones and fine conglomerates deposited in channels and longitudinal bars within braided streams, interbedded with sandy mudstones deposited on interfluvial floodplains.

The sandstones and pebble conglomerates are in lenticular and plane beds, with planar, tangential, and trough cross-beds in sets up to 1 m thick. Of special interest are the numerous sets of avalanche foresets, cut and-fill structures, and well preserved channels and bars with up to 2 m of relief. Paleocurrent directions are generally toward the southwest, but have a very wide ranging distribution. These deposits have been interpreted (Hubert el al., 1978) as representing relatively wide and shallow ephemeral rivers of high gradient and low sinuosity, floored by braided channels with intervening bars, and carrying a coarse bedload of pebbly sand.

Red, planar-laminated, sandy mudstone in lenticular and planar beds is interbedded with the sandstones and conglomerates, and comprises about 15% of the section exposed here. It also is present as large clasts and blocks within some of the channel deposits. The mudstones are considered to be overbank floodplain deposits. Many of the mudstone beds have grayish calcareous horizons in their upper portions, with small green patches and lenses. The calcareous material is nodular, and increases in abundance upwards through an individual bed commonly to a laminated or brecciated calcareous crust at the top. Tubular structures that thin and radiate downwards through these beds are thought to be rhizomorphs or root casts. These calcareous horizons have been interpreted (Hubert, 1978) as paleosol horizons with well developed caliche profiles.

STOP 8: WEST PEAK OF HANGING HILLS

West Peak is supported by the upper, glaciated surface of the Holyoke Basalt, a low TiO_2, quartz normative tholeiite that is the second and most voluminous of the three Early Jurassic extrusive units in the Hartford Basin. It spread over a distance of at least 3500 km^2, and it has a thickness of >100 m, for a cumulative volume of at least 350 km^3. Its feeder system is unknown, but geophysical evidence suggests that the central segment of the axial fault zone may have been the feeder (de Boer, 1992). The exposure here shows cross sections of differentially weathered polygonal cooling columns. Differential weathering indicates more rapid cooling along the joints. Sets of NNE-trending joints and minor faults intersect the cooling joints and become more abundant towards a major (steeply west dipping) fault zone east of the peak. Downdip motion of the hanging (western) block along this fault zone and at least eight similar and parallel fault zones further east between Meriden and Berlin, is responsible for the apparent sinistral offset of the Holyoke ridge in the central part of the Hartford Basin. Similar motions account for the relatively abrupt deepening of this basin to the northwest (Wenk, 1984, 1989; de Boer and Clifton, 1988).

The topography of the glaciated Hartford Basin can usually be seen clearly from this site. To the west lies a minor valley underlain by the New Haven Arkose, and to the northeast one can see the depression occupied by the Connecticut River valley, underlain by the Portland Formation. Directly south are the outlines of the Sleeping Giant, a massive diabase sill with Talcott affinity, and East Rock and West Rock, large dike-sill complexes which encircle the city of New Haven.

STOP 9: MEANDER PLAIN FACIES OF NEW HAVEN ARKOSE

This description is based on recent work by McInerney (1993). About 130 m of the Upper Triassic New Haven Arkose are exposed in this roadcut, in a succession with up to 22 couplets of fluvial channel sandstone-conglomerate and floodplain mudstone, with numerous caliche horizons in the floodplain mudstones.

The most prominent feature at this stop is the alternation of sequences of light-colored sandstone and conglomerate and darker, fine-grained units. These sequences represent a variety of channel deposits formed by high sinuosity shallow rivers on a broad, intermittently flooded, vegetated floodplain. The channel deposits contain structures formed as bedforms or bars within the rivers. Lateral accretion surfaces, trough cross-bedding, low angle cross-bedding, ripple cross-lamination, and burrows are abundant. The channel units generally fine upward and often contain basal concentrations of pebbles or irregular mudstone clasts derived from cut bank collapse. The floodplain mudstones are characteristically bioturbated by plants and invertebrates. A complete range of partially to wholly bioturbated deposits can be observed.

McInerney (1993) defined 8 types of architectural-depositional units here: isolated channel sheet, amalgamated channel sheet, isolated channel ribbon, minor channel sheet, muddy sandstone (crevasse splay and levee),"U"-shaped fill units, massive mudstone, and caliche units. Each of these represents distinct depositional sub-environments in the meandering river environment. Paleocurrent directions are variable, as expected on a meander plain, but indicate a dominant flow direction to the south and southeast. Petrology of the channel sandstones and conglomerates indicate that some of the detritus was derived from the west side of the basin.

STOP 10: ALLUVIAL PLAIN FACIES OF BASAL NEW HAVEN ARKOSE, THE "GREAT UNCONFORMITY"

Since its discovery in 1890, the angular uncomformity between Late Triassic conglomeratic arkose and steeply tilted Paleozoic mica schist exposed in Roaring Brook has been one of Connecticut's most famous outcrops, often called the "Great Unconformity." Extrapolation of the structure of this single exposure along the entire western margin of the basin led Barrell (1915) and many later investigators to conclude that basin subsidence was controlled primarily by fault displacements along its eastern boundary. Therefore, the western boundary was inferred to be a hinge zone and the basin as a whole was considered a half graben. Subsequent studies, however, have revealed that much of the western margin of the basin also is bounded by faults, some of which apparently were syndepositionally active. The Roaring Brook exposures are bounded both to the east and west by E-dipping, NNE-trending normal faults (Fritts, 1963).

Davis (1898) used this exposure to advance his hypothesis that initial sedimentation in the basin took place on a peneplained basement surface, with detritus entering the basin from both sides. He emphasized that some clasts in the Triassic arkose immediately above the unconformity were locally derived: "...the pebbly sandstones contain fragments of quartz and schist, some of which may be identified as corresponding to the crystalline rocks in place in their neighborhood; and this gives assurance that the sandstones were made from the ruins of the foundation rocks on which they lie." (Davis, 1898, p. 20). Noting angular clasts of quartz and feldspar in the arkose at this locality, Rice and Foye argued for their local provenance from the abundant quartz veins and pegmatite dikes in the underlying schist "The arkose is composed of very local material. Fresh fragments of feldspar, two or three inches in diameter, still retain bits of tourmaline adhering to them. Their probable source may be found in pegmatite dikes only a few yards away." (1927, p. 135).

However, advocates of the broad-terrane hypothesis have repeatedly denied the existence of a western provenance or locally derived sediment at the Roaring Brook locality. Longwell (1933, p. 112) states: "No fragments of the schist are recognizable in the sediments above it." Agreeing with Longwell's assertions, Wheeler (1937) also concluded that the faults at this locality were postdepositional. Krynine (1950) petrographically compared quartzose clasts with pegmatite vein quartz at this locality, and concluded that none of the sediment was of local origin; he also suggested that the arkose was derived from the eastern crystalline highlands, 27 km to the east.

Recent examination of the Roaring Brook exposures confirms the presence in the arkose of small, but abundant fragments of mica schist which closely resemble the underlying mica schist. Furthermore, it seems unlikely that angular and unweathered clasts of quartz and feldspar in the arkose could have been transported across the basin without appreciable abrasion or disintegration. Easterly-directed paleocurrents obtained from the Roaring Brook exposures (Hubert el al., 1978) support a western provenance for some of this sediment.

ROAD LOG

Mileage (see Figure 8 for route)

FIRST DAY

0.0	Mileage starts at parking lot entrance of Holiday Inn, Cromwell, CT; turn left onto Sebethe Dr.
0.1	Turn left onto CT-372E.
0.4	Turn left onto I-9IN.
1.4	Turn off at Exit 22 onto CT-9S; keep right towards Middletown.
6.9	Stoplight on CT-9S at Middletown under Connecticut River bridge; continue south on CT-9 and note small outcrops of Jurassic Portland Fm. redbeds for next 3 miles.
10.1	Stop on right shoulder at merge sign just past Exit 11 (Randolph Rd.).

OVERVIEW OF EASTERN BASIN (5 minutes)

Continue south on CT-9 across border fault.

10.8	Roadcuts in Brimfield Schist (Ordovician) on the Bronson Hill Anticlinorium.
11.6	Turn off at Exit 10 onto Aircraft Rd.

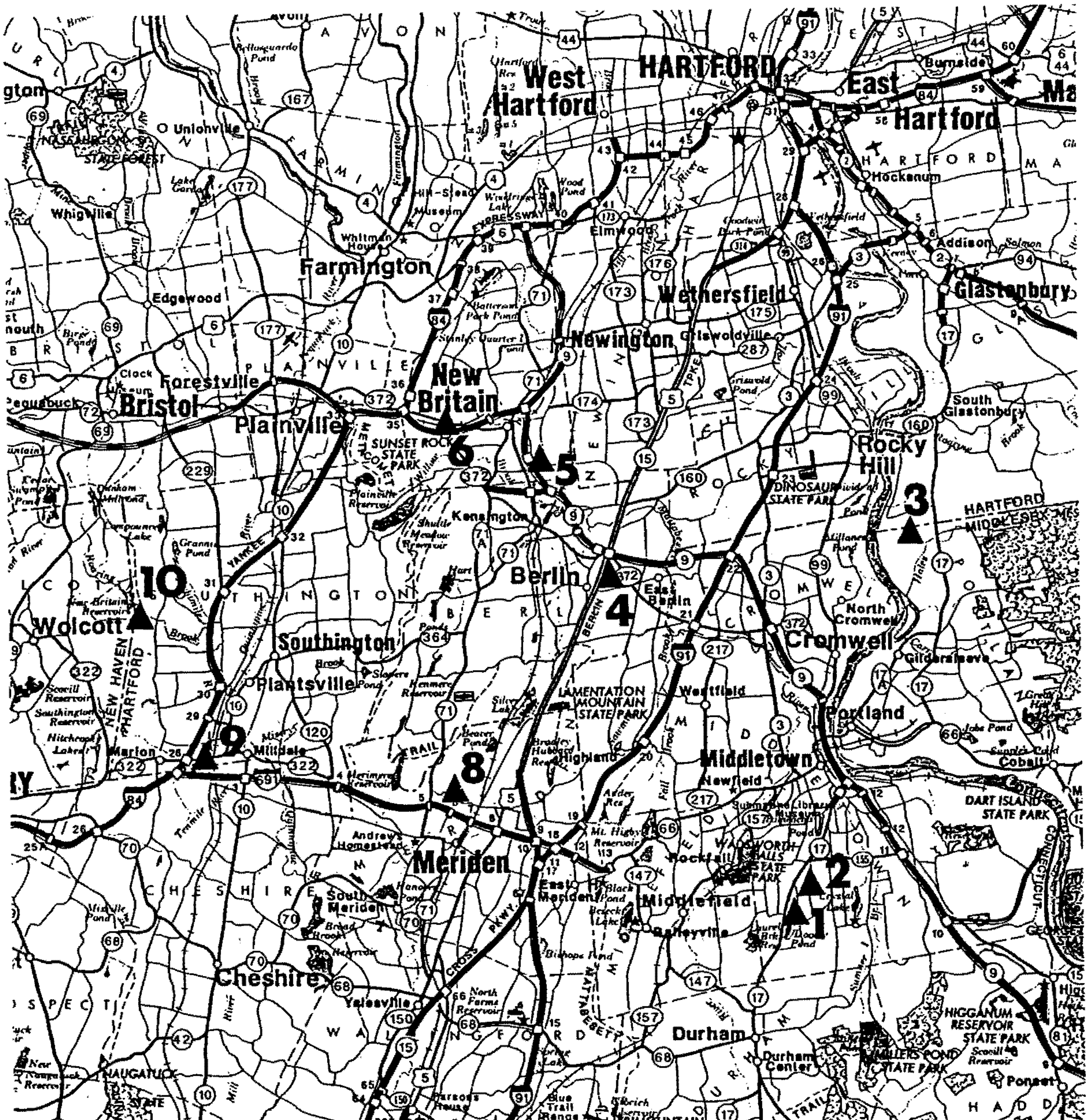

Figure 8. Route map of field trip through central Connecticut, taken from Connecticut State Highway Map.

12.2 U-TURN at stoplight at bottom of exit ramp.

12.4 Return to CT-9N; overview of basin as you proceed north.

14.1 Turn off at Exit 11 towards Randolph Rd.; note metamorphic rocks in cuts to right (east); you are descending the fault line scarp of the eastern border fault.

14.4 At bottom of exit ramp turn left (W) onto Randolph Rd.; note low outcrop of Portland Fm. sandstone by union hall just before passing under CT-9 overpass; you have just crossed the eastern border fault.

16.6 Turn left at stoplight onto CT-17S.

17.1 Turn left onto Coleman Rd. just past Stonegate Apts.; note outcrops of conglomerate along road to left.

18.1 Turn left onto Kelsey St.

18.3 Pull off across from small lane to left just over the crest of the hill; follow Lane to north across field and through woods to west-facing ledges of conglomerate.

STOP 1: UPPER FAN FACIES OF PORTLAND FM. (30 minutes)

18.4 Turn left at Y onto Maple Shade Rd.

19.1 Pull off just past house #275 on left; enter field to left through gate along stone wall; outcrops of sandstone are located both ahead (west) behind the house and to the right (north) below the road.

STOP 2: MID-FAN FACIES OF PORTLAND FM. (30 minutes)

19.3 Turn left and continue south on Maple Shade Rd.

19.6 Turn right onto Randolph Rd.

21.8 Turn right onto CT-9N.

25.1 Turn left at second stoplight (just before bridge), entering Middletown and onto CT-66E and CT-17N.

25.3 Turn right at stoplight just past church in Middletown; follow CT-66E and CT-17N over bridge.

26.3 Stoplight in Portland; continue straight ahead on CT-17A.

29.4 Turn left onto CT-17N.

30.0 Ascending eastern fault line scarp of basin.

31.4 Crossing ridge of Ordovician Brimfield Schist; Devonian Glastonbury Gneiss supports hills to the east, with line of famous pegmatite mineral quarries between them.

32.2 Turn left onto Old Maid's Lane; note low roadcut of schist to right (east) at turn; you will cross the border fault and enter the basin within 500 m.

32.6 Turn left onto gravel road just before third tobacco barn on left; proceed past orchard over hill and into gravel pit. Walk north across gravel pit to moundlike outcrop.

STOP 3: EASTERN LAKE MARGIN FACIES OF PORTLAND FM. (30 minutes)

Retrace route back through Portland to Middletown via CT-17S, CT-17A, and CT-66. LUNCH at Wesleyan University Science Tower. Return to CT-9N under bridge over Connecticut River.

0.0 Reset mileage at stoplight entering CT9 under bridge in Middletown; turn left onto CT-9N.

4.8 Bear left on CT9N towards New Britain.

5.4 Roadcuts in East Berlin Formation: equivalent units will be seen at the next stop.

6.0 View to left (south) of Mt. Higby, supported by Holyoke Basalt.

7.3 Turn off at Exit 21 onto CT-372W.

7.7 Roadcuts through Hampden Basalt and East Berlin Fm.

8.2 Turn left at stoplight towards US-5/CT-15S.

8.3 Turn left again at second stoplight.

8.5 Turn right onto US-5 and CT-15S.

8.9 U-turn at stoplight back onto US-5/CT-15-N.

9.4 Pull off highway to right by school bus route sign, just before entry ramp onto CT-9S towards Middletown; proceed on foot up south side of entry ramp towards CT-9S to examine roadcut; **do** not try to cross entry ramp!

STOP 4: LACUSTRINE-FLUVIAL CYCLES IN EAST BERLIN FM. (30 minutes)

Proceed north on US-5/CT-15.

9.7 Exit right towards CT-9N and follow signs to CT-9N towards New Britain.

10.1 Turn left onto CT-9N.

12.9 Pull off to right shoulder just past Exit 25 (Ellis St.) and beneath overpass at 35 mile marker; beware of entering traffic on ramp!

STOP 5: FAULTED AND MINERALIZED HAMPDEN BASALT (15 minutes)

Continue on CT-9N.

13.5 Exit left at Exit 28 onto CT-72W.

16.9 Turn off at Exit 34.

17.2 Turn right onto Crooked St. and proceed to stoplight at intersection with CT-372.

17.4	Turn right at stoplight onto CT-372E and pull into parking area on right just past Sunoco
17.5	Carefully cross street to large exposure to east of Mobil station.

STOP 6: MID-BASIN FACIES OF SHUTTLE MEADOW FM. (30 minutes)

Return to Holiday Inn in Cromwell via CT-72E, CT-9E, and I-91S; end of first day.

SECOND DAY

0.0	Reset mileage at Holiday Inn parking lot; exit lot to left onto Sebethe Drive; turn left at stoplight onto CT-372E, and left again at second stoplight onto I-91S.
3.0	View of Mt. Higby (Holyoke Basalt) ahead.
5.6	View to right (west) of Hanging Hills of Meriden.
19.9	Turn off at Exit 10 onto CT-40 towards Hamden.
21.7	Pull off onto right-hand shoulder of highway at overpass by long roadcut of sandstone and conglomerate.

STOP 7: BRAID PLAIN FACIES OF NEW HAVEN ARKOSE (45 minutes)

Continue ahead to end of highway.

22.6	Turn left onto CT-10S; keep right for immediate turn.
22.7	Pull off to right into front parking area of New Haven Savings Bank, pull through lot and U-turn back onto CT-10N, and immediately turn right back onto CT-40 towards I-91.
25.2	Exit left onto I-91N.
37.8	Turn off at Exit 17 towards I-691.
38.8	Turn off at Exit 68W onto I-691W.
39.6	Hanging Hills in foreground. Note several long roadcuts in New Haven Arkose for next few miles.
43.4	Turn off at Exit 4 to CT-322 towards Southington.
43.7	Turn left onto CT-322E and cross over I-691.
44.6	Turn left into park just before pond, go around pond to right.
44.9	Continue straight ahead at stop sign.
45.0	Turn left at second stop sign.
45.1	Gate at foot of Hanging Hills under I-691 overpass; continue ahead into fault line gorge.
46.4	Turn left at dam and cross over dam. Keep right at Y in road (<u>not</u> towards Castle Craig).
48.2	Stop in parking lot at top of mountain by towers. Proceed south on foot to cliff edge; <u>be very careful!</u>

STOP 8: WEST PEAK OF HANGING HILLS; OVERVIEW AND LUNCH STOP (one hour)

Retrace route back to I-691.

52.7 Turn left onto I-691W.

55.7 Pull off highway onto right-hand shoulder at roadcut just before Exit 2 to I-84E.

STOP 9: MEANDER PLAIN FACIES OF NEW HAVEN ARKOSE (30 minutes)

55.8 Continue west and exit just ahead at Exit 2 onto I-84E.

57.8 Turn off at Exit 30 towards Marion Ave.

58.0 Turn left onto Marion Ave.

58.8 Turn right onto Frost St.

59.5 Turn right onto Mt. Vernon Rd.

61.4 Turn left onto Roaring Brook Rd., just past sharp right curve.

61.6 Stop in cul-de-sac at end of Roaring Brook Rd. Proceed on foot from power supply box to west of house #112 along path -100 m north to exposures along Roaring Brook.

STOP 10: ALLUVIAL PLAIN FACIES OF BASAL NEW HAVEN ARKOSE AND THE "GREAT UNCONFORMITY" (45 minutes)

Retrace route back to I-84 and enter I-84E towards Boston. END OF TRIP.

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Trip E

Tectonics, Wall-Rock Alteration and Emplacement History of the Lantern Hill Giant Quartz Lode, Avalonian Terrane, Southeastern Connecticut

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INTRODUCTION

Southern New England was subjected to regional extension during Late Triassic to Early Jurassic. Evidence for this deformation is present in the existence of basin-fill sediments and flood basalts in continental failed rift basins of the Newark terrane (e.g. the Hartford, Deerfield and Pomperaug basins) (Figure 1) (Rodgers, 1985; Zen et al., 1983). The largest of these rift basins is the N-S trending Hartford basin that traditionally has been interpreted to possess the structure of a half graben (e.g. Rodgers, 1985) with the "hinge" located along the western border and a significant "listric" normal fault making up the eastern border.

Earliest sedimentation in the Hartford basin has been presumed by some geologists to have been related to the beginning of Newarkian extensional tectonics and the opening of the Atlantic Ocean. The oldest unit in the Hartford basin is the New Haven Arkose (Krynine, 1950; Rodgers, 1985). The New Haven Arkose has been dated palynologically to be 219 Ma [Late Triassic] (Cornet, 1977, pers. comm., 1994). Extensional tectonics associated with the break up of Pangea and the opening of the modern Atlantic Ocean must have been in operation by that time, but when did it begin?

The distribution of regions exhibiting Mesozoic extensional stress in southern New England is not well known. Down-dropped and preserved blocks such as the Hartford and Pomperaug basins have escaped erosion during the past 220 million years. Late high-angle faults that displace the high-grade metamorphic rocks of the highlands outside the rift basins are attributed to Newarkian tectonics (e.g. Eberly, 1985).

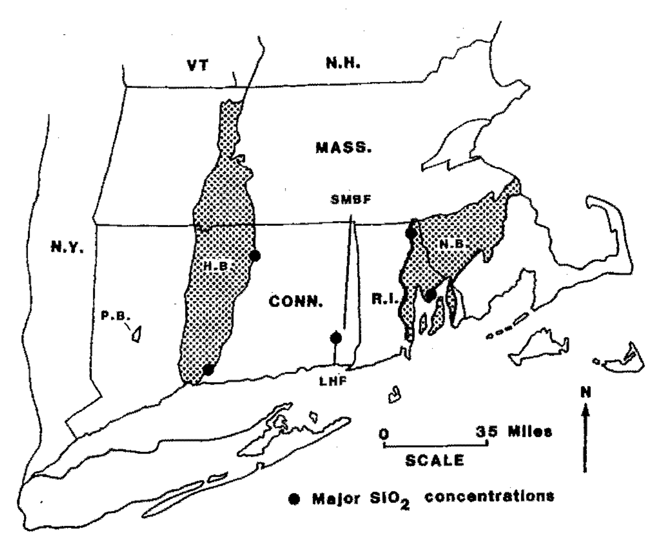

Figure 1. Major basins and major silica concentrations in southern New England. H.B. and P.B. denote the Mesozoic Hartford and Pomperaug basins; N.B. denotes the Carboniferous Naragansett basin. LHF and SMBF refer to the Lantern Hill fault and the Snake Meadow Brook fault respectively.

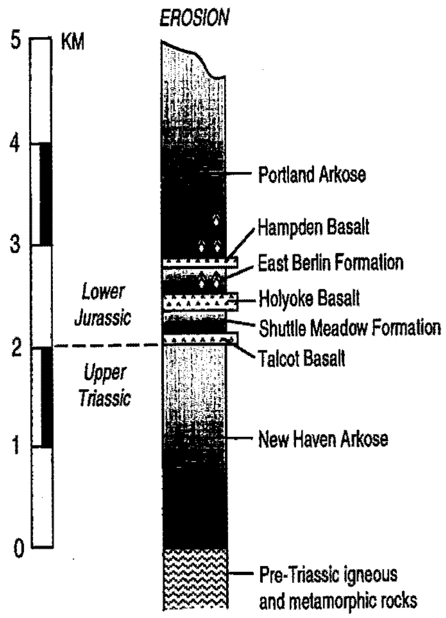

Figure 2. Stratigraphic column for Mesozoic rocks of central Connecticut with generalized thicknesses of units. Hydrothermal quartz clasts from the North Branford lode are reported in sedimentary units above the Holyoke Basalt, indicating that the source of the quartz is that age or older. (modified from Hubert et al., 1978)

The New Haven Arkose is hydrothermally silicified near the western border fault of the Hartford basin (Clifton, 1987; de Boer, pers. comm., 1989) and in the small western outlier known as the Cherry Brook basin (Figure 1). Silicification in the Cherry Brook basin appears to be incorrectly interpreted as a silcrete paleosol by McDonald and Textoris (1984). Mesozoic rocks along the eastern border of the Hartford basin are not known to be silicified. Instead, clasts of hydrothermal quartz are found in the sedimentary formations above the Holyoke Basalt (i.e. the Early Jurassic East Berlin Formation and higher in the section) (Figure 2) (Russell, 1922). These are derived from a giant quartz lode, similar in many ways to that of Lantern Hill, in the footwall of the eastern border fault near the town of North Branford. The North Branford lode was formed along a fault and exhumed. Its age and that of the fault must predate the East Berlin Formation. New radiometric age data indicate that silicification along the Lantern Hill (LH) fault occurred at 238 Ma (Middle Triassic: *Anisian*) (Altamura and Lux, 1994), and therefore correlation to the North Branford lode and tectonics of the Hartford basin may be possible.

Other lodes and post-orogenic quartz veins occur along the eastern border fault in Connecticut and in New Hampshire and may be related. North of the North Branford lode near the Town of Middletown, a 240.6 $\pm$2.0 Ma plateau age ($^{40}Ar/^{39}Ar$) characterizes hydrothermal muscovite from late quartz veins that intrude a Permian rare-metal pegmatite body (the "White Rocks pegmatite") in the footwall of the eastern border fault of the Hartford rift basin (Altamura and Lux, 1994, in preparation). Biotite separates from elsewhere in the same pegmatite district (the Middletown pegmatite district) have yielded K-Ar ages of 240 $\pm$5 Ma (Brookins and Armstrong, 1980). It is generally accepted that pegmatites from Portland, Connecticut, part of the same pegmatite district, were emplaced from a magmatic source at about 255$\pm$5 Ma (Brookins and Armstrong, 1980). Pegmatite muscovite cooling ages were determined to be 255 $\pm$3 Ma for separates from near the village of Cobalt, Connecticut, and 244 $\pm$5 Ma for those from near Portland, Connecticut (Brookins and Armstrong, 1980). Pegmatite biotite separates from the Strickland Quarry, also near Portland, yielded 241 $\pm$5 Ma (Brookins and Armstrong, 1980). Muscovite in the pegmatites should have been blocked (earlier) at higher temperatures than coexisting biotite. The exact geological significance of the muscovite age determination for quartz veins that intrude the White Rocks pegmatite cannot be determined from existing age data. Further study is warranted. It should be noted that the age determination for vein-muscovites is similar (within error bars) to crystallization ages obtained for the Lantern Hill lode. A detailed study of silica deposits along the eastern border fault of the Hartford rift basin, and a test for genetic correlation to the Lantern Hill fault is differed to a future investigation.

Field Trip Area

The field area is included on the Old Mystic 7.5' quadrangle in southeastern Connecticut. This report focuses on 363 acres that were formerly the property of the U.S. Silica Company of Connecticut and includes the entire map area of the Lantern Hill quartz lode (Figure 3).

Southeastern New England is predominantly underlain by crystalline rocks characterized as a complex of abundant granitic

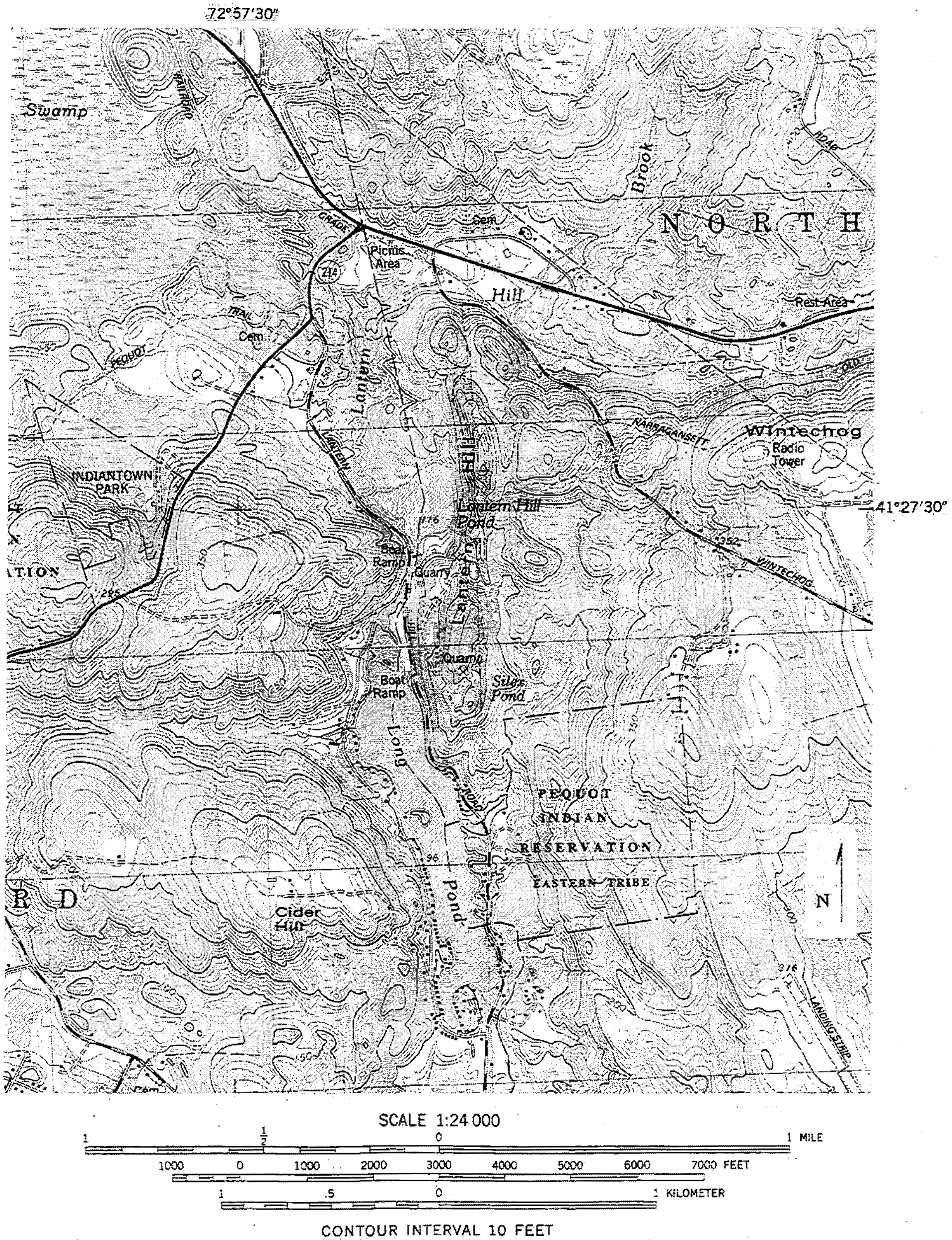

Figure 3. Topographic map showing the Lantern Hill quartz lode field area, North Stonington and Ledyard, Connecticut. The linear ridge labeled Lantern Hill and the two knobs at the northern end are underlain by resistant quartzose rock. (modified from the Old Mystic 7.5' quadrangle, U.S.G.S.)

intrusions (gneisses) and metasedimentary and metavolcanic rocks that are Late Precambrian to Middle Paleozoic in age (e.g. Rodgers, 1985, Hermes et al. 1994; Zen et al., 1983). These rocks indicate a period of intrusion and probably of metamorphism, and folding at about 620 Ma (Rodgers, 1980; Wintsch and Aleinikoff, 1987). Rocks of this terrane can be traced from near New Haven Harbor, eastward to Rhode Island and eastern Massachusetts and then northward (Percival, 1842; Rodgers, 1985; Hermes et al., 1994; Zen et al., 1983). These rocks have been referred to as the Avalonian composite terrane, because it actually consists of several fragments (e.g. Wintsch et al., 1992). These rocks were subsequently intruded by granitic plutons of the Permian Westerly and Narragansett Pier granites (Rodgers, 1985).

The Lantern Hill lode stands as a N-S trending topographic high in the center part of the study area. The entire study area was subjected to glacial weathering and erosion during the Wisconsinan ice advance of the last ice age. Glacially polished surfaces, striae, roches moutonnée, till and stratified drift are features that can be observed.

PREVIOUS WORK

Goldsmith (1985) mapped the bedrock geology of the Old Mystic and Mystic quadrangles. He mapped the Lantern Hill lode as vein quartz and identified the zone of hydrothermally altered country rock adjacent to the lode on the west. No detailed investigation has focused solely on the Lantern Hill lode. Percival (1842) provided the first clear description and interpretation (unusual for him) of the lode and alteration zone as part of his manuscript: *The Geology of Connecticut.* Loughlin (1912) included useful information on the lode in his report: *The Gabbros and Associated Rocks at Preston, Connecticut.* Rodgers (1970) compiled a list of some quartz lodes in New England, including the Lantern Hill lode, in *Tectonics of the Appalachians,* and suggested the possibility that they might be related. Altamura (1987) reported the results of the first radiometric age study (K-Ar) on the lode. Altamura and Gold (1993a; 1993b; 1994) and Altamura (1994, in preparation) characterized the mineralogy and geochemistry, metasomatism, structural geology and economic geology of the lode. Altamura and Lux (1994) provided the results of a detailed $^{40}Ar/^{39}Ar$ investigation of the lode, metasomatic rocks and unaltered country rocks of the area.

PURPOSE

The purpose of this field trip is to consider the tectonic character of the Lantern Hill fault and quartz lode, and wall-rock alteration, in an effort to learn more about petrogenesis and emplacement history. These considerations will include placing this feature into the plate tectonic framework of the southern New England Appalachians.

This field excursion presents the results of research that characterizes deformation along a highly silicified portion of the Lantern Hill fault. This is a regional fault, whose major tectonic activity during the Middle Triassic, preceded, by some 19 Ma, first graben-fill sedimentation in the Hartford failed rift basin (Altamura, 1987; Altamura and Lux, 1994). The Lantern Hill lode and fault zone record evidence for rifting associated with the embryonic opening of the Atlantic Ocean (Altamura and Lux, 1994). Other giant lodes that are comparable in size, and associated with regional silicified faults, occur elsewhere in southern New England (Figure 1) and may be genetically related to the Lantern Hill lode (e.g. Rodgers, 1970; Altamura, 1987).

ACKNOWLEDGMENTS

The topics of this paper are part of a Ph.D. research project by the author at Penn State with D.P. Gold as faculty research advisor. Field work on this project has benefited from a number of friends and colleagues. R. Hunnisett of U.S. Silica Co. was continually helpful in my work within the open-pit quarry. U.S. Silica Company provided access to the mine property and a new topographic base map (1992) of the study area (1:1,200 scale) that was flown specifically for this project. Field excursions to the study site with H. Ohmoto during the 1991 field season and D.P. Gold, R. Goldsmith, J.G. McHone, R. Parizek and R.P.Wintsch during the 1992 field season were enlightening and very much appreciated.

H. Belkin opened his laboratory to me for fluid inclusion and SEM work. S. Sylvester assisted with the XRF analyses. D. Coburn assisted with the field collection of samples ear-marked for analytical work. D.R. Lux collaborated with all the new radiogenic isotope work. R. Malek assisted with the specific gravity determinations, and S. Raeburn with the probe work. J. Webb gave assistance in interpreting some of the geochemical data. D.P. Gold reviewed the ms. and N. McHone edited it. The author is responsible for any errors.

I wish to thank the College of Earth and Mineral Sciences at Penn State, the Connecticut Geological and Natural History Survey, Paul D. Krynine Fund, Sigma Xi and The U.S. Silica Company for financial support. I would also like to thank the Mashantucket-Pequot tribe for allowing our field party access to the Lantern Hill area for this field trip.

The Lantern Hill Fault

The Lantern Hill fault is one of several silicified fault zones in southern New England that trend along the eastern side of the orogenic axis of the Paleozoic Appalachians. The Lantern Hill [LH] fault transgresses the Proterozoic Z Avalonian terrane and extends north 15 kilometers from Long Island Sound to the Town of North Stonington, Connecticut. There it steps to the east about 4 km in an en echelon fashion and continues north again to the Connecticut-Massachusetts stateline as the Snake Meadow Brook [SMB] fault (Figure 4) (Altamura, 1987; Connecticut Geological and Natural History Survey, 1990). The LH and the SMB fault zones have been proposed as en echelon components of a

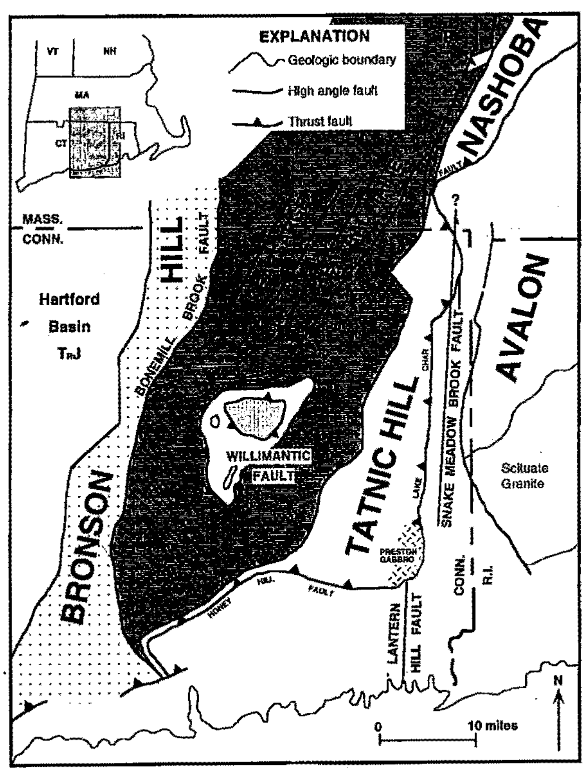

Figure 4. Terrane map of eastern Connecticut and adjacent areas in Massachusetts and Rhode Island. Note the location of the Lantern Hill fault with respect to the terrane boundary between Avalon and Tatnic Hill - Nashoba belt (Gander terrane). The Lantern Hill lode occurs at the interesection of the Lantern Hill fault with this boundary, the Honey Hill fault.

regional fault system that cuts the entire state of Connecticut from south to north (Altamura, 1987), a total distance of some 45 miles. To the south, the LH fault zone may be extended to correlate with Mesozoic basins of the continental shelf of the modern Atlantic Ocean that have been interpreted from geophysical data (Hutchinson et al., 1986). A projection of the SMB fault northward into Massachusetts is on strike with the silicified Wekepeke fault and may represent a further northward continuation of the LH-SMB fault system.

Over most of its strike length the north-south trending Lantern Hill fault is expressed topographically as a valley and displaces the east-west trending tectonic grain of the Avalonian anticlinorium. Displacement of these orthogonally-intersected units by the Lantern Hill fault includes a left lateral component of 0.4 km. To the south, the Lantern Hill fault is mapped striking into Long Island Sound (Figures 1 and 4). At its northern end, 400 m north of Lantern Hill, near North Stonington, Connecticut, this fault displaces the NNW-dipping (243°-24°) Late Paleozoic Honey Hill fault [HHF] that has been correlated with the Lake Char fault by Rodgers (1985). From here the LH fault is difficult to trace northward across the HHF into the upper plate, at the location of the main body of the Silurian (Dixson, 1982) Preston Gabbro (see Figure 4). The Honey Hill fault is characterized by mechanically tough Paleozoic mylonites and ultramylonites that may have, in combination with the Preston Gabbro, acted as a resistant boundary, forcing the eastward, en echelon side-step of the Lantern Hill fault to the SMB fault (Altamura, 1987). The Snake Meadow Brook fault has the same trend as the Lantern Hill fault (Figure 4) and is silicified over much of its length. Indeed quartz veins that may be related to the Lantern Hill lode are found adjacent to mylonites along the trace of the Lake Char fault zone (i.e. Honey Hill fault) and suggest localized precipitation of quartz along it.

Figure 5. Ductile folds preserved by relic foliation in metasomatized wall rocks within the fault zone. Similar ductile folds are found in metasomatic blocks and horses elsewhere within the quartz lode, but are not found in host rocks outside of the fault zone. Fold axes plunge moderately to steeply to the west and southwest. Deformation within the ductile regime occurred prior to silicification.

Figure 6. Veining and replacement processes were the two modes of emplacement of silica into the Lantern Hill fault zone. Shown are veins and fractures associated with brittle deformation within the fault zone. The cliff is about 60 feet high. Veins of the complex are cut by brittle faults and joints that are in turn cut by late-stage quartz veins.

A 1.6 km-long giant quartz lode is located within the Lantern Hill fault beneath this intersection with the mylonitized boundary between the Preston Gabbro (Gander terrane) and Avalonian units. The NNE-trending Lantern Hill giant quartz lode is approximately 71 m thick and flares significantly at its northern extent near the fault contact with the upper plate rocks (Rodgers, 1985).

The type locality for the Lantern Hill fault zone is on Lantern Hill in North Stonington, 7 kilometers north of the village of Old Mystic, Connecticut, in the Old Mystic, Connecticut, 7.5 minute quadrangle. Lantern Hill is the largest of four prominent hills (Figure 3) that are underlain by a quartz vein complex referred to as the Lantern Hill giant quartz lode (Altamura, 1987; Altamura and Gold, 1993a; 1993b). The Lantern Hill lode is by far the largest silicified zone along the trace of the LH-SMB fault system, although both faults are characterized by the presence of hydrothermal quartz (Altamura, 1987). The Lantern Hill quartz lode consists of a sheeted vein complex of relatively homogeneous and milky white quartz. The lode is approximately 1.6 km long and 60 meters thick (U.S. Silica Company, unpublished data). The second and southernmost of the two largest hills that make up the lode is locally known as Long Hill (Figure 3)(NB, the knobs of Long and Lantern hills are combined under the label Lantern Hill on U.S.G.S. 7.5 minute topographic map).

Several phases of deformation are recorded in exposures throughout and adjacent to the lode and to the south along the Lantern Hill fault zone. Preserved in the bedrock is a record that includes deformation in both ductile (Figure 5) and brittle (Figure 6) regimes. On the basis of his quadrangle-scale bedrock mapping, Goldsmith (1985) proposed that displacement along the Lantern Hill fault increases from south to north. He inferred stratigraphic throw (vertical displacement) of approximately 100 m in the Mystic River area some 10 km south of Lantern Hill. To the north at Lantern Hill he suggests approximately 420 m of stratigraphic throw. But these estimates are only approximate, because lithologic contacts of the moderately dipping host metamorphic formations (i.e. marker units) on either side of the fault are difficult to place accurately. In addition the attitude of the LH fault was not constrained by Goldsmith. Apparently he considered the fault to be subvertical. Interpretation of drill data supplied to the author by U.S. Silica Company suggests that the fault zone, defined by the relatively planar lode, may not be vertical at Lantern Hill, but dips to the west at an average dip of 51°. It should be noted that all speculations on movement are based on separations, not slip.

Goldsmith (1985) indicates a west-down sense of displacement for the LH fault, a sense based on normal dip-slip motion on a fault that lies a kilometer or so to the east of Lantern Hill. This is probably a far too simple story, as an analysis of mesoscopic-scale fabrics exposed in the Lantern Hill quartz lode and surrounding area indicate a complicated kinematic history that includes dip-slip (normal and reverse) as well as strike-slip (sinistral and dextral) faulting episodes.

The primary goal of the tectonic phase of this investigation of the LH fault zone was to discern the multiple stress and motion indicators in a time sequence for the critical zones through analysis of veins, mesoscopic-scale faults, and joints for kinematic and stress indicators. In addition to the LH fault, five faults in southern New England have experienced similar pervasive quartz mineralization (Figure 1). These include those along the eastern border fault of the Hartford rift basin and its northward continuation (e.g. the North Branford and Manchester (Connecticut) and Mine Ledge (New Hampshire) quartz lodes respectively) and the western border fault of the Narragansett basin (Diamond Hill lode). The Mount Hope quartz lode near Bristol, Rhode Island, although not associated with a mapped fault, has strong similarities to members of this list. In addition, in New Hampshire, the Pinnacle fault zone, Gaza fault, Lantern Hill fault (yes another one), Silver Lake fault, Ammonoosic fault zone, and minor faults are reported to have experienced pervasive silica mineralization (Robinson, 1988). Only the geology of the Lantern Hill fault zone (Connecticut) will be examined for this field trip.

The Lantern Hill Quartz Lode

Natural cliffs of the Lantern Hill quartz lode and quarry exposures provide a fortuitous opportunity to collect structural data in the heart of the Lantern Hill fault zone, where elsewhere along strike glacial drift covers all but the margins of the zone. Differential weathering between the quartz lode and the surrounding schists and gneisses has resulted in approximately 119 m of relief to yield the highest elevation along Connecticut's coastal slope.

The main silica ore zone, as exposed in the quarry, is 70 m thick, and is composed of sheeted vertical veins that range from 10 cm to 60 cm thick. Most of these veins, and minor fault planes that bound and cut veins, trend north to northeast. Veins of the lode are generally oriented at a slight oblique angle to the trend of the LH fault valley although a number of veins are parallel to the trend of the fault valley (Figure 7) (Altamura, 1987). Other minor veins are present in stockwork patterns within the metasomatized wall rock. Nine generations of quartz veins throughout the lode have been distinguished from cross-cutting relationships.

Minor subvertical faults occur within the lode and in the adjacent metamorphic rocks on either side of the Lantern Hill fault. The majority of these trend north-northeast and dip steeply east or west. Slickenlines indicate principally vertical dip-slip

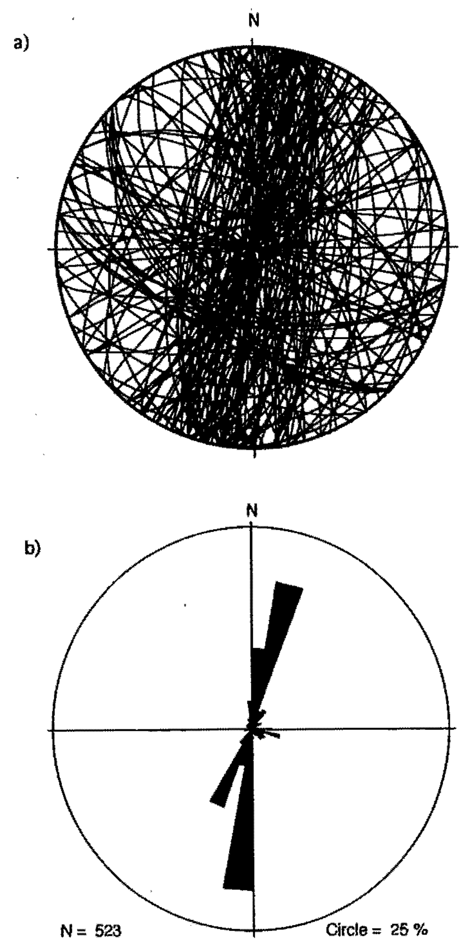

Figure 7. a) Equal area stereogram depicting veins as great circles on the lower hemisphere. Note the vast majority of veins trend NNE and are steeply dipping. Veins with other orientations are common in the stock-work veins that make up the wall rock; b) rose diagram representing vein attitudes. Note that petals contain a right-hand rule component.

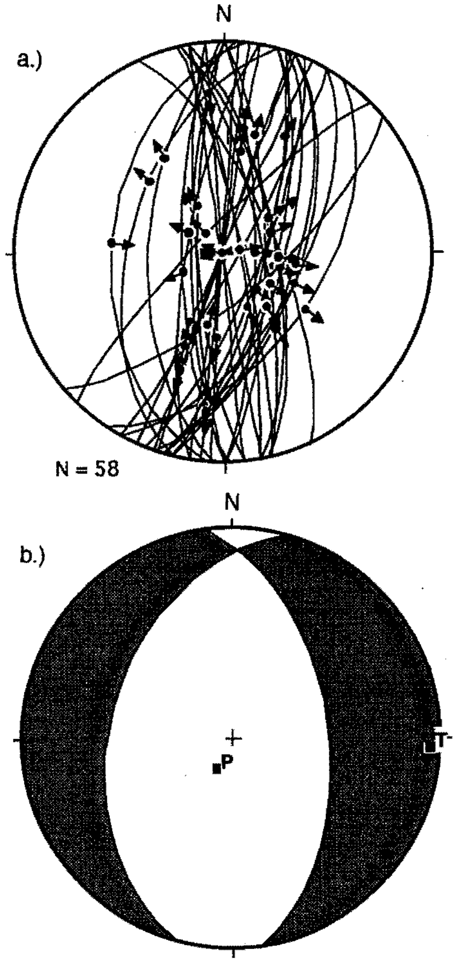

Figure 8. a) Equal area stereogram representing fault planes from the Lantern Hill fault zone as great circles on the lower hemisphere. Slip vectors for faults are indicated by the arrows. The population is dominated by normal faults, although a few reverse and strike-slip faults were found. The dominant trend is NNE and dips are moderate to steep; b) Equal area stereogram showing the fault plane solution for Lantern Hill faults represented above. The principal compressive stress orientation (P) is steeply dipping to the SSW, and the orientation of maximum tension (T) is horizontal and ESE.

(normal) motion (Figure 8). However, evidence of strike-slip movements after the main stage ore formation occurs within the Lantern Hill lode. Overall stress release for the fault system, as indicated from vein and fault attitudes with consideration to the c-axes of quartz crystals lining veins and minor normal faults, was roughly WNW-ESE (280°-100°) during a period of regional vertical uplift and subhorizontal extension.

Lack of exposures make it difficult to determine the attitude of the LH fault. However, interpretation of drilling data for the lode that were obtained from the U.S. Silica Company suggests that the dip of this silica ore body, and therefore the fault zone, may be between 40° and 75° to the west. The fault dip appears to be steepest (75°) at the southernmost extent of the lode and shallowest (40°) to the north (Blodgett, U.S. Silica mine manager, pers. comm., 1986). Net horizontal displacement along the LH fault includes 700 m of a sinistral strike-slip component and an apparent rotation of dip of metamorphic units of some 15° between the hanging wall and the footwall. The steeper dip occurs in the footwall or eastern block. A representative cross-section across the silicified zone is provided in Figure 9.

An early stage of ductile deformation preceded an overprinting by brittle and extensional deformation associated with emplacement of silica. The contemporaneity of muscovite $^{40}Ar/^{39}Ar$ cooling ages on either side of the Lantern Hill fault zone may indicate that its main displacement occurred under ductile conditions prior to the closure of argon blocking temperatures, or that late faulting was dominated by strike-slip motion. Alternatively, the amount of dip-slip faulting during brittle faulting (i.e. that below the closure temperature of argon in muscovite) may not have been sufficient to displace $^{40}Ar/^{39}Ar$ isochrons with error bars of ± 1 - 2 Ma. A significant net dip-slip component to the Lantern Hill fault during brittle deformation is not indicated.

Close spatial relationships between quartz veins and mesoscopic-scale faults of the Lantern Hill lode suggest that mineralization and faulting were coeval during the early Mesozoic and may have been responsible for activity on the Lake Char/ Honey Hill fault. Temperatures of homogenization obtained from fluid inclusions from earliest veins to latest veins that make up the complex range from 360°C to 165°C, respectively. No evidence of boiling of hydrothermal fluid was found in the more than 40 specimens analyzed for fluid inclusions. Assuming a geothermal gradient of 40°C/km at the time of mineralization, these data suggest that the depth of vein formation and contemporaneous faulting may have been between 4 and 8 km. From study of

Avalonian metamorphic units elsewhere in Connecticut, Wintsch et al. (1992) suggests that at 238 Ma, wall rock of the Lantern Hill lode may have been at about 5 km depth and a pressure of 180 MPa.

If so, then 5 km has been eroded to form the current exposures. Projecting the Paleozoic Honey Hill fault to the south above the Lantern Hill lode results in a reconstruction to 238 Ma and positions the fault as a plane about 2,000 feet above the highest elevation of Lantern Hill today. This reconstruction utilizes a representative fault dip for the Honey Hill of 35°, taken from local regional foliation. A considerable thickness of Gander terrane (including perhaps the Preston Gabbro) probably existed above the lode at 238 Ma. The intersection of the high-angle Lantern Hill fault with the Honey Hill fault results in an intersection lineation that is thought to have been an avenue of highest fluid mobility. Projecting the lode to its intersection with that projected surface results in a flared juncture similar to that boundary as mapped by Rodgers (1985). I envisage the original shape of the lode to have been that of an inverted "New Jersey" road barrier with a pinched keel and flaring at the top (Figure 10).

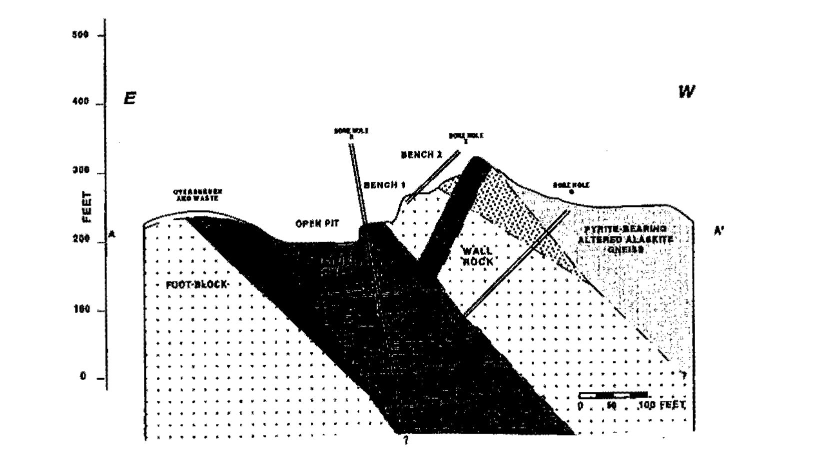

Figure 9. Representative cross-section across the Lantern Hill fault zone at Long Hill near North Stonington, Connecticut. The cross section was prepared using 1:1,200 scale lithologic map data and bore hole logs provided by the U.S. Silica Company. Dark shades used for the cross-section indicate zones of highest silicification; stipple patterns are used to represent unaltered to altered wall rock. Note the zone indicated as pyrite-bearing.

Hydrothermal-fluid migration probably was constrained by the paucity of fractures that propagated across less porous higher yield strength mylonite of the Honey Hill fault. The thickening of the lode is interpreted to represent ponding of ascending hydrothermal fluids at this boundary.

Only minor quartz veins, similar in composition and silica replacement attributed to hydrothermal activity associated with the Lantern Hill lode, occur locally within the "Gabbro" (Loughlin, 1912; Sclar, 1958; Goldsmith, 1985) and along the northward continuation of the Honey Hill fault, the Lake Char fault (Dixon, 1982; Goldsmith, 1985). From these observations I interpret that the Lantern Hill fault zone continues beneath the Preston Gabbro following the mega-intersection lineation of the Lantern Hill

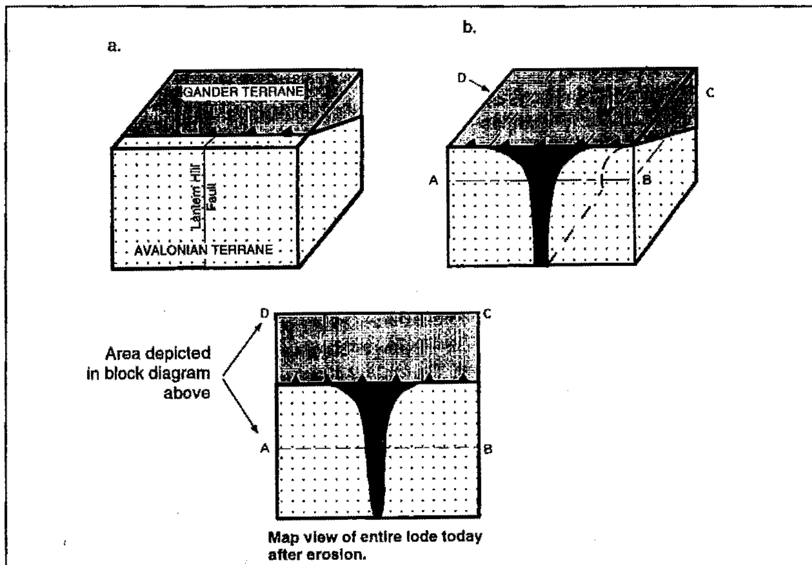

Figure 10. Block diagrams of the Lantern Hill fault a) prior to, and b) after mineralization, and cartoon geologic map showing the geometric relationships of the silicified zone (shaded pattern) with respect to the Avalonian and Gander terranes today. Ascending silica-bearing fluids are envisaged to have ponded at the terrane boundary resulting in a quartz body with the shape of an inverted New Jersey barrier. It is speculated that the Lantern Hill lode continues northward under the Honey Hill fault zone.

fault with the Honey Hill fault.

Quartz precipitated from aqueous fluids preserved in fluid inclusions that provide temperatures of homogenization from 340°C to 165°C, cooling through time. Fluid flow was largely controlled by a fracture system dissecting high-grade metamorphic rocks in a failed rift setting. Fluids migrated along the Lantern Hill fault zone, entering wall rocks through fractures and along grain boundaries. Vein formation and metasomatic alteration of wall rocks occurred. There is an incompatibility between the very "clean" aqueous fluids trapped in fluid inclusions (0.2 wt. % NaCl equivalent) and apparent reactivity of fluids as indicated from wall-rock alteration.

Based on solubility of quartz constrained by T_h of 340-165°C and overall salinity determinations for all quartz of the lode, an estimated 1.3 x 10^9 km^3 of aqueous fluid flowed through the fracture system to result in the volume of quartz equivalent to proven reserves (Altamura and Gold, 1994). This value is only one order of magnitude less than all the water in the Earth's oceans, and it seems likely that a circulation system that recycled hydrothermal fluid must have been in operation. In such a system, silica-bearing fluids would have to migrate up and along the strike of the Lantern Hill fault zone, precipitating quartz. In such a planar circulation system, recharge could have been either north or south of Lantern Hill - as well as beneath it. The heat source responsible for this geochemical system is speculated to have been a mantle diapir related to changing tectonic conditions as Alleghenian compression was replaced by Newarkian extension.

The mechanism of silica precipitation could have been due either to drop in temperature or pressure of silica-bearing fluids (Ohmoto, pers. comm., 1991). The presence and dominance of rather large common quartz crystals within the lode might suggest that it is unlikely that temperature dropped rapidly. With the propagation of faults, pressure would have dropped in pulses, yet temperature could have been held relatively constant in the system for long durations of tectonic activity. As the fault zone became more permeable, a change from lithostatic pressure to hydrostatic pressure could have occurred.

Evidence for a change from lithostatic to hydrostatic conditions may be recorded in the fluid inclusions in quartz that show a drop in temperature of ~100°C for main-stage silica mineralization, suggesting that mixing of cooler, perhaps meteoric fluids, entered the fault system at that time. Nine dominant vein sets make up the lode. Fluid inclusions from the oldest vein sets (QV_1-QV_5) record highest T_h and youngest veins (QV_6-QV_9) record the lowest.

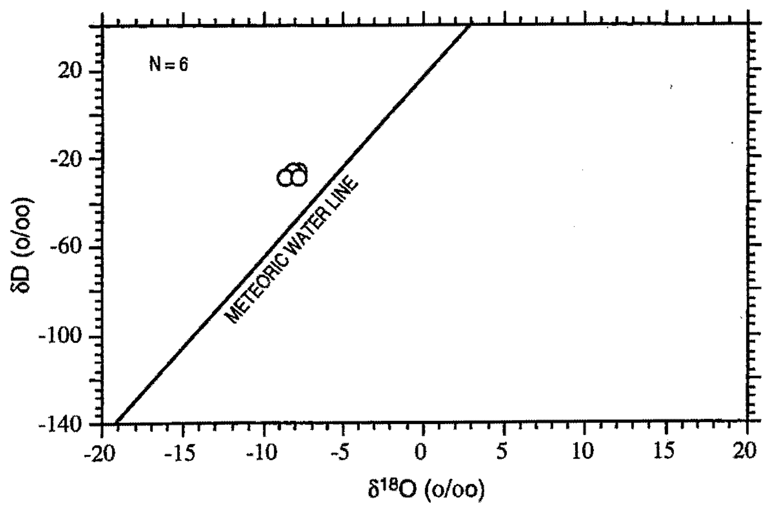

Figure 11. dD-$d^{18}O$ characteristics of aqueous fluid within primary fluid inclusions in a late-stage quartz vein [QV_8] of the Lantern Hill quartz lode, North Stonington, Connecticut. Note that data plot very close to the meteoric water line and indicate that meteoric water was a significant component in late-stage hydrothermal fluids. Fractionation of oxygen isotopes between quartz host (+11 o/oo) and fluid inclusion water was probably responsible for moving the $d^{18}O$ of fluid inclusion water off of the meteoric water line.

If this indeed was the history of the hydrothermal fluids, then stable isotope information (e.g. $d^{18}O$ and dD) may shed some light on this problem. Preliminary $d^{18}O$ and dD data of fluid inclusion waters from a late-stage vein [QV_8] suggest a meteoric water contribution, as isotopic ratios are similar to meteoric water and fall nearly on the meteoric water line on a discriminatory plot (Figure 11). A detailed stable isotope investigation is deferred to a future investigation. The mechanism for precipitation of silica at the Lantern Hill lode may not only have been a consequence of a drop in pressure, but also a drop in temperature.

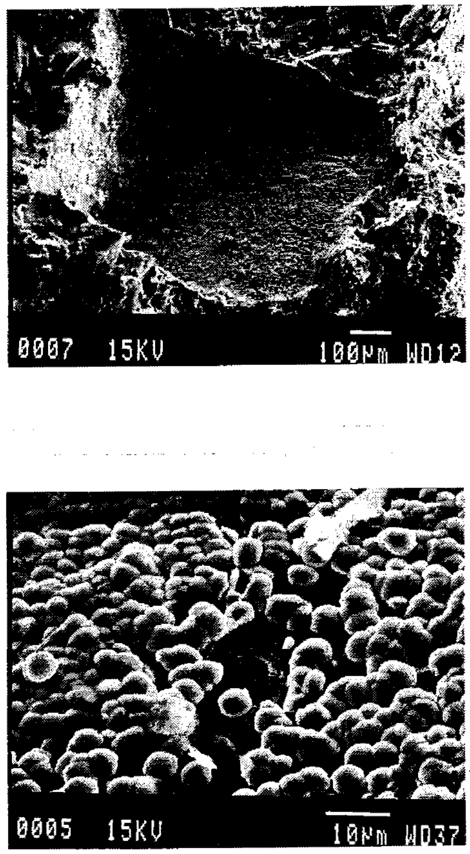

Figure 12. a) Cubic mold relic of pyrite in wall rock of the Lantern Hill lode. b) Walls of cubic mold are lined with a botryoidal iron oxide-hydroxide mineral, perhaps goethite. Matrix is altered alaskite.

Metasomatic Alteration

The Lantern Hill quartz lode transgresses Upper Proterozoic "basement" rocks of the Avalonian terrane that here includes felsic and mafic gneisses, mica schist and quartzite. Based on 1:1,200-scale mapping and core logs for shallow bore holes, an estimated 31 million cubic yards of hydrothermal quartz were emplaced along the fault zone and resulted in the Lantern Hill lode. Enormous volumes of fluid must have circulated through the fault zone at this time and reacted with wall rocks. Some of this quartz may represent a product of alteration of wall-rock feldspars; released K_2O could be the source of hydrothermal muscovites that have been analyzed for radiogenic isotopes as part of the age dating investigation of the Lantern Hill fault (e.g. Altamura, 1987; Altamura and Lux, 1994). The most significant alteration was to the Hope Valley Alaskite Gneiss that makes up much of the country rock in the study area (Goldsmith, 1985).

Hydrothermal quartz precipitated from aqueous fluids that ranged in temperature from about 370°C to 180°C, assuming hydrostatic conditions (435°C to 205°C if lithostatic conditions), apparently cooling through time. Fluid flow was largely controlled by a fracture system dissecting high-grade metamorphic rocks in a failed rift setting. Fluids migrated along the fracture zone, entering wall rocks through fractures and along grain boundaries. A transect from unaltered alaskite through metasomatized alaskite into the ore zone, shows the following key assemblages progressively veinward: 1) magnetite-bearing alaskite; 2) metasomatized alaskite with sericite, kaolinite, and pyrite; 3) metasomatized alaskite with sericite, kaolinite, and geothite pseudomorphs after pyrite, and 4) quartz veins.

Broad zonation halos and selective alteration along fractures suggest that within a planar zone of weakness, H_2S-bearing solutions permeated and altered country rock. Alaskite gneiss reacted with feldspars and magnetite to produce, respectively, sericite and pyrite. Reactions that produced sericite from microcline and albite are a potential source of some lode silica. Volumetric calculations indicate insufficient silica from this mechanism and the necessity for an external supplemental source of silica (Altamura and Gold, 1993c).

The mineralogical changes during alteration were largely transformation of the feldspars to clay minerals (mixed layer Na-smectite(?)/illite), and impregnation of the rock with quartz. These changes are clearly reflected in the losses and gains of elements calculated. The quartz precipitation is shown by the gain of a considerable amount of Si, and formation of sericite (clay minerals) by the increase in the concentration of water. During alteration of the feldspars, most of the Ca, Mg and Sr disappeared, but more than half of the Na, K, Rb, Ba and Y was retained, presumably in the clay minerals. A substantial loss of Fe^{2+} occurred during the alteration, as magnetite ($FeO.Fe_2O_3$) was altered to early hydrothermal pyrite and then to iron oxides and hydroxides during a later oxidizing phase, as recorded by goethite(?) pseudomorphs after pyrite (Figure 12). Overall, the solutions causing the metasomatic alteration were hydrous and carried large amounts of dissolved silica, but very little else (Altamura, 1994, in preparation).

Gold

Giant quartz lodes can be thought of as two types: giant quartz veins and giant gold-quartz veins. Wall-rock alteration adjacent to giant quartz veins is generally sericitic and wall rock alteration adjacent to giant gold-quartz veins commonly is propylitic (Hemley and Jones, 1964). Wall-rock alteration adjacent the Lantern Hill lode is sericitic and is essentially "barren" with regard to Au (.015 ppm) (Altamura and Gold, 1994). The cost of mining and processing the low concentrations of Lantern Hill gold would not be economic.

Kinematic Model for the Lantern Hill fault

Hydrothermal fluids entering the Lantern Hill fault zone during Middle Triassic deformation would have lowered the mean stress on the fault system. P_{fluid} in the LH fault zone at the time of silicification of the zone may have ranged from hydrostatic to lithostatic pressure. Any fluid pressure would have counteracted normal stress on the fault zone allowing for a greater likelihood of brittle failure.

At the time of faulting a decrease in P_{fluid} on the system would have occurred. Brecciation and fracturing due to faulting would have increased the permeability in the fracture zone, allowing for mobility of silica-bearing hydrothermal fluids into the fault zone. In addition there would be an increased likelihood of dissolving of finely crushed quartz. Episodic decrease of pressure is one of the two possible mechanisms (NB, episodic temperature changes is the other) by which multiple generations of silica may have precipitated within the lode. Precipitation of quartz may have aided in reducing the permeability of the fracture zone, allowing for the build up of P_{fluid} to

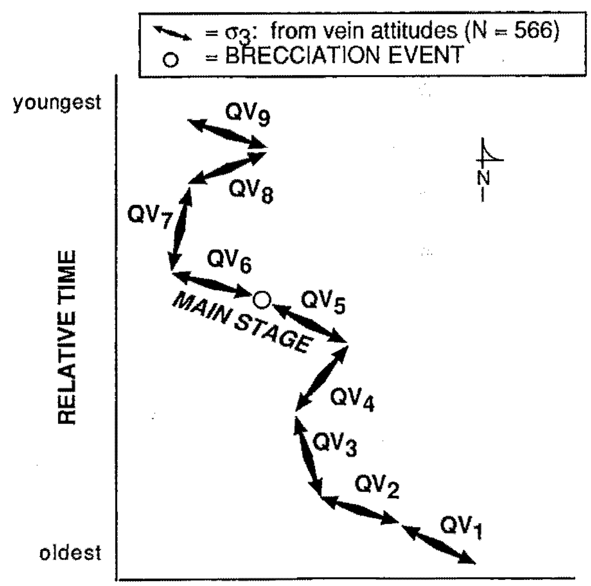

Figure 13. Chronologic extensional stress summary for quartz veins [QV] of the Lantern Hill fault zone, described in text. Note the pre-main-stage brecciation event that affects only the earliest veins.

again occur, leading to a stress state and fluid conditions whereby the process could repeat itself. Crack-seal and stockwork veins, faulted and brecciated veins, and silicified breccia in the Lantern Hill lode may represent the evidence of this interaction of hydrothermal fluids and tectonic activity. Activity along the Lantern Hill fault may have been associated with reactivation of the Lake Char/Honey Hill fault at this time.

Timing of Brittle-Regime Deformation and Silicification

The Lantern Hill quartz lode is a complex, massive, sheeted vein-network confined within a N-S trending fault zone. Silica was deposited by replacement and by filling (e.g., veins). The origin of the dissolved silica is hypothesized to be a product of hydrolysis of feldspar in wall rock along the length and depth of the Lantern Hill fault zone. Silica migrated up and along strike to a site of deposition near a major terrane boundary (Gander: upper plate/Avalon: lower plate), the Lake Char/Honey Hill fault zone (Figure 10). Little hydrothermal quartz is present in the upper plate.

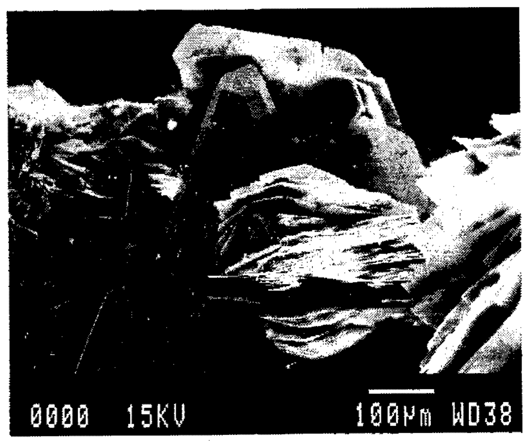

Figure 14. Scanning electron microscope photomicrograph of intergrown euhedral quartz and muscovite crystals lining a vug within a main-stage quartz vein (QV_6) of the Lantern Hill quartz lode. Muscovite yielded a 238 Ma plateau age and fluid inclusions in quartz yielded mean temperatures of homogenization of 215°C (see text for discussion). (Photo by H.E. Belkin, U.S. Geological Survey)

Nine vein sets (QV_1-QV_9), delineated on the basis of cross-cutting relationships, account for most of the silica within the lode and stockwork veins in wall rock. Veins vary in orientation, thickness, texture and mineralogy. Where quartz crystal c-axes can be observed, they are normal to the veins, and s_3 orientation is considered to be perpendicular to quartz veins. QV_5 and QV_6 represent the main stage of silica emplacement. A chronodynamic model for these veins is portrayed in Figure 13.

The oldest veins (QV_1-QV_5) are characterized by primary fluid inclusions with a mean temperature of homogenization (T_h) of 340°C. The T_h for QV_6 is 215°C and is 165°C for late-stage veins. A temperature correction due to pressure of formation, based on estimated depth of emplacement, causes an upward revision of the highest temperature of the hydrothermal fluids to 370-435°C depending on whether hydrostatic or lithostatic conditions prevailed. Main- and late-stage veins are revised to 230-265°C and 180-205°C respectively. No evidence for boiling was encountered.

Metamorphic quartz and muscovite (locally) of the Hope Valley Alaskite Gneiss [HVAG] are preserved within the metasomatic aureole, but feldspars were altered to (?) mixed layer illite and magnetite to pyrite/pyrrhotite. Illite yielded a plateau age of 236.4±2.6 Ma, whereas metamorphic muscovites within the metasomatic zone yielded ages of 255 Ma (see Altamura and Lux, 1994). Metamorphic muscovites in HVAG from a site ~1 km from the quartz lode yielded 255 Ma ages (Altamura and Lux, 1994).

Illitization of the feldspar within the quartz lode is inferred to have occurred at the time of silica mineralization, as illite and main-stage hydrothermal muscovite (Figure 14) yield the same age (~238 Ma). The small grain size of illite may have permitted some diffusion of argon after initial crystallization of this mineral, but because its plateau age is similar to that of larger diameter hydrothermal muscovites, its closure with respect to argon must have occurred over a short period of time, if not at the time of its crystallization.

Primary muscovite within metasomatically altered and unaltered HVAG ~1 km from the study area yield the same age (~255 Ma). Apparently exposure to hot hydrothermal fluids, on the order of 370°C and perhaps higher, was insufficient to reset primary muscovite within the metasomatic aureole (Altamura and Lux, 1994).

Hydrothermal muscovite in QV_6 is syntectonic and was emplaced along with quartz during brittle deformation along steeply dipping extensional fractures at 238 Ma. Since unaltered metamorphic muscovites in HVAG provide age determinations of ~255 Ma (NB, Alleghenian cooling ages), absolute ages of hydrothermal muscovites constrain both Lantern Hill mineralization and associated faulting. The age of silicification and coeval faulting was Early Middle Triassic (*Anisian*) - some 19 Ma prior to first sedimentation into the Hartford failed rift basin (see Cornet, 1977) some 35 miles to the west. Hydrothermal muscovite from quartz veins within the SMB fault have yielded a similar age (Altamura, 1987), suggesting a correlation with the Lantern Hill fault. The timing of silicification and coeval faulting along the Lantern Hill fault may be associated with earliest Newark rift tectonics associated with the embryonic opening of the Atlantic Ocean.

Economic Geology

Excavation of quartz from the Lantern Hill lode was underway at least by 1872 (Kimball, 1987). A small long-abandoned quarry can be found along Lantern Hill Road near Long Pond and numerous old prospect excavations can be found on Lantern Hill. As recently as 1994 the Lantern Hill lode at Long Hill was excavated in an open pit by the U.S. Silica Company of Berkeley Springs, VA, a division of Rio Tinto Zinc. Quarrying of the Lantern Hill lode was conducted to extract high-purity silica for such uses as foundry sand, aquarium sand, swimming pool filters, sandbox sand, and architectural aggregate (Harrison and Altamura, 1994). Lantern Hill quartz lode silica makes up the concrete faces of the J.F. Kennedy presidential library near Boston (Hunnisett, personal. communication, 1992). Lantern Hill silica has also been used for making glass (unpublished records of the U.S. Silica Co.).

Road Log

The region is still generally rural and has long been known for its beautiful scenery. Recent developments include a large casino and associated hi-rise hotel on the Mashantucket-Pequot Indian Reservation in Ledyard.

The strike length of the Lantern Hill fault is covered on the U.S.G.S. Old Mystic topographic map (1:24,000 scale), partially reproduced in Figure 15.

0.0 mi	Begin road log near the main entrance to parking lot of the Radisson Hotel and Conference Center, Cromwell, CT. Turn left onto Connecticut Route 372 east.
1.5	Turn left onto Connecticut Route 9 south.
26.6	Turn left onto Interstate 95 east.
47.7	Take exit 90. Turn left at bottom of ramp onto Connecticut Route 27 north.
49.1	We will stop in the village of Old Mystic (still on Route 27) to obtain provisions for lunch.

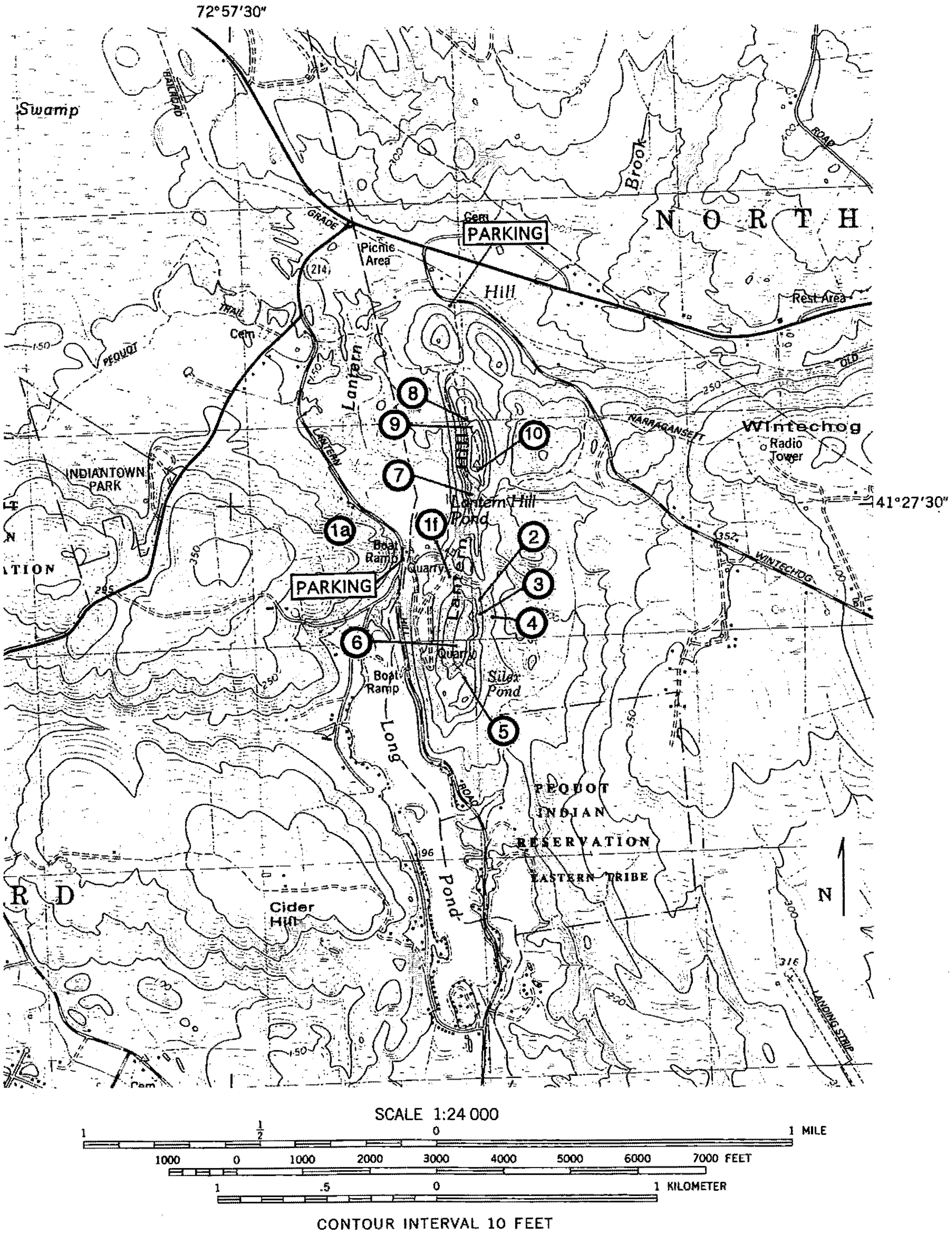

Figure 15. Base map of the Lantern Hill quartz lode study area, North Stonington, Connecticut, showing location of field-trip STOPS 1-10 and parking areas.

49.4	Turn right onto Connecticut Route 184 east.
50.1	Turn left onto Lantern Hill Road (note street sign).
54.1	Part I. Long Hill Excursion (STOPS 1-6 on Figure 15).

The entrance to the former U.S. Silica Company open-pit quarry and plant is on the right approximately 200 feet south of Lantern Hill Pond. Convenient parking for STOPS 1-7 is along Lantern Hill Road immediately south of the entrance gate

Permission to park and enter the property is needed. All field trip STOPS are on the property of the Mashantucket-Pequot tribe. David Holahan, Mashantucket-Pequot Reservation, P.O. Box 3060, Indiantown Road, Ledyard, CT, may be contacted for permission to enter the property.

STOP 1a - 1f. TRANSECT ACROSS THE METASOMATIC AUREOLE OF THE LANTERN HILL LODE

The 1.6 km long Lantern Hill lode comes in contact with several host rock units as it transgresses perpendicularly meta-igneous and meta-sedimentary units of the Avalonian terrane. The Hope Valley Alaskite Gneiss (~620 Ma, Proterozoic Z) is the dominant wall-rock unit adjacent to the fault. It can be traced along a series of outcrops, from unaltered rock outside the lode, to altered rock of the contact aureole, and into the core of the lode. The purpose of STOPS 1a - 1f is to follow the progression of metasomatic changes as one approaches the central core of the lode (Figure 16). This traverse is described in the text.

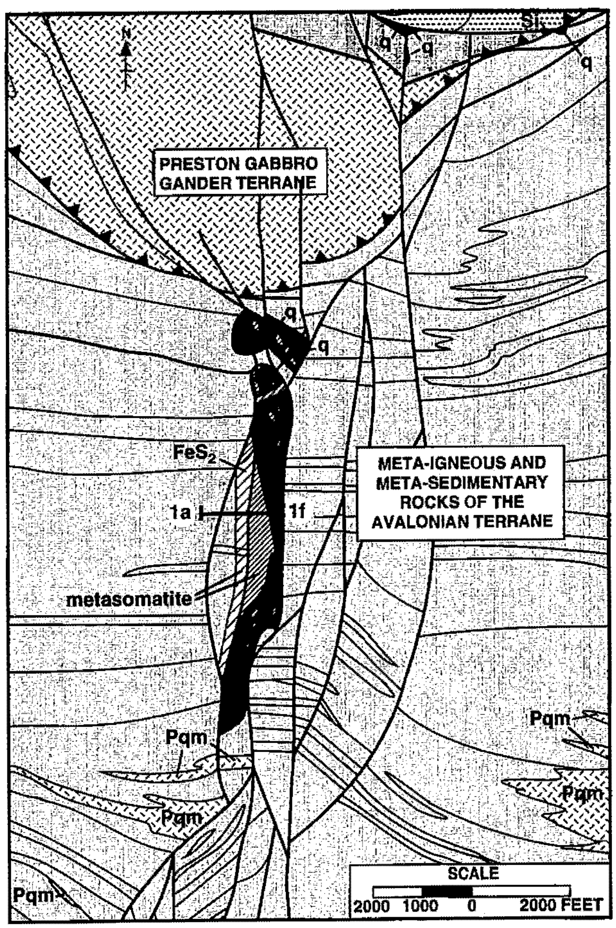

Figure 16. Generalized bedrock geological map of the study area showing fundamental geologic units of interest for the geochemistry and geochronology investigations. Note: q = quartzoze rock; Si = silicified rock, and Pqm = Permian quartz monzonite of Goldsmith (1985).

STOP 1a. UNALTERED ALASKITE

(Location: west side of Lantern Hill Road across from the entrance to the quarry and plant.)

Alaskite is used to designate an albite-rich granitic rock that contains but a few percent dark minerals. At this locality the Hope Valley Alaskite Gneiss (Rodgers, 1985) is a light-pink to gray, medium- to coarse-grained gneiss. It is composed of approximately equal amounts of quartz, alkali feldspar and plagioclase feldspar with trace amounts of magnetite. The alkali feldspar was identified by XRD to be microcline. The plagioclase was identified petrographically to be An_8. Locally muscovite is an accessory mineral in the alaskite.

Outcrops to the west, near the top of this hill, were sampled and used as the "anchor" (unaltered end member) for a geochemical traverse into the Lantern Hill lode (Altamura, 1994, in preparation). The Hope Valley Alaskite Gneiss here strikes east-west and dips north, and sampling was conducted parallel to the strike of the formation.

STOP 1b. ALTERED(?) ALASKITE

(Location: 300 feet west of STOP 1a on the opposite side of Lantern Hill Road in the loading dock area of the former ore-bagging plant.)

Alaskite gneiss at STOP 1b can easily be recognized in hand sample. Evidence for some alteration may be present in the form of "pinking" of the microcline and therefore the overall appearance of the gneiss.

The attitude of gneissic foliation at this stop is 310°-25°, but is somewhat variable. An orthogonal pair of joint sets is present (175°-68° and 270°-80°). The face of the outcrop reveals a fault that has an attitude of 205°-60° with slickenlines characterized by a rake of 70° south. This fault is inferred to be a normal fault based on tectonic setting.

STOP 1c. METASOMATIZED ALASKITE

(Location: approximately 250 feet east along the base of the same slope on which STOP 1b. occurs and adjacent to the ore loading building)

At STOP 1c both feldspars are significantly altered to clay minerals (kaolinite, illite and Na-montmorillonite (?)). Mineral cleavage in feldspars is noticeably less preserved. Magnetite present in the alaskite at STOP 1b is absent, and instead euhedral (cubic) pyrite is present. This is interpreted to indicate a sulfur fugacity in the ascending hydrothermal solutions sufficiently high to react with wall-rock magnetite in the alaskite to form pyrite and pyrrhotite. Comparison of total iron (as Fe_2O_3) contents between unaltered alaskite (STOP 1a) and its pyrite-bearing metasomatic equivalent here reveals an increase of 0.2 weight percent and may indicate that iron was introduced or mobile during metasomatism. The deposition of pyrite would require hydrothermal fluids of a reducing nature.

STOP 1d. METASOMATIZED ALASKITE AND "THE CRUSHER FAULT"

(Location: approximately 250 feet east of STOP 1c to an outcrop at the base of the primary crusher assembly that looms some 50 feet above.)

The alaskite here appears a bleached tan and gray color with brassy cubes of pyrite that are easy to find. Crystal faces are smooth and do not exhibit striations. Relic gneissosity is well preserved.

Joints exposed at this site define two (045°-90° and 295°-50°) dominant sets. Two small normal faults that are approximately 1 meter apart have an attitude of 336°-70° and dip-slip slickenlines. Faults are interpreted to be east-side down normal faults. These faults are considered to be late, as they are well-developed and are not significantly mineralized by either quartz or pyrite. They provide evidence for the existence of the regional east-down normal fault mapped by Goldsmith (1985).

STOP 1e. RUSTY-METASOMATIZED ALASKITE

(Location: approximately 150 feet east of STOP 1d "as the crow flies". However we need to ascend to reach the outcrop and will walk up the quarry access road that can be seen to the north from STOP 1d. This gravel road trends NNE for about 350 feet and then "switches back" to the south. Continue on this road about another 350 feet until the truck platform for the primary jaw crusher (the scaffolding is visible from below at the previous STOP) appears to the right of the road. STOP 1e is the outcrop across the quarry road to the east of the crusher.)

This outcrop is noticeably iron-stained. In freshly broken samples of metasomatite, cubic molds, now partially filled with a botryoidal iron mineral (goethite (?)) (Figure 12), are apparent. Unaltered pyrite is present in this outcrop as well.

Aureole chemistry

A transect from unaltered alaskite through metasomatized alaskite to the ore zone, illustrates the following key assemblages progressively veinward: 1) magnetite-bearing alaskite; 2) altered alaskite with sericite, kaolinite, and pyrite; 3) metasomatite with sericite, kaolinite, and geothite pseudomorphs after pyrite, and 4) quartz veins. Detailed sampling for geochemical analyses was conducted along this traverse and is reported in Altamura (1994b). The results of whole-rock major and trace element data acquisition reveal information concerning wall-rock alteration and element mobility (see Altamura, 1994b). K_2O, Na_2O, Rb, Sr, Ba, and Y show veinward losses; Si, S, H_2O (as LOI), Mn and V show veinward gains, and Al_2O_3, TiO_2, and Zr were relatively constant.

Both mass and volume increases of 24% occurred as a result of alteration, and wall rock was impregnated with silica. Alternatively some SiO_2 may represent alteration of wall-rock feldspars. Released K_2O could be the source of hydrothermal muscovites that have provided 238 Ma $^{40}Ar/^{39}Ar$ ages. There is an incompatibility between the very "clean" aqueous fluids trapped in fluid inclusions (0.2 wt. % NaCl equivalent) and apparent reactivity of fluids as indicated from wall-rock alteration. Based on solubilities constrained by fluid inclusions T°C and salinity determinations, an estimated 10^9 km^3 of aqueous fluid flowed through the fracture system to precipitate the volume of silica equivalent to proven reserves, using silica solubility drops of 300 ppm.

Veining

Gneissic layering of the Hope Valley Alaskite Gneiss remains preserved at this site as well as along the entire transect thus far. Stockwork-quartz veining is abundant at this STOP, especially nearest the core of the lode that is less than 50 feet away. This STOP is located on the hanging wall of the lode. Wall-rock quartz veins range in thickness from a few mm to ~ 1 meter. Veins vary in orientation and texture, but are composed almost exclusively of quartz, and variation in mineralogy is rare.

A classification scheme was developed to better understand these veins. This vein classification relies on cross-cutting relationships, orientation, vein thickness and texture as key discriminators. Nine (9) dominant vein sets were identified in this way.

The 9 vein sets (QV_1-QV_9) account for most of the veins of the lode and wall rocks. Figure 13 shows the change in orientation of vein sets with respect to relative time. The five oldest vein sets (QV_1-QV_5) have been recognized only in the metasomatite. QV_6 makes up the center of the lode and represents main-stage silicification. QV_7-QV_9 are post main stage and cut QV_6 and earlier veins.

The mean temperature of homogenization (T_h) for aqueous fluid inclusions of QV_6 is 215°C. A temperature correction due to the influence of paleopressure is based on an estimated depth of emplacement of 5 km, an estimate derived from data of Wintsch et al. (1992), Wintsch (pers. comm., 1994) and temperature-correction tables presented by Roedder (1984). This causes an upward revision of hydrothermal fluid temperature to 230-265°C, depending on whether hydrostatic or lithostatic conditions are assumed.

The oldest veins (QV_1-QV_5) are characterized by primary inclusions with a T_h of 340°C. Temperature correction to T_h due to pressure obtained for primary fluid inclusions results in trapping temperatures (T_t) of 370-435°C for hydrostatic and lithostatic conditions, respectively. Assuming that hydrothermal fluids entering the fault zone were nearly at thermal equilibrium with host rocks, then these early veins formed at a significantly earlier time and at a greater depth than main-stage veins. An estimated depth of emplacement of 7 km at 250 Ma is proposed based on *P-T-t* data for the Avalonian terrane (Wintsch et al., 1992).

An older vein [QV_0] that occurs furthest out in the aureole may be related to hydrothermal activity associated with the development of the lode. This quartz vein is about a meter across and composed of "bull" quartz with coarse-grained books of muscovite. This quartz contains two types of fluid inclusions: two-phase H_2O and three-phase CO_2-H_2O (Figure 17). The coexistence of aqueous and CO_2-bearing inclusions in the sample vein may represent evidence of immiscibility within precipitating solutions. Two-phase aqueous inclusions homogenized at 315°C. In three-phase inclusions, CO_2 vapor homogenizes with CO_2 liquid at 31°C and total homogenization occurs at 306°C. Note that these T_h have not been corrected for pressure. Based on *P-T-t* data, Wintsch et al. (1992) concluded that Avalonian terrane rocks of south-central Connecticut had cooled down from an Allegenian thermal event to about 350°C at 250 Ma. Muscovite in QV_0, dated by the $^{40}Ar/^{39}Ar$ method, yielded a 255 Ma age. This age determination represents an upper constraint for the age of this vein. QV_1-QV_5 and QV_0 appear to represent an early stage of silcification along the Lantern Hill fault.

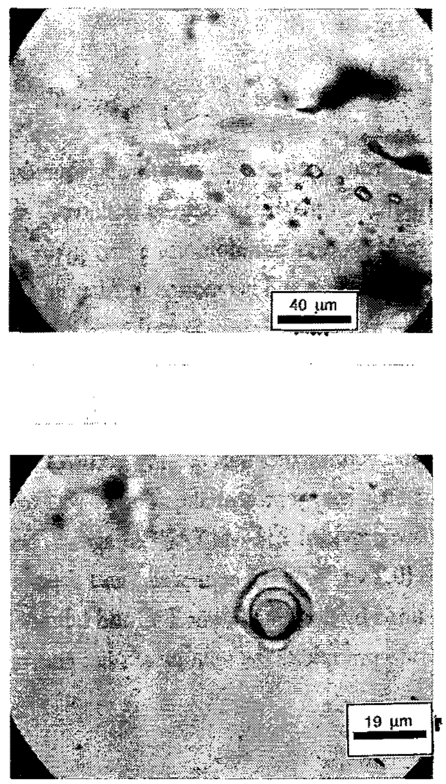

Figure 17. Photomicrographs showing two-phase H_2O and three-phase CO_2-H_2O primary fluid inclusions in a quartz vein [QV_0] in the hanging wall of the Lantern Hill giant quartz lode. a) Two-phase H_2O inclusions (small vapor bubble) and three-phase CO_2-H_2O inclusions (large vapor bubble) and b) Enlargement of three-phase CO_2-H_2O inclusion showing, from interior of inclusion outward: CO_2 vapor, CO_2 liquid, H_2O liquid and quartz matrix.

If QV_0 is related to hydrothermal activity associated with the formation of the lode, then it indicates that earliest hydrothermal solutions into the fault zone were CO_2-bearing, whereas later and ultimately colder solutions were characterized by aqueous fluids only. It seems plausible that CO_2 may have exsolved from later solutions as the fault system became more open. Country rock in contact with the Lantern Hill giant quartz lode is strongly sericitized and silicified, but lacks the propylitic alteration associated with giant gold-bearing quartz lodes.

Late-stage veins (QV_7-QV_9) were considerably colder than main-stage veins and are estimated to have been emplaced at a depth of 4 km. Temperature of crystallization was determined to be between 180 and 205°C, for hydrostatic and lithostatic conditions respectively.

QV_1-QV_5 are the oldest veins and were precipitated from hydrothermal fluids on the order of 370-435°C. QV_6 is part of main-stage silicification and formed from solutions that were 230-265°C. QV_7-QV_9 are the youngest veins and formed from hydrothermal fluids that were 180-205°C. Apparently veins formed from hydrothermal solutions that cooled through time. A significant temperature drop is indicated during main-stage silicification.

Wall-rock $^{40}Ar/^{39}Ar$ Study

The metasomatic aureole preserves metamorphic quartz and muscovite (locally), but feldspars are altered to sericite (e.g. illite) and magnetite to pyrite/pyrrhotite. Illite from the metasomatite provided a plateau age of 236.4±2.6 Ma. Metamorphic muscovites within the metasomatite yield ages of 255.7±1.9, 256.4±1.7 and 255.5±2.0 Ma. Metamorphic muscovites from a site in the host rocks ~1 km from the quartz lode yield the following ages: 255.1±2.0 and 253.9±2.0 Ma (Altamura and Lux, 1994; in preparation).

We interpret the sericitization of the feldspar in wall rocks of the quartz lode to have occurred during lode emplacment. Both illite and newly formed QV_6 muscovite yield the same age (~238 Ma) within the margin of analytical error (±~2 Ma). Primary muscovite within the altered host rock and in unaltered HVAG ~1 km from the lode yield the same age (~255 Ma). Apparently exposure to hot hydrothermal fluids that hydrated feldspar was insufficient to reset primary muscovite within the lode.

Because stockwork quartz veins (QV_1-QV_5) provide T_f of 370°C and perhaps higher, they are assigned an age of emplacement of 250 (perhaps 255) Ma or later when host rocks were about that temperature. The closure age for sericitization (i.e. illitization) as determined from the $^{40}Ar/^{39}Ar$ study is 238 Ma. If that age represents a crystallization age for the illite, then QV_{1-5} must be older, because pre-existing illite could not have been closed at a temperature of about 370°C. Early stockwork veining may not have been responsible for the final wall-rock alteration recorded by illite. It is unknown when sulfidation of wall rock occurred with respect to veining and wall-rock alteration.

STOP 1f. MAIN ORE ZONE OF THE LANTERN HILL LODE (99.45 wt. % SiO_2)

(Location: approximately 50 feet east and via the first quarry road to the north along the base of STOP 1e.)

The core of the lode is exposed in the south-facing wall of an early quarry. At this locality the lode consists of a set of sheeted NNE-trending and subvertical quartz veins (up to 1 m thick) that occur in a rose pattern and breccia layers. This site is the last STOP in the series that make up the metsomatic transect, albeit - we are no longer in alaskite metasomatite. At other locations in the core of the lode, a metamorphic fabric still is discernable in the alignment of green illite and chlorite. Lode emplacement occurred by vein filling and impregnation but also by replacement (STOP 3).

Rocks of the lode were subjected to coeval and late-stage faulting (see Figure 8). The lode is dominated by mesoscopic-scale normal faults but strike-slip and reverse faults can be found. Breccia (STOPS 3, 5 and 8) occurs in various degrees of lithification by hydrothermal silica and varies from nonpermeable massive breccia to partially silicified breccia to very friable breccia (i.e. "soft" quartz).

STOP 2. STOCKWORK QUARTZ VEINS, BENCH QUARRY WALLS

(Location: from STOP 1e. proceed south along the quarry road for 275 feet to junction with another quarry road that enters from the west. Continue up hill along the quarry road that winds toward the southeast, along the base of the outcrop, some 100 feet to a "Y" intersection in the road. The high road leads to Bench 2 and 3 of the open pit and the low road leads to Bench 1. Take the low-road for 900 feet to STOP 2.)

Stockwork veins and metasomatite are exposed in the east-facing quarry walls along the road side and in the most recently active quarry faces of Bench 1. Based on bore hole logs made available to the author by the U.S. Silica Company, the quartz lode dips steeply to the west, making these wall-rock exposures part of the hanging wall. At the base of the west wall is a N-S trending lithified breccia zone. Quartz breccia in the lode is variably iron-stained depending on degree of silicification of cataclasite. Less lithified and more permeable breccia is invariably more iron-stained, and is identified as the yellow breccia unit on the litho-structural map of the lode (Altamura, 1994b).

When the author last visited this site in 1993, two sinistral strike-slip faults were exposed immediately adjacent to the active quarry face: 007°-90° (rake = 25° south) and 010°-90° (rake = 27° south). The fault trending 007° contained a considerable amount of fault gouge (silica rock flour). A white rock flour was packed against eudehral stubby quartz crystals (2.5 cm in diameter) that were preserved in sags present along the exposed fault surface (Figure 18). Apparently unevenesses, or asperities, along the fault surface were ground into rock flour that was transported into the sags.

Figure 18. Faulted quartz vein (QV_6) exposed within Lantern Hill quartz lode. Quartz crystals are preserved within sags on the fault surface, but were destroyed on converging asperities. A 5 cm white paste of quartz fault gouge surrounds quartz euhedra.

Careful examination of preserved, approximately 5-cm, quartz crystals revealed that the up-side faces were encrusted with a film of cryptocrystalline bluish-green material. Whereas rock flour was unconsolidated and could be washed off the surface of quartz crystals, bluish-green material was firmly attached to crystal faces. Examination of the material using a scanning electron microscope revealed a well developed crystal structure (Figure 19). The material is a precipitate and not a particulate as was the rock flour. A qualitative X-ray spectrum, in combination with morphology and color indicates that this mineral is plumbogummite [$PbAl_3(PO_4)_2(OH)_5.H_2O$].

This faulted quartz vein is assigned to the main-stage of silica mineralization [QV_6]. Fluid inclusion analyses yielded a T_f of 230-265°C that is also characteristic of QV_6. Petrographic study revealed the presence of a birefringent daughter mineral within the fluid inclusions. Quartz crystals were cracked and fresh surfaces were analyzed using a scanning electron microscope. Some fluid inclusion chambers were fractured to gain access to the daughter material. The daughter material displays a platy hexagonal form (Figure 20). Qualitative X-ray data in combination with crystal morphology suggest that this mineral is kaolinite. Some alumina must have been mobile in hydrothermal fluids, but was not volumetrically significant.

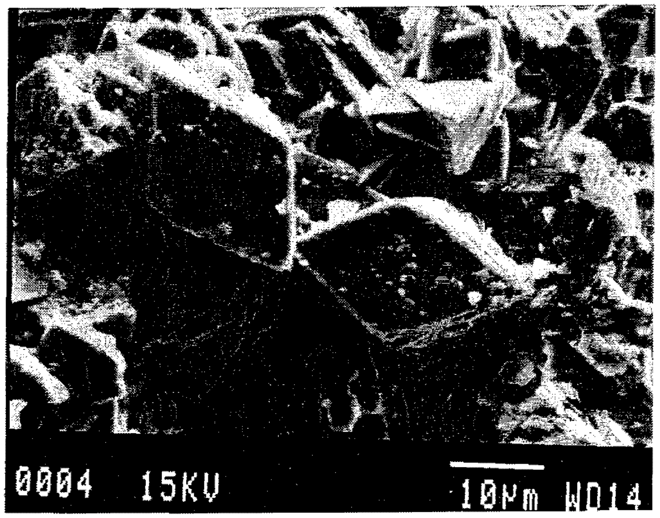

Figure 19. Scanning electron microscope photomicrograph of plumbogummite crystals from a quartz vein (QV_6) within the Lantern Hill quartz lode. Plumbogummite occurs as a crust on quartz euhedra.

The geological history recorded at this outcrop is: 1) fracturing parallel to the regional trend of the Lantern Hill fault; 2) vein filling [QV_6] at 238 Ma; 3) precipitation of plumbogummite at an elevation higher than the outcrop, from fluids of a markedly different chemistry than that from which quartz precipitated, and which flowed through open-pockets of the vein, and 4) faulting that granulated quartz-vein asperities into rock flour that collected within sags of the fault surface.

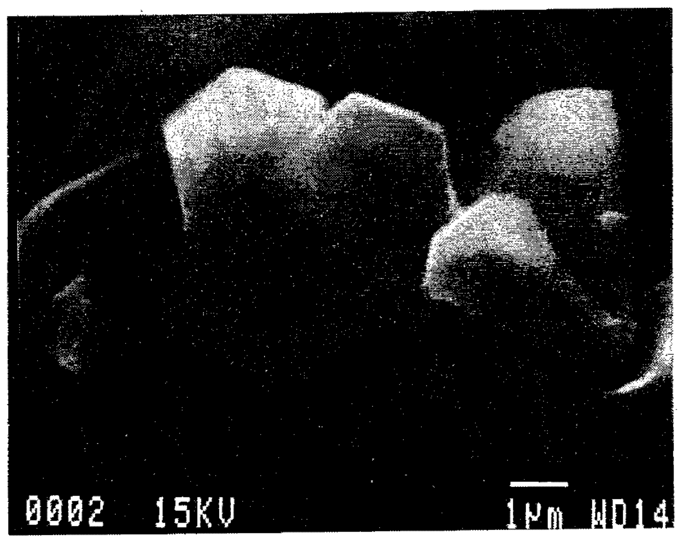

Figure 20. Scanning electron microscope photomicrograph of kaolinite daughter minerals within a broken fluid inclusion in quartz from the Lantern Hill quartz lode.

During quarry operation in 1994, excavation occurred simultaneously at this location and also within the open pit. Exposures in the quarry face at this locality consisted of much silica ore that was incompletely replaced, thereby containing subordinate green color impurities. Silica ore, nearby but from the level of the open pit, was of higher grade. Some was blended with ore from Bench 1 in order to meet specifications for high purity silica products.

At the mesoscopic or even megascopic scale, it is difficult to deduce whether the lode consists of massive veins or completely replaced wall-rock screens and horses. Comb structures are present locally and indicate open vein growth. Microscopically relic metamorphic quartz retains an undulous extinction under cross-polars, whereas hydrothermal (or replacement) quartz does not. The quartz lode was not subjected to significant post-emplacement metamorphism that might have affected main-stage silica deposition (equivalent to QV_6).

QV_6, impregnated silica and in situ replacement silica from the sericitization of wall -rock feldspars, represents the main-stage of silica mineralization at the Lantern Hill lode. The relative proportions of silica derived from one or more of these processes is unknown, but evidence for all is represented. Neither pyrite and pyrrhotite nor empty cubic molds are found within high-grade silica ore of the core of the lode. This observation may represent evidence that sulfur replacement of wall-rock was not part of main-stage silica mineralization. It may have been part of an earlier hydrothermal event that occurred under reducing conditions and a higher F_{S2}.

Hydrothermal activity associated with illitization of wall rock (238 Ma), perhaps accompanied by the alteration of pyrite and pyrrhotite to goethite(?), occurred later, perhaps under more oxidizing conditions. It is interesting to note that where illitization is pervasive, no unaltered pyrite or pyrrhotite is found. These sulphides are found in alaskite that is altered predominantly to kaolinite and Na-montmorillonite(?). Mesoscopically these two metasomatic varieties of the alaskite are easily distinguishable, due to the distinctly greenish color of illite-bearing metasomatite. Alternatively, the alteration of pyrite may have occurred more recently as the deposit rose into the zone of weathering.

STOP 3. LATE-STAGE CROSS-VEINS LINED WITH MEGACRYSTIC QUARTZ EUHEDRA

(Location: about 20 feet west of STOP 2, but directly above on Bench 2. Follow quarry road back to the north to the "Y" intersection in the road described earlier. Take the high road this time; hike uphill 430 feet from "Y" intersection of Bench 1 and Bench 2 to another "Y" branch in the road - this one with Bench 2 as the low road and Bench 3 as the high road. Take the high road for approximately 450 feet to STOP 3: west quarry wall.)

A single quartz vein that trends 340° and dips steeply easterly [QV_8] is exposed at this location. The vein is approximately 0.5 m thick and is typified by a planar vuggy character. Vugs are lined with hexagonal terminated crystals that range in length from 5 cm to 20 cm and in diameter from 1 cm to 4 cm. These are the largest crystals found anywhere within the Lantern Hill lode and are a distinguishing characteristic of QV_8. The prismatic habit of individual quartz crystals is distorted, probably due to interference with closely packed neighbors and yields surfaces with stubby crystal forms dominated by the pyramidal terminations. This particular vein can be traced throughout the quarry. Small caves (vugs), the largest approximately 2 m x 1 m x 1 m, can be seen on Bench 3 by looking up at the west quarry wall from bench 3 (Figure 21).

Figure 21. Outcrop on bench 3 of the former U.S. Silica open-pit quarry on Long Hill. Shown are crystal-lined vugs within a late-stage quartz vein [QV_8]. The common quartz crystals were removed by collectors shortly after this photograph was taken. Note: wooden grid is 10' x 10'.

Quartz of this vein is mostly milky and exhibits crystal zonation that is marked by an abundance of primary fluid inclusions. Fluid inclusions were analyzed and T_h of 165°C was obtained. Correction for paleopressure results in a T_t of 180-205°C for QV_8. Fluid inclusions are two-phase aqueous inclusions and salinity determinations are 0.45 weight percent equivalent NaCl. Observations on cracked crystals under the scanning electron microscope reveal a few small cubic crystals within broken fluid inclusion chambers and on immediately adjacent outside surfaces. These crystals may have been halite that precipitated upon release of the solution from its ancient tomb.

Fluid from these primary inclusions, collected from carefully selected and prepared specimens of QV_8, was extracted and oxygen and hydrogen isotopic ratios were measured. The isotopic signature of this fluid is similar to meteoric water (Figure 11). The $d^{18}O$ of inclusion fluids ranges from -7.8 to -8.6 o/oo and dO^{18} of host quartz [QV_8] ranges from +10.7 to 10.8 o/oo. dD of inclusion fluid ranges from -26 to -29 o/oo. A slight variation of isotopic ratios from meteoric water is hypothesized to have been caused by interaction of inclusion fluids with quartz making up the walls of the inclusion chamber.

The surface of the vein is iron-stained, perhaps evidence that these vugs are rather permeable and may serve as ground-water channelways. This would apply to any of the vuggy veins and permeable breccia zones. Ground-water flow would probably be directed in the orientation of veins and breccia zones (340° and 000° strikes), not dissimilar to the path that hydrothermal fluids took long ago.

STOP 4. FLOOR OF THE U.S. SILICA CO. QUARTZ QUARRY

(Location: from STOP 3, go 450 feet north on quarry road to the "Y" intersection of Bench 2 and 3; continue north 430 feet to the "Y" interesection of Bench 1 and 2. Continue north about 275 feet past the primary crusher and STOP 1e; take a right on the quarry road that leads to STOP 1f. Follow this road south downhill into the main pit for about 750 feet to the most recently excavated face.)

This site is in the core of the silica ore zone, where the highest grades of silica were obtained. Discrete veins are discernable in quarry walls or in blocks on the quarry floor. Much of the material appears massive and originated as thin veins or complete replacement of wall-rock screens and horses. Undulous extinction of relic metamorphic quartz is used to distinguish quartz of different origins. Muscovite intergrown with quartz from this site (Figure 14) provided a plateau age of 238 Ma. Rare green patches may represent evidence of wall-rock contamination in this part of the lode.

At the base of a segment of the west wall approximately 100 feet from the quarry face, an iron-stained granular texture zone is interpreted as a silicified breccia. A photomicrograph (Figure 22) illustrates the cataclastic nature of the rock.

Along the east wall some 400 feet north of the quarry face, a rock unique to the lode contains the record of a geological history that involves at least three distinct events. A dense, gray microcrystalline variety of quartz is exposed in the wall of the quarry at this site. A quartz vein lined with 1 cm long euhedral quartz crystals with hexagonal terminated tops was rooted to a cataclasite of quartz breccia. Druzzy quartz was surrounded by the microcrystalline quartz.

Figure 22. Cataclasized hydrothermal quartz from the Lantern Hill fault zone as shown in thin section under cross-polars. Length of photomicrograph is about 8 mm..

The geological history portrayed is one of hydrothermal quartz precipitation, followed by shearing, in turn followed by quartz precipitation of quartz crystals into fractured cataclasite, then precipitation of microcrystalline quartz. Both breccia fragments in the cataclasite and druzzy quartz euhedra contain aqueous fluid inclusions that provide T_h of 270°C.

Many N-S trending fractures can be seen in the quarry walls throughout the pit. Some of these are faults - mostly normal faults. Upon leaving the open pit, walk through an area bounded by two high walls, one making up the west wall and the other the east. Careful observation with the sun in the right position reveals that slickenlines occur on these surfaces. The eastern fault is dip-slip and has an attitude near the base of the high wall of 175°-89° with slickenlines that have a rake of 90°, but the surface is curviplanar and toward the top the fault plane dips steeply east. It is assumed to be a normal fault based on tectonic setting. The western fault has an attitude of 175°-79° and slickenlines that are subhorizontal. The sense of this fault is uncertain, but displacement of metamorphic units on either side of Lantern Hill require a component of sinistral strike-slip faulting. It may be that this is one such fault.

STOP 5. OLD SILEX MINE, SOUTH END OF LONG HILL

(Location: 1,000 feet south of STOP 4, following the abandoned quarry road on the eastern side of Bench 1.)

A considerable amount of quartz was quarried at this site. Silex, a finely ground form of quartz for use as filler, was excavated here long ago, and this quarry has been referred to as the Silex mine (Loughlin, 1912). Loughlin suggests that this site was actively excavated in 1905, and records of the U.S. Silica Co. indicate that excavation also occurred at this quarry around the time of World War II.

A considerable amount of "soft" quartz was reported to have been extracted from this locality by quarrymen. According to files of the U.S. Silica Co., the zone of friable quartz was considered to be significant because of its relatively low iron content (a requirement for glass sand) and ease in rod milling and processing for glass sand and flour. Little friable quartz remains, most having been mined out. In the quarry walls, a punky yellow-stained quartz breccia remains. During active operations in 1992, a friable quartz zone was intersected by the quarry face on Bench 1 approximately 50 meters south and above STOP 4. Friable and brecciated zones of the quartz lode are due to cataclasis of both vein and massive quartz.

Quartz veins that are 15-20 cm thick, trending N to NNE, steeply dipping and exhibiting free growth or comb structure are present in quarry walls and adjacent bedrock. These are characteristic main-stage veins (QV_6). Thin (5 mm) green muscovite veins (180°-77°) are exposed in the western quarry wall. Similar muscovite veins from STOP 4 yielded a 238 Ma age. N- (000°-74°; rake = 44° N) and NNE- (020°-90°; rake = 44° N) trending oblique faults displace rocks at this locality. In addition a 030°-90° joint set is pervasive in bedrock along the eastern quarry wall.

STOP 6. PERMIAN OR OLDER QUARTZ VEIN [QV_0] IN METASOMATIZED ALASKITE.

(Location: approximately 500 feet NNW of STOP 5. Follow the quarry road out of the south end of the old Silex mine area. Take a right uphill at the first "Y" interesection. The locality is approximately 100 feet on the right.)

The alaskite here is green and contains molds of goethite(?) pseudomorphs after pyrite. QV_0 is approximately a meter across and composed of "bull" quartz with coarse-grained books of muscovite. Locally muscovite is not restricted to the vein, and is present in the wall rock of the vein.

The quartz contains two types of fluid inclusions: two-phase H_2O and three-phase CO_2-H_2O (Figure 17) and were described earlier. Both muscovite in QV_0 and in vein walls were dated by the $^{40}Ar/^{39}Ar$ method and yielded a 255 Ma age (Altamura and Lux, 1994; in preparation). This age determination represents an upper constraint for the age of the vein. QV_0 seems to record evidence that earliest hydrothermal solutions intruded into the fault zone during the Permian (or earlier), long before main-stage silicification occurred in the Middle Triassic.

*** LUNCH: OVERLOOK OF QUARRY ON EASTERN HIGH WALL***

(Location: follow quarry road out of main pit to the primary crusher, turn left and follow the Bench 1 quarry road south. Proceed around the open pit at the Bench 1 level to the footwall of the deposit. Cross the ridge of a talus pile that separates the open pit from a small wetland, proceeding north; a well marked path into the woods will appear. Follow this path 700 feet or so to a quarry overlook on the eastern high wall. Note glacial polish on much of this surface.)

Continue north on Lantern Hill Road.

56.0	Turn right onto Connecticut Route 214 north
56.2	Turn right onto Connecticut Route 2 east.
56.4	Turn right onto Wintechog Hill Road.
56.6	STOPS 7-10 (Figure 15).

Convenient parking is on the right at an unpaved dirt pull-off. The trail (an old cart road) up to Lantern Hill can be seen to the right.

*Part II: Lantern Hill Excursion (STOPS 7-10)**

*(for our GSA field trip, we will hike through the forest to Lantern Hill from Long Hill and will shuttle participants back to their parked vehicles on Lantern Hill Road as necessary)

STOP 7. "SADDLE CLIFF" SILICIFIED BRECCIA

(Location: within the ridge between Long Hill and Lantern Hill is a topographic saddle east of and adjacent to Lantern Hill Pond. Jutting up as a small, steep hill is what I have informally named "saddle cliff". From the lunch STOP, proceed along the wooded path approximately 700 feet, where a path uphill to left can be taken for 40 feet to STOP 5.)

Saddle cliff has been mapped primarily as yellow breccia. This cataclasite consists of granulated vein quartz, that includes broken crystals, with breccia fragments that range in size from rock flour to a few cm across. Veins classified as QV_6 as well as high-angle normal faults and pervasive joints cut the breccia. The yellow color of the rock is due to surface coating on clasts and does not persist into the clasts themselves.

Retrace route to the main path and follow it around to the south in order to reach the base of Saddle Cliff to observe further evidence of brittle deformation, this time in the form of a very tight stockwork pattern.

METASOMATIZED PLAINFIELD FORMATION ON LANTERN HILL

STOP 8. METASOMATIZED SCHIST

(Location: from Saddle Rocks continue on the same wooded trail along the flank of the western talus slope of the lode. Please take care of your footing on the steep slopes. A trail entering from the east is refer to as the Narragansett Trail. At this point on the trail Lantern Hill Pond will be on the left (west) some nearly 200 feet below. About 350 feet north of Saddle Cliff is a clearing in the woods that allows a clear view of Mount Lantern Hill (our final STOP of the day) looming up to the east some 100 feet up. Note the joint control of exposed cliffs. This is a good example of roche moutonée with the plucked cliff facing south.

Continue on this same trail northward along the western flank of the lode. After about 800 feet look eastward to the high cliffs for a pervasive layering or parting in the bedrock of the lode. This rock is proposed as a metasomatite on the basis of considerable mineral impurities, with an inherited layered fabric (primary foliation). This layering strikes E-W and dips moderately to the north, consistent with regional foliation. After hiking for a total of about 1,325 feet (from the view of the roche mountonée) to a long abandoned quarry road, a path rises to the right (south) toward the summit of Lantern Hill. The path to the top of a hill is followed by a saddle and then a second hill, the summit, referred to here as Mount Lantern Hill. Take the westernmost of two parallel trails that ascend the hill, the one that flanks the western edge of the hill for about 700 feet. The area of STOP 8 begins with the bedrock on which the trail passes on the southern side of this saddle.)

Although this rock is pervasively iron-stained, a fresh sample will reveal that it is layered and consists of mineral grains that are visible to the naked eye. The rock is a deep green color in places. Green platy minerals, that include chlorite and illite make up the rock. It is best classified as a quartz-chlorite schist. Empty cubic casts that are filled or lined with iron-oxide-hydroxides are very common, more so than in the metasomatite of the Hope Valley Alaskite Gneiss on the hanging wall at Long Hill. The size of cubic casts (almost 1 cm) at this locality can exceed any found elsewhere in metasomatite associated with the lode. Presumably these casts are relic of secondary pyrite and probably minor pyrrhotite (as at Long Hill) that resulted from the sulfidation phase of wall-rock alteration of the Lantern Hill lode. The overall rock is commonly ramified by numerous quartz veinlets that parallel and transgress the foliation.

The best candidate for the protolith of this rock is the schistose member of the Plainfield Formation that has been mapped nearby in the host metamorphic units outside the lode (Goldsmith, 1985). In addition to similar petrofabrics, both the Plainfield schist and the quartz-chlorite schist occur in lithologic (stratigraphic(?)) succession with the quartzite member of the Plainfield Formation. STOP 9., about 30 feet along the trail to the south, will be to a site where the metasomatized equivalent of the Plainfield quartzite is exposed.

Elsewhere within the core of the Lantern Hill lode any incorporated wall rock is almost totally replaced by quartz. The preservation of recognizable schist and quartzite members of the Plainfield Formation is striking. Within the quarry walls at Long Hill small horses of alaskite, in various stages of alteration, could be recognized here and there near the central part of the lode (e.g. Bench 1). Whole rock and trace element analyses that were obtained on unaltered Plainfield schist outside of the fault zone and the metasomatite, may shed some light on the alteration process.

Chemical analyses were obtain for unaltered Plainfield schist outside of the quartz lode and the schist metasomatite at this locality (Table 1). Analysis of these data using the isocon method of Grant (1986) indicates that most elements were relatively immobile: MnO_2, P_2O_5, Na_2O, CaO, TiO_2, K_2O, MgO, FeO, and Al_2O_3. Silica and LOI increased, while Fe_2O_3 decreased. If these two analyses are representative, then it is consistent that hydrothermal fluids silicified and hydrated the metasomatite. Ferric iron may have been removed from the system at the time of wall-rock alteration, or perhaps later as part of the process of near surface weathering. Iron-stained outcrops may have derived some iron from local wall-rock alteration.

Foliation of the quartz-chlorite schist strikes about 085° and dips steeply south where first encountered on the north-facing slope of the saddle. To the south foliation steepens to vertical. After crossing a faulted quartz vein (255°-45°; rake = 90°), the foliation dips to the north. Regional foliation throughout the study area typically is 270° with a moderate dip angle. Why does the attitude of the foliation differ at this STOP compared to regional foliation as portrayed by Goldsmith (1985)?

In studying the gabbros and associated rocks near Lantern Hill, Loughlin (1912) reports that agreement between foliation and bedding was found in small or minor folds, but none were observed in principal folds, as no axes of the latter were exposed. He recognized miniature folds or undulations oblique, or in many places normal, to the general strike and

Table 1. Chemical composition of unaltered and metasomatized schist and quartzite members of the Plainfield Formation.

	Schist (unaltered)	*Schist (altered)*	*Quartzite (unaltered)*	*Quartzite (altered)*
SiO_2	80.26	56.65	91.20	91.73
TiO_2	2.10	0.74	0.20	0.24
Al_2O_3	14.74	7.46	4.38	3.17
FeO_2	6.69	2.99	0.41	1.24
Fe_2O_3	6.42	2.03	0.68	0.27
MnO	0.24	0.09	0.02	0.03
MgO	4.01	1.68	0.45	0.97
CaO	0.90	0.04	0.04	0.36
Na_2O	0.39	N.D.	0.45	0.14
K_2O	3.87	1.69	1.39	0.65
P_2O_5	0.13	0.02	N.D.	0.05
LOI	3.56	2.45	0.98	1.05
Subtotal	99.69	99.44	100.19	99.91
Ba	735	382	272	235
Sr	71	2	3	17
V	281	79	31	18
Y	36	15	4	1
Rb161		101	98	31
Zr	327	173	84	137
Ni	89	20	3	tr
Total Sx	N.D.	0.02	N.D.	N.D.

LOI, lost on ignition; major elements and x in weight percent; trace elements in ppm; tr, trace amounts; N.D., not detectable. Note: unaltered samples were collected in the field area, but outside of the Lantern Hill fault zone, and altered samples were collected from wall-rock horses within the quartz lode.

dip. Perhaps this change in foliation attitude at this STOP is an example of a principal fold that Loughlin (1912) did not find. The southerly dip of the foliation is certainly normal to the attitude of regional foliation elsewhere in the Old Mystic quadrangle. Alternatively, the metasomatite is cut by a fault quartz vein and movement along this fault might have tectonically rotated the block into its abnormal orientation. In this latter case it may represent a horse similar but probably larger than those that can clearly be observed in the faces of the open-pit quarry on Long Hill. The deformation could possibly have resulted from a combination of both processes: folding under high grade conditions followed by brittle fracturing associated with main-stage emplacement of the quartz lode.

STOP 9. METASOMATIC QUARTZITE

(Location: 30 feet south along the same trail; crossing the faulted quartz vein. Quartzite makes up the bedrock over which the trail passes. Watch your footing!)

The mineralogy and texture of the quartzite show no significant difference to that of Plainfield quartzite outside of the lode and contact aureole. A diagnostic biotite parting, typical of the Plainfield quartzite found throughout southeastern Connecticut, is present in this quartzite (Goldsmith, pers. comm., 1992) and is recognizable at the mesoscopic scale.

Both unaltered host-rock and metasomatic quartzite were analyzed for whole rock and trace element compositions (Table 1). Al_2O_3 and TiO_2 apparently were immobile during metasomatic alteration; CaO, MgO, FeO and LOI were increased; K_2O, Na_2O, and Fe_2O_3 show loses. MnO_2, P_2O_5 and SiO_2 intersect the isocon, indicating little change in terms of mass balance. Overall chemistry and mineralogy of the quartzite seems to only have been modestly changed. The schist metasomatite, on the other hand, was significantly silicified and biotite was altered to chlorite.

Relic metamorphic fabric in the quartzite metasomatite shows evidence of ductile deformation that is not reported in Plainfield quartzite outside of the fault zone (e.g. Goldsmith, 1985). Foliation defined by biotite partings within the quartzite is folded such that mesoscopic-scale folds plunge moderately to vertically WSW. Relic foliation within the quartzite varies along a sequence of outcrops from 245°-80° to 270°-90°, but may be 345°-80° near folds. A steepening and overturning of foliation might be accounted for by late low-angle normal faults along the Honey Hill fault, but the development of folds with vertical fold axes seems unlikely.

Folds and variations of foliation as described above are not consistent with mapping of host rock units outside of the lode (Goldsmith, 1985; this study) and may represent evidence of deformation along the Lantern Hill fault zone while in the ductile regime. One such fold is truncated by a vertical boundary, and drag sense along the boundary indicates a sinistral displacement.

Lithologies, that include the Plainfield Formation, are sinistrally displaced about 2,000 feet on either side of the quartz lode. How much of this component of net displacement was due to brittle deformation associated with Mesozoic activity along the Lantern Hill fault, and how much displacement might be associated with earlier ductile deformation is not known.

STOP 10. "HARD" MASSIVE QUARTZ ATOP LANTERN HILL

(Location: 400 feet south of STOP 9 to the summit area of Lantern Hill. A U.S. Geodetic Survey marker, embedded in the white quartzose rock, is approximately 480 feet above sea level. The highest point of Lantern Hill is 100 feet to the north at 491.5 feet. Steep cliff faces make up the east and western margin of the hill at this location, so proceed with caution.)

Bedrock at this locality is mapped as white quartzose rock and white quartzose rock with iron impurities (Altamura and Gold, 1993b, Altamura, 1994b). Lithology was mapped at 1:1,200 scale as part of the mining plan for the U.S. Silica Company and the names of the units reflect the economic interests of the mining company. Highest-grade ore was the white quartzose rock unit. Specifications for most of the products prepared from Lantern Hill silica required a low concentration of non-quartz minerals, especially iron-bearing phases. The summit of this hill would have held significant value to the mining company had it not been for its popularity as a wilderness trail, and for the fine panoramic views of the ocean (e.g. Fisher's Island and Block Island sounds), Montauk and Orient points on Long Island and surrounding coastal lowlands. Due to topographic and geologic conditions at Lantern Hill, the biological habitat at Lantern Hill is somewhat distinct. Interesting plant populations characterize the area, including stands of pitch pines near the summit of Lantern Hill. The historical record indicates that the following rare plants have been found: *Junais debilis, Asplenium montanum* (1908) and *Carex nigromargrnata. Xyris smalliana* is also reported from here (N. Murray, pers. comm., 1991).

The Lantern Hill lode formed by the processes of replacement of wall rock and fracture filling. Hydrothermal fluids migrated into the Lantern Hill fracture zone following rather continuous fractures related to a specific tectonic event and also along microscopic fractures and grain boundaries, removing much of a host's original alkalis by converting feldspar to sericite (illite). Sericitization must have preceded much of the quartz replacement/remobilization, because where wall rock is completely altered to a 99% SiO_2 rock, feldspar must have been altered and largely removed from its position in the gneiss in order for the empty space to be filled by hydrothermal quartz.

Open fractures provided plenty of room for additional precipitation of quartz. The character of deposition resulted in veins that vary from vuggy to massive. In completely replaced wall rock, massive veins can be difficult to discern in outcrop, but are apparent under the petrographic microscopic. Recall that by and large quartz of the Lantern Hill lode is nearly all milky white, baring the presence of mineral impurities of such keen interest to the former quarrymen.

Aside from the water that hydrated feldspars, the only other constituents that have left evidence of their presence in the lode besides SiO_2 are F (sericite and larger hydrothermal muscovite grains), NaCl (fluid inclusions), Al_2O_3 (rare kaolinite as daughters in fluid inclusions), K_2O (hydrothermal muscovite in vugs and vein feldspar reported by Louglin (1912)), sulfur (in pyrite), and Pb and PO_4 (in plumbogummite). Isocon diagrams attest to the mobility of some of these constituents for those rocks analyzed. The Lantern Hill lode is essentially "barren" with respect to precious and base metals when compared to the giant gold-bearing quartz lodes out west. Measurable amounts of trace Au, Ag, Hg, Te, Cd, Sb, As, Bi and Se are present in host and wall rocks as well as in the lode and may have been mobile. Highest gold assays were for the pyrite-bearing metasomatite that yielded 0.020 ppm.

Whole rock chemical analyses were conducted on specimens from this locality and along a traverse to the east. SiO_2 content of quartzose rock at this site is 98.64 weight percent. At the base of the cliff, recognizable tan to pink Hope Valley Alaskite Gneiss crops out. Alaskite chemical composition at the first encounter with host

Figure 23. Joints in massive quartzose rock atop Lantern Hill. The intersection of this joint pair breaks the massive quartzose rock into diamond-shaped lithons.

rock is similar to that of alaskite at STOP 1a (i.e. unaltered alaskite). This west-east sequence of outcrops from the lode to alaskite makes up a rare traverse that roughly parallels the strike of the foliation in the alaskite. Mesoscopically, microscopically and chemically, no alteration is indicated to alaskite immediately adjacent to the lode or further to the east. The eastern border of the lode at this locality must be truncated by a late high-angle fault that is covered by the talus slopes of the hill. Goldsmith (1985) maps a regional N-S trending fault through this location.

Massive quartzose rock atop Lantern Hill is equated with main-stage silica mineralization, because hydrothermal quartz that replaced feldspars in wall rock is pervasive, and occurs within the core of the lode, along with vuggy veins and pockets assigned to QV_6. Massive quartz and QV_6 are cut by a pervasive joint set (165°-85° and 195°-85°), sharing an angular relationship such that the rock is broken into lithons that resemble elongate diamonds or sigmoids in cross-section (Figure 23). Slickenlines have not been observed on the surfaces of these fractures. In the absence of evidence of shearing on this joint pair, they are considered to represent mode I joints that developed due to stress fields of different orientations. Both joint sets transgress hydrothermal quartz and are in turn cut by quartz veins. They must have formed during the time of silicification of the Lantern Hill fault.

Over broad regions joint sets are remarkably consistent in their orientation (e.g. Engelder and Geiser, 1980), and joints have been used to understand regional paleostress (e.g. Engelder and Geiser, 1980, Hancock, 1985). Joint development in perturbed stress fields near faults were considered by Rawnsley et al. (1992) who report that joints propagating in perturbed stress fields such as that near a fault will curve to follow the directions of the stress field trajectories. At Nash Point near Wales, England, these workers discovered continuous exposures of joints that reflected this phenomenon of perturbation of joint trend in proximity to mesoscopic-scale faults.

Curved joints and faults can be observed in the Lantern Hill lode and must reflect changes in the paleostress field (Figure 24). The lode is a part of the Lantern Hill fault zone, and the likelihood that joint patterns curve and are perturbed by faults is high. Broad exposures of this jointed surface are rare within the lode and observation of joints tracing into a fault on this surface has not occurred. But the presence of mode I joints of varying orientations is indicative that the stress field changed through time. In the core of the fault zone, it seems likely that this perturbation in the stress field was due to proximity to a fault(s). Curved fractures are noted elsewhere in the lode and a sigmoidal pattern was adapted in the map interpretation (Altamura and Gold, 1993b, 1994b).

Figure 24. Joints cutting massive quartz of the Lantern Hill quartz lode, North Stonington, Connecticut. Note the curvature or change in orientation of joints in proximity to a dense zone of fractures. Curved joints may record evidence of stress perturbation in proximity to a fracture zone.

Additional support for this model of joint formation comes from late-stage quartz veins that cut the perturbed joints at this site. There are two vein sets at this locality that cut this brittle strain, but do not show evidence of having been jointed. These veins have attitudes respectively of 210°-80° and 020°-90°. The former is oblique to the joints and the general NNE-trend and the latter is subparallel to the fault zone. The veins are about 5 cm thick, massive and lack comb-structure. The relative ages of these veins is known with respect to massive replacement quartz [and QV_6] and the perturbed joints, but not with respect to pervasive post-main-stage veins mapped elsewhere in the lode [QV_7-QV_9]. Specimens of these veins were studied for fluid inclusions in an effort to use temperature as a discriminating factor to place them within the chronodynamic scheme worked out for the dominant veins of the lode. The 210°-80° veins contain primary two-phase aqueous fluid inclusions that homogenize between 200-225°C, placing them within the T_h range of main-stage mineralization. The 020°-90° vein was unique among all Lantern Hill quartz veins studied in that no measurable fluid inclusions were found.

Mount Lantern Hill Lookouts

(Location: from the summit of Lantern Hill, there are two very nice lookouts that offer panoramic views: the southernmost edge of the summit and the easternmost edge of the summit.)

The southernmost provides a view down the Lantern Hill fault valley (Altamura, 1994b) to Long Island Sound and Long Island (Figure 25). The eastern edge of the hill offers a view of Block Island, Rhode Island and its sound. Other directions provide interesting views as well. It is little wonder that legends and reports indicate that native American Indians and early American colonists used Lantern Hill as a lookout and a signal station.

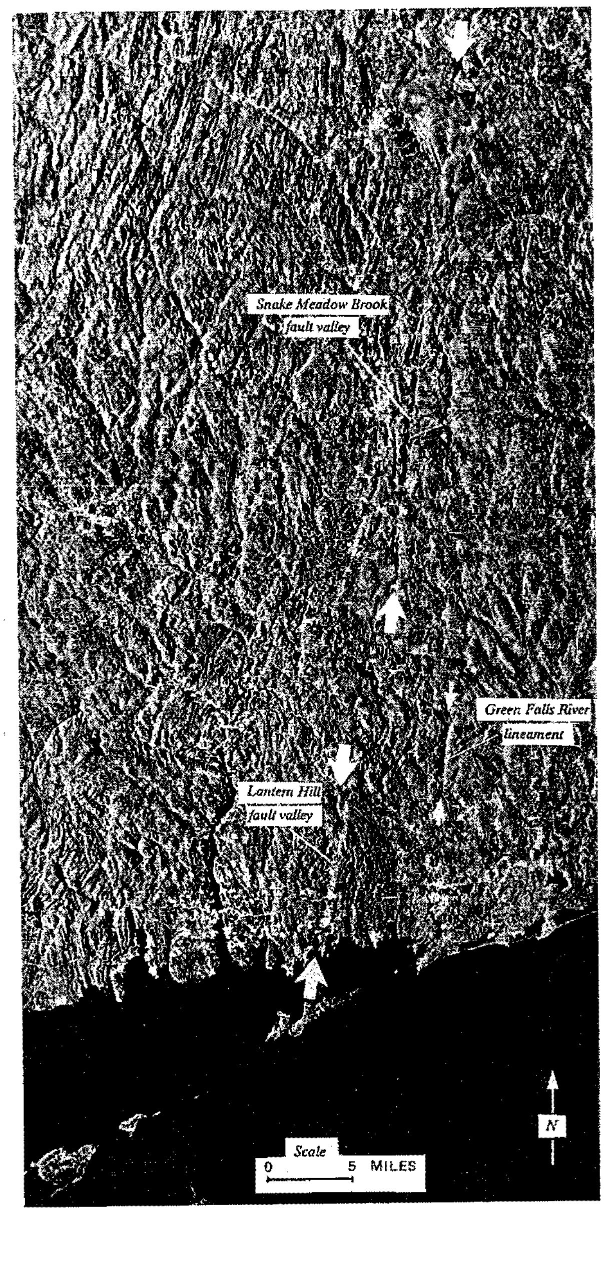

Figure 25. Synthetic aperture radar (SAR) image of eastern Connecticut, showing the Lantern Hill and Snake Meadow Brook fault valleys (Altamura, 1987). Parallel lineament to the east, here identified as the Green Falls River lineament, may represent a genetically related fracture zone.

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