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TABLE OF CONTENTS

WERE EARLY JURASSIC DINOSAURS GREGARIOUS? REEXAMINING THE EVIDENCE FROM DINOSAUR FOOTPRINT RESERVATION IN HOLYOKE, MASSACHUSETTS

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INTRODUCTION

Our understanding of dinosaurs has changed radically since the “Dinosaur Renaissance” of the 1970s (see Bakker, 1975). Before that time, these animals were considered evolutionary dead-ends which gave way to the more advanced mammals after the end-Cretaceous extinction event. However, as Bakker pointed out in his analysis of such disparate lines of evidence as bone histology and predator-prey ratios, many dinosaurs appear to have been much more like the mammals in terms of their metabolism than had been previously thought. At the same time, our view of the social lives of dinosaurs also changed. No longer were they viewed as solitary animals, rather, many became viewed as social or gregarious animals. But what evidence is there that we should change our view of dinosaurs and their social lives?

First, we must more define what it means to be gregarious. Gregariousness is the tendency of some animals to gather into structured social groups for a common purpose, such as greater defense against predators, increased access to food, increased breeding efficiency, protection and rearing of young, or increased migration efficiency (Currie and Eberth, 2010). As Ostrom (1972) pointed out, not all natural accumulations of animals constitute a gregarious assemblage. For example, frogs in a pond are not gregarious because the pond dictates where the animals live. Typically, for living animals, the social interactions inherent in gregarious behavior are obvious. For extinct animals, however, gregariousness can only be inferred from their fossil record. Where body fossils (e.g., bones and teeth) are abundant, gregarious behavior is often inferred from deposits consisting of a single taxon, called monospecific bone beds (Horner, 2002). Unfortunately, many regions, such as the Late Triassic- through Early Jurassic-aged Hartford Basin of Connecticut and Massachusetts (Fig. 1) are nearly devoid of body fossils due to the effects of soil and groundwater chemistry (McDonald, 1992, 1995; Getty and Bush, 2011).

In places with a depauperate body fossil record, trace fossils—such as footprints and trackways—may be used as a proxy for skeletal material and can be used to infer gregarious behavior. Indeed, trace fossils may be better indicators of gregariousness than body fossils since they are sedimentary structures that reflect the behavior of the living animals. In contrast, bone beds may result from other factors, including serial predation (predators killing their preferred prey species in the same location over time) and postmortem sorting and accumulation by biological or sedimentological processes (Rogers and Kidwell, 2007).

Gregariousness may be inferred from trace fossils by a suite of characteristics (Lockley, 1991). Among the most obvious of these is trackway parallelism, since herding animals will leave nearly parallel trackways as the herd moves from place to place. Indeed, Edward Hitchcock, in his seminal studies of fossilized footprints, considered parallel trackways as evidence of gregariousness as early as 1836 (see historical account, below). However, parallel trackways also may result from animals walking next to

Figure 1. Location and geologic context of Dinosaur Footprint Reservation (DFR).
A: Bedrock geology of the northern portion of the Hartford Basin. The star incates the location of DFR within the basin. **B:** Simplified stratigraphic column of Hartford Basin rocks showing the position of DFR immediately above the Granby Tuff. **C:** Google Earth image of the land surrounding DFR. Note the position of DFR immediately to the east of U.S. Rte. 5 and additional outcrops along the west bank of the Connecticut River (labeled "beds by river." **D:** Oblique view of the "Ostrom bed" at DFR. **E:** View to the south of beds cropping out along the Connecticut River. These beds, which are not part of the reservation, also preserve dinosaur tracks. Panels A and B modified from Collette et al. (2011) and reproduced with permission of Atlantic Geology.

a physical barrier, such as a shoreline (e.g., Ostrom 1972; Lockley, 1991). Since physical barriers can create the illusion of group behavior by funneling animals along a predetermined path, other criteria must also be considered. These other criteria include the constant spacing between both straight and turning trackways that results from members of a group walking next to each other without colliding (Currie, 1983). Additionally, similar speeds calculated from trackways can also be used as evidence for group behavior since the animals will tend to move at the roughly the same velocity in order to remain in a coherent group (Martin, 2006, p. 433). Finally, because track depth will vary with time (tracks made in wet mud will be deeper than those made in the same substrate after it has dried in the sun a while), depth may be used to infer or reject hypothesized gregariousness (Lockley, 1991). Thus, Getty (2005, p. 173) rejected the hypothesis that two trackways of carnivorous dinosaurs were made by a group because the nearly overlapping and subparallel trackways had significantly different depths and preservational quality.

Based on both body and trace fossil evidence, examples of gregariousness are now known in many groups of herbivorous dinosaurs, including ankylosaurs (e.g., McCrea and Currie, 1998; McCrea, 2000), ceratopsians (e.g., Currie and Dodson, 1984; Lockley and Hunt, 1995; Ryan et al., 2001; Qi et al., 2007), ornithopods (e.g., Cotton et al., 1998; Scherzer and Varricchio, 2010), and sauropodomorphs (e.g., von Huene, 1928; Bird, 1944; Lockley et al., 2002; Castanera et al., 2011). Additionally, gregariousness has been inferred for carnivorous dinosaurs, the theropods, but it has been more controversial, with some researchers supporting the hypothesis (e.g., Currie and Eberth, 2010) and others challenging it (e.g., Roach and Brinkman, 2007). The main thrust of arguments such as those put forth by Roach and Brinkman is generally one of degree, with those researchers arguing that carnivorous dinosaurs were unlikely to have formed complex cooperative groups, as seen in mammalian carnivores, to takedown of large prey. Instead, they argue that theropod dinosaurs were more like most modern diapsid reptiles (the group to which dinosaurs belong), which tend to form only loose associations during feeding.

The focus of our field trip is Dinosaur Footprint Reservation (DFR) in Holyoke, Massachusetts, the site from which Edward Hitchcock first (and unwittingly) proposed gregariousness in dinosaurs back in 1836. This site is often referred to in discussions of dinosaur social behavior, but the evidence remains controversial for some because the tracks are almost universally accepted as having been produced by large, approximately 6 m long, carnivores (e.g., Olsen et al., 1998; Smith and Farlow, 2003; Rainforth, 2005; Lucas et al., 2006, but see Weems, 1987, 1992 for a dissenting opinion). Additionally, ripple marks suggest the presence of a nearby shoreline at the site, but previous arguments for gregarious behavior have either not taken these structures into account or simply dismissed them (e.g., Hitchcock, 1836, 1848, 1858; Ostrom 1972). Thus, the details of the dinosaurs' interactions with their environment have yet to be fully explored. The purpose of this field trip is to reexamine the evidence for gregarious behavior at DFR by integrating the footprint and sedimentological evidence available at the site.

HISTORICAL ACCOUNT

Edward Hitchcock first described dinosaur footprints at what is now DFR in 1836. A large track that Hitchcock collected from the locality (specimen number 15/3 at the Beneski Museum of Natural History, Amherst College) was designated as the type specimen of the ichnospecies *Ornithichnites giganteus* (the tracks are now called *Eubrontes giganteus*, see taxonomic discussions in Olsen et al. [1998] and Rainforth [2005]). Hitchcock also noted that four *O. giganteus* trackways were oriented in the same direction and nearly paralleled each other, which he argued was an indication of group behavior in the trackmakers. However, he believed that birds or bird-like animals made the tracks, later citing the discovery of *Archaeopteryx* for support (Hitchcock 1865). (Cope [1867a,b, 1870] first argued that Connecticut River Valley footprints were those of dinosaurs).

When first described by Hitchcock in 1836, *E. giganteus* trackways were known only from DFR. However, he continued to find additional examples, reporting *E. giganteus* at six localities in 1848 and at least eleven localities in 1858. He continued to argue for the gregarious nature of the trackmakers, and although he suggested that additional sites preserved parallel trackways (1848, p. 170), the only specific example provided was DFR. The exposure appears to have been enlarged by 1858 based on his description of "several" parallel trackways that were intersected by others. Interestingly, Hitchcock suggested that the animals must have been traveling obliquely to the shoreline, but he did not say why he thought a shoreline was present. Neither did he discuss the possibility that the shoreline may have affected the animals' direction of travel.

After Hitchcock's death in 1864, local geologists continued to cite the parallel trackways at DRF as evidence of gregarious behavior (e.g., Bain and Meyerhoff, 1963), but most workers on Hartford Basin dinosaur tracks (e.g., Lull, 1915, 1953; Olsen and Padian 1986; Olsen et al., 1998; Olsen and Rainforth, 2003; Rainforth, 2003) focused their efforts on much-needed revisions to Hitchcock's taxonomy. However, in 1960s, the trackways at DFR caught the attention of John Ostrom of Yale University, who conducted an in-depth study at the site. Ostrom (1972) identified 22 trackways that he attributed to *E. giganteus*, 19 of which were approximately parallel on a single bedding plane hereafter referred to as the "Ostrom bed." Additionally, Ostrom identified six small theropod trackways oriented in different directions than the much larger *Eubrontes*. Although he noted the presence of ripple marks, Ostrom discounted them as an indicator of a physical barrier because he did not observe any preserved on the main track-bearing bed. Ostrom further argued against the presence of a path-directing physical barrier because smaller theropod trackways were oriented obliquely to the large theropod trackways. Thus, he concluded that the trackway parallelism, in the absence of a physical barrier, was best explained by gregarious behavior of the trackmakers. Ostrom's study is now entrenched in the literature as an oft-cited example of gregarious behavior (e.g., Colbert, 1989; Lockley and Matsukawa, 1999; Lingham-Soliar et al., 2003).

Recently, some researchers (e.g., Coombs, 1990; Olsen, 2002; Roach and Brinkman, 2007) have revived speculation that the ripple marks indicate a shoreline and have consequently questioned the interpretation that the trackways at DFR represent gregarious behavior, suggesting instead that the parallel trackways represent shoreline-paralleling behavior. However, these authors did not conduct a detailed examination of the site in order to support their arguments. Another study, by Smith et al. (1996), showed that parallel trackways occurred on beds 69 m above the Ostrom bed, suggesting that long term trends affected animal behavior. Finally, Getty (2004) identified numerous herbivorous dinosaur tracks that had not been recognized previously.

GEOLOGICAL CONTEXT

The Ostrom bed (Fig. 1D) and some of the surrounding land is owned and administered by the Trustees of Reservations as DFR. The reservation, located at 42º 14.5´N latitude, 72º 37.4´ W longitude, is accessible by taking U.S. Rte. 5 north from Holyoke or south from Northampton, Massachusetts. The sedimentary rocks exposed at the outcrops were deposited in the Hartford Basin, an asymmetrical half graben that opened up during the deposition of the Newark Supergroup during the breakup of Pangea in the Late Triassic and Early Jurassic. The rocks at DFR are part of the lower Portland Formation and were deposited in a shallow lacustrine to playa settings (Olsen et al., 1998) in a monsoonal climate strongly influenced by Milankovich cycling (Olsen, 1986). The Ostrom bed lies almost immediately above the Granby Tuff (Fig. 1B), which crops out on the west side of Rte. 5 to the north and south of the pull-off for the reservation. A 30+ m thick stratigraphic section is discontinuously exposed and extends ~ 100 m to the east, where the youngest beds dip at approximately 13º to the east below the Connecticut River (Fig. 1C, E). Footprints are preserved in virtually all of the exposed beds.

The Ostrom bed consists of gray, fine-grained, micaceous sandstone. Sedimentary structures on this bed include faint (oscillation?) ripples (cf. Ostrom, 1972; Fig. 2A). In many places the Ostrom bed is eroded, exposing a rippled layer approximately 1-2 cm below (Fig. 2B). These ripples are typically straight-crested to slightly sinuous and are symmetric to slightly asymmetric in cross section. Thus, their shape is consistent with oscillatory wave generation, rather than unidirectional currents (Nichols, 1999, p. 54). Similar but more sinuous ripples, occur to the east on a remnant of a bed approximately 13 cm above the Ostrom bed (Fig. 2C-D). Additional structures on the Ostrom bed include elongate and irregularly ovate to circular depressions with a maximum diameter of 2.2 cm (Fig. 2.E). These structures are often called "raindrop impressions;" however, based on their large size and irregular shape, we consider them to be molds of weathered away concretions or pebbles.

Sedimentary structures suggesting subaerial exposure and desiccation are rare. Desiccation cracks occur on a single bed high in the section; none occur on or near the Ostrom bed itself. Raindrop impressions have not been observed on any beds, but rare evaporite pseudomorphs were observed by Bain and Meyerhoff (1963), although these authors did not indicate on which beds the structures were observed.

Besides dinosaur tracks, no other fossils were observed on the Ostrom bed. Fossilized wood is abundant high in the section (Fig. 2F), along with rare examples of vertical *Skolithos* burrows (Fig. 2G). Horizontal burrows are restricted to the rippled layer immediately below the Ostrom bed. These burrows range from 1-5 mm in diameter (Fig. 2B, H) and are of variable length. They are subhorizontal and appear to have a minor vertical component based on their abrupt, yet rounded, terminations. The burrows are concentrated in the troughs of the ripples, although they cross-cut some ripple crests. They are not uniformly distributed across the surface; sediments at the north end of the site are only slightly disrupted (bedding plane bioturbation index 1 of Miller and Smail, 1997, see Fig. 2A), whereas exposures at the south end of the site are more highly disrupted (bedding plane bioturbation index of 3, see Fig. 2H). The concentration of the burrows in ripple troughs suggests that the animals were collecting organic detritus that accumulated in low-lying areas, whereas the variability in bioturbation from the north to south end of the site may reflect a water depth gradient.

Overall, the sedimentology and trace fossils at the site are consistent with a shallow, saltwater aquatic setting with occasional subaerial exposure. The relative rarity of invertebrate bioturbation may have resulted from high salinity in the lakes. Additionally, dysoxic or anoxic bottom conditions may have developed at times if the water became stratified with dense hypersaline water on the shallow lake bottoms. Such conditions have been inferred for the lakes in the underlying East Berlin Formation (Gierlowski-Kordesch and Rust, 1994) and elsewhere in the Portland (Sime and Getty, 2009).

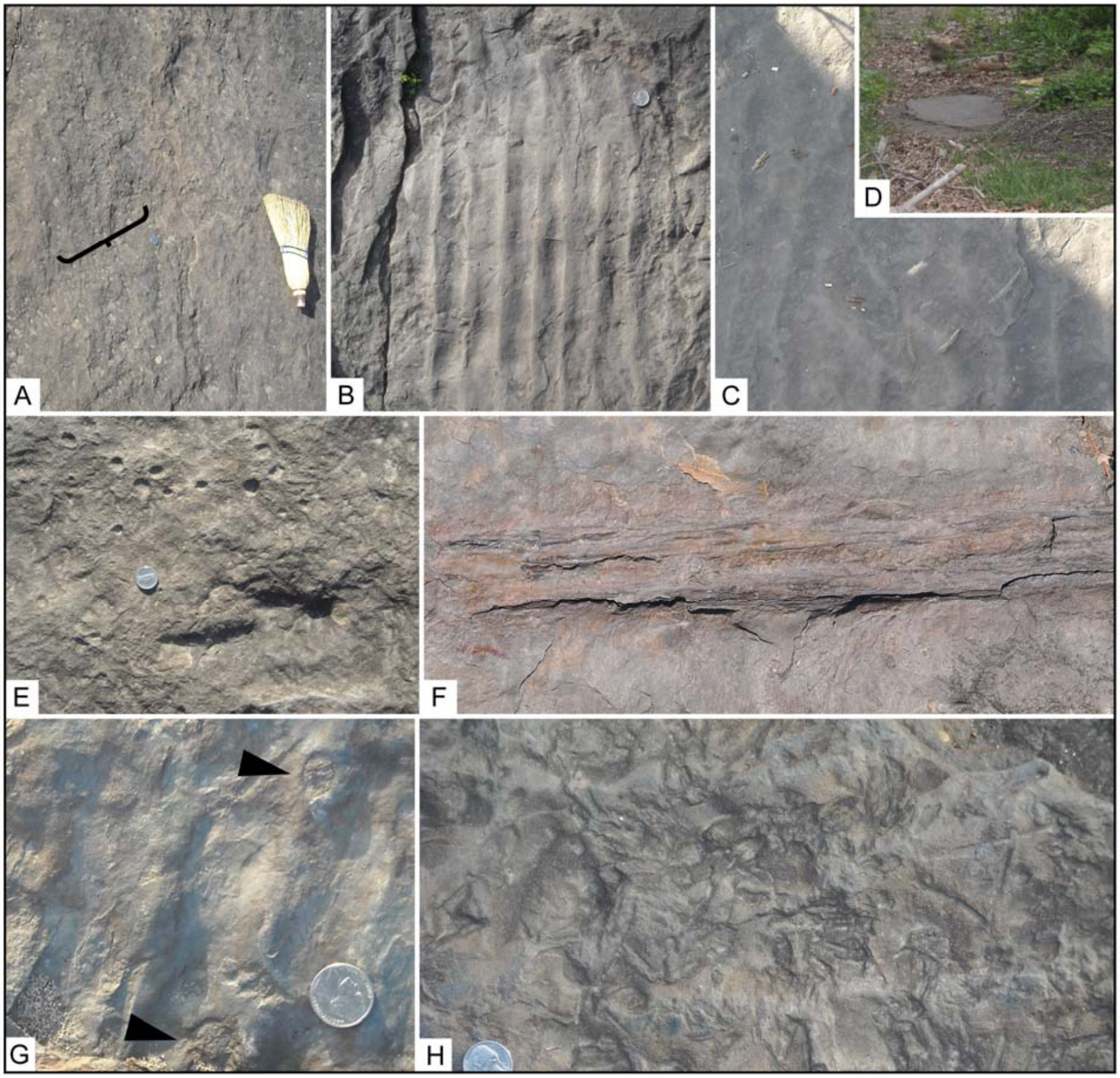

Figure 2. Sedimentary structures and non-vertebrate fossils from DRF and the surrounding area. **A:** Poorly preserved ripple marks on the Ostrom bed are indicated within the bracketed region. **B:** Well-developed, straight to slightly sinuous oscillation ripples are preserved on a bed discontinuously exposed 1-2 cm below the Ostrom bed. This photograph was taken at the north end of the site, where few horizontal burrows are preserved. **C:** Sinuous ripples preserved on a bed 13 cm above the Ostrom bed. **D:** View looking to the east from the Ostrom bed showing the remnant layer on which the ripples in C are preserved. **E:** Ovate pits on the Ostrom bed. Note the theropod track below and to the right of the coin. **F:** Fossilized wood is preserved on multiple beds by the Connecticut River. Field of view in this image is approximately 0.8 m. **G:** *Skolithos* burrows, indicated by arrowheads, occur on some beds by the river. **H:** The same rippled bed as in C. This exposure, from the south part of the site, shows intense bioturbation concentrated in the troughs of ripples. Scales: coin = 2.1 cm; whisk broom = 28 cm.

METHODS

The footprints at DFR were re-examined in the summer of 2010 to test the gregariousness hypothesis for the *Eubrontes* trackmakers and to provide a comprehensive survey of the non-*Eubrontes* tracks at the site. Efforts were first focused on the Ostrom bed, which was divided into two sections (north and south, separated by a large, comma-shaped, vegetated region) that were examined separately. Footprints were first located and outlined with chalk. Due to the highly weathered nature of most of the footprints, each section was examined during both the morning and late afternoon, when low-angle light struck the surface from different angles. Wherever possible, tracks were identified to the ichnogenus level using the definitions found in Olsen and Padian (1986) for the crocodylomorph track *Batrachopus*, Olsen et al. (1998) for the theropod ichnogenus *Eubrontes*, and Olsen and Rainforth (2003) for the small ornithischian *Anomoepus*. Theropod tracks smaller than *Eubrontes*, such as *Grallator* and *Anchisauripus*, were lumped together and referred to as "small theropod tracks" due to the poor preservation of most examples.

After the footprints were identified, their positions were recorded using a Leica Flexline TS/02 total station. Each footprint was characterized by two points: one proximal (usually the metatarsal-phalangeal pad behind digit four) and one distal (usually the tip of digit three). The data were transferred from the total station into Autocad ® software to produce a preliminary map. The map was then entered into Adobe Illustrator ® to produce a final map on which the dinosaur tracks are represented by idealized outlines for clarity.

Additional data collected in the field included compass bearings for ripple crests, for all *Eubrontes* trackways on the Ostrom bed and on beds examined by Smith et al. (1996), and for all other trackway types in the north end of the Ostrom bed. Compass bearings were measured by running a tape measure down the midline of the trackway and measuring the orientation of the tape. When necessary, the orientations of isolated tracks were recorded from a line drawn down the middle of the track, through the middle digit. The data were analyzed using PAST ® software.

RESULTS

A total of 787 dinosaur tracks, including *Eubrontes*, *Anchisauripus*, *Grallator*, and *Anomoepus* were identified, as was one manus-pes set of the crocodylomorph ichnogenus *Batrachopus* (Fig. 3). This contrasts strongly with the 134 dinosaur tracks reported by Ostrom (1972). The new map of the site can be seen in Figure 4A and the relative proportions of track types at the site is given in Figure 4B. Details for the different track types are discussed below.

The orientation of 106 ripple marks were measured on the bed immediately below the Ostrom bed, four orientations were measured on the Ostrom bed, and 27 were measured on the surface 13 cm above the Ostrom bed. All ripple crest orientations, regardless of bed, show the same east-west trend (Fig. 5A).

A total of 231 *Eubrontes* tracks were identified on the Ostrom bed, almost twice as many as the 117 tracks reported by Ostrom (1972). Assuming that none of the animals walked across the surface more than once, the isolated tracks and complete trackways represent 53 animals. Forty of those tracks/trackways, or 75% of the total, trend approximately to the west (Fig. 5B). Significantly, the other

13 trackways are not randomly oriented either, but are clustered in a group to the east. Thus, the large theropods that crossed the Ostrom bed have two preferred orientations approximately 180º apart. Regardless of direction of travel, most of the *Eubrontes* trackways are relatively straight or meander only slightly. However, a single trackway appears to show an animal stopping and then taking a turn to the right (Figs. 3I-J, 4A, arrowed trackway).

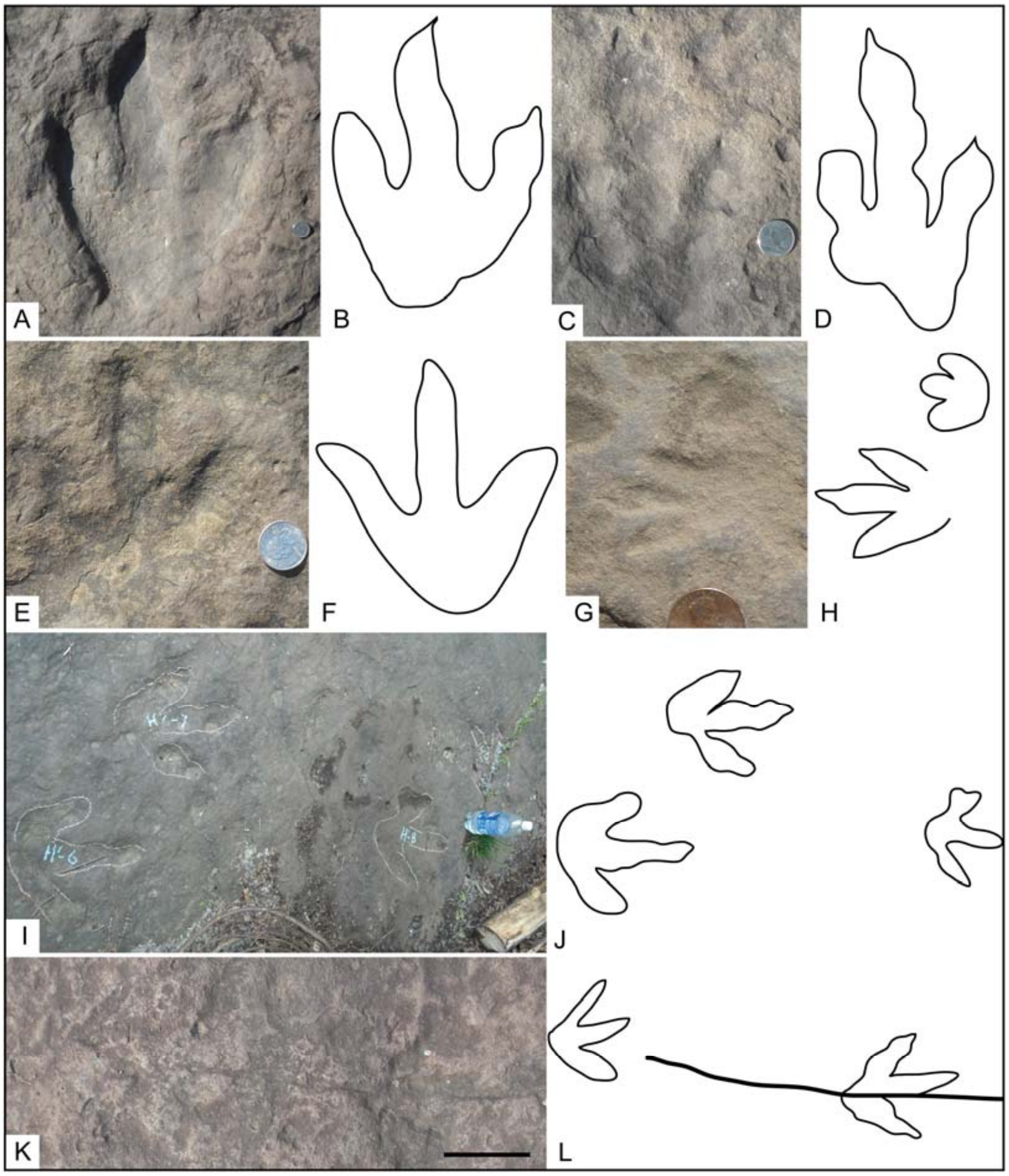

Figure 3. Vertebrate trace fossils preserved on the Ostrom bed. **A:** *Eubrontes giganteus* track, produced by a large theropod. **B:** Interpretive drawing of A. **C:** Smaller theropod track (*Anchisauripus* isp.). Note the pads on the toes. **D:** Interpretive drawing of C. **E:** *Anomoepus scambus* track, produced by a small herbivorous ornithischian. Note the wide angle between the toes, stubby middle digit, and blunt claws that are characteristic of this ichnospecies. **F:** Interpretive drawing of E. **G:** A single manus-pes set attributable to the crocodylomorph ichnogenus *Batrachopus*. The manus is smaller and at the top of the image; the pes is larger and at the bottom. The pes is missing the first digit. **H:** Interpretive drawing of G. **I:** Oblique view of the turning *Eubrontes* trackway. **J:** Interpretive drawing of I. **K:** Close-up of the *Anomoepus* trackway with a trail drag. **L:** Interpretive drawing of K. Coin in A = 2.1 cm, in C and E = 2.4 cm, and in G = 1.9 cm, bottle in I is approximately 20 cm tall, scale bar in K is 10 cm.

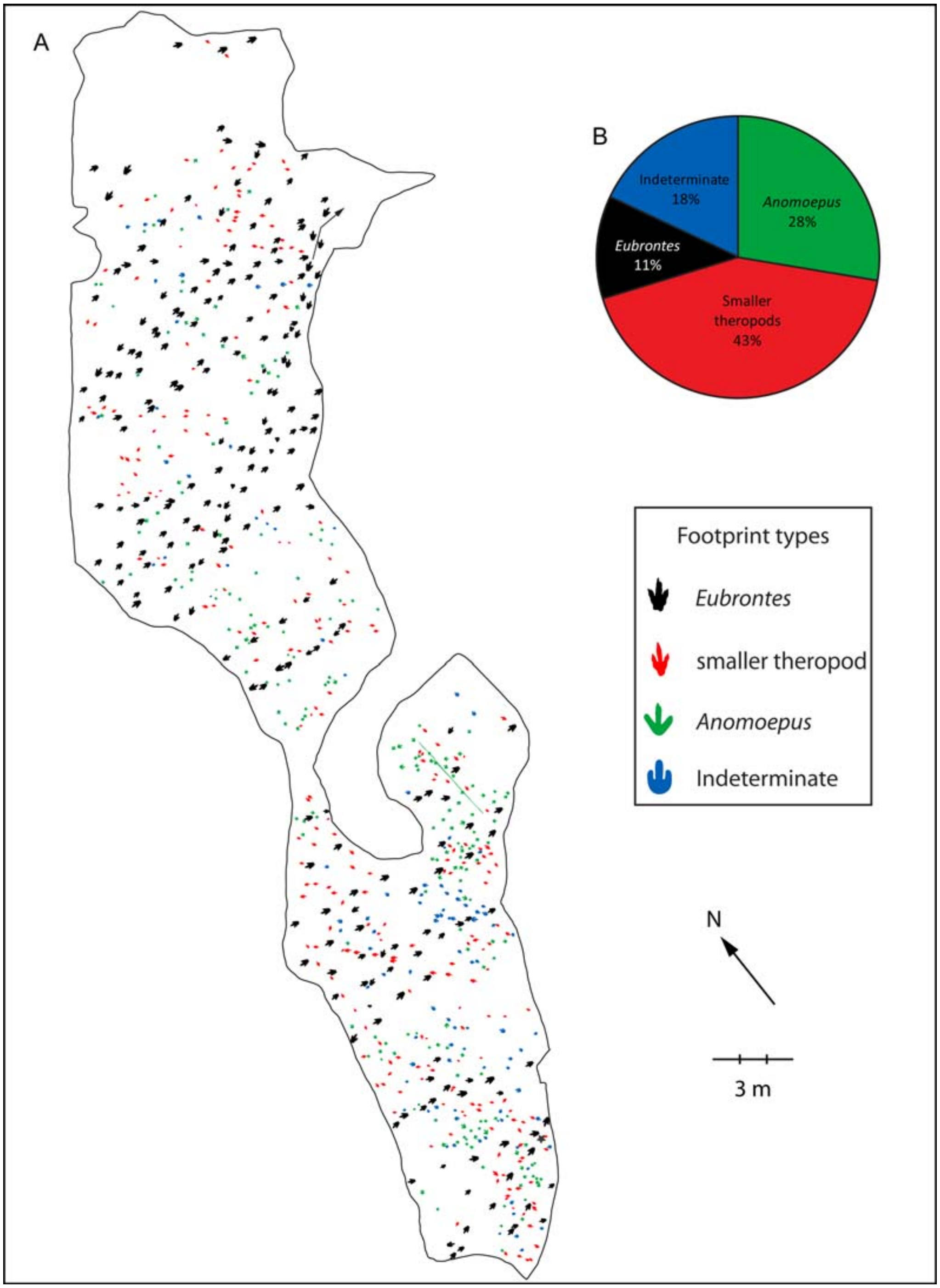

Figure 4. Details of the dinosaur tracks from the Ostrom bed. **A:** Color-coded map showing all 787 dinosaur tracks identified. The position, size, and orientation of individual tracks are accurate, but idealized track symbols were used for clarity at this scale. The long green line indicates the position of an *Anomoepus* tail drag. Also note the bent arrow on the track surface, which indicates the position and direction of movement of a large theropod that stopped and then turned to the right. The small star in the lower right of the map indicates the position of the *Batrachopus* manus-pes set. **B:** Relative proportions of trackmakers. These data were calculated as the number of individuals per taxon, assuming that each trackway and isolated footprint was produced by a unique individual, divided by the total number of individuals across taxa.

In addition to the *Eubrontes* on the Ostrom bed, 49 *Eubrontes* trackways were identified on 20 beds by the Connecticut River, and bearings were recorded for these as well. These data were combined into a single rose diagram (Fig. 5C) because most beds had only one trackway. The combined data show a similar trend to the Ostrom bed, with many trackways headed to the west and fewer trackways headed to the east. However, the trackway orientations are more widely distributed, with some trackways headed to the northeast and northwest. None, however, are oriented to the south. Significantly, a few trackways within a few meters of each other are nearly parallel.

A large number of other tracks were identified on the Ostrom bed, most of which had not been previously reported. These include 248 small theropod (both *Grallator* and *Anchisauripus*) tracks, 205 *Anomoepus* tracks, 103 unidentified tracks, and a single crocodylomorph manus-pes track set attributable to *Batrachopus*. Again assuming that no animal crossed the track surface more than once, these trackways represent 197 small theropods, 127 ornithischian herbivores, and a single crocodilian. Orientations of 86 smaller theropod trackways at the north end of the site show a statistically significant ($p < 0.05$, Chi squared test) northwest-to-southeast trend (Fig. 5D). The 44 *Anomoepus* trackways at the north end of the site show greater trackway concentrations to the east and west. However, *Anomoepus* trackways also trend to the north and south, and neither Chi-squared nor Raleigh tests indicated statistically significant non-random distributions (Fig. 5E). As Getty (2004) noted, a single *Anomoepus* trackway bears a long, sinuous tail drag (Figs. 3K-L, 4A).

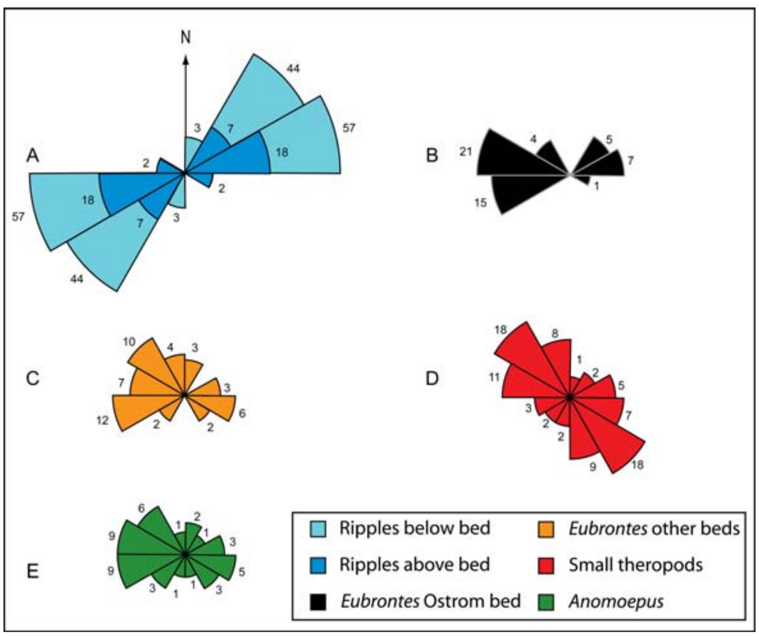

Figure 5. Equal area rose diagrams of ripple and track orientations. **A:** Orientations of ripples below and above the Ostrom bed. Note that the two sets of data overlap and suggest a northeast-southwest trend to the paleoshoreline. **B:** Orientations of *Eubrontes* trackways on the Ostrom bed. Note the bimodal distribution with the two groupings nearly 180º offset. Note also that these groupings are trend to the east and west, approximately parallel to the shoreline. **C:** *Eubrontes* orientations from 20 beds by the Connecticut River (upsection from the Ostrom bed). Note the bimodal distribution toward the east and west. The trend is less strong due to the amalgamation of multiple beds and changes in shoreline orientation higher in the section. **D:** Distribution of small theropod (*Anchisauripus* and *Grallator*) tracks on the northern part of the Ostrom bed. Note the strong bimodality that is oriented oblique to the shoreline. **E:** Orientations of *Anomoepus* tracks on the Ostrom bed. Greater numbers of trackways are oriented to the east and west; however, trackways are oriented north and south as well, and bimodality was not statistically significant.

DISCUSSION

Ostrom (1972) reported that ripple marks did not occur on the bed that he mapped, and therefore he rejected these structures as indicating a physical barrier (i.e., shoreline) at the site. However, faint ripples do in fact occur on the Ostrom bed. More significant than these poorly defined ripples is that fact that well-developed oscillation ripple marks occur only 1-2 cm below and 13 cm above the Ostrom bed, and that all of the ripples have similar orientations to the northeast and southwest. Considering that oscillation ripples tend to form subparallel to the shoreline (Johnson and Baldwin, 1996, p. 263), and that ripple crests of the same general orientation occur below, on, and above the Ostrom bed, it is reasonable to assume that the general trend of the paleoshoreline at DFR was also from the northeast to southwest, at least during the deposition of those beds. Thus, it is conceivable that the shoreline acted as a physical barrier that constrained the dinosaurs' movement at this site.

A bimodal distribution is considered strong evidence of shoreline-paralleling behavior (Lockley, 1986). The fact that the bimodal distribution at DFR encompasses all of the *Eubrontes* trackways on the Ostrom bed, with the two trends approximately paralleling the inferred shoreline, strongly suggests that the shoreline did in fact dictate which direction these large theropods traveled. This assertion is reinforced by the similar east-west trend of many of the *Eubrontes* trackways at the top of the section (Smith et al., 1996; thus study). Bimodality is not as pronounced for these trackways, but this is not surprising considering that shoreline patterns may have changed in the time that it took for the intervening beds to be deposited. Although many beds higher in the section do show ripple crest trends to the east and west, there is more variation in orientation among beds (personal observations), and thus there may have been many minor changes in shoreline orientation. Further, the depositional setting appears to have changed slightly higher in the section. Rare mud cracks suggest that desiccation may have been greater higher in the section, and *Skolithos* burrows, which are known only in playa settings in the underlying East Berlin Formation (Gierlowski-Kordesch, 1991), have been identified.

Although the data from the Ostrom bed strongly supports the hypothesis that the animals were engaging in shoreline-paralleling behavior, this alone does not falsify the hypothesis that the animals were gregarious because, as Lockley (1991) has pointed out, shorelines will direct groups of animals in the same way that they direct individuals. Some lines of evidence seem to support the gregariousness hypothesis, whereas others don't. For example, most of the trackways for which velocities have been calculated indicate relatively slow, steady progression (Lepore, 2006), which could indicate gregariousness. Additionally, three nearly parallel and overlapping trackways (including the animal that turns, see fig. 4A) traverse the site oblique to the orientation of all the trackways that they cross. These trackways may represent a small group of animals that crossed the surface at the same time.

From an ecological perspective, Weems (1987, 1992, 2003) argued that gregariousness in the dinosaurs at DFR would have been unlikely if they were theropods because the animals were larger than their potential prey items (the tracks of large prosauropods, such as *Otozoum*, are never found in association with *Eubrontes* tracks, from which it is inferred that these animals occupied different habitats [Rainforth, 2003]), and because potential prey left fewer footprints in the Hartford Basin, suggesting that they were rarer (e.g., see Wright, 1997). Traveling in large groups, therefore, would have been disadvantageous because any individual would have to have shared its prey with its cohorts, thereby decreasing its potential food intake. Weems accepted at face value that the dinosaurs at DFR were gregarious, and argued instead that gregariousness indicated the trackmakers were herbivores. While we reject the contention that the *Eubrontes* trackmaker was an herbivore, as have most other researchers based on osteological comparisons of the tracks to skeletal material, we note that the argument put forth by Weems has some merit and we suggest that the theropods crossed the Ostrom bed as isolated individuals, or at most in small groups. Further research, outlined in the following section, is designed as an additional test of the hypothesis that the *Eubrontes* trackmaker was gregarious.

The orientations of the trackways of other dinosaurs at DFR are quite interesting—smaller theropod trackways are preferentially oriented northwest and southeast, offset from both the ripple crest and *Eubrontes* orientations. If this trend proves to be consistent on the south end of the site, it may indicate that small and large theropods utilized the ecosystem differently. Specifically, movement towards and away from the water may support the hypothesis that Early Jurassic theropod-dominated faunas were based on piscivory, as proposed by Olsen (2010) for the Hartford Basin and Milner and Kirkland (2007) for the St. George Dinosaur Discovery Site in Utah. The more random orientation of the herbivorous ornithischians suggests that the lake was not as important at defining their patterns of movement as it was for carnivores.

CONCLUSIONS AND FUTURE WORK

The dinosaur tracks at Dinosaur Footprint Reservation in Holyoke, Massachusetts have been re-examined for evidence of gregarious behavior in the light of sedimentary structures present at the site. Oscillation ripple mark crest orientations below, on, and above the main Ostrom bed were used to constrain the approximate shoreline orientation, which was approximately northeast to southwest. *Eubrontes* tracks preserved on the Ostrom bed represent 53 large theropods exhibiting a bimodal distribution oriented approximately to the east and west, roughly parallel to the inferred shoreline. This distribution, along with evidence that the distribution continues on surfaces stratigraphically above the Ostrom bed, strongly suggests that the physical barrier produced by the shoreline dictated the orientation of these animals' movement. Trackways analyzed to date indicate that smaller theropods trended oblique to the shoreline and that small herbivores show no strong preferred orientation. The orientations of small theropods relative to the shoreline may support the hypothesis that the Early Jurassic dinosaur fauna of the Hartford Basin was based on piscivory.

Shoreline-parallel behavior by the *Eubrontes* trackmakers does not falsify the hypothesis that the animals were gregarious, since a physical barrier will direct the orientations of groups just as it directs the paths of individuals. The gregariousness hypothesis can be further tested by considering whether parallel *Eubrontes* trackways occur at other localities and in other sedimentary facies in which such physical barriers as shorelines were transient or absent altogether. Framed in this way, the absence of parallel trackways in other sedimentary environments, such as playa-lake facies, would constitute strong evidence that the parallel trackways at DFR are a facies-specific accumulation rather than evidence of gregariousness. Significantly, detailed studies at Dinosaur State Park in Rocky Hill, Connecticut by Ostrom (1972) and Farlow and Galton (2003) have failed to find a preferred orientation in the *Eubrontes* trackways at that site. We are currently examining additional sites to test the facies specificity of parallelism in *Eubrontes* trackways.

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Road Log:

Time to arrival at stop: 40 minutes.

0.0	Trip begins outside the Hartford Marriot Downtown, at 200 Columbus Blvd. Hartford, CT 06103
0.03	Heading north on Columbus Blvd., turn right onto Grove St.
0.05	Turn right at fork and merge onto I-91 N.
34.7	Take exit 17A to merge onto MA-141 E (Easthampton Rd) toward Holyoke.
0.5	Turn left onto US Rte. 5 (Northampton St).
2.2	Pull off into parking area at destination.
0.0	Turn left out of parking area to head on US Rte 5 south (Northampton St).
2.0	Turn right onto Hampden St
0.2	Merge onto I-91 S via ramp to the left
34.8	Take exit 31 for State St
0.3	Merge onto CT-2 W
0.02	Turn left onto Columbus Blvd
0.1	Hotel will be on the left

Total mileage: 74.9

THE HARTFORD BASIN FROM THE HANGING HILLS TO THE SOUND

Brian Skinner, Geology and Geophysics Department, Yale University, New Haven, CT
Leo Hickey, Geology and Geophysics Department, Yale University, New Haven, CT
Anthony R. Philpotts, University of Connecticut, Storrs, CT
Jay Ague, Geology and Geophysics Department, Yale University, New Haven, CT

Overview of the Hartford Basin of the Connecticut Valley

(Adapted from The Geology of the Connecticut Valley, by Nicholas G. McDonald, in Field Trip Guidebook No.1, The Geological Society of Connecticut, 2010)

One of the prominent geologic features of southern New England is the Mesozoic (Triassic-Jurassic) sequence of terrestrial sedimentary rocks, flood basalts, and diabase intrusions that comprise the lowland of central Connecticut and west-central Massachusetts. This broad, elongate lowland extends from the Massachusetts-Vermont border in the north to Long Island Sound in the south, encompassing an area of some 1,300 square miles. Bordering the lowland are rugged hills known geographically as the Eastern and Western Uplands. The uplands are comprised of metamorphic and igneous rocks of Proterozoic and Paleozoic ages. The structures, ages, and lithologies of the rocks in the uplands are markedly different from rocks in the lowland.

The Connecticut Valley, so named because the Connecticut River drains most of the region, is subdivided into two primary basins, the short, narrow Deerfield Basin to the north in Massachusetts, and the expansive Hartford Basin in the south (Figure 1.1). Two smaller, isolated basins containing Mesozoic strata and basaltic flows lie in western Connecticut—the Cherry Brook Basin (Platt, 1957) and the Pomperaug Basin (Hobbs, 1901). Three parallel, northeasterly trending diabase dikes of early Jurassic age transect the crystalline rocks of the Eastern and Western Uplands; these have been petrographically and chemically correlated with the three flood basalts that are exposed in the southern half of the Hartford Basin (Philpotts and Martello, 1986). The nomenclature and stratigraphy of the sedimentary and volcanic rocks of the Hartford Basin are shown in Figure 1.2.

The Hartford Basin is a member of a series of similar early Mesozoic basins that extend for about 1,200 miles along the eastern margin of North America, from Nova Scotia to South Carolina, and that approximately parallels the structural trend of the Appalachian orogen (Figure 1.3). These basins formed through lithospheric extension in the Late Triassic and Early Jurassic continental rifting that ultimately created the Atlantic Ocean. There is a great deal of lithologic and structural similarity among the basins, and they contain similar fossils. Because they share a common chronology and structural history, the rocks in the Hartford Basin and the other basins are assigned to an all-inclusive lithostratigraphic unit, known as the Newark Supergroup (Froelich and Olsen, 1984).

The story of the Hartford Basin began about 200 million years ago when Pangea was beginning to fragment. During the latest Triassic and Early Jurassic, the region of the Hartford Basin was marked by wet and dry climatic intervals that alternated with warm and dry climatic intervals favoring the formation of sand dunes and seasonal rivers.

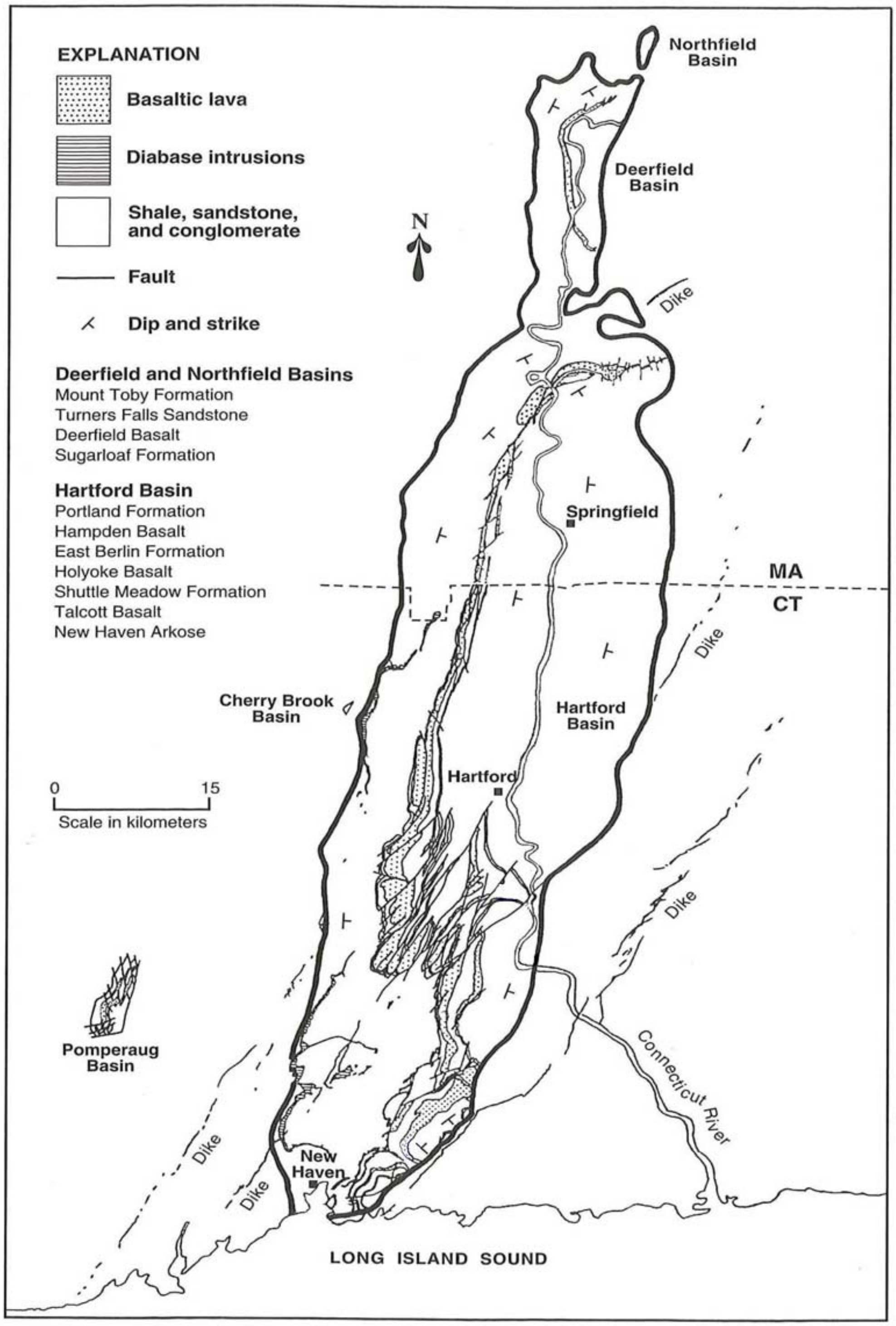
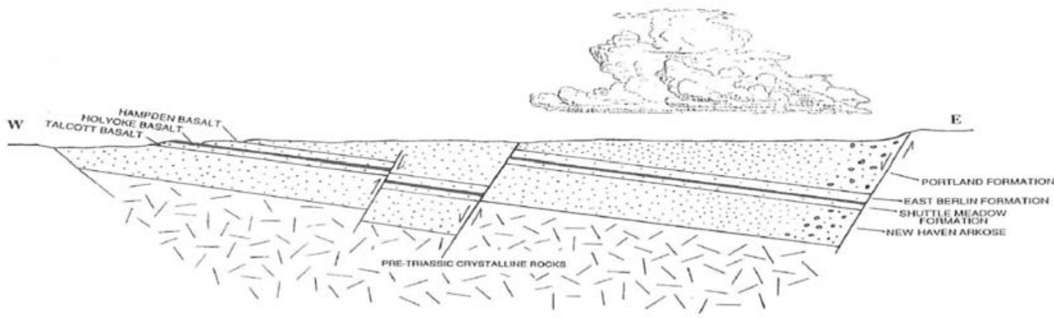


Figure 1.1 Generalized geologic map of the Connecticut Valley Lowland region of southern New England showing the location of the major early Mesozoic depositional basins and the Stratigraphy of Triassic and Jurassic sedimentary and igneous rocks. The Middleton Basin is shown on Fig 1.3. (from McDonald, 2010)



UNITS	THICKNESS (m)	AGE		DESCRIPTION
Portland Formation	2000	Pliensbachian-Hettangian	Lower Jurassic	Mostly red, cyclical, shallow water lacustrine clastics; some gray and black lacustrine (some deep water) clastics with minor limestone; red fluvial and alluvial clastics
Hampden Basalt	60	Hettangian	Lower Jurassic	Tholeiitic basalt flows
East Berlin Fm.	150	Hettangian	Lower Jurassic	Mostly red, cyclical shallow water lacustrine clastics; some gray and black lacustrine (often deep water) clastics with minor limestone; minor red fluvial and alluvial clastics
Holyoke Basalt	100	Hettangian	Lower Jurassic	Tholeiitic basalt flows
Shuttle Meadow Fm.	100	Hettangian	Lower Jurassic	Mostly red, cyclical shallow water lacustrine clastics; some gray and black lacustrine (often deep water) clastics with minor limestone; minor red fluvial and alluvial clastics
Talcott Basalt	65	Hettangian	Lower Jurassic	Tholeiitic basalt flows
New Haven Arkose	2250	Hettangian-Late Carnian	Upper Triassic-Lower Jurassic	Red and brown, coarse, soil-modified fluvial clastics

UNITS	THICKNESS (m)	AGE		DESCRIPTION
Mt. Toby Fm. Turners Falls Sandstone	2000	Hettangian	Lower Jurassic	Mostly red, coarse alluvial clastics; lateral equivalent of Turners Falls Ss. Mostly red, cyclical, shallow-water lacustrine clastics; some gray to black, cyclical, deeper-water lacustrine deposits
Deerfield Basalt	50	Hettangian	Lower Jurassic	Tholeiitic basalt flows
Sugarloaf Fm.	1700	Norian?	Upper Triassic	Red, buff, and minor gray, coarse to fine fluvial and alluvial clastics

Figure 1.2 Stratigraphic nomenclature in the Connecticut Valley.
***Top*:** Idealized east-west cross-section through the Hartford Basin.
***Table*:** Currently recognized formations in the Hartford and Deerfield Basins (from McDonald, 2012; modified from Olsen, et al., 1989)

The stop in Hamden to view the New Haven Arkose is far out on a flood plain, a considerable distance from the faulted eastern margin of the basin. The stop in the Portland Quarry, is in the Portland Formation, and is close to the eastern margin of the basin, where alluvial fans, sand dunes, and seasonal river deposits are found.

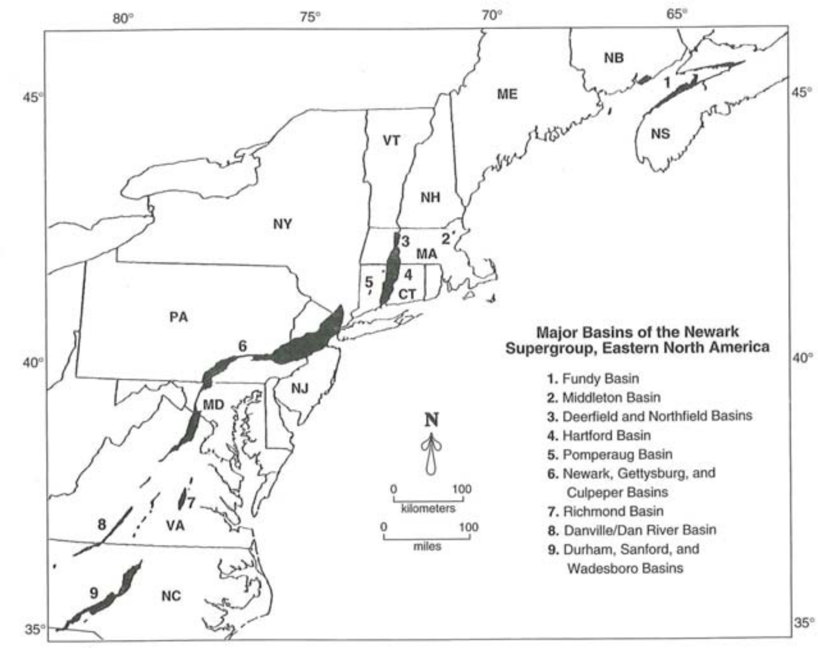


Figure 1.3 Exposed early Mesozoic (Newark Supergroup) basins of eastern North America. These basins formed as a result of lithospheric extension during the early phases of the continental rifting episode that ultimately created the Atlantic Ocean (from McDonald, 2010; after Unger, 1988).

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Field Stop 1: The Tilcon Traprock Quarry, North Branford.
Reprinted from Geological Society of Connecticut Field Guide #1, Chapter II, by A.R. Philpotts, 2010.

The Holyoke Basalt at the Tilcon Traprock Quarry, North Branford: The geology, petrology, and history of one of the world's largest flood-basalt eruptions.

Anthony R. Philpotts
Emeritus Professor of Geology and Geophysics
University of Connecticut

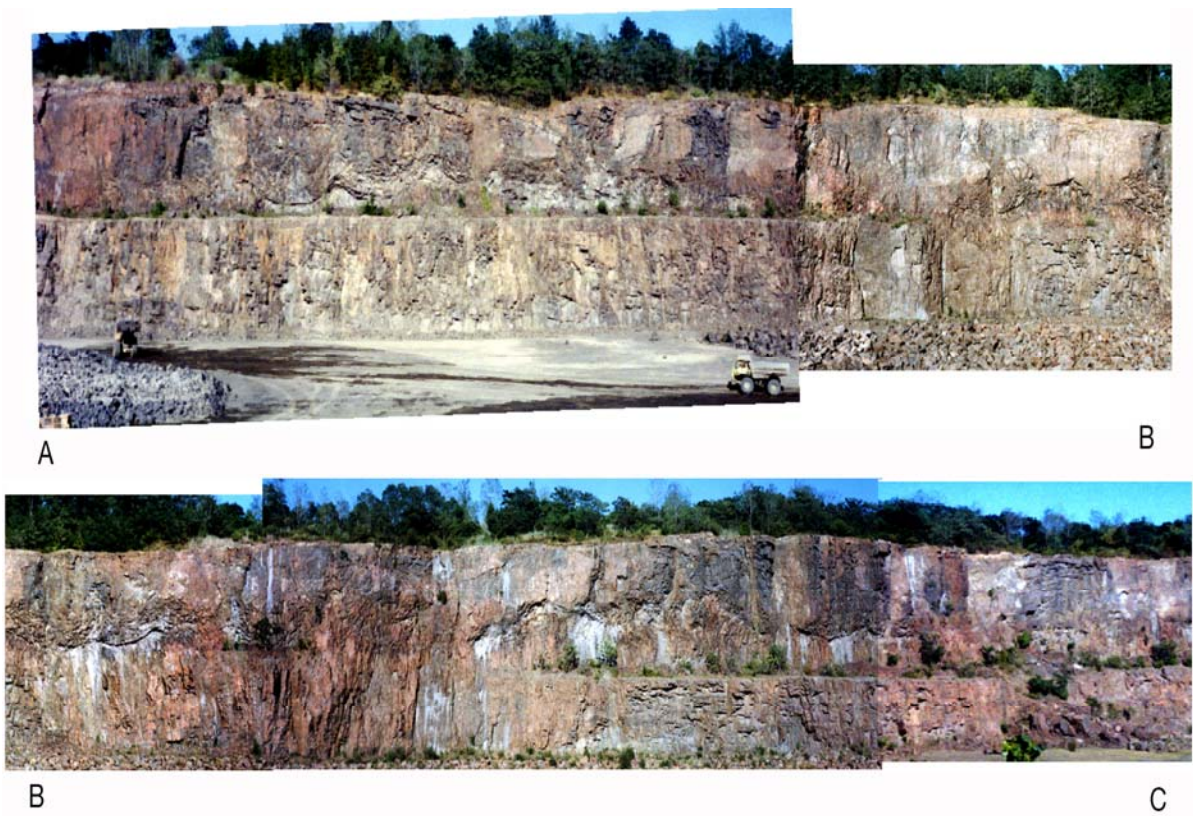

Figure 2.1 Two photographs A/B and B/C fit together to provide a panoramic view of part of the 8-km-long east wall of the North Branford quarry. Horizontal plane shown mid-way through the photographs is a quarried surface 20 m above the quarry floor, at approximately 120 m above the base of the 200-m-thick flow. The cuspate boundary seen in the quarry face marks the division between cooling joints that propagated down from the surface and up from the base of the flow. In the quarry wall beneath this boundary are meter-thick horizontal sheets of ferrodiorite that are slightly darker than the buff colored basalt. These sheets formed from liquid expelled from the lower part of the flow by crystal-mush compaction during solidification of the lava.

The North Branford quarry, operated by Tilcon of Connecticut Inc., is one of the largest trap-rock quarries in the world. It supplies crushed stone for roadbeds, asphalt, and concrete aggregate for use in Connecticut and New York States. The eastern face of the quarry, which purports to be the longest face of any trap-rock quarry in the world, provides a spectacular 8-km long vertical exposure through the central part of the Holyoke flood-basalt flow, the middle and thickest of the three Mesozoic basalt units in the Hartford Basin. Because of its great thickness, the Holyoke flow cooled slowly and differentiated to produce horizontal sheets of ferrodiorite and granophyre (granite) in its central part. This exposure provides an ideal site to witness the products of igneous differentiation where we have a clear understanding of the differentiation mechanisms with none of the complications normally associated with deep-seated plutonic bodies.

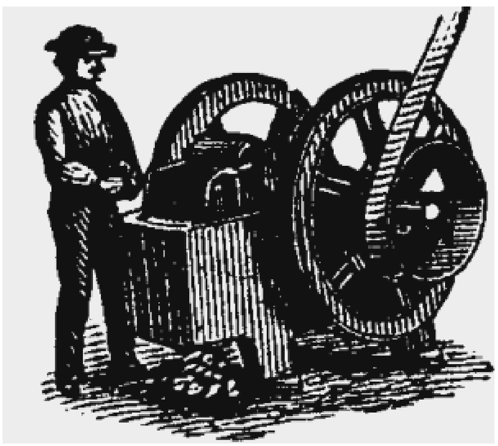

Figure 2.2 Eli Whitney Blake's stone crusher and advertisement from 1879. *Source*: www.gutenberg.org.

THIS NEW
BLAKE'S STONE AND ORE BREAKER AND CRUSHER.

For breaking hard and brittle substances to any size. Endorsed by the leading Mining, Manufacturing, and Railroad corporations in the United States and Foreign Countries. First Premium wherever exhibited, and hundreds of testimonials of the highest character.

A NEW SIZE FOR PROSPECTING AND LABORATORY USE.

☞ All Stone Crushers not made or licensed by us, containing vibratory convergent jaws actuated by a revolving shaft and fly-wheel, are infringements on our patent, and makers and users of such will be held accountable.

Address
BLAKE CRUSHER CO., New Haven, Conn.

The North Branford Quarry was opened by the New Haven Trap Rock Company in 1914 and has remained in continuous operation ever since, despite several changes in ownership. On opening, it produced 2000 tons of crushed stone per day; this has now risen to ten times that. The stone is crushed with a manganese steel gyratory crusher which is a descendent of the "jaw crusher" invented by Eli Whitney Blake, the nephew of Eli Whitney who invented the cotton gin and mass production of muskets. Today, 65-ton-quarry trucks bring the freshly blasted rock from the quarry face to the crusher where it begins a 2-mile-long trip on conveyor belts to other crushers and sorters. The major products are crushed stone for road and rail beds, and aggregate for asphalt and concrete. Basalt is, without doubt, the most desirable stone for all of these purposes, because of its durability and density. Its durability is a direct consequence of its texture, which is determined by the way in which the basaltic magma crystallized.

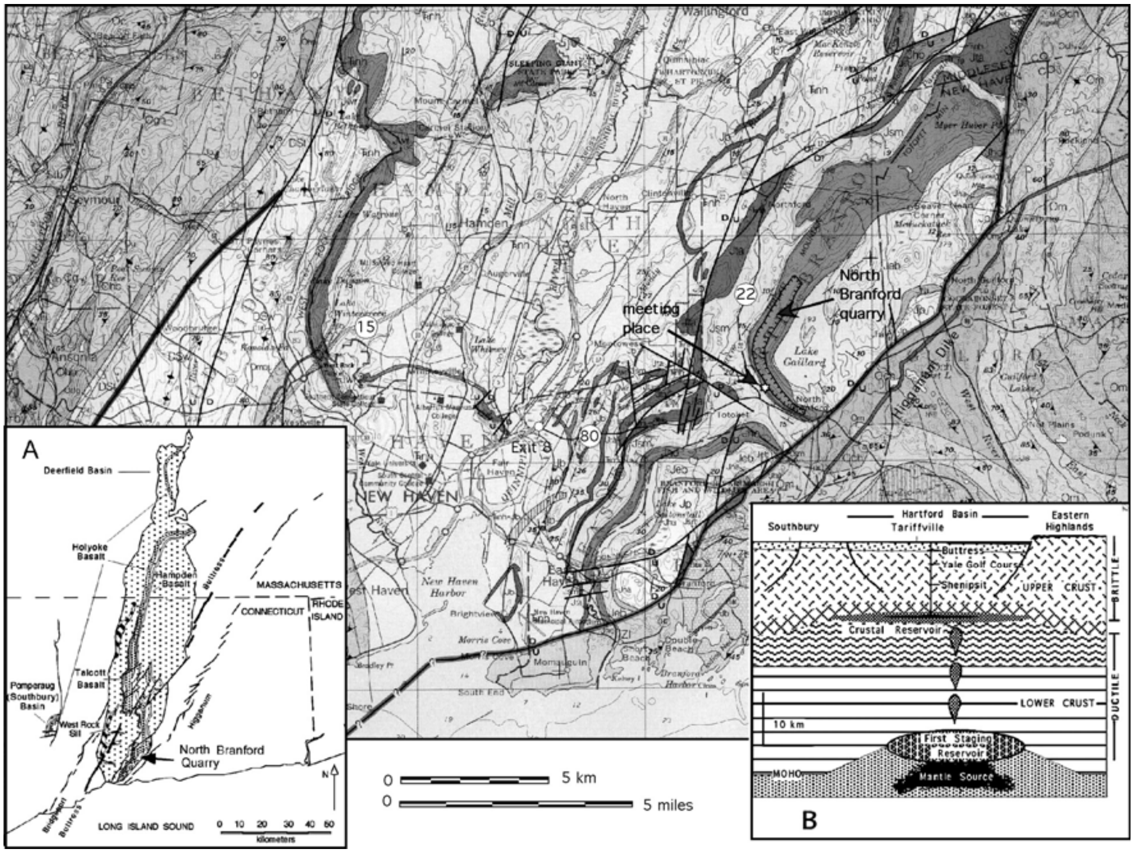

Figure 2.3. Geological map of the southeastern part of the Hartford Basin (Rodgers, 1985) and location of the North Branford quarry. ***Inset A*:** Distribution of the three volcanic units in the Hartford Basin –Talcott, Holyoke, and Hampden—and their associated dikes (Higganum, Buttress, and Bridgeport) and sills. ***Inset B*:** Hypothetical cross section through the basin at the time of the Holyoke eruption.

The North Branford quarry is located in the extreme southeastern part of the Mesozoic Hartford Basin, the southern end of the quarry actually terminating against the basin's eastern border fault. The quarry is in the Holyoke basalt, the middle of the three volcanic units in the basin. The lava erupted from the Buttress dike (Fig. 2.3A), which like the earlier Higganum and later Bridgeport dikes trend obliquely across the basin. These dikes formed perpendicular to the extension that eventually rifted apart Pangea to form the Atlantic Ocean.

The Holyoke basalt is the thickest of the three volcanic units in the basin, forming a single 200-m-thick flow in the North Branford area. This flow can be traced northward to the Vermont border and westward to the Pomperaug Basin. The volume of the flow is estimated to have been in excess of 1000 km^3. This gigantic sea of lava would have taken over 100 years just to solidify and much longer to cool to a temperature where steam was no longer gushing from fractures in its surface.

The entire 200-m-thickness of the flow is exposed at North Branford, with the base of the flow being exposed on the road leading from the quarry weighing station into the quarry itself, and the top of the flow being exposed along the shore of Lake Gaillard immediately to the east of the quarry. The flow erupted and cooled as a single unit. This is evidenced by the fact that only one generation of bubbles rose toward the top of the flow. If multiple magma pulses had inflated the flow, multiple layers of bubble accumulation would be found, and they are not. Fluvial sediments rapidly covered the flow, so there was no time for the friable scoriaceous and ropy surface features seen along the shore of Lake Gaillard to be eroded. Water from these sediments would have percolated down through fractures in the crust of the flow and promoted cooling.

When flood basalts solidify they generate characteristic fractures that propagate down from the surface and up from the base of the flow, as seen in the example from Iceland in Figure 2.4. Fractures propagating up from the base form extremely regular polygonal joints, which are referred to as the colonnade and the less regular ones extending down from the surface are referred to as the entablature. Because cooling is more rapid from above, the colonnade and entablature meet at a level, which is usually one-third the height of the flow.

The Holyoke also has a well-developed colonnade and entablature, but it is a little more difficult to see because of many additional tectonic joints. The colonnade/entablature boundary, however, is always evident and forms a prominent line along the entire east face of the North Branford quarry. This boundary, however, is at a height of 60% of the flow. This is one of the most important facts about the Holyoke flow, because it indicates that as it solidified, material was transferred from the roof to the floor of the solidifying sheet and hence displacing the final solidification level (colonnade/entablature boundary) upward.

Figure 2.4 *Left*: Colonnade and entablature joints in the 15-m-thick basalt flow at Aldeyjarfoss, Iceland. The colonnade/entablature boundary is one-third the height of the flow. ***Right*:** The colonnade/entablature boundary in the Holyoke basalt flow in the North Branford quarry, Connecticut. This cuspate boundary is at 60% of the height of this 200-m-thick flow.

Despite the fine-grain size of the basalt throughout the Holyoke flow at North Branford, this body of magma did not simply freeze *in situ*. It was an actively convecting body of magma that underwent extreme differentiation to eventually produce coarse-grained ferrodiorite and fine-grained granite (granophyre). As you stand on the floor of the quarry, picture yourself in the middle of a huge sheet of magma in which you can look up and see the roof slowly solidifying downward to form the cuspate colonnade/entablature boundary. The question is what was going on in this magma sheet at the time? Photomicrographs (Fig. 2.5) of samples through the flow provide answers.

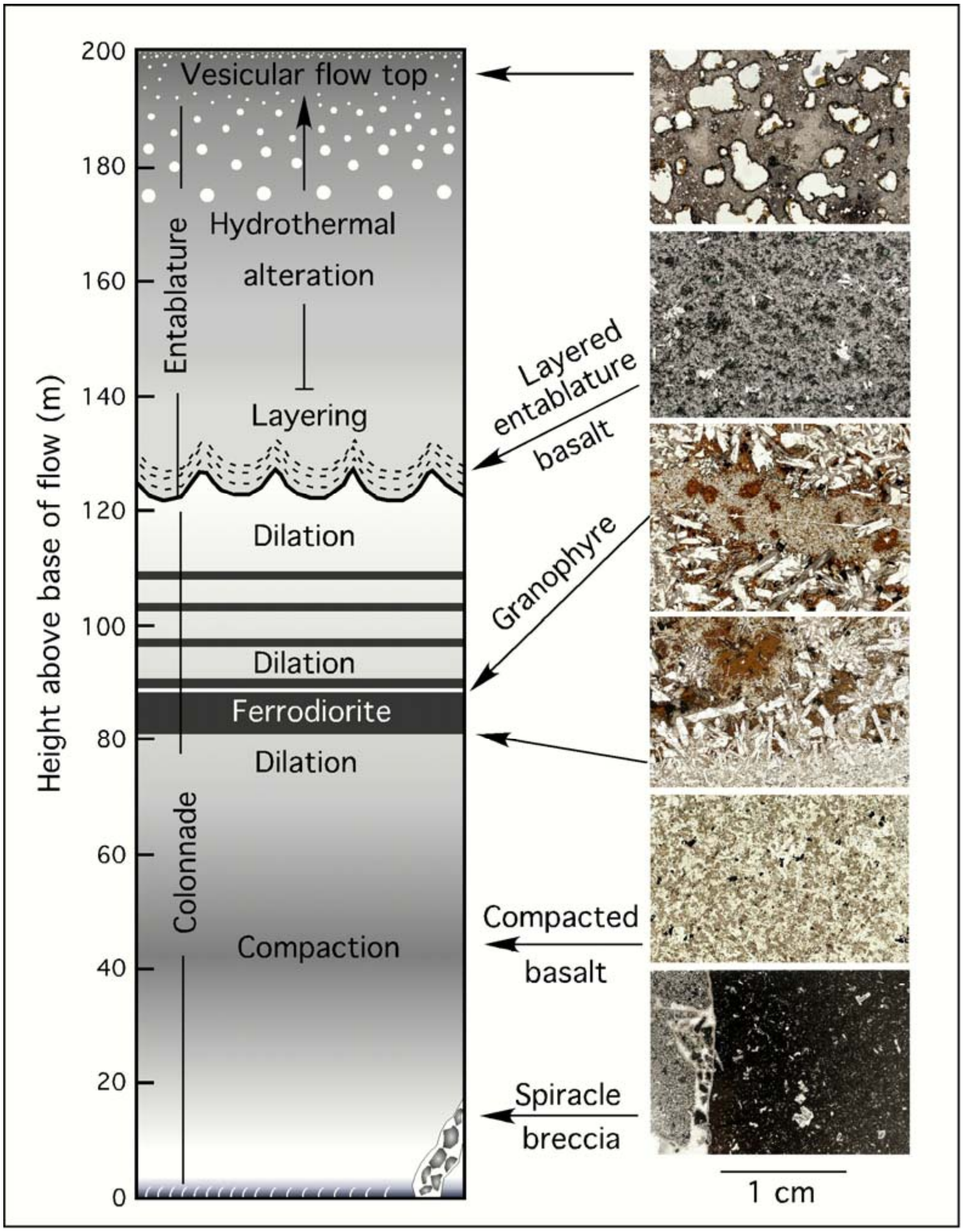

Figure 2.5 Photomicrographs of the different rock types in the Holyoke flow at the North Branford quarry under plane-polarized light and at a magnification given by the 1-cm scale bar.

***From top to bottom*:**

a. Vesicular flow top with altered plagioclase phenocrysts in a devitrified matrix.

b. Basalt of the entablature with small ophitic clusters of pyroxene and plagioclase crystals separated by dark patches of mesostasis. The ophitic clusters and patches of mesostasis form a horizontal layering.

c. Sheet of fine-grained granophyre in the upper part of a ferrodiorite sheet.

d. Lower contact of coarse-grained ferrodiorite sheet with fine-grained basalt.

e. Basalt of the colonnade, showing clusters of small plagioclase crystals surrounding patches of granular pyroxene and equant magnetite grains.

f. Spiracle breccia (*left*) with fragments of crystallized basalt and fragments (*right*) with quenched glassy margins.

Water percolating down through prominent fractures in the crust of the Holyoke flow produced “cold fingers” that promoted crystallization in the roof zone. These “cold fingers” distorted the isotherms into the cuspate shape now seen on the colonnade/entablature boundary. Dripping instabilities developed from the base of the cusps, with dense crystal mush sinking to the floor. This also dispersed crystals throughout the entire thickness of the flow, which is why the basalt is fine-grained throughout, despite the great thickness of the flow.

Although the fine-grained basalt in the quarry appears to be homogeneous, chemical analyses reveal a significant variation with height (Fig. 2.6). In rising from the floor, the basalt steadily becomes enriched in elements that enter early crystallizing minerals (e.g. MgO) and depleted in elements that enter the melt (TiO_2).

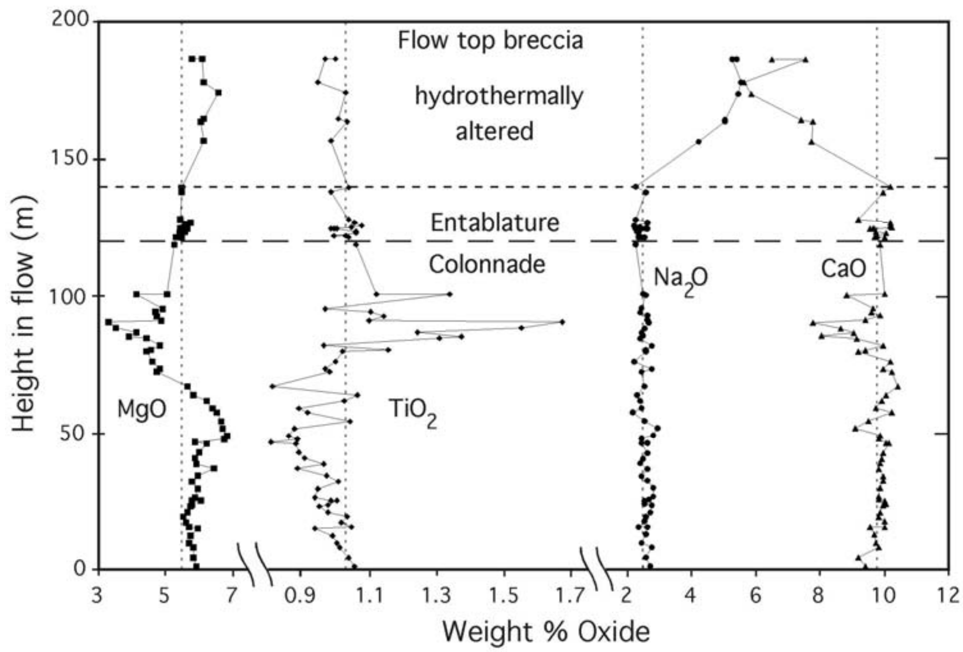

Figure 2.6. Chemical profiles through the Holyoke flow in the North Branford quarry.

About 75 m above the base of the flow, this trend reverses, and the basalt passes rapidly into a 10-m-thick sheet of coarse-grained ferrodiorite, which is depleted in MgO and enriched in TiO_2. Numerous decimeter-thick sheets of ferrodiorite occur at a regular 1-m spacing above this, continuing up almost to the colonnade/entablature boundary. Toward the top of most ferrodiorite sheets, thin centimeter-thick sheets of granophyre occur. At the colonnade/entablature boundary, the composition of the basalt is the same as at the chilled base of the flow. Above this, however, the basalt becomes severely altered and is depleted in CaO and enriched in Na_2O, presumably as a result of hydrothermal alteration brought about by water circulating through the crust of the flow as it cooled.

Calculations show that there is a mass balance through the enrichment and depletion profiles in the colonnade, and that the variation can be explained by the upward expulsion of residual liquid from a compacting pile of crystal mush. The degree of compaction in this crystal mush has actually been determined by quantitatively measuring textural anisotropy in oriented thin sections of the basalt. The maximum compaction measured in this manner is 12%, which occurs in samples with the lowest TiO_2. The liquid that was expelled from the mush dilated and eventually ruptured the basalt in the center of the flow to form the ferrodiorite sheets. Because these sheets were formed from liquid that contained few crystal nuclei, the resulting rock is coarse-grained, in contrast to the juxtaposed basalt which is fine-grained (Fig.2.4, 2.5). There are no rocks with compositions intermediate between the basalt and the ferrodiorite. The reason for this is that for compaction to take place, an interconnected network of crystals must exist. Melting experiments show that this occurs in the Holyoke basalt when it is one-third crystallized. The melt that is expelled at this degree of crystallization has the composition of the ferrodiorite.

The origin of the granophyre poses a problem. Small amounts of interstitial granophyre can be found in any sample of Holyoke basalt, and it is even more abundant in the ferrodiorite. The granophyre sheets typically occur near the top of ferrodiorite sheets and are composed of a fine-grained intergrowth of alkali feldspar and tridymite with minor fayalite and hedenbergite-rich pyroxene. Their association with the ferrodiorite suggests that they are derived from it. However, both of these rocks were formed from liquids, but there are no rocks with compositions intermediate between the ferrodiorite and the granophyre. How, then, could the ferrodiorite liquid change into the granophyre liquid without leaving a trace of intermediate compositions?

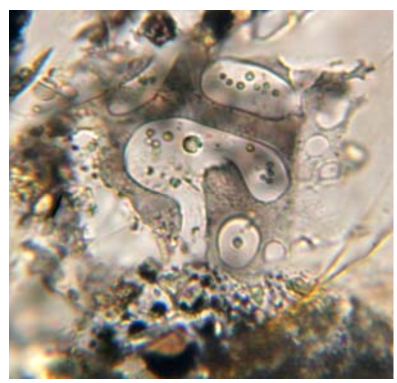

Figure 2.7. Photomicrograph under plane polarized light of glassy immiscible Si-rich (clear) and Fe-rich (brown) droplets in the mesostasis of the basalt in the entablature of the Holyoke flow. The composition of the Si-rich glass is identical to that of the granophyre sheets in the Holyoke basalt. The Fe-rich glass has the composition of an Fe-rich pyroxene with additional ilmenite and apatite. On crystallizing, this liquid could go undetected on the margins of crystals and in interstitial patches. Any resemblance of the large Si-rich droplet to Snoopy is purely coincidental. Width of field 66 mm.

A possible solution can be found in the entablature of the Holyoke flow, where rapid quenching preserved glasses with evidence of liquid immiscibility (Fig. 2.7). When the basalt in the entablature was approximately 75% crystallized, the residual liquid entered a 2-liquid field and split into immiscible fractions, one having an iron-rich pyroxene composition and the other a granophyre composition. In the center of the flow, where cooling was slower, this same immiscibility could have allowed a granophyric liquid to separate and accumulate toward the top of the ferrodiorite sheets. The conjugate iron-rich liquid would have sunk and accumulated amongst the crystals in the ferrodiorite where it would have crystallized to iron-rich pyroxene, magnetite, ilmenite, and apatite, which are abundant in the base of the ferrodiorite sheets (Fig. 2.5).

The following references may be useful if you want to read more about the information presented above.

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Field Stop 2: Summit of East Rock, New Haven.

This stop is on the top of the East Rock sill, a favorite viewing spot for sights of New Haven and Long Island Sound.

As we drive up to the summit, notice the columnar jointing in the walls of the road cuts. The joints dip steeply west because the sill dips gently east. The top of the sill has been glacially polished but long public use of the site has obliterated much of the evidence.

Looking south we can see Long Island if the day is clear. Long Island is the terminal moraine of the last ice advance. Long Island Sound is now saline but prior to the sea entering it was filled with the fresh water of Lake Connecticut. The shoreline of Lake Connecticut reached the level of the playing field immediately below the viewing point. Looking to the southeast you can see the lobe of a large delta that formed in Lake Connecticut from sediment carried by the Mill River that today runs along the west and south sides of East Rock before entering New Haven Harbor.

In the distance, to the south and south west, the Western Uplands can be seen emerging from beneath the Triassic and Jurassic fill of the Hartford Basin. The dip slope of the emerging uplands is the Triassic unconformity. Looking further west, the West Rock sill is visible. Both the East Rock and West rock sills are intruded into New Haven Arkose.

Field Stop 3: New Haven Arkose, intersection Whitney and Sherman Avenues, Hamden.

The stop is a large road-cut in the New Haven Arkose, an Upper Triassic alluvial sequence exposed on the entrance road to the Sports Facility of the York Hill Campus of Quinnipiac University, 0.2 miles SW of the intersection of Whitney and Sherman Avenues, Hamden. The rocks we examine are approximately 600 m above the base of the New Haven Arkose, which is about 2000 m thick in this part of Hamden.

The bedforms and lithologies in the outcrop (cross-bed sets of pebbly sandstone and conglomerate) are the deposits of a braided stream system. At this outcrop you will see lenticular cross-bedded sandstone and conglomerate channel-deposits interbedded with red sandy mudstones and siltstones that formed in the adjacent floodplain. The floodplain deposits, especially the red to maroon mudstones, are often overprinted by paleosol profiles. The paleogeographic distribution of the channel and braid-bar belts and the intervening flood-plain areas in the region is shown in Figure 3.1.

In the mudstone sequences that lie between the channel deposits, you will see the development of at least incipient paleosol profiles. These are often indicated by the development of vertical joints, known as peds, which occur in the upper portion of the mudstone. Peds were produced by the action of burrowing organisms and plant roots. Paleosol profiles are superimposed on all types of sediment, regardless of lithology, and represent the gradual overprinting of soil features on the primary depositional signal.

You can observe conspicuous bands of grey-green calcareous caliche as concretionary layers and horizons, and the infilling of old root cavities. The existence of old root cavities indicates that although the environment was seasonally dry, the floodplain supported a forest.

Near the base of the hill, at the northernmost part of the outcrop, you can see that the sandstone and shale appear baked and change color from red to a bluish-blackish hue. These sediments have been baked by a basaltic dike that is a branch of the Buttress Dike, which fed the Holyoke basalt. The actual contact is now covered by fill.

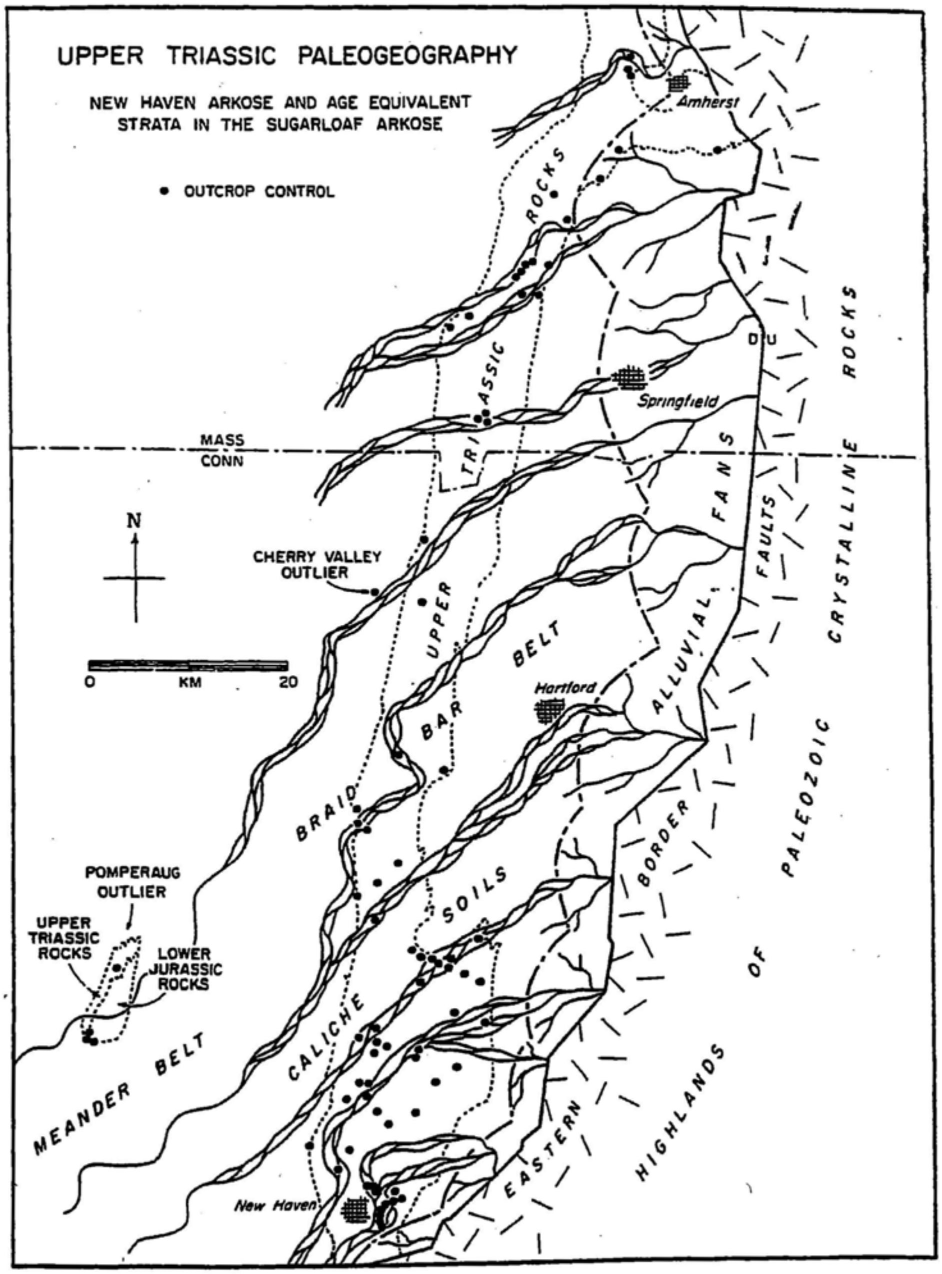

Figure 3.1 River morphology and dispersal patterns generalized for Late Triassic time in the Hartford Basin and Pomperaug outlier (after Hubert et al. 1978)

References

Hubert, J.F., Reed,A.A., Downdall, W.L., and Gilchrist, J.M., 1978, Guide to the Mesozoic Redbeds of Central Connecticut, State Geological and Natural History Survey of Connecticut, Department of Environmental Protection, Guidebook 4, 129 p.

Field Stop 4: Lighthouse Point, New Haven.

This stop is on the Light House granite gneiss, here somewhat broken and shattered due to its proximity to the Eastern Border Fault. The granitic gneiss is Proterozoic in age.

Standing on the shore and looking northwest, we see Morris Cove in the foreground to the right. On the northern side of the Cove stands Forbes Bluff, a small diabase sill that intrudes the New Haven Arkose. The Eastern Border Fault that separates the Hartford Basin from the Eastern Uplands, reaches Long Island Sound at Morris Cove, and passes close to the southern side of the Cove. The fault continues southwesterly beneath New Haven Harbor. Along the strike of the fault is a deep, steep-walled valley that has been determined seismically to reach a depth of 600 feet beneath sealevel. The valley was presumably carved out glacially by removal of fault gouges and breccias. The deep valley is now filled with sediment.

Beyond Forbes Bluff you can see East Rock and West Rock, both diabase sills that intrude into New Haven Arkose. The two sills dip gently east, having the prevailing dip of the Late Triassic and Early Jurassic arkoses they intrude. On the far horizon, metamorphic rocks of the Western Uplands can be seen emerging from beneath the Triassic unconformity, which has the same dip as the overlying sediments and sills. The easterly dips are due to movement on the Eastern Border Fault, which is a normal fault, west side down.

The city of New Haven occupies the level area in the center of the view. It is built on sediment deposited in a great delta that formed as sediment carried by the Mill River entered glacial Lake Connecticut.

Field Stop 5: Talcott Basalt, Meriden.

This stop is a superb exposure of the Talcott Basalt, the earliest of the three flood basalts preserved in the Hartford basin. The outcrop is a cut made for the parking lot behind the Target store at the west end of Cold Spring Avenue, Meriden.

The Talcott is the lower of the two great basalt flows that form the Hanging Hills. The sheer cliff of the upper two-thirds of the Hanging Hills is the Holyoke basalt. The Shuttle meadow Formation forms a swale between the lower, terrace forming Talcott flow and the upper cliff-forming Holyoke flow. The Talcott lies on the top of the New Haven Arkose.

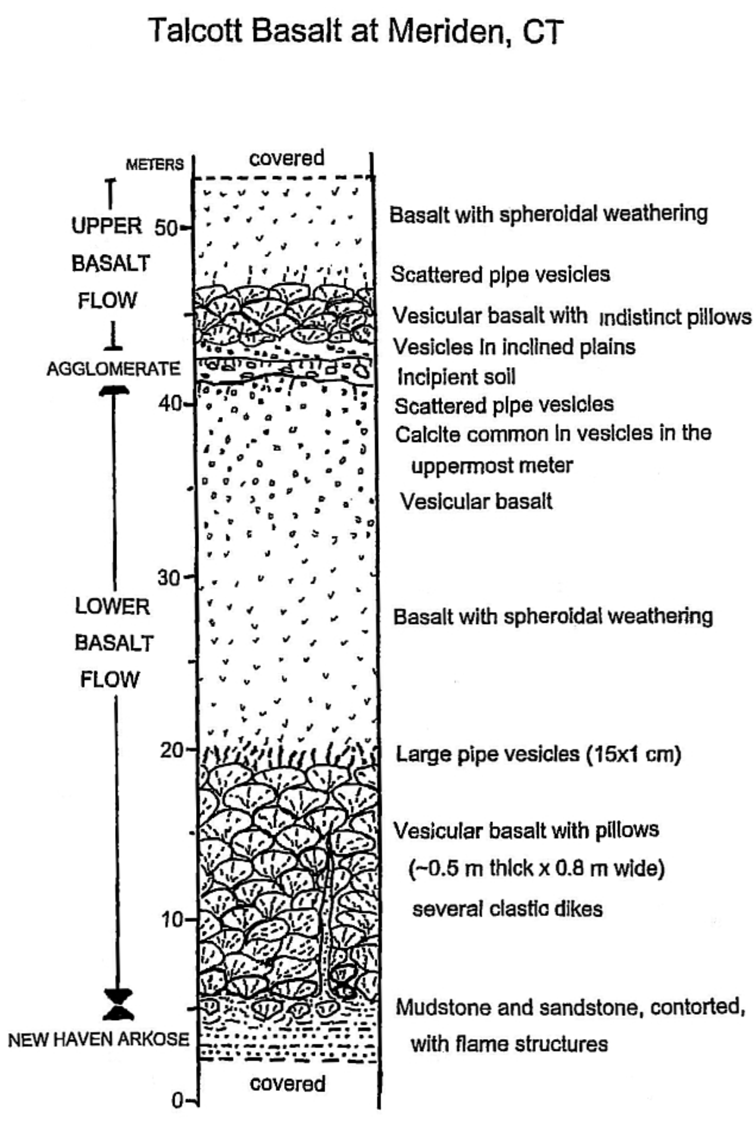

Figure 5.1. Measured section of the Talcott Basalt and uppermost New Haven Arkose behind the Target store, Meriden. (composite from Hickey and MacClintock, 2005, and Hubert et al., 1978)

The Triassic-Jurassic boundary is located in the uppermost New Haven Arkose, within a meter of its contact with the Talcott. The Talcott represents a beginning of basaltic volcanism in the Hartford Basin and tholeitic volcanism of remarkably similar composition and vesicularity began within a 21,000 year time span at the beginning of the Jurassic Period in the eastern North American rift basins, from Culpepper, Virginia to the Fundy Basin in Nova Scotia (Olsen and Gore, 1989, p.11).

The Talcott basaltic flow is about 67 m thick in this area (Hanshaw, 1968, p. 2), but only the lower two-thirds of the flow is exposed in the cut. The Talcott consists of two flows, separated by 0.5 to 0.8 m volcanic agglomerate. The Target cut exposes a continuous section of approximately 47 m, including the entire lower flow, 0.8 m of agglomerate, and the lower 12 m of the upper flow. Pillow structures and pipe vesicles are especially well developed in the lower flow, as are occasional clastic dikes (Fig. 6.1).

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

A thin agglomerate layer resting on top of the lower flow records a pause in volcanic activity, with reworking of the volcanic terrain. A zone of weathering at the top of the agglomerate appears to represent the remains of an incipient soil. This was overridden in turn by the upper flow, which vaporized the water in the soil and produced prominent vesicles in the basal half-meter. Indistinct pillows are present in the lower few meters of the upper flow, but these are hard to distinguish from the spheroidal weathering that is found higher in the upper flow.

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Field Stop 6: The Hanging Hills, Meriden. (Provided the site is accessible)
(Description after the late John Rodgers, Guidebook for the 77th Annual Meeting of the New England Intercollegiate Geological Conference, 1984, pp.398-400).

The stop is the East Peak of the Hanging Hills. The base of the tower that stands on the peak is 955 feet above sea level. We stand approximately on the top of the Holyoke basalt flow, which holds up all the highest hills in the Connecticut Valley except those of the West Rock sill, Sleeping Giant of Mt Carmel, and the Barndoor intrusive complex of Avon, and Granby.

The characteristic double jointing of the Holyoke flow is particularly well displayed here. The older columnar joints show narrow altered (silicified?) selvages, which weather in slight relief above the rock face and the joint itself: the younger systematic joints cut undeflected across the columnar joints, implying that they had been entirely healed. Large boulders of this rock showing the characteristic jointing patterns are scattered over the countryside from the Hanging Hills to long Island Sound. The well known “Judges Cave” on top of West Rock in New Haven is a group of such boulders, so placed as to provide some shelter: in 1661, three regicides (Dixwell, Goffe, and Whalley) hid there from the agents of Charles II, following restoration. The regicides are now remembered by having 3 of New Haven’s main road named after them.

The Talcott flow forms the bench below, between us and the lake in Hubbard Park. Between the Talcott and Holyoke flows is a sedimentary unit consisting of maroon colored, shaley siltstone, and cross-bedded sandstone, of the Shuttle Meadow Formation. The Hampden basalt flow, the uppermost of the three flood basalts, forms low ridges in the country to the northeast, beyond the dip slope on the Holyoke flow, but they are not clearly visible from East Peak. The city of Meriden, spread out below to the southeast, and all the country to the south, is underlain by the New Haven Arkose, and is drained by the Quinnipiac River to Long Island Sound at New Haven Harbor.

In the Hanging Hills the Holyoke flow and the strata above and below strike nearly east-west and dip gently north, in strong contrast to their normal north-south strike and gentle easterly dip elsewhere in the Connecticut Valley. The change in strike and dip is evidently associated with the particularly intense faulting in the Meriden region and especially with the large sinistral (left) offset in map pattern caused by the largest of these faults. (Rodgers mentions in an aside that “Sinistral offset does not prove sinistral strike slip, of course”).

Looking to the east, across Lake Merimere, we see South Mountain, on which the Holyoke flow is displaced only a little to the left from what we are standing on: half hidden behind South Mountain is Cathole Mountain, on which the Holyoke flow is displaced still more. The largest fault (or group of faults) then offsets the Holyoke flow 8 miles to the northeast, to the north end of Lamentation Mountain, the northernmost of the north-south mountains in the middle distance, where the Holyoke flow resumes its normal strike and dip. The main Hartford line of the New Haven Railroad and the Berlin Turnpike (visible at the foot of the mountain) go through the gap between, which is the lowest divide into the Connecticut River drainage basin (175 ft). Chauncey Peak, the south end of Lamentation Mountain, is slightly offset from the rest. A series of faults offset the Holyoke flow and various roads go through the gaps so formed, I-95 goes through a gap south of Higby Mountain, Route 66, the New Haven Railroad to Middletown, Route 17 and Route 80 all pass through fault gaps in the Holyoke flow.

Behind Higby and Lamentation Mountains, and off to the northeast are the Eastern Uplands, separated from the Connecticut Valley by the Eastern Border Fault. The Eastern Uplands were the chief source of the Triassic and Jurassic sediments in the Hartford basin. Due east of our view point one can make out the break in the Uplands at Middletown, where the Connecticut River turns away from the Connecticut Valley to find its way through the Uplands to the Sound.

In the opposite direction, the Holyoke flow extends west to West Peak (1,024 ft above sealevel) and then turns abruptly north, resuming its normal strike and dip. Thence it extends north for many miles, though broken and somewhat offset by faults, forming Talcott Mountain west of Hartford and reaching Mt. Tom and Mt. Holyoke, on opposite sides of the Connecticut River in Central Massachusetts—these can be seen from the top of the tower on a very clear day.

Beyond West Peak is a valley underlain by New Haven Arkose, and behind that the Western Uplands. To the south, however, the West Rock sill appears, first as low hills within the valley, then higher and higher in front of the Western Uplands until Mount Sanford reaches the skyline and hides them. Out in the valley southwest of Mount Sanford is the large mass of Sleeping Giant, an irregular sill or stock higher in the New Haven Arkose than the West Rock sill and thought to be nearly above the main basement feeder dike.

From the latitude of the Hanging Hills south, the hills upheld by the Holyoke lava flow and the Sleeping Giant and West Rock sills reach heights that decline steadily southward, reaching sealevel around New Haven Harbor; the slope is about 45 ft per mile (8 meters per kilometer). From the Hanging Hills north, however, no peaks on the Holyoke flow reach 1,000 feet until Mt. Tom (1,200 ft.) and the Holyoke Range; the slope from West Peak to Mt. Tom is only about 4 feet per mile (less than a meter per kilometer). The sloping hill-top surface to the south is continuous with the surface beneath the Cretaceous rocks on Long Island. For this reason, from the Hanging Hills Long Island appears higher than any of the hills between—and the area south of the Hanging Hills therefore represents the Fall Zone; even the highest hills to the north have been reduced by erosion well below this surface. One can there imagine that when erosion was going on, Cretaceous rocks still reached inland as far as Meriden.

Field Stop 7: Brownstone quarries, Portland, Connecticut .
(Adapted from Geological Society of Connecticut Field Guide No.1, Chapter 3, by Peter M. LeTourneau)

The Portland brownstone quarries are on the eastern side of the Hartford Basin, less than 3 kilometers from the Eastern Border Fault. The quarries and nearby exposures illustrate the diversity of the sedimentary environments at the rift margin (Figures 7.1 and 7.2). Within 1 kilometer of the quarries, boulder conglomerate, organic-rich, laminated black shale, and fluvial sandstone may be observed. The quarries occupy a position on the western edge of the Crow Hill Fan complex that has its depositional center on the top of Crow Hill in the town of Portland. The eastern edge of the fan complex can be observed along Rte 17 in Portland near the golf course, where boulder conglomerate is exposed in the road cuts and natural outcrops. The southern edge of the fan complex is observed along the north side of Rte 66 in the vicinity of the miniature golf course where thinning and fining wedges of the fan conglomerate lithosomes are observed. The Connecticut River laps against the northern flank of the fan complex in the vicinity of Petzold's Marina, where a long stratigraphic section of interbedded alluvial fan and deep water lacustrine deposits can be seen.

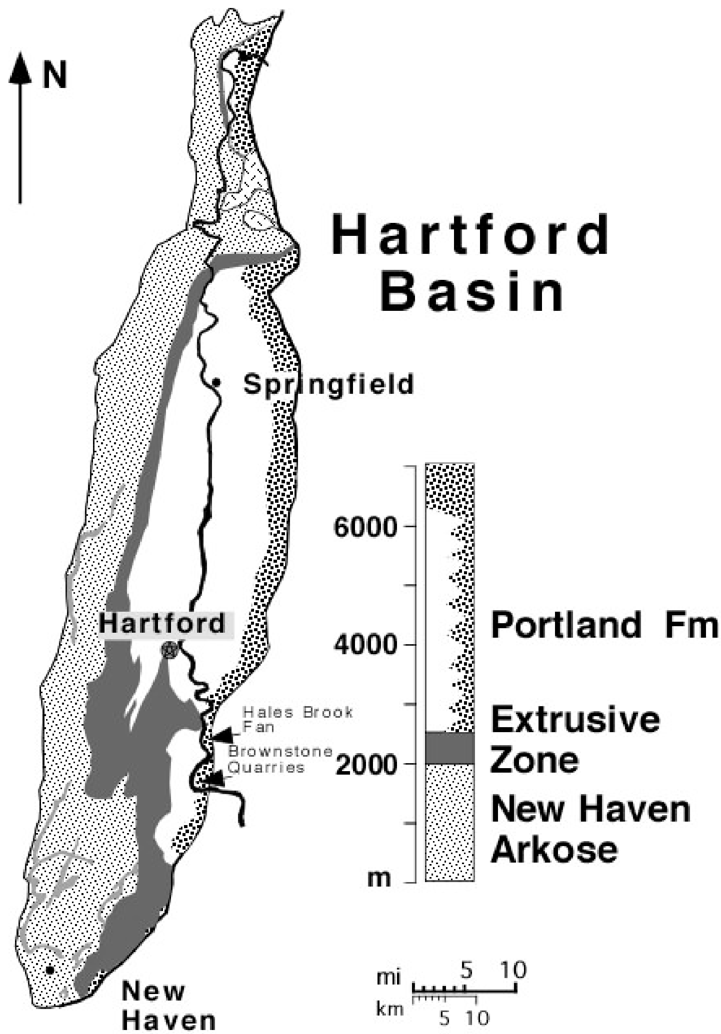

Figure 7.1 Summary of Hartford basin stratigraphy and location of the Portland brownstone quarries (from LeTourneau, 2010).

The Portland brownstone quarries consist of two large, water-filled pits. The main (northern) pit consists of the Middlesex quarry on the north side of the promontory and the Brainerd quarry on the south side. South of Silver St, which bisects the quarries, are the Shaler and Hall workings. On the extreme northern side of the Middlesex quarry, and well above the water line, are the modern workings of the Meehan quarry. A definitive and captivating account of the history of the brownstone quarries can be found in Guiness (2003), and a discussion of the rocks in the context of the paleogeographic distribution of alluvial fan complexes along the rift margin in central Connecticut (Figure 7.3) can be found in LeTourneau (1985).

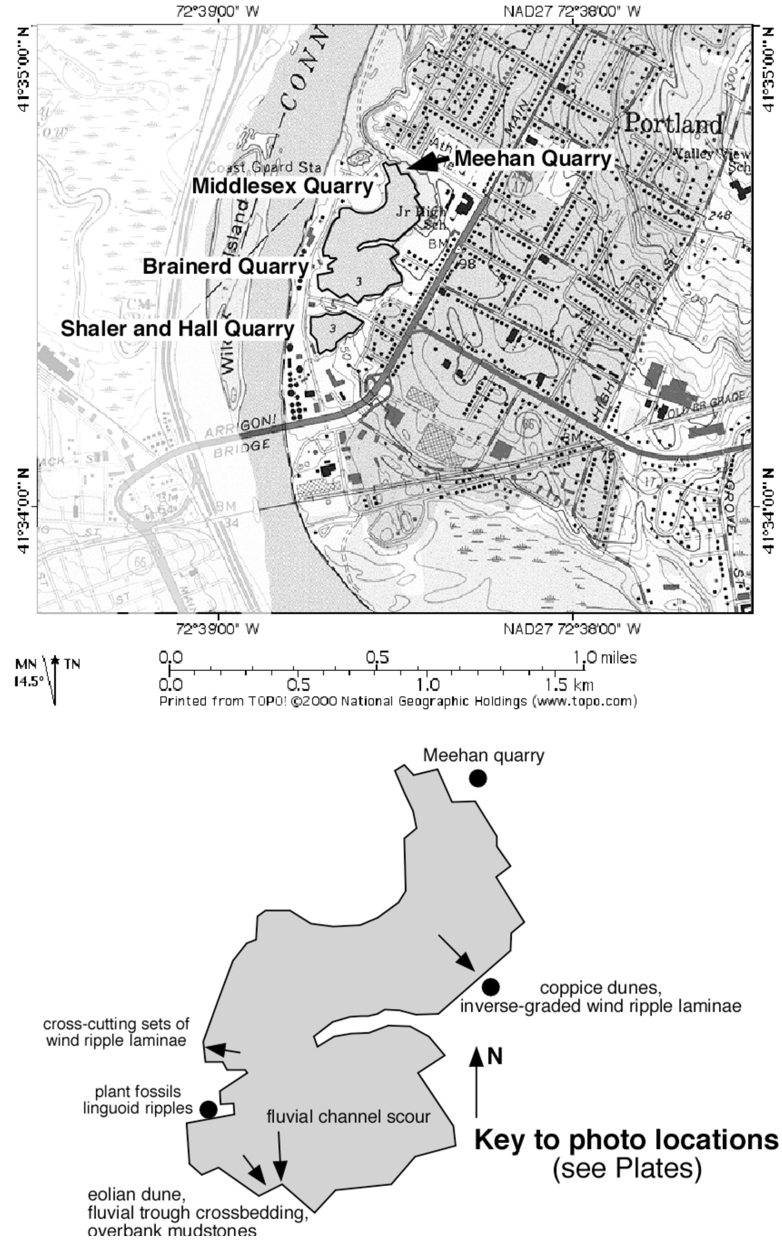

Figure 7.2 Location of the Portland brownstone quarries (from LeTourneau, 2010).

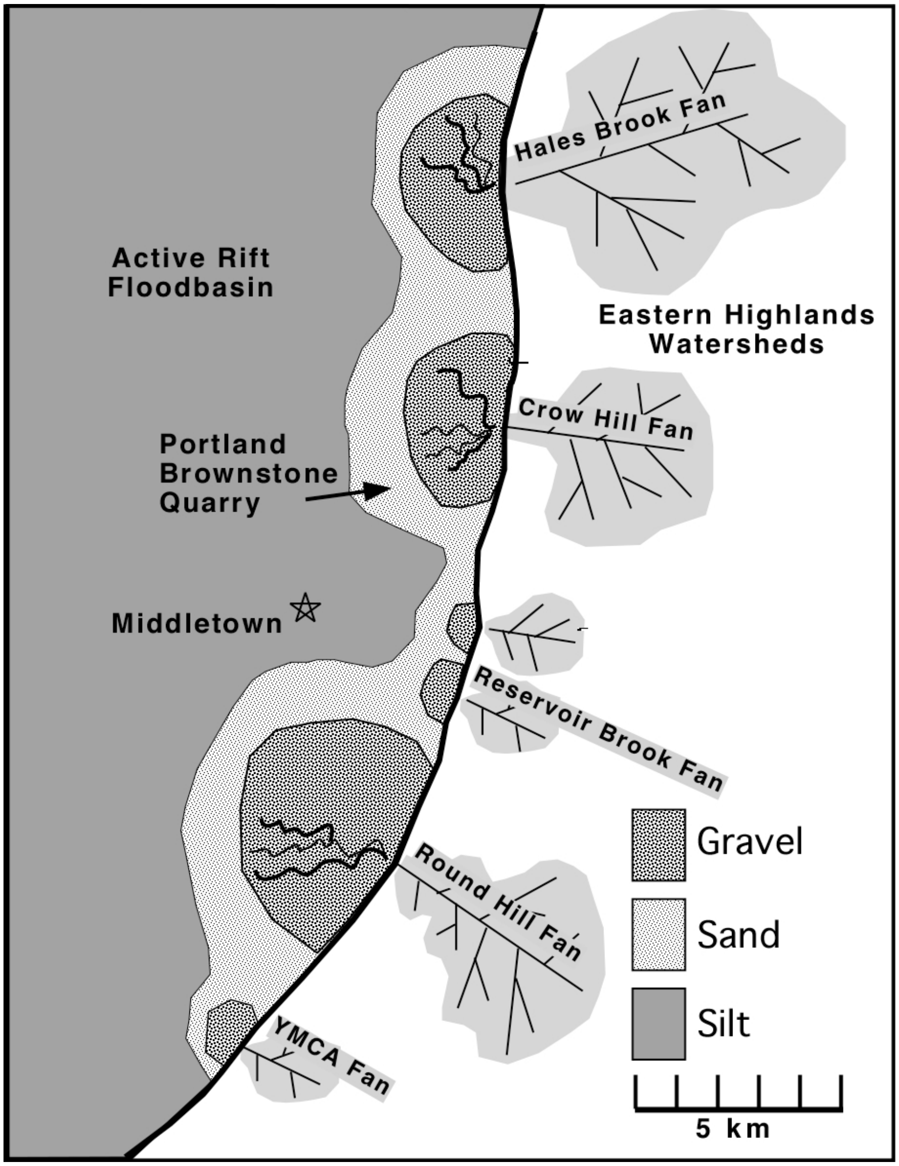

Figure 7.3 Paleoenvironmental reconstruction (from LeTourneau, 2010).

The quarries are in the Portland Formation, the uppermost stratigraphic unit in the Hartford basin. The rocks are arkosic sandstones ("Portland arkose"), containing quartz and feldspar grains with calcite and hematite cement. Long-regarded as a fluvial deposit, the quarry exhibits many features indicative of stream or river environments, including sand and gravel channels, gravel bars, and overbank sand and silt. Evidence of episodes of desiccation includes mudcracks and dinosaur tracks.

Careful examination of sedimentary features reveals that eolian deposits are a significant component of the sequence. These features, which have been discussed by LeTourneau and Huber (2006), include sedimentary features attributable to sand sheets, low angle dunes, and linear "coppice" dunes. The eolian beds were apparently preferred for building stone because of their uniform grain size and even texture. Fluvial beds about 15 m thick alternate with eolian beds, indicating possible cyclic climatic control on deposition.

A close view of the large east wall of the Middlesex pit reveals an abundance of inverse-graded, low angle, inclined planar stratification, indicative of migrating wind-ripples (pin-stripe lamination). In addition, several enigmatic, large-scale, convex-up dune forms can be observed. Letourneau and Huber describe these unique sedimentary structures as “coppice dunes”, formed around clumps of plants. Evidence for the coppice origin of the features includes complex internal stratification with root traces, and inverse-graded wind ripple lamination. A modern model for the Portland brownstone eolian deposits is the Stovepipe Wells dune field in Death Valley, California. There a relatively thin sheet of dune sand overlaps alluvial fan and fluvial deposits on the side of the extensional basin. Small coppice dunes anchored by plants are found in the interdune area.

Eolian sedimentation in the upper part of the Portland Formation at Portland was promoted by both favorable a paleolatitudinal position, deposition within the dry climate interval of a 405 ky climate cycle, and proximity to fan-related sources of sand. The deposits formed at about 20^0 paleolatitude, on the arid side of the estimated 10^0 arid-humid climate boundary based on the evaporation-minus precipitation models of Crowley and North (1991).

<u>A brief history of The Stone that Shaped America</u>

No building material is more closely associated with American cities than brownstone from the Mesozoic basins. Brownstone was a highly fashionable and desired building material during the late 1800’s. The golden age of urban development coincided with the peak desire for the warm brown stone, and as a result brownstone buildings are common in cities along the eastern seaboard.

The Connecticut River flows through a central valley that contains thick deposits of red beds and in places along the river, cliffs of red and brown arkosic sandstone can be observed. The Portland Brownstone quarry is located on the banks of the Connecticut River and early settlers noticed that the fine-grained sandstone was good for making mill stones or for structural use in buildings. In Portland, the chance occurrence of fine eolian sandstone and a ready means of water transport were responsible for the growth of a regional industry and a national trend in architectural fashion. Other red-bed quarries are found elsewhere in the Hartford Basin and in the rift valleys of eastern Virginia, north-eastern Pennsylvania, and New Jersey, but none rival the fine quality and color of the stone found in Portland.

Stone was produced from the quarry beginning in the late 1700s and by the late 1800s the brownstone industry was employing hundreds of workers from Wales, Germany and elsewhere. Stone was cut in the quarry then placed in storage areas for “seasoning”, which allowed natural ground water in the stone to seep out, slowly firming the stone in the process. Unseasoned stone, if used immediately, was subject to peeling and cracking as the water in the rock froze in the winter months. The cut and seasoned stone was hauled by horse and cart, and later by steam-powered cable winches, to barges on the riverbank. The sail-powered barges, called “brownstone schooners” hauled the stone to cities all over the east coast, with most of the stone destined for New York City. Almost every city on the east coast of the United States has fine examples of buildings made of Portland Brownstone, and fine examples can even be found in San Francisco and Denver.

During the years of its peak use as building stone, demand forced the main quarry to ship material of less than the highest quality, including “unseasoned” stone. As the price of Portland Brownstone increased, it was used as a “facing” stone, or thin veneer, a practice that allowed water to seep behind the bedding planes, causing the stone to rapidly peel or break apart. Portland stone quickly gained a new, but undeserved reputation as an unstable building material. But buildings that were made of high quality stone that was properly seasoned and correctly placed are sound and beautiful more than 100 years later. A notable example is the Villard house, now the front of the Helmsley Palace hotel on Madison Ave, in Manhattan.

The Portland quarries closed when a catastrophic flood of the Connecticut River in 1938 inundated the quarries and all the equipment. The end of the quarries was actually in sight at the time of the flooding because the popularity of the stone was in decline. In 1994 a new chapter opened when the Meehan quarry was opened adjacent to the Middlesex quarry. Stone from the new workings are mainly used to repair older classic buildings. The old quarries were designated a National Historic Landmark in 2000, and the Portland Brownstone quarries are now used as a park and recreational area.

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ROAD LOG
THE HARTFORD BASIN FROM THE HANGING HILLS TO THE SOUND: ARKOSE PLAINS AND FLOOD BASALTS

Total	Elapsed	Description/Directions
0.0	0.0	Leave Marriott Hotel/ Hartford Convention Center. Drive north on Columbus Street.
0.2	0.2	Turn right on entrance to I-91 South.
0.6	0.4	Colt Firearms Company headquarters with its Turkish tower on the right.
3.2	2.6	Crossing a meander cut-off of the Connecticut River.
8.9	5.7	Road cut in the Early Jurassic East Berlin Formation.
11.8	2.9	Road rises to follow the south trending ridgeline of the Hampden Basalt.
16.8	5.0	Crest of Holyoke Basalt ridge. The swale to the south is underlain by the Early Jurassic Shuttle Meadow Formation.
18.7	1.9	The edge of the flat here is formed by the contact between the Early Jurassic Talcott Basalt and the Late Triassic New Haven Arkose.
20.2	1.5	Exit 16 to the Wilbur Cross Parkway (DO NOT TAKE!) Driving south on the New Haven Arkose. The high ridge to the east is the Metacomet Ridge, formed by the Holyoke Basalt flow.
25.9	5.7	Take Exit 14 on right. Turn right at base of ramp on East Center Street.
27.3	1.4	Turn left on to Woodhouse Avenue and CT 150 South.
28.2	0.9	Cross under I-91. In 1.3 miles the road crosses the Muddy River, an important post glacial stream in the area, and then two small hills formed by Jurassic gabbros.
30.6	2.4	Turn right at intersection on to CT 150 South (Woodhouse Avenue).
31.4	0.8	Junction of CT 22 South. Bear left on Woodhouse Avenue, now CT 22. The low rise ahead lies on the Talcott Basalt.
31.6	0.2	Northford and the Junction of C 17. Turn right on Middletown Avenue and then bear left immediately on to CT 22 South and Forest Road. The road now follows the valley of the Farm River, which cuts into the Shuttle Meadow Formation, between the Talcott Basalt to the right and the Holyoke Basalt ridge on the left.
36.1	4.5	Turn left into the Tilcon North Branford Quarry and **Field STOP 1**.

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36.1	0.0	Leave entrance to Tilcon North Branford Quarry and turn left on to Ct 22 South and then immediately turn right on CT 80 (Foxon Road).
36.4	0.3	Main New Haven water filtration plant on left.
37.6	1.2	Road cuts through several faulted outcrops of the Talcott Basalt
37.8	0.2	Cross North Haven Town Line. For the next 1.5 miles the road traverses a complexly faulted and poorly exposed sequence consisting of New Haven Arkose, Talcott Basalt, and the Shuttle Meadow Formation.
39.3	1.5	From here westward the bedrock consists of the New Haven Arkose with gabbro dikes and small lacccoliths.
39.8	0.5	Highway cuts a gabbro dike.
41.8	2.0	New Haven Arkose and a gabbro dike on the left.
42.2	0.4	Outcrop of New Haven Arkose on the right.
43.1	0.9	Overpass of I-91 and CT 17, keep straight on what is now Middletown Avenue (CT 17 South).
43.6	0.5	Bridge over the Quinnipiac River.
44.0	0.4	Turn right on to Ferry Street, cross RR bridge and turn right on State Street (US 5 North).
45.2	1.2	Turn left at the traffic light on Ridge Road, start up the hill formed by the Jurassic gabbro of the East Rock sill.
45.5	0.3	Bear left at traffic light on to Davis Street.
45.6	0.1	Turn left on to East Farnum Drive and the entrance to East Rock Park.
45.7	0.1	Turn left on to the summit access road.
46.5	0.8	**Field STOP 2.** EAST ROCK SUMMIT. View of New Haven, the Sound and Long Island from the top of the East Rock Sill.

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46.5	0.0	Leave STOP 2 and retrace route to East Rock Park entrance at East Farnum Drive and Davis Street.
47.7	1.2	Turn right on to Davis Street.
47.8	0.1	Turn right on to Ridge Road.
47.9	0.1	Bear left at fork, staying on Ridge Road.

48.1	0.2	Turn left at the traffic light on State Street (US 5 North). The road proceeds north along the western shore of Glacial Lake Quininipiac, on the right. This lake had been filled and drained by 5,400 yrs ago and a mixed hardwood forest grew on it (Tiffney and Pierce, 198?). The slope on the left is underlain by the New Haven Arkose.
50.0	1.9	Mound on left marks the site of the Stiles Clay Pit, a major producer of Brick for New Haven and Yale University. The varved clays that represented the initial fill of the lake were succeeded by deltaic and flood plain deposits and then by brackish water swamp deposits about 2,250 years ago, when a hardwood forest was replaced by brackish-tolerant Spartina.
51.1	1.1	CT 40 crosses State Street on the overpass. Take first left after this on to Devine Street.
51.2	0.1	Left entrance ramp on to CT 40 East.
52.1	0.9	Outcrop of New Haven Arkose in cut with numerous shallow braided channels and extensive caliche development.
53.0	0.9	Exit right to CT 10 North (Whitney Avenue).
53.9	1.0	Turn left on to Shepherd Avenue.
54.3	0.4	Turn left into parking lot of Farmington Canal Trail and **Field STOP 3.** NEW HAVEN ARKOSE AND THE BUTTRESS DIKE AT THE ENTRANCE TO THE YORK HILL CAMPUS OF QUINNIPIAC UNIVERSITY.

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54.3	0.0	Turn right onto Shepherd Avenue from parking lot of Farmington Canal Trail
54.5	0.2	Turn right on to Ct. 10 South (Whitney Avenue).
55.2	0.7	Right exit on to the ramp for Ct 40 East.
56.2	1.0	New Haven Arkose outcrop again, best views on the left side.
58.5	2.3	Cross the Quinnipiac River and keep right on exit ramp to I-95 South. We will be travelling south on the east side of Glacial Lake Quinnipiac. The East Rock Sill is visible on the right (west).
62.5	4.0	Cross the Quinnipiac River and ascend a low rise formed by the Wisconsin-age delta of the Mill River, which dammed the Quinnipiac to produce Lake Quinnipiac and which forms the hill on which Fair Haven is built.
64.2	1.7	Keep left on to I-95 North.
65.5	1.3	Cross the Pearl Harbor Memorial Bridge, now being replaced by a massive new structure, and take Exit 50, to the right, on to Frontage Road.
65.7	0.2	Turn right on Woodward Avenue.

67.2	1.5	Turn right into Fort Hale Park and overview of the New Haven Harbor, Morris Cove, and Lighthouse Point.
67.2	0.0	Exit parking lot and turn right on to Fort Hale Park Road. The hill ahead and to the right is a Jurassic gabbro intrusion into the New Haven Arkose.
67.4	0.2	Turn right on Townsend Avenue.
68.2	0.8	Cross the Eastern Boundary Fault, Morris Cove on the right. The preglacial course of the Quinnipiac River crosses the road here. Street becomes South End Road.
68.5	0.3	Bear 45 degrees to the right here on to Lighthouse Road.
68.6	0.1	Pardee-Morris House on left. Burned by the British July 5, 1779, rebuilt 1780.
69.0	0.4	Enter Lighthouse Point Park AND **FIELD STOP 4**. LIGHTHOUSE POINT AND THE EASTERN BOUNDARY FAULT. (Lunch)

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69.0	0.0	Exit park. Turn right on Lighthouse Road.
69.5	0.5	Turn left (45°) onto South End Road.
70.2	0.7	South End Road becomes Townsend Avenue.
71.0	0.8	On the right is the 200 year-old Townsend House in a style known as Cottage Gothic. In the 19th Century this property extended downhill to the shore of the estuary. The Townsends were New Haven bankers and early investors in Colonel Drake's Pennsylvania Rock Oil Company.
71.3	0.3	Fort Wooster Park and Beacon Hill on the right. The hill is supported by a Jurassic gabbro intrusion. This was a sacred burial ground of the Quinnipiac Indians, who met the English colonists who settled New Haven on the shore below the site on April 24, 1638. It is also arguably the earliest land in America reserved for the Indians, having been set aside in November of that year. As late as the middle of the Nineteenth Century, members of the dwindling tribe returned to this hill in the summers.
71.8	0.5	Turn left onto Frontage Road, enter I-95 South and keep right across the Pearl Harbor Memorial Bridge.
73.2	1.4	Keep right and then left on to I-91 N.
90.0	16.8	Exit 18 on the right for I-691 West. Hanging Hills formed by the Holyoke Basalt lie to the north (right).
92.5	2.5	Turn right on Exit 6 to Lewis Avenue, stay in center lane.
93.7	1.2	Turn left in the outer lane on to Lewis Avenue north, stay in middle or right lane past entrance to Meriden Mall, then move left.

94.0	0.3	Turn left on Kensington Avenue.
94.5	0.5	Turn left on Chamberlain Highway.
94.6	0.1	Turn right on Coldspring Avenue.
94.8	0.2	Turn left into Target parking lot and **Field STOP 5**. TARGET LOCALITY. Talcott pillow basalt in contact with the New Haven Arkose.

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94.8	0.0	Leave Target parking lot by the south entrance.
95.1	0.3	Cross the Chamberlain Highway on to entrance ramp for I-691 West.
95.9	0.8	Long outcrop of the upper New Haven Arkose on the right.
97.0	1.1	Take ramp on right to Exit 8, turn left at base of ramp on to West Main Street.
98.5	1.5	Turn left into Hubbard Park entrance road.
98.7	0.2	Turn right past the pond.
98.9	0.2	Turn left on to Reservoir Avenue.
100.0	1.1	Turn left at dam on to Percival Park Road and proceed up the dip slope of the Holyoke Basalt.
100.9	0.9	Keep right to East Peak.
101.5	0.6	Parking lot for Castle Craig and **Field STOP 6** HOLYOKE BASALT AT THE HANGING HILLS OVERLOOK, MERIDEN.

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101.5	0.0	Leave Castle Craig parking lot retrace route to the entrance to Hubbard Park.
104.5	3.0	Turn right on West Main Street.
105.6	1.1	Turn right on ramp to I-691 East. The road runs east along the upper contact of the New Haven Arkose. The Hanging Hills formed by the Holyoke Basalt lie to the north (left).
110.6	5.0	Cross under I-91. I-691 now becomes CT 66 East Washington Street.
111.6	1.0	The road starts up a low ridge underlain by the Talcott Basalt. The swale ahead is formed by the Shuttle Meadow Formation.
112.2	0.6	Ridge top of the Holyoke Basalt. The road descends a long dip slope of the Holyoke Basalt.
113.3	1.1	The swale with small lakes covers the East Berlin Formation.

113.5	0.2	The road crosses the Early Jurassic Hampden Basalt, the highest of the three Hartford Basin flows.
117.3	3.8	Wesleyan University on the right.
117.5	0.2	CT 66 East turns left at Main Street.
117.8	0.3	Follow Main Street (CT 66) north to the ramp for the Arrigoni Bridge.
118.4	0.6	Cross the Connecticut River on the high bridge. View of the Bronson Hill anticlinorium ahead and across the Connecticut River. To the right on the ridgeline is Isenglass Hill, whence came the urananite crystals from which B. B. Boltwood of Yale University determined the first U/Pb date in 1906.
118.8	0.4	Turn left on Sidon Street.
119.0	0.2	Turn right on Brownstone Street.
119.3	0.3	On right is Old Portland Quarry flooded in the Great Hurricane of 1936.
119.6	0.3	Arrive at **Field STOP 7** PORTLAND ARKOSE QUARRY, EAST PORTLAND.

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119.6	0.0	Leave STOP 7.
120.2	0.6	Turn left on Sidon Street.
120.6	0.4	Turn right on CT 66 West (Main Street).
	0.7	Turn right on to ramp for CT 9 North. The bedrock here consists of the Portland Arkose.
	6.5	Take Exit 20, on the right to I-91. The country from here to Weathersfield (Exit 24) is underlain by highly faulted basalt flows and sedimentary rocks. Mount Tom can be seen in the far distance to the north, with the plain of Glacial Lake Hitchcock in front of it.
	12.8	Exit29 left to Columbus Street and the Convention Center. Turn left on Columbus Street.
	0.2	Arrive at the Marriott Hotel and **END OF TRIP** on right.

TEMPERATURE-TIME PATHS TIE THE TALES OF TWO FORELANDS: THE NARRAGANSETT AND HARTFORD BASINS

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INTRODUCTION

One of the largest impediments to tying the tectonic history of stratified foreland rocks to their hinterland is the lack of time constraints on processes in the hinterland. The fossil record in many foreland settings is precise enough to reveal the times and rates of deposition and subsidence, but parallel temporal information dating the deformation and the exhumation of rocks in the hinterland is generally lacking. This obstacle is being overcome in southern New England, where 7.5 minute quadrangle mapping has been supplemented by many detailed geochronologic, structural, and petrologic studies. In this field trip we integrate the data from these temporal, thermal, and structural studies to develop pressure-temperature-time paths. These allow us to correlate metamorphic and deformational events in the hinterland with sedimentary and metamorphic events in the sedimentary foreland. On this trip we examine rocks near the Bronson Hill terrane in north-central Connecticut (Figs. 1, 2) that were involved in both Carboniferous and Jurassic events. For the Carboniferous event we will show evidence for regional, SE-verging, ductile thrusting that $^{40}Ar/^{39}Ar$ cooling ages show is out-of-sequence (Wintsch et al., 1993). The timing of this thrusting correlates well with the times of both the deposition and metamorphism of sediments in the Narragansett basin (Skehan, 2008) and suggests that these Pennsylvanian sediments were deposited in the foreland of those thrusts. For the Jurassic, we will show evidence for the local provenance of the clasts in the Portland arkose, demonstrating that these sediments were in a foreland setting adjacent to the same Bronson Hill rocks. Thus the rocks of the Bronson Hill terrane were the hinterland both for the Pennsylvanian sediments in the Narragansett basin to the east and for the Mesozoic arkoses in the foreland setting of the Hartford and Pomperaug basins to the west.

TEMPERATURE-TIME PATHS AND GEOLOGIC SETTING

Our approach to interpreting the geology of southern New England is guided by a comparison of contrasting temperature-time (T-t) paths derived from thermochronologic analysis of minerals in the rocks of Rhode Island and eastern and western Connecticut. Integrated, these analyses define the times and rates of cooling of metamorphic terranes (Fig. 3; Wintsch et al., 2003b). The boundaries of discontinuities in amphibole cooling ages outline thermotectonic terranes, and the cooling ages of these domains define the times of exhumation, from which relative motions can be inferred. By combining these results with kinematic indicators that span a range of metamorphic grade, we have identified several discrete thermotectonic terranes in eastern Connecticut and inferred their SE transport direction (Wintsch et al., 1992; 1993; 2003a; black arrows, Fig. 1).

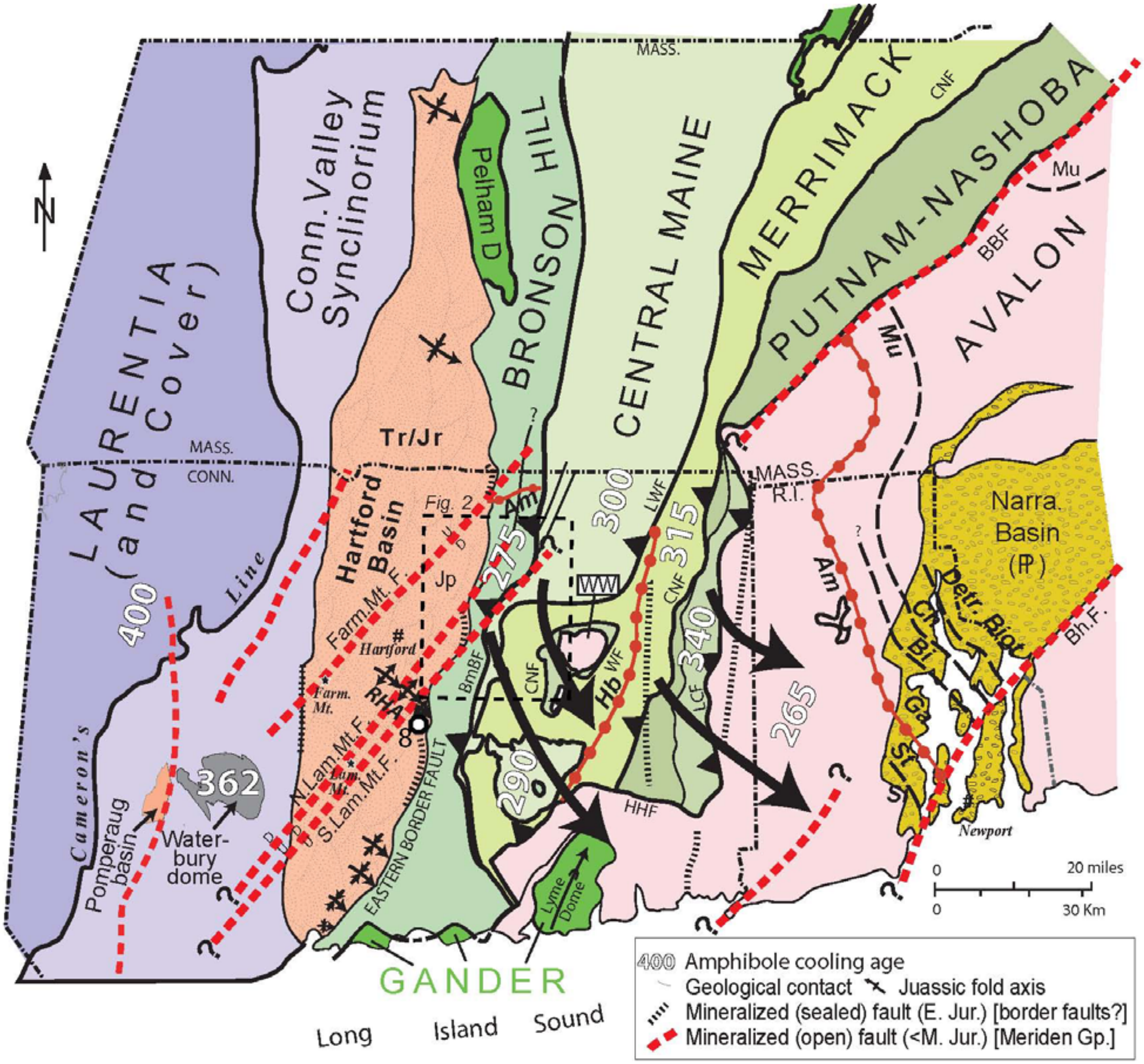

***Fig. 1**. Terrane map of southern New England modified from Wintsch et al. (1998) and Walsh et al, (2007). Bold black lines with teeth on the west side mark ductile faults at the boundaries of thermotectonic terranes. White numbers indicate approximate amphibole $^{40}Ar/^{39}Ar$ cooling ages of the terrane indicated (from Wintsch et al., 1993; Dietsch et al., 2010). Bold black arrows show the direction of late Paleozoic motion of the Gander cover terranes (Putnam-Nashoba, Merrimack, Central Maine, Bronson Hill) over each other and over the Avalon terrane. Beaded red lines show amphibole (Am) resetting isopleths, with Am on the high temperature side. Similarly, long dashed black lines show resetting isopleths of muscovite (Mu), both from Attenoukon (2009). Dashed red lines are brittle faults inferred to be post-Jurassic in age because some cut early Jurassic sediments. Lamentation block defined by the northern and southern Lamentation Mountain faults from Davis (1898). Farmington Mountain fault from Simpson (1966). Post Early Jurassic anticlines and synclines are indicated with black arrows, with the Rocky Hill anticline (RHA, Resor and de Boer, 2005) identified. Mineralized fault zones are indicated with black dashed lines. Metamorphic zones in the Narragansett (Nara) basin include detrital biotite Detr. Biot; chlorite, Ch; biotite, Bt; garnet, Ga; staurolite, St; and sillimanite, Si. Other abbr.: BBF, Bloody Bluff fault; BhF, Beaverhead fault; BmBF, Bonemill Brook fault; CNF, Clinton-Newbury fault; HHF, Honey Hill Fault; Jp, Jurassic Portland Formation; LCF, Lake Char fault; LWF, Lake Wangumbaug fault; WF, Willimantic fault; and WW, Willimantic window.*

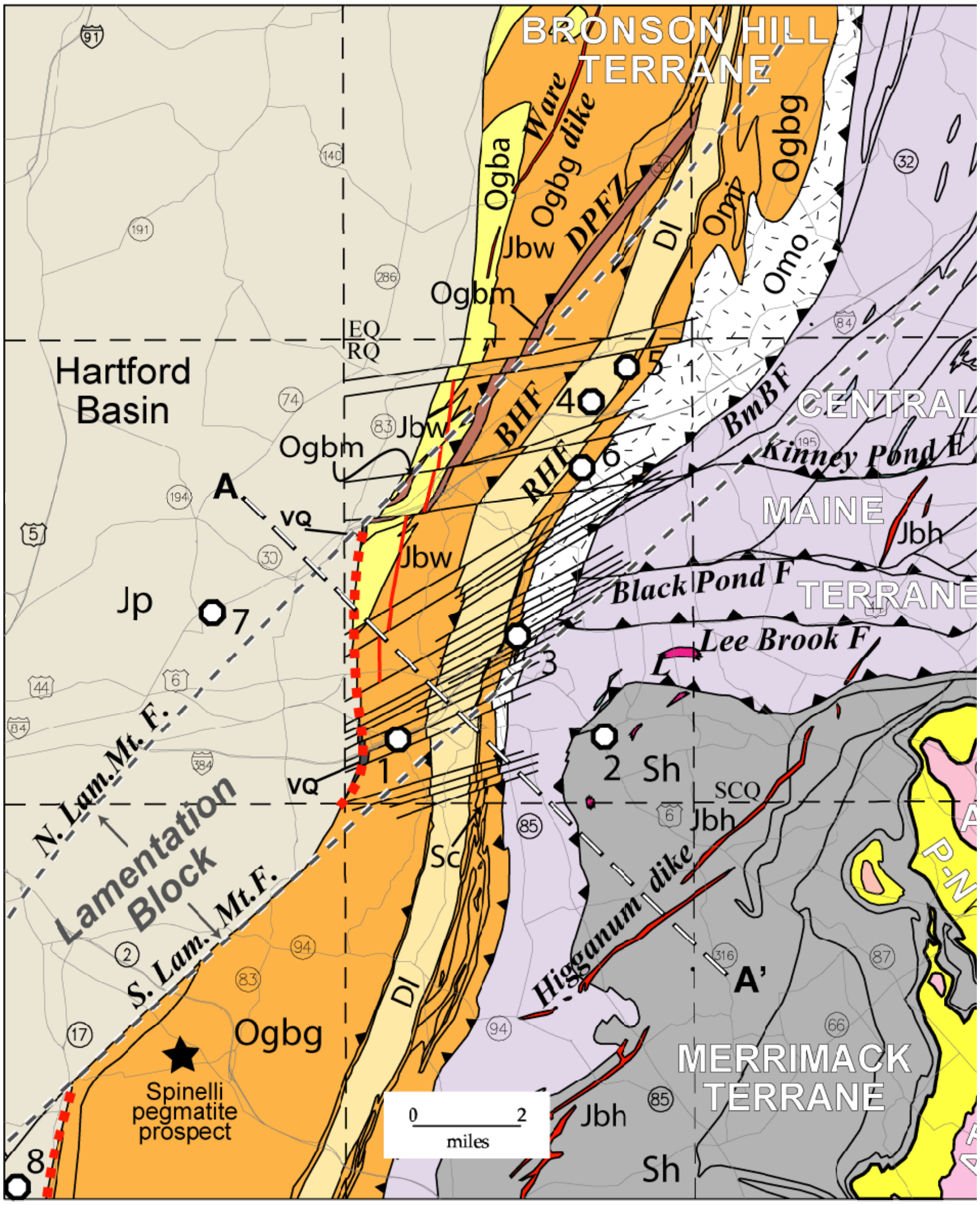

Fig. 2. *Bedrock geologic map of the Rockville and adjacent quadrangles from Wintsch et al. (2011). Contacts between all shaded units are ductile faults, active throughout the Permian, Triassic, and possibly Cretaceous. Position of the Lamentation Mt. faults extrapolated from Davis and Griswold (1894) and interpreted here to be Late Jurassic. Solid lines are brittle faults interpreted to be Cretaceous. Field trip stops are circled and numbered 1-8. For abbreviations, see text and Wintsch et al. (2011).*

In western Connecticut, however, a single T-t path can satisfactorily describe the post-Early Devonian thermal history of all the rocks of marginal Laurentia (Fig. 3). This is significant because it shows that all of the deformation or prograde metamorphism in these rocks occurred prior to the early Devonian (Sevigny and Hanson, 1993;1995). Thus, by the late Paleozoic these rocks were cold (Fig. 3) and strong enough to act as a backstop against most Alleghanian deformation and metamorphism. Consequently, this Alleghanian deformation was restricted to the still warm and even warming rocks to the east, where shortening caused the stacking of thrust nappes (Fig. 1).

Southeastern New England

Geochronologic evidence for Devonian (Acadian) metamorphism in rocks of the Bronson Hill arc (or Kingston arc of Aleinikoff et al., 2007) in Massachusetts is strong (Tucker and Robinson, 1990), but evidence in the same rocks east of Hartford is remarkably lacking, in spite of intense geochronologic study (Aleinikoff et al., 2002; Wintsch et al., 2003b). In fact several lines of evidence suggest that prograde metamorphism in these rocks occurred during the late Paleozoic Alleghanian orogeny. Evidence derived from rocks in the vicinity of this field trip include (1) monazite extracted from a pegmatite in the Littleton schist on the north side of Bolton Notch (1 km w of Stop 3) produced whole-grain U-Pb ages of ~305 Ma (Coleman et al., 1997). A similar pegmatite is exposed at Stop 4; (2) whole grains of sphene from the Fitch Formation from just south of Bolton Notch yielded a U-Pb age of ~305 Ma (Coleman et al., 1997); and (3) metamorphic sphene grains replacing magmatic sphene from the Glastonbury Gneiss near the Spinelli prospect (Stop 4, Wintsch et al., 1998a) are no older than ~290 Ma (Wintsch et al., 2005a). Exposures of similar Glastonbury Gneiss are present at Stop 1. Together these data indicate that prograde metamorphism and high-grade fabric development in the Bronson Hill rocks in the Rockville area are not Acadian but are Pennsylvanian and Permian in age, which we infer to be Alleghanian (Fig. 2). Permian cooling ages of amphibole in these rocks (Wintsch et al., 2003a) define the last time amphibolite facies metamorphic conditions prevailed in the Rockville area, and other thermochronometers show rapid cooling to ~200°C in the Late Triassic (Fig. 2). Amphibole ages from terranes to the east grow progressively older, from ~275 Ma in the Bronson Hill terrane to ~340 Ma in the Putnam-Nashoba terrane (Fig. 1).

The prograde metamorphism was caused in part by the loading of several thrust nappes verging to the SE (Wintsch et al., 1993; 2001; 2003a; 2011). Evidence for this ductile motion on this trip is present in the form of NW plunging stretching lineations that are associated with kinematic indicators (especially S-C fabrics and deformed boudins) indicating top to the SE motion. NE trending pegmatites and veins further support SE vergence. First the fractures opened to the SE, allowing emplacement of the vein material, and then top to the SE rotation of veins and pegmatites shows subsequent SE motion reactivating foliation planes. Such evidence is common in rocks of the Rockville quadrangle, especially a progressively C-W rotated pegmatite (stop 4) and even sheath folds (stop 6).

In SE New England prograde Alleghanian metamorphism is demonstrated by the metamorphism of Late Pennsylvanian (Cantabrian to Stephanian C, ~306-303 Ma, Lyons, 1984; Wagner and Lyons, 1997) sediments of the Narragansett basin (Rhode Island Formation) that lie unconformably upon Avalonian rocks in Rhode Island and adjacent Massachusetts (Fig. 1). The depositional age of these sediments overlaps with the time of cooling and SE thrusting of the terranes in eastern Connecticut (Wintsch et al., 1993). Peak metamorphic temperatures of these sediments (e.g. Burks et al., 1998; Skehan, 2008) were reached at ~280 Ma (Dallmeyer and Takasu, 1992), similar to the time of peak metamorphism of the rocks of this trip (Fig. 2). The overlapping times thrusting and cooling in eastern Connecticut with the times of depositions and metamorphism in Rhode Island are consistent with the deposition of these sediments in the foreland of eastern Connecticut's thrusts, followed by the overriding of the sediments by those thrusts (Wintsch and Sutter, 1986).

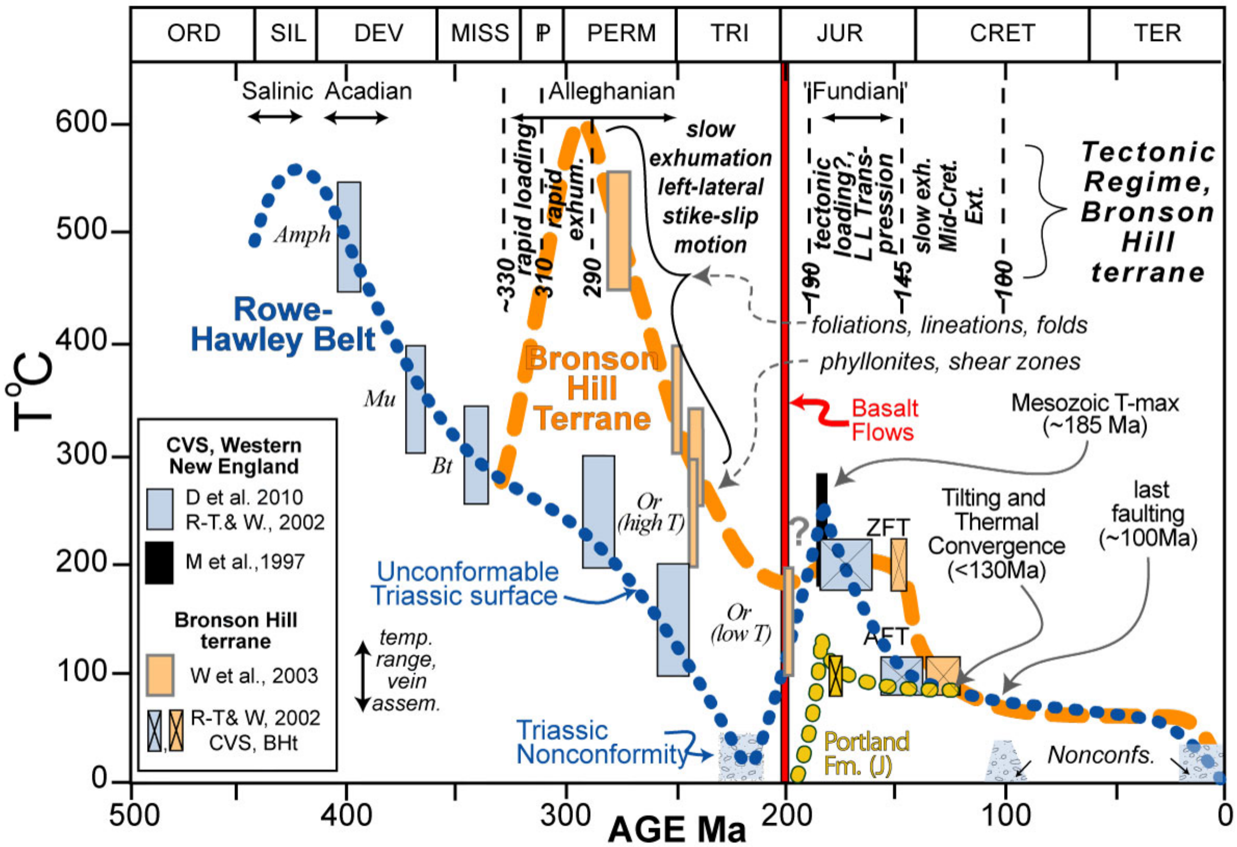

__Fig 3.__ A comparison of the thermal history of the rocks at the unconformity under Late Triassic sediments the Connecticut Valley synclinorium (CVS) in western New England (compiled in Roden-Tice and Wintsch, 2002) with that of rocks in the Bronson Hill terrane in the southern Rockville quadrangle area (modified from Wintsch et al., 2003a). Changes in the tectonic regime of the Bronson Hill terrane in the Rockville area are indicated. Temperature estimates of mapped structures applied to the BH curve (dashed arrows) leads to estimates of their relative and absolute timing. Shaded boxes refer to the closure ages of the minerals: Am, amphibole; Mu, muscovite; Bt, biotite; Or, orthoclase; ZFT, zircon fission track; AFT, apatite fission track. Abbr.: M et al., 1997, Merino et al., 1997; D et al.2010, Dietsch et al., 2010; R-T & W 2002, Roden-Tice and Wintsch, 2002; Nonconfs, Nonconformities; W et al., 2003, Wintsch et al., 2003.

Southwestern New England.

In contrast to the late Paleozoic history of SE New England, the rocks of western Connecticut (Fig. 1) cooled much earlier, with peak temperatures reached before the Middle Devonian (e.g. Dietsch et al., 2010; Fig. 3). Prior to this cooling, these rocks had a complicated Taconic, Salinic, and Acadian metamorphic and deformational history (e.g. Sevigny and Hanson, 1993;1995; Dietsch et al., 2010; Hibbard et al., 2010). By the middle of the Devonian, however, these rocks shared a rather uniform cooling history constrained by $^{40}Ar/^{39}Ar$ cooling ages of amphibole, micas, high- and low-temperature K-feldspar (Clark and Kulp, 1968; Dietsch et al., 2010; Kunk, pers. comm.) and by the Late Triassic unconformity (Fig. 3) on the western margin of the Pomperaug basin and on the south-western margin of the Hartford basin (Fig. 1).These metamorphic rocks project under the Triassic non-conformity in south-central Connecticut and so must have shared the thermal overprint of the overlying Mesozoic sediments (Roden-Tice and Wintsch, 2002). Accordingly, the T-t curve for these rocks is carried into the Mesozoic as marked by the blue dotted line of Fig. 3. The interpolated thermal history of the Jurassic rocks of Stops 7 and 8 is shown by the yellow dotted line in Fig. 3. Because the rocks of western Connecticut (Laurentian margin rocks) were cold and thus strong by the Late Paleozoic, they acted as a buttress against which the thrust nappes east of the Hartford basin were repelled (Fig. 4).

Mesozoic.

By Early Jurassic the assembly of Pangea was complete. However, during the Permian the approach of Gondwana had shifted from counter-clockwise with generally a dextral shear margin to a clockwise motion, causing the north-south shortening and sinistral, transpressive displacement along the Appalachian margin (Withjack and Schlische, 2005).

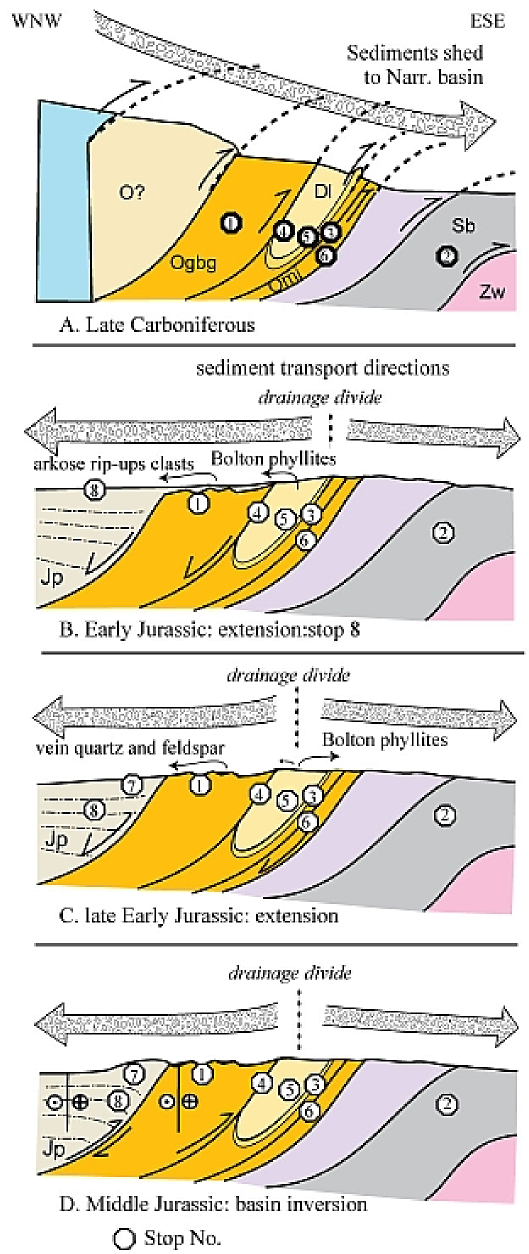

Fig. 4. *Schematic cross sections of the Rockville area, (A-A', Fig. 2). (**A**). At ~275 Ma, crustal shortening against the Laurentian buttress (blue) cause thrusting that shed sediments to the SE. (**B**). At ~195 Ma, in transtension thrust faults were reactivated with normal motion. Arkosic sediments deposited upon the Glastonbury Gneiss provided arkose pebbles and minimized clasts of gneiss (Stop 8). (**C**). At~192 Ma, Glastonbury Gneiss was well exposed and provided pebbles of gneiss and pegmatite (Stop 7). (**D**). At ~ 180, Ma basin inversion reactivated the border fault as a thrust that produced drag folds against the fault and NW trending folds in the arkose (Fig. 1; Stop 8).*

Local evidence for the Permian to early Mesozoic reversal of shear sense in the hinterland comes from northward tilting of the rocks of SE New England such that the grade of metamorphism increases significantly to the south (Wintsch et al., 2003b; Walsh et al., 2007; Hermes et al., 1994; Skehan, 2008). Support for the idea that N-S shortening persisted into the Mesozoic comes from several sources. The most compelling argument for sinistral transpression comes from the widespread inversion of many Mesozoic basins (e.g. Withjack et al., 1998; Schlische et al., 2003; Withjack and Schlische, 2005). This evidence comes from the development or tightening of folds in Mesozoic sediments, normal faults reactivated as reverse and strike-slip faults, from anomalously steep strata dips near border faults (Wise, 1992), from small-scale overturned thrust faults (Wintsch et al. 2003a; Wise and Hubert, 2003), as well as from calcite twins and local axial plane cleavage (e.g. Withjack et al., 1998; Withjack and Schlische 2005).

In the Hartford basin in general and in the Rockville area of this field trip in particular, this Jurassic transpressive motion was accommodated by a reactivation of the Eastern Border fault. This fault became an oblique sinistral strike-slip fault, and motion caused steepening of Mesozoic bedding as the west-dipping fault was reactivated with reverse motion (Wise, 1992). Other map-scale structures include NW trending fold axes in Mesozoic beds of the Hartford basin and NE trending strike-slip faults (Fig. 1). Outcrop-scale structures in southern New England also support this interpretation. These observations include south-verging small scale thrusts in Connecticut (e.g. Stop 3) and in Massachusetts (Wise and Hubert, 2003), and strike slip faults with shallow slicken-sides (e.g. Stop 4a). More detail regarding these brittle faults and joints is given in Wintsch et al. (2003a; 2011).

Contrasting Carboniferous T-t Paths.

These two opposing tectonic settings between the beginning of the Carboniferous and the end of the Jurassic were caused by reversals of motion sense and alternating extension, transpression, and shortening along the plate boundary. The different crustal positions of the various terranes across southern New England are revealed in the contrasting P-T-t paths of their constituent rocks. Here we examine the Temperature-time paths and explores potential tectonic interpretations in the context of the overall collision of Gondwana with Laurentia.

The contrasting thermal histories of Laurentia and Bronson Hill belts shown by the T-t paths of Fig. 3 allow large-scale tectonic relations to be deduced. The only assumption that must be made is that the fault that separated Laurentia from Bronson Hill rocks during the later Paleozoic was west dipping, as is the present listric eastern border fault that now underlies the Hartford basin (Fig. 1; Rodgers, 1985). Recognizing that there is a ~20 m.y. delay between peak pressure and peak temperature, the T-t paths can be understood as proxies for loading and exhumation. Thus throughout the late Paleozoic, the Laurentian margin was cooling and was being slowly exhumed following early Paleozoic metamorphism. However, the Bronson Hill terrane was being heated and actively loaded during most of the Carboniferous. These opposite trends (Fig. 6) can be reconciled if rocks of eastern Laurentia were actively loading the Gander cover terranes activating the west dipping Eastern Border fault. Thrust motion would explain exhumation of the Laurentian hanging wall rocks as well as loading of Gander footwall rocks (Fig. 7).

Exhumation of the thrust nappes in eastern Connecticut began in the east with the Putnam-Nashoba terrane (340 Ma). Exhumation followed progressively to the west in an out-of sequence series of thrusts, each moving to the SE over the previously exhumed nappe (Fig. 1). This pattern of motion can be reconciled with right-lateral transpression, considering that composite Laurentia was cold in the

Carboniferous and would have acted as a 'buttress' that accommodated very little internal strain. Thus successive terranes hitting the buttress were progressively extruded south and east over each other and over the approaching Avalon terrane (Fig. 4). The period between ~280 and 220 Ma was a time of transition. Plate-scale tectonics show that right-lateral transpression evolved to left-lateral motion. Carboniferous thrust motion ceased because all terranes reached thermal equilibrium by the middle Mesozoic.

Mesozoic T-t Paths

Beginning in the Late Triassic sediments began to accumulate on Laurentian crystalline rocks in the Hartford and other basins (Olsen, 1997). These are now exposed above unconformities under the Pomperaug basin and along the southwestern margin of the Hartford basin. Triassic and Jurassic beds thicken and coarsen to the east, suggesting that faulting was syn-depositional in response to normal motion on the EBF. Studies of the sedimentology and paleocurrent directions suggest a dominant westward transport direction (Weddle and Hubert, 1983; Hubert et al., 1992; 2001). ^{40}Ar/^{39}Ar thermochronology of detrital white micas and K-feldspar shows that virtually all of the sediments in the Hartford and Pomperaug basins, both Triassic and Jurassic, yield Permian cooling ages (Hubert et al. 2001; Blevins-Walker, 2008; Kunk and Burton, unpub.) These Permian ages rule out a western Laurentian provenance for the sediments, because cooling from Acadian metamorphism leaves those rocks with dominantly pre-Carboniferous mica cooling ages (e.g. Clark and Kulp, 1968; Fig. 3). However, Permian muscovite cooling ages are completely consistent with a provenance east of the Hartford basin in Gander cover rocks, where erosion of rocks cooling from Alleghanian metamorphism has left uniform Late Permian muscovite cooling ages (e.g. Wintsch et al. 1993; 2003b). Many of the K-Ar whole rock ages of sediments in the Newark basin are also younger than the surrounding Acadian gneisses (Abdel-Monem and Kulp, 1968), suggesting a considerable detrital component from the east (van Houten, 1988).

Further evidence for provenance comes from detrital zircons found in arkoses collected in the Hartford (south end, Stop 7) and Newark basins (Fig. 8; Aleinikoff et al. 2011). In the central Hartford basin the lower Jurassic Portland Formation contains zircons with ages consistent with a derivation from the Bronson Hill Ordovician and Silurian arc rocks, Acadian intrusive and migmatitic rocks, and the ubiquitous Grenville-age suite (Tucker and Robinson, 1990; Aleinikoff, et al., 2007). What is unforeseen is the very small population of Neoproterozoic zircons from nearby Gander and Avalon terranes, and the absence of late Paleozoic (Alleghanian) zircons that are well known from these terranes (e.g. Walsh et al., 2007), but not identified in most Gander cover terranes (Wintsch et al., 2007). In contrast, the presence of both Neoproterozoic and Alleghanian populations in the Newark basin (Fig. 5, ruled and stippled peaks) demonstrates that Gander and or Avalon rocks were exposed in the Early Jurassic, and present in the watershed of the rivers entering the Newark basin. Together these results show that the watersheds draining into the Hartford basin were isolated from Gander/Avalon rocks by some drainage divide. In contrast, the sediment filling the Newark basin did have a significant contribution from the east where the Avalon/Gander terranes would have been exposed. The western transport of these sediments may have reached as far west as the Jurassic Navajo sandstone (Aleinikoff et al., 2010). These terranes are exposed in extreme southern New England and can be identified by their magnetic characteristics to the south (Fig. 6). The plausible paths of rivers feeding into the Newark basin are indicated by the beaded arrows.

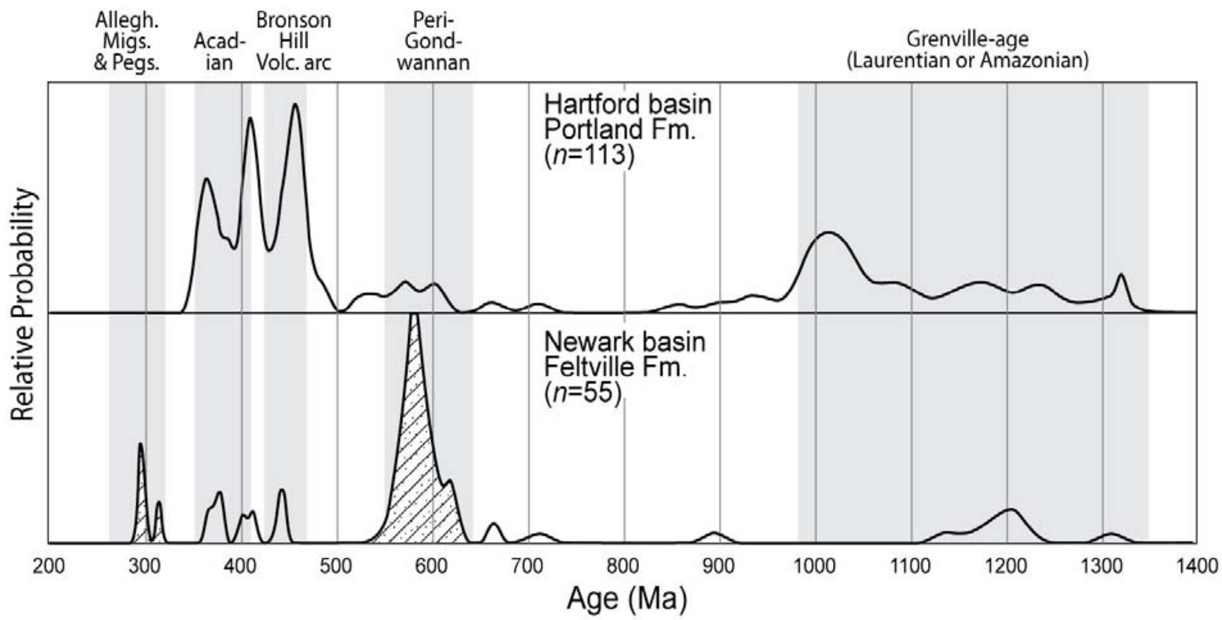

***Fig. 5**. A comparison of the ages of detrital zircons from lower Jurassic sediments in the Hartford basin Portland Fm. (Stop 7) and sediments of similar age in the Newark basin (see Fig. 6) on relative probability plots. The ages of the zircons in the Hartford basin sample can be reconciled with Bronson Hill magmatic rocks, and the ubiquitous suite of Grenville-age rocks. Zircons from the Feltville Formation differ in that they contain a large population of grains with Neoproterozoic ages and a distinct population of grains of late Paleozoic age (ruled and stippled peaks). The latter are characteristic of zircons derived from Gander and high-grade Avalon terrane rocks (e.g. Walsh et al. 2007).*

The significance of these arguments to the T-t paths of Fig. 3 is two-fold. First, the Mesozoic paleoslope of the sediments in the Harford and Pomperaug (and probably in part the Newark) basins was generally to the W and SW, except in the earliest Jurassic, when ponding prevailed. This general western slope shows that for all but 1 m.y. (~200 Ma) of the ~20 m.y. of sedimentary record preserved, sediment flux from the east outpaced basin subsidence, such that the Mesozoic slope was generally W or SW, and Laurentia can be considered to have been continuously loaded from ~210 to 185 Ma. This is reflected in the positive slope of the Early Jurassic T-t curve of Fig. 3. Second, the eastern provenance of the vast majority of the sediments indicates that Gander cover rocks were continuously exhuming, apparently by erosion alone, throughout Mesozoic sedimentation. Thus the slope of the orange T-t curve of the Bronson Hill terrane (Fig. 3) was always negative, indicating continuous cooling (and exhumation), however slowly, throughout this time.

Mesozoic sediments accumulated in these basins from early Norian until at least the Sinemurian (Olsen, 1997) from ~210 to 190 Ma. However, fission track ages of detrital apatites are universally reset, and fission track ages of zircons have been reset in all Triassic and in most Jurassic sediments in the Hartford and Deerfield basins (Roden and Miller, 1991). Low-temperature steps of $^{40}Ar/^{39}Ar$ age spectra from K-feldspars derived from the Deerfield basin have also been reset (Hubert et al., 2001). These results show that loading brought ambient basin temperatures to > 150°C, suggesting an additional thickness of Jurassic sediments of 5 km or more over the present exposure level (Roden-Tice and Wintsch, 2002). This thickness is consistent with a sedimentation rate of ~0.6 mm/yr (Schlische et al., 2003, Fig. 2) if sedimentation persisted at that rate until 185 Ma.

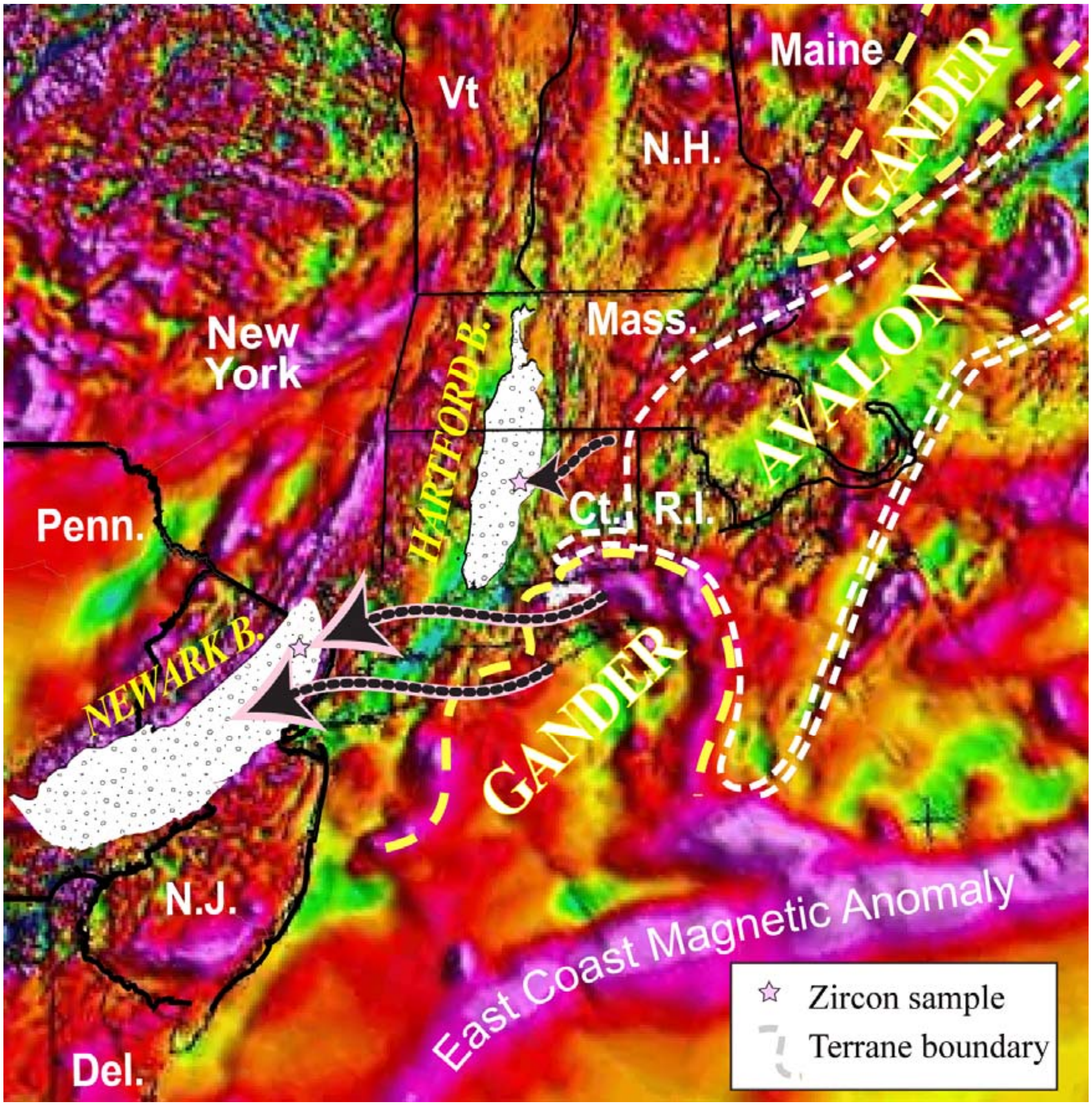

Fig. 6. *An aeromagnetic map of northeastern USA showing the East Coast Magnetic High in warm colors, and areas of low magnetic intensity in cool colors. The locations of the Hartford and Newark basins are superimposed in white dotted regions. Regions underlain by the Gander and Avalon terranes are identified, with boundaries shown in bold dashed lines. Black arrows rimmed in pink show the likely provenance areas*

The oldest Jurassic fission track ages in the Jurassic Portland Formation are ~180 Ma (Roden-Tice and Wintsch, 2002), ages that place a minimum age on the time of maximum loading, and a minimum age on the time of inversion from deposition to exhumation. This age is consistent with the ~185 Ma K-Ar age of authigenic illite (Merino et al., 1997) and with the arguments of Hubert et al. (2001) for the time of maximum diagenetic temperature. Accordingly, we take 185 Ma as the time of peak temperature (and close to the time of peak loading) of Jurassic sediments. Consistent with this, the thermal histories of the basal Triassic (Norian) unconformity and the early Jurassic sediments form nested T-t curves in Fig. 3.

The transition from deposition to erosion of Mesozoic sediments in the Hartford basin occurred at ~185 Ma, the time of inversion in the Fundy and most Mesozoic basins (Withjack et al., 2009). This Early Jurassic age is also compatible with inversion and north-south shortening in the Hartford (Fig. 8) and Deerfield (Wise et al., 1992) basins, and with the strike-slip faults exposed in stops 4 and 5. The steep easterly dips of some arkosic sediments (Wise, 1992) could also be explained by post Early Jurassic reverse motion on along the EBF. The north-south shortening event recorded in these structures and the coincidental beginning of exhumation can both be explained by sinistral transpressive motion along the Appalachian margin (Fig. 5, inset). This left-lateral event is named informally the "Fundian" (Wintsch et al., 2003a; 2011) because it is in response to reversals in regional plate-scale motions (Fig. 5).

The above interpretations modify and update those of Wintsch et al. (2003a). However, the details of the structures and their orientations given there have not changed and still provide the data base for the discussion and interpretations given here. Some new details of the individual stops are given below with each stop description.

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ROAD LOG.

Follow I 84 E to Buckland St. Exit 62. Turn L at Buckland (N). Cross Buckland Hill Mall Road. Turn L (W) on Pleasant Valley Road. Take the next left into the Park and Ride. Consolidate cars.

Mileage

Meet at 7:45 am. Leave hotel at 8 AM sharp.

0.00	Set odometers to 0 at corner of Columbus and Grove Streets immediately after exiting Hotel.
0.05	Turn right onto highway and follow signs to I-84 east; cross Connecticut River.
1.5	Merge with I-84 East and immediately move out of right lane to avoid HOV entrance
2.2	Traffic from I-91 merges on left
4.0	Exit 59: Bear right onto I-384E
9.3	Low overgrown outcrop on right is conglomerate of Portland Formation (Jp). Eastern Border Fault of Mesozoic Hartford Basin is in ravine just past outcrops and before exit ramp.
9.4	Exit 4 from I-384E; turn right at end of exit ramp onto Wyllys Street
9.5	Turn left onto Spring Street Extension. Turn around and park cars near intersection on north side of road. Stop 1 is around corner on the entrance ramp to I-384E.

STOP 1. **GLASTONBURY GNEISS** (Ogbg)**,** Bronson Hill Terrane, Rockville Quadrangle (Aitken, 1955; Wintsch et al., 1998a)

The purpose of this stop is to show the large range of Paleozoic structures preserved in the Glastonbury gneiss, from Late Ordovician magmatic features through early Alleghanian ductile strain to late Mesozoic brittle deformation. The most conspicuous structure is the penetrative schistosity cut by pegmatites. However, these rocks are unusual because they preserve some of their Ordovician magmatic history, in spite of an Alleghanian overprint of ~600˚C and a model pressure of >12 kb equivalent to > 35 km depth (Wintsch et al., 2003). Most conspicuous are the K-feldspar phenocrysts that still preserve Carlsbad twinning and a perthitic texture (Wintsch et al. 2005). Amphibole (hastingsite) grains are also present and visible with a hand lens. Zircons from this unit are magmatic and U-Pb dating indicates an age of ~450 Ma (Aleinikoff et al., 2002). In addition to zircon, sphene also preserves a late Ordovician to early Silurian age with oscillatory and sector zoning compatible with magmatic crystallization (Aleinikoff et al., 2002). Magmatic structures still preserved are the dark biotite and amphibole bearing xenoliths present in several locations. Their aspect ratios are typically < 1:5, indicating low to moderate sub-solidus strain.

The penetrative lineation plunging a few degrees toward the N is strong enough that these rocks may be considered an “L > S-tectonite” in which the lineation is more strongly developed than the schistosity. The lineations defined primarily by quartz and biotite anastomose around cm-size orange phenocrysts of K-feldspar (Fig. Stop 1.1) but the lack of conspicuous foliation makes the down-lineation view mimic a magmatic texture because the phenocrysts seem to float randomly in an apparently massive matrix. Such structures are more common in orthogneisses than in metasedimentary rocks because they have not inherited a strong foliation parallel to bedding. The biotite streaks, epidote aggregates, and quartz rods that define the lineation here and regionally were produced by the reaction K-feldspar + amphibole +H_2O = biotite + epidote + quartz (+sphene). The magmatic reactants destroyed by this syntectonic reaction are stronger than the metamorphic products, and this is an example of reaction weakening (Wintsch et al., 2005). As the reaction progressed, the relatively strong K-feldspar and amphibole were destroyed by dissolution, and the new biotite and quartz grains segregated into streaks and ribbons that made the mineral assemblage weaker and better able to accommodate more strain. Some of the K-feldspar dissolved in this reaction was precipitated as metamorphic K-feldspar in local inter-crystalline extensional sites (Wintsch et al., 2005). As the rock grew too strong to deform by dissolution creep, K-feldspar also precipitated in monomineralic veins (Fig. Stop 1.2) that open perpendicular to the SE extension. The veins demonstrate that the composition of the local metamorphic fluid was relatively alkaline and supersaturated in K-feldspar.

Fig. Stop 1.1. An oblique down plunge NE view of the penetrative lineation in the Glastonbury Gneiss. Large orange grains are relic K-feldspar phenocrysts. Weak foliation dips moderately NE (right).

Fig Stop 1.2. A K-feldspar vein cutting the fabric in the Glastonbury Gneiss. The mono-mineralic nature demonstrates that this vein is sub-solidus, and probably hydrothermal.

Fig. Stop1.3. A zoned quartz and K-feldspar vein cuting the L>S fabrics in the Glastonbury gneiss. The lack of plagioclase demonstrates a non-eutectic composition and a likely hydrothermal origin for this vein.

Zoned pegmatitic veins also cut this fabric. These quartz-K-feldspar veins can be strongly zoned, with pure quartz cores and essentially pure K-feldspar margins – a zoning pattern well-known in larger, and formerly commercial pegmatites, such as the Spinelli prospect (Fig. 2). The margins of these veins are sharper than those of the feldspar veins, showing that they post-date essentially *all* ductile deformation in the bulk rock, consistent with the few pegmatite dikes that cut these rocks with a strike is ~070. The ages of the syntectonic sphene by-product show that this reaction occurred between in pulses at ~290, 270, and 260 Ma. Biotite and quartz locally show S-C structures consistent with top to the south-east sense of shear. Thus this lineation shows a SE extension that occurred in the Permian, probably between 270 and 260 Ma.

Several later brittle faults and fractures cut these gneisses. Slicken lines show that most are normal, with some RL and some LL, but regionally the LL faults dominate. Some fractures are mineralized with epidote, suggesting lower greenschist facies conditions at the time of opening. The slickensides and rare cherty quartz or zeolite in the faults show that they were active at <~100°C. Strikes are near N and NE, parallel to the basin margin and Meriden group regional faults respectively. Joints in this outcrop share the regional orientations of joints in the Glastonbury Gneiss (Ogbg) throughout the area (Fig. 2), that is N, NE, and E. The 100°C temperature estimate suggests a Late Jurassic or Early Cretaceous age.

9.5 Return to cars. Turn right off Spring Street Extension and right onto I-384E, passing outcrops of STOP 1.
10.8 Passing outcrops of Bolton Schist (Dl).
12.3 Bolton Notch, passing cuts of Bolton Schist, Fitch and Clough Formations. Keep right and follow Rte. 6E.
12.6 Passing road-cut of Middletown complex (Omi)
13.5 Road-cut of Southbridge Formation
14.6 Coventry town line
15.0 STOP 2. Road-cuts on west side of Rte. 2. Park at north end (beginning) of outcrop because we will turn left at the adjacent intersection. Walk to south end of outcrop.

RTE. 6 HAS HEAVY TRAFFIC THAT TRAVELS FAST! BE CAREFUL!

STOP 2. HEBRON GNEISS, Merrimack terrane, Rockville quadrangle (Aitken, 1955; Wintsch et al., 1998a).

The purpose of this stop is to show the rock types and both ductile and brittle structures in the structurally lowest unit in the Rockville area. The local name for these rocks is the Hebron Formation (Sh, Fig. 2) that are part of the Merrimack terrane of regional extent in eastern Connecticut correlating with the Oakdale and Paxton formations in Massachusetts (Merrimack terrane, Fig. 1), and the Berwick and Elliot formations farther north in New Hampshire. The rocks at this exposure include greenish layers of quartz-plagioclase-hornblende-biotite granofels alternating with brownish quartz-plagioclase-biotite schist, each from 1 t o10 cm thick. In southern Massachusetts, near Webster, this same unit is at lower grade and contains alternating layers of quartz-muscovite-chlorite schist and quartz-muscovite-dolomite schist, reflecting protoliths of mudstone and calcareous mudstone, respectively.

The layers in this outcrop are highly attenuated, folded and boudinaged. Some of the folds are intrafolial. The rheological contrast between the calc-silicate granofels and schist layers is highlighted by the zone of high strain localized at a calc-silicate-schist contact. It occurs at chest height opposite the sign indicating the left turn lane. Extensional shear bands that dip parallel to a NNW-trending stretching lineation indicate top-to-the south sense of shear. Several pegmatite veins intrude the schist and granofels and record several stages of deformation. Early pegmatites are strongly attenuated, foliated, folded, and boudinaged. Locally coarse-grained quartz-plagioclase veins can be traced into layers as thin as one cm, where the texture is aplitic. Some of these attenuated pegmatite layers are deformed by asymmetric folds overturned to the south by extensional shear bands and small scale thrust faults. Most of these structures indicate a top-to-the south sense of shear, although some small folds have north vergence. In contrast, thicker and coarser grained plagioclase-quartz-biotite (beryl) pegmatites are not foliated but are boudinaged with quartz filling the boudin necks. Some contain subhedral plagioclase crystals up to 7 cm long with bands of randomly oriented biotite inclusions that seem to outline several stages of plagioclase growth. A hornblende age of 281 Ma from the hornblende granofels here (Kunk and Wintsch, unpub.)

proves that this outcrop reached middle amphibolite facies in the Alleghanian, and places a minimum age on the ductile deformation. Sh rocks east of the Willimantic dome contain only one generation of concordant pegmatites, and there amphibole and sphene ages are Pennsylvanian. Thus these structures are pre-Alleghanian and most likely Acadian. In the rocks at this outcrop there is a second generation of pegmatite and younger structures deforming them. We interpret these to be Alleghanian; thus these rocks show the effects of both the Acadian and Alleghanian orogenies.

The outcrop is also cut by brittle faults and joints. Part of the evidence that the Jurassic faults affected this area is that they trend 050° and show down on the NW side displacement. These as well as other fractures and brittle faults are unmineralized (but note the limonite filling of one open joint at the north end of the outcrop) and were therefore formed at temperatures less than 100°C. The apatite fission track age of 130 Ma records the time of cooling through ~100°C. Thus the maximum age of these brittle structures is Early Cretaceous or younger.

15.0 Return to cars and turn left immediately onto South Street.
REMEMBER TRAFFIC AND SAFETY CONCERNS.
15.5 Turn left onto Swamp Road
15.8 Turn left onto Brewster Road (no street sign at the time of writing the road log).
16.8 Drumlin-shaped deposit of thick till on left side)
17.3 Turn left onto Rte. 44 (17.7: top of “drumlin”).
18.9 Bolton Lake on right
19.2 Pass road cut on right (Stop 3a).
19.3 Park just beyond first road cut and “school bus stop ahead” sign. Stop 3b is just ahead. Turn around walking E to STOP 3A.

TRAFFIC ON RTE. 44 IS USUALLY HEAVY AND THE SHOULDER HERE IS NARROW. KEEP OUT OF TRAFFIC AND INSIDE THE WHITE LINE!

Stop 3A. MONSON GNEISS, (Omo) Bronson Hill Terrane (Aitken, 1955; Wintsch et al., 1998a).

The purpose of this stop is to examine the lithology of the Monson orthogneiss, and variable high grade and low grade structures in it. This gneiss contains plagioclase, quartz, and biotite, with accessory epidote, sphene, and locally amphibole, probably relict from its late Ordovician crystallization. Here the rocks are strongly layered and foliated. The strong layering is defined by bands of plagioclase + quartz with variable amounts of biotite, and the foliation is defined by the preferred orientation of this biotite. The layering is further accentuated by very thick biotite-free layers of quartz and feldspar, interpreted to have been pegmatites, but now strongly attenuated by the very high strain in these rocks.

This gneiss is cut by unfoliated pegmatites, and by quartz veins. The former formed at temperatures of 500-600°C in the early Permian (see Stop 1). If a ~300°C estimate for the temperature of formation of these quartz veins is accurate, then they formed in the early Triassic (orange curve, Fig. 3). The strong foliation cut by these structures must then be late Pennsylvanian. The biotite-rich folia have been reactivated with muscovite + chlorite, reflecting lowest greenschist facies conditions (Early Triassic, Fig. 3). Quartz is a by-product of the alteration of biotite to chlorite, and quartz lenses anastomose through these folia, further evidence of continued reaction and displacement in these zones.

All these rocks are cut by gently N-dipping cataclasitic thrusts, some also containing now devitrified pseudotachylite. While the displacement is probably small, SE-verging slickensides suggest shortening at temperatures <~200°C in the middle Jurassic (Fig. 3). Thus these structures provide evidence for a Middle Jurassic, transpressive “Fundian” event (Fig. 3).

Finally, the rocks and all these structures are cut by several unmineralized (and thus probably Cretaceous) joint sets. One set striking 040-055° is subparallel to the highway and defines the outcrop surface. These joints also parallel the South Lamentation Mountain fault which we think passes just south of this outcrop, and thus provide evidence that this fault does propagate into the crystalline rocks to the NE. A still later set, striking 075-090°, is parallel to the late cross faults of Fig. 2 and is interpreted to reflect this very late deformation. The latter set may be rusty, showing that they have been open to recent ground water movement.

Walk west, past the cars to the next outcrop on the N (right) side of Rt 44, to STOP 3B.

STOP 3B, MIDDLETOWN COMPLEX (Omi), Bronson Hill terrane, (Aitken, 1955; Wintsch et al., 1998a).

The purpose of this stop is to show the very strong deformation in the Middletown complex and the younger overprint of these structures by Mesozoic structures. The rock here is a muscovite-chlorite schist, locally differentiated into 1 mm alternating layers of pinstriped mica and quartz. We interpret this structure to be a >300°C greenschist facies mylonitic schist, probably of early to middle Triassic age (Fig. 3). There are tectonic inclusions of amphibolite in these schists, showing first that the strain is very high (mylonitic), and second that the Middletown Complex (Omi, Fig. 2) may be cut by these mylonites. Low angle thrust faults dip gently north, where deformation of the hanging wall forms north vergent folds. These are interpreted to have formed at ~200°C, a temperature warm enough to allow this ductility and faulting, consistent with the brittle-ductile transition. These structures show N-S shortening, and support the interpretation of the Fundian shortening event. Local displaced quartz veins in some of these faults show LL displacement. As such, they are middle Triassic and provide more evidence for the LL, transpressive Fundian event. Joints dip moderately to steeply N, some rusty, and some showing displacement as normal faults. This again is reflective of the late Cretaceous 070° family of cross faults. These rocks contrast with the plagioclase-rich rocks present at Stop 3A, and demonstrate that at identical temperature and pressure conditions, mica-rich rocks are much weaker and can deform ductily, while the former are fracturing and deforming by cataclasis.

19.3 Return to cars and proceed west on Rte. 44.
19.5 Turn right (north) onto Quarry Road
19.7 Note trenches at the base of the hill to the left. They mark the former site of extensive quarries for the WNW dipping quartzite visited on this trip as Stop 5.The quarries could not expand down dip but expanded along strike, for almost a mile. In the 1700s, this quartzite was used for headstones and building stones, but evolved by ~1810 to larger scale operations, when properties of resistance to moisture, stain, and acid made it useful for door stoops in homes and table tops in chemistry labs and hospitals. By 1820 slabs were exported as far as Philadelphia, Baltimore, Wash. D.C., and even New Orleans. Percival (1842) noted that the quarries then active were not likely to remain commercially viable because the quartzite dipped west under Box Hill, and keeping the quarries dry and stable would be prohibitively expensive.
20.4 Vernon town line. Quarry Road becomes Bolton Road.
22.2 Turn right onto Reservoir Road.
24.2 Turn left onto Rte. 31 and pass under I-84.
24.8 Turn right onto Rte. 30
25.1 Tolland town line
25.3 Turn right at traffic light onto Industrial Park Drive West.

25.5 Turn right onto Gerber Road.
25.8 Outcrops of Littleton Schist on both sides of road. Just beyond is former cul-de-sac turn-around. Continue through the turn-around; large positive-pressure air dome structure ahead is Star Sports Complex.
26.2 Drive past lower parking lot and turn right into upper parking lot. Drive past "Do not enter" sign to the right of the air dome and around to the back of the complex. NOTE: *this is private property and we have permission (established ahead of time) to enter the restricted area. If you follow this road log at another time, make sure you make appropriate arrangements ahead of time so as not to ruin this site for future access.*
26.4 STOP 4

STOP 4. BOLTON SCHIST (LITTLETON SCHIST), Bronson Hill terrane, Rockville Quadrangle (Aitken, 1955; Wintsch et al., 1998a). Star Hill Sports Complex, 100 Gerber Road.

The purpose of this stop is to view the argillaceous rocks of the Bolton schist. This unit has received extensive study in part because of its abundant euhedral staurolite crystals, locally reaching 3 cm in length, and in part in the study of metamorphism and deformation (e.g. Bell et al., 1997; Busa and Gray, 1992; 2005; Hickey and Bell, 1999; Berg and Moecher, 2005; Spear et al., 2008). The Bolton schist is the youngest of three units, the oldest being the Clough Quartzite, conformably overlain by the calcareous Fitch Formation. Fossils in rocks thought to correlate with the latter (Elbert et al., 1988) date it at early Devonian, and the unit is identified on Fig. 2 as Dl.

At this stop we are on the right-side-up east limb of the overturned Bolton Syncline. The basal Clough quartzite non-conformably overlying the Middletown complex (Stop 5) would lie a few hundred meters east of this outcrop. Local evidence for the synclinal structure comes from the repetition of the Clough quartzite on both NW and SE sides of this schist. However, in the Rockville quadrangle quartzite on the NW flank is rare, and the Fitch is absent, probably because they were cut out by the Box Hill reverse fault (BHF, Fig 2) that places the Glastonbury orthogneiss (Ogbg, Fig. 2) against these metasediments. There is no sedimentary structural evidence at this outcrop that this unit is right-side-up. Indeed, any bedding has been so transposed by the dominant foliation that bedding, to say nothing of topping directions, cannot be recognized. Even quartz veins form rootless intrafolial folds (Fig. Stop 4.3).

This stop provides outcrop-scale and cm-scale evidence for late Paleozoic motion to the SE. S-C fabrics are well developed all along this fresh > 100 m cut. Relatively late pegmatites are rotated to the SE (Fig. Stop 4.1), and early pegmatites are strongly attenuated and boudined (Fig. Stop4.2). Thin pegmatites ~25 cm thick may be stretched and necked to zero thickness, reflecting high ductility. Thicker pegmatites are less attenuated, and boudins are produced by fractures, subsequently filled with quartz. Necking is limited to the vicinity of the quartz veins, where muscovite schist is drawn into the limited regions of extension (Fig. Stop 4.2). The hinge of an intrafolial folded quartz vein plunges NW, suggesting rotation due to high strain. Grain-scale S-C fabrics are pervasive and generally show a SE motion sense, similar to the rotation of veins and pegmatites.

Unusually fresh rock is exposed in these outcrops, which are typical of the rather monotonous gray-weathering, muscovite-biotite-quartz-plagioclase-garnet+/-staurolite +/- kyanite schist. Staurolite is present in most outcrops (Fig. Stop 4.4) and may be more common in schists adjacent to quartz veins; kyanite is not common. The staurolite-kyanite grade of these rocks is certainly Alleghanian because hornblende ages as close as 4 km south are reset to 263 Ma (Wintsch et al., 2003b). Moreover, isotopic monazite and sphene ages (Coleman et al., 1997; Spear et al., 2008) record ages no older than Pennsylvanian and matrix grains as young as mid-Permian, confirming that the age of prograde metamorphism is Alleghanian. Finally, a monazite from a pegmatite similar to those exposed here (Figs.

Stop 4.1; 4.2) but from the north side of Bolton Notch yielded a Late Pennsylvanian U-Pb age (Coleman et al., 1997). If these pegmatites are of the same generation, then high-grade metamorphism is Late Pennsylvanian, and reactivation of the surrounding schistosity is Permian.

Several generations of foliation are present. Older foliations are preserved as inclusion trails in garnets, and younger foliations are also preserved in staurolite porphyroblasts (Hickey and Bell, 1999). The dominant and composite matrix foliation here dips moderately WNW and has been reactivated by several strain events. Muscovite in this foliation is intergrown with biotite and locally with chlorite, showing it to have been last recrystallized at lower greenschist facies conditions. From Fig. 3 the latter conditions are shown to be Late Permian. Late Permian muscovite $^{40}Ar/^{39}Ar$ ages (Wintsch et al., 1993) and matrix monazite ages (Spear et al., 2008) show that this reactivation involved the recrystallization of these minerals, and probably most minerals. The strain accommodated in this foliation is so high that early formed structures are highly attenuated. Early pegmatites are now sub-parallel to the schistosity, and early-formed quartz veins are transposed into the dominant foliation and form intrafolial folds.

The cuts passed on Gerber Road (Stop 4A) on the way to this stop expose a few brittle normal faults (Fig. Stop 4A.1) Some striking ~180 with steep W dips also show slickensides plunging ~30° SSW. Such motion is normal, but with a component of LL displacement. If the slickensides developed at ~ 200°C, then these faults provide evidence for the change from LL transpression to trans-extension in the Late Jurassic (Fig. 3). Still later joints strike between 060-090 and dip moderately S. They are rusty, thus open to ground water, and developed in the middle Cretaceous.

26.4 Return to cars and drive around to front of air dome and back onto Gerber Road.

26.7 **Stop 4A**. Cuts similar to those just visited. Ductile structures and deformed veins again show top to the SE sense of motion. Exclusively at these cuts are the drill holes bored during construction of this cut that show a reverse sense of offset. These have been found before in the hanging wall of the Honey Hill fault (Block et al., 1979) and probably reflect the release of elastic strain energy in the rock. With such a varied and complicated history, it is surprising that the Bolton schist could store so much strain to show this much reverse motion. We can only speculate that minor crenulation of the mica folia in this schist stored that strain energy lingering since mid Mesozoic, until construction of this cut. The cuts by the sports complex may be too new to show this displacement.

27.1 Turn left onto Industrial Park Drive

27.3 Turn left at right (northeast) at light onto Rte. 30

28.5 Turn right at light (where Rte 30 turns left) onto Mountain Spring Road and proceed up steep hill.

29.0 Turn left into small driveway by the out building of the Burgundy Hills quarry (currently operated by Better Stones and Gardens of East Hartford). *This is private property also and we have permission (established ahead of time) to enter the area. If you follow this road log at another time, make sure you make appropriate arrangements ahead of time so as not to ruin this site for future access.*

29.6 Burgundy Hill Quarry pit.

Fig. Stop 4.1. *Near vertical pegmatite dragged right (SE) and boudinaged.*

Fig. Stop 4.2. *A strongly attenuated pegmatite with quartz filling the neck and limited filling of the neck region with adjacent schist.*

Fig. Stop 4.3. *High strain in the schist produced an intrafolial fold from a strongly attenuated quartz*

Fig. Stop 4A.1. *A distant view of two Mesozoic faults: one dipping to the right (E) marked by the 20 cm long yellow notebook, and the other fault dipping to the left (W) with slickensides suggesting left-lateral motion.*

Fig. Stop 4.4. *A close-up view of the garnet and staurolite (long rod-shaped grain) porphyroblasts. Note the muscovite-rich folia wrapping around the porphyroblasts.*

STOP 5: CLOUGH QUARTZITE (Sc), MIDDLETOWN COMPLEX (Omi), Burgundy Hill Quarry. Rockville Quadrangle (Aitken, 1955; Wintsch et al., 1998a).

***Fig. Stop 5.1**. North-facing view of the noncomformable contact between the mafic amphibolite of the Middletown complex (lower right) with the basal yellow-orange-weathering conglomeratic Clough quartzite. Yellow hammer is 40 cm long.*

***Fig. Stop 5.2.** A down-plunge view of the quartz-pebble conglomerate at the base of the Clough quartzite. The pebbles are stretched and plunge gently NNW. Pen is 15 cm long.*

The purpose of this stop is to show the Clough quartzite, and the faulted, nonconformable contact separating it from the underlying Middletown complex. This quarry exposes the base of the Clough quartzite on the right-side-up limb of the Bolton syncline, where it is a quartzite and quartz-pebble conglomerate. Most quartz pebbles here are <5 mm in diameter and do not define the strong NW lineation that larger pebbles do in the southern part of the Rockville quadrangle. The quartzite is interlayered with muscovite-rich schists from 1 to 10 mm think. These typically contain burgundy red garnets from 0.1 - 2.0 cm in diameter. They have euhedral outlines in the schists but are anhedral and flat facing the quartzite layer from which they cannot crystallize. The quartzite tends to break along these muscovite-rich layers, sometimes into square meter sized slabs, bestudded with burgundy red garnets.

Brittle structures cut all rocks in this quarry. Late brittle faults with slickensides, but only minor displacement, are both RL and LL. They tend to strike ~340 and 070°, dipping E. The sense of offset is consistent with these faults constituting a conjugate set of faults, but more work is needed to test this. Joints are common in the quarry. They commonly strike ENE and NNW and are rusty. Another conspicuous rusty joint strikes ~010° and dips east. The 070 strike is parallel to the late cross faults of Fig. 2, and is interpreted to reflect this deformation. The 010 striking joints could be related to reactivation of this strike as a bordering fault; however, the dip is probably wrong for this interpretation.

One of the more significant discoveries of our mapping of these rocks is the preservation of amphibolite with relic gabbroic textures as tectonic inclusions in the typically pinstriped amphibole orthoschist that is here seen to be an L-S- tectonic. These gabbroic textures preserve a lath-like outline of plagioclase grains in a matrix of mafic minerals, now mostly amphiboles. They were exposed in boudin-like structures on the SW facing cut of the lower quarry, where blocks are surrounded by the pervasively well foliated and lineated pinstriped amphibolite. The best exposure of this rock is in low, glacially polished and unquarried outcrops 100 m NE of the quarry behind stockpiles of gravel. Other exposures of more schistose rock lie just east of the upper quarry, where the dominance of garnet stimulated their mining for abrasives over 100 years ago. The metamorphic fabrics and differentiation into bands led previous mappers to interpret the protolith of these rocks as metavolcanics. However, the preservation of relic gabbroic textures here, and more extensive exposures of them on strike in the Ellington Quadrangle (Collins, 1954; Wintsch, unpub.) show that the protolith of these rocks is a slowly cooled, probably intrusive gabbro, and not a volcanic rock.

29.6 Return to cars and drive out of quarry.
29.9 At end of quarry access road, turn right onto Mountain Spring Road.
30.5 Turn left at stop light (base of steep hill) onto Rte. 30.
31.6 Passing Tolland Industrial Park
32.1 Turn left (south) onto Rte. 31.
32.6 Pass under I-84.
33.8 Turn right onto Dockerel Road and park on right side of road.
Walk back down the hill to Rte. 31 and Stop. 6.

STOP 6: Middletown Complex (Omia), Rockville Quadrangle (Aitken, 1955; Wintsch et al., 1998a).

The purpose of this stop is to examine the strongly foliated and lineated orthoschists of the Middletown complex and to examine the evidence for top to the SSE motion (Fig. Stop 6.1) in the earliest Permian (Fig. 3). These cuts expose both mafic and felsic metaigneous rocks, a conspicuous meter thick unmetamorphosed tonalitic dike in the center of the road cut and sheath folds at the NW end.

Fig. Stop 6.1. These pistachio-green boudins formed as epidote veins dipping ESE. Ductile deformation that produced the strong foliation and NW plunging amphibole lineation rotated the veins CW top to the SE during waning metamorphic conditions. With continued deformation during waning metamorphic conditions, quartz and then calcite precipitated in the boudin necks.

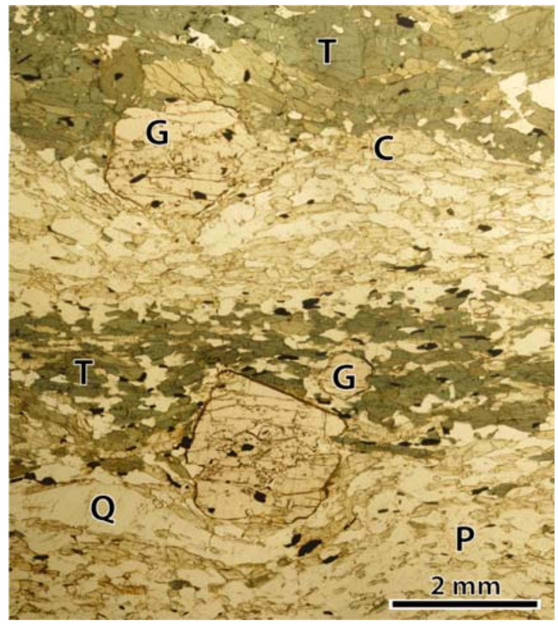

Fig. Stop 6.2. A photomicrograph showing bands of tschermakitic amphibole (T) interlayered with cummingtonite + plagioclase bands (C). Garnet (G) porphyroblasts are subhedral. Quartz is a product of the garnet-producing reaction.

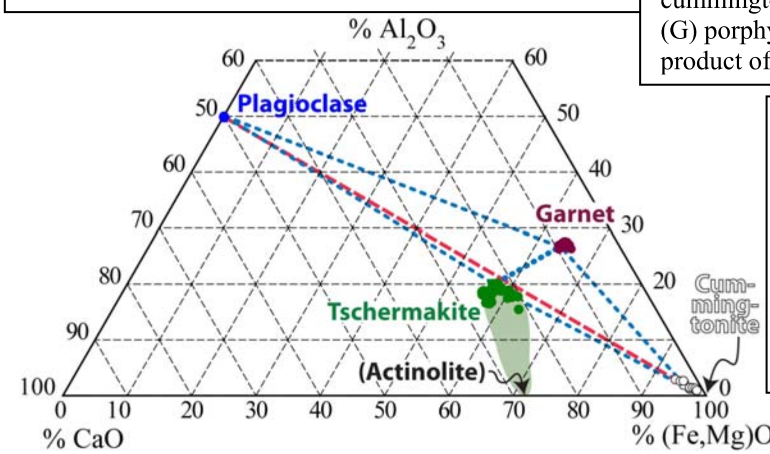

Fig. Stop 6.3. An ACF ternary diagram showing the compositions of plagioclase, tschermakitic amphibole, cummingtonite, garnet, (from Fig. Stop 6.2). Blue dotted tie lines join the compositions of high-grade minerals; red dashed tie line joins the retrograde reaction products plagioclase + cummingtonite.

The most abundant rock type exposed in this cut is the mafic amphibole schist. It dominates the structurally higher parts of Omi in general and the western (higher) part of this exposure. Both amphibole and plagioclase schists are penetratively foliated and lineated, and locally host a pin-striped layering parallel to foliation that dips gently WNW. Layering in the amphibolite is defined by different modes of tschermakitic amphibole, cummingtonite, plagioclase, quartz, and locally epidote. Grain diameters are typically 100-300 µm, but small plagioclase grains can be up to 600 µm. Garnet is locally common in some layers and may reach up to 2 cm in diameter to produce a garnet amphibolite. Epidote is a common accessory interstitial mineral whereas biotite and calcite are present only locally. Garnet porphyroblasts from 2 to 20 mm in diameter are more common at the east end of the outcrop and typically form sigma-type structures, indicating top to the SSE motion sense. Electron microprobe analyses of the minerals identified in Fig. Stop 6.3 show the green pleochroic amphiboles to be tschermakitic hornblendes (T), the colorless amphiboles to be cummingtonite (C), the garnets (G) to contain ~10 mole % grossularite, and the An content of the plagioclase to be 30-35%. Hornblende grains are zoned with Si, Mg and Na decreasing from core to rim, and Al, Ti, and K increasing from core to rim. This zoning pattern is suggestive of prograde growth, an inference confirmed by Holland and Blundy (1994) edenite-richterite thermometry. Calculated temperatures for these samples range from ~650°C to ~690°C from core to rim. Using the amphibole-garnet-plagioclase-quartz thermobarometer (Berman, 1991), we calculate pressures and temperatures to be ~6.0 Kbars and ~630°C. In combination with previous one-dimensional thermal modeling (Wintsch et al., 2003, Wintsch et al., 2005), these results allow us to conclude that these rocks experienced prograde deformation during the late stages of the Alleghanian Orogeny, ~275 million years ago (Stewart et al., 2012).

At the SE end of the cut toward the Monson gneiss the structural base of the amphibolite unit is interlayered with a plagioclase-quartz schist. Interlayering of decimeter thin layers is conspicuous, and demonstrates that layering is transposed into the NW-dipping foliation. Thin amphibolite layers are boudinaged, reflecting the greater strength of amphibolite over plagioclase-quartz schist. Quartz-tourmaline veins in the boudin necks (middle greenschist facies and Late Permian?) recorded a SE stretch before being mined out by rockhounds.

Veins and dikes are common in this outcrop. The meter-thick dike opposite Dockerel Road is unfoliated and nearly undeformed, whereas thinner, strike-parallel, plagioclase-quartz veins cut the amphibolite at a high angle but are dragged and thinned to the SE, showing ductile reverse motion sense. This same SE vergence is demonstrated by beautiful chains of layered, pistachio-green epidote granofels boudins. On first glance the chains appear to be parallel to the foliation and layering, but with study they are seen to climb to higher structural levels to the east. When restored to their former single layer, the boudins converge to the form a steeply cross-cutting epidote-granofels vein. If interpreted as veins, the clock-wise rotation of the veins caused by the large amount of top to the SE strain becomes apparent. In this context, the thin concordant plagioclase veins can also be interpreted as early veins transposed into virtual parallelism with the foliation. Finally, the degree of strain can be understood to be immeasurably high by the occurrence of sheath folds. They are very difficult to identify, but we have found them in the amphibolite at both ends of the outcrop. When viewed down lineation, they appear as ellipsoids 1-2 cm thick, 2-4 cm wide, and infinitely long. Their axes plunge ~330, parallel to the penetrative amphibole and mica lineations in these rocks.

Taken together, then, the significance of this and the last outcrop is that the amphibolites are not part of a formation (Rodgers, 1985) but part of a complex of orthoschists. High strain imposed a penetrative foliation and lineation in these rocks, a strain that consistently shows top to the SSE. Pennsylvanian crystallization ages of sphene from the Glastonbury gneiss (Aleinikoff et al., 2002), and the Fitch formation (Coleman et al., 1997), together with the completely reset amphibole ages of central Connecticut (Wintsch et al., 2003b, here 279 Ma) show that this deformation was Alleghanian.

The final stage in the ductile deformation of these rocks is present in the form of sigma-shaped quartz tails on garnet porphyroblasts and of rare drag folds with east-dipping axial planes that fold the foliation and lineations in the amphibolite. Both structures show a W or NW vergence, for a reversal in motion sense. They are interpreted here to reflect the extensional deformation of the later Triassic, when the Hartford basin began to accumulate sediments. Finally NE striking joints, commonly filled with small amounts of zeolite, show the latest deformation here to be Cretaceous. Return to cars.

33.8 Return to Rte. 31 and turn left (north).
35.0 Pass under I-84 and turn left onto I-84 west.
35.8 Outcrops on west are Glastonbury Gneiss
38.7 Outcrops along entrance ramp to I-84 east (on left across highway) expose schist. This is the westernmost exposure crystalline rocks along the highway.
39.0 Approximate location of Eastern Border Fault. A heavily silicified zone is exposed along a former railroad grade a few hundred meters to the south.
39.8 Gerber Scientific on right. Years ago highway construction exposed conglomerate of the Portland Formation along the low cut adjacent to the highway that has since grown over and/or been covered by soil. This constrains the location of the border fault in this area. If you look to the left as you pass you will notice a low line of hills that marks the edge of the crystalline terrane to the east and the trace of the border fault.
41.2 Conglomerate and conglomeratic sandstone exposures of Portland Formation along highway from here until the exit ramp. These are braided stream deposits from Jurassic alluvial plains.
42.1 Turn off highway onto Exit 62/60 to Buckland Street
42.5 Bear right following signs to Buckland Street.
42.9 Turn right onto Pleasant Valley Road at light.
43.2 Turn right (south) at light onto Buckland Street
43.5 Pass under I-84 and get into middle lane preparing to turn left after the entrance to I-84 East.
43.7 Turn left onto Redstone Road ; Take an immediate left into the parking area behind Extra Mart. *Be aware that these cliffs may still have some loose rock; you may not want to stand too close.*

STOP 7. PORTLAND FORMATION (Jp), Manchester quadrangle, Colton, 1955.

The purpose of this stop is to show the subarkosic arenites and their pebble compositions. This is also the site from which a sample was collected for detrital zircon dating. Sediments here were are alluvial plain deposits of braided streams that overlie stream deposits with overbank facies (exposed to the west along I-291) so they are near the transition to meandering stream depositional facies (probably a lower gradient). To the east, along the highway (we just passed them) are coarser braided stream facies and possibly some distal alluvial fan facies. Note that the meandering stream facies are farther west but, because of the easterly dip, stratigraphically older. The distal alluvial fan deposits are farther east but stratigraphically younger. One might ask whether we are looking at changes in the depositional environment caused by geography or changing tectonics.

This sequence of slightly graded beds have coarse pebbles at their base, and some have pebble-filled channels at their base (Fig. Stop 7.1, shown with images of Stop 8). Spot counting of the pebble lithologies shows that they are dominated by vein minerals: quartz> K-feldspar> K-feldspar + muscovite (almost 50% of some gravels). Up to 25 % of the pebbles are gneissic and could be derived from the Glastonbury gneiss. Quartzites and gray muscovite-rich phyllites present in stops 4 and 5 constitute less than 10% of the pebbles in most gravels. Veins similar to the pebbles present here are common in the Glastonbury gneiss, suggesting that it was the main source of the pebbles deposited here. Quartzites that should be durable during transportation are not common. Zircon ages (Fig. 5) suggest that the sand-sized fraction came from a wider range of lithologies but very little from the Avalon and Gander terranes.

43.75 Return to cars and turn right onto Redstone Road.

43.8 Turn right (north) onto Buckland Street. After passing under I-84 get into left lane.

44.3 Turn left onto Pleasant Valley Road

44.6 Turn left onto I-84 entrance. Note that the Penney's warehouse (on right after turn; behind Hooters) is built on a local delta-top plain (elev. 140-160'). The area is underlain by sand and gravel deposits that in a nearby gravel pit, exposed spectacular delta topset and foreset bedding before being mined. These sediments are referred to as the Hockanum River delta deposits (Stone et al, 2005) of Late Wisconsinan age. The delta formed by meltwater stream deposition along the eastern margin of glacial Lake Middletown, an early pre-cursor to glacial Lake Hitchcock in the Hartford Basin. As Lake Middletown lowered, glacial Lake Hitchcock developed and was impounded behind a massive sediment dam that blocked the Connecticut River valley in Rocky Hill and Cromwell. The spillway for Lake Hitchcock became stabilized on bedrock in a 2 mi. long, 800' wide channel along the New Britain-Newington town line to the west. Glacial Lake Hitchcock had an approximate 4K yr. lifespan, from about 15,500-11,500 (radiocarbon) yrs. BP and stretched from central Connecticut northward, well into northern Vermont and New Hampshire. In Connecticut and southernmost Massachusetts, the lake was drained by 13,500 yrs. BP due to the initiation of glacio-isostatic rebound that caused the Rocky Hill dam to break and the lake floor to be exposed.

45.0 Traffic entering from left; merge and move into left lane.

45.8 Bear left onto I-84 entrance ramp. Outcrops of conglomeratic Portland Formation on both sides of ramp are braided stream channel facies. Note fining upward sequences and lack of overbank deposits.

46.2 Merge with I-84 west.

47.5 Coming over the hill consider the low area where the city of Hartford was built (elev. ~50'). The city was built on bottom sediments of glacial Lake Hitchcock. The area is underlain by varved clay and silt, annual layers that settled out in the glacial meltwater lake. The lake level lowered over its history, but at its stable level (approximately 115 ft. in the vicinity of the city) would have been about at the 6th floor of the major buildings (the high stand would be at about the 10th floor). A faint shoreline sand (beach) deposit is recognized (Colton, 1965) under the highway sign at the merge with I-384 (at mile 47.9; note slight decrease in the slope at that point). It has nearly the same elevation (120-150') as the top of the Hockanum River delta plain at the Penney's warehouse. The hill we just came over (called Sunset Ridge) is a drumlin and would have been an island in the lake before the Hockanum River delta filled in the near shore area of the lake connecting the island back to the mainland.

49.8 Passing left exit for I-91 south. Migrate into left lane after passing this exit.

50.9 Bear left at exit 55 onto Rte. 2 south.

55.4 Take exit 7 (another left exit) onto Rte. 17 south toward Portland.

61.1 Turn right onto Old Maid's Lane. Outcrop on left is chlorite-bearing gneiss of the Collins Hill Formation (Ordovician). Eastern Border Fault just ahead. Flat area ahead is underlain by sand and gravel that forms the surface of the Cromwell deltaic deposits built into the southern basin of glacial Lake Middletown. These deposits formed the eastern side of the Rocky Hill dam of glacial Lake Hitchcock. Looking west, the other side of the glacial lake dam can be seen at the same level across the valley, today breached by the Connecticut River and its terraces.

61.5 Turn left just before tobacco barn onto unpaved access road to gravel pit.

61.8 Gate to former Dufford Construction Company's gravel pit: if gate is open drive in. If gate is locked, walk about 0.5 mile to large outcrop on left in mined-out area *This is private property and we have permission (established ahead of time) to enter the area. If you follow this road log at another time, make sure you make appropriate arrangements ahead of time because you may get locked in if you do not.*

STOP 8. Portland Formation (Jp) fanglomerate, Glastonbury quadrangle, (Herz, 1955; Olsen et al., 2005)

The purpose of this stop is to examine the clasts in this fan-delta conglomerate, and to note the dip of the bedding. The stratigraphy, sedimentology, and paleontology of this outcrop have been thoroughly described by Olsen et al. (2005). They discuss the subdivision of these rocks, constituting the Hales Brook fan into several beds, both conglomeratic and fossil fish bearing shales. Those interested in more information should refer to this work.

Fig. Stop 7.1 Typical gravel filled channel at the base of one of many braided stream beds exposed here. Most pebbles are vein minerals, and very few pebbles are slates or phyllites. Book is 20 cm. long.

Fig. Stop 8.1. View of the south-facing dip slope of the conglomerate. The steep dip is a result of post Early Jurassic folding along the Rocky Hill anticline (Resor and de Boer, 2005)

Fig. Stop 8.2. Gray slates and phyllites (Bolton schist equivalents) dominate the clast composition here, but vein quartz and quartz + feldspar (large clast, center-left) are present.

Fig. Stop 8.3. Clasts of (Clough?) quartzite (pale gray) form an important minor clast population, but amphibolite (center, black) is rare.

There are two observations to make here in support of the themes of this trip. First, the provenance of these rocks varies significantly from Stop 7. Reconnaissance pebble counts show that slates, phyllites, and schists are the dominant population of the clasts (Fig. Stop 8.2). Quartzite forms a second but lesser component, and vein quartz is present. Amphibolite, arkose rip-up clasts, and basalt (Figs. Stop 8.2; 8.3) form very minor components. Missing is a major component of rocks derived from the Glastonbury gneiss and most rocks from east of the Bolton syncline. From this we can infer that there was a major divide on the east side of the Clough quartzite (Fig. 4B), such that rivers east of this structure flowed to the east. It is also likely that the Glastonbury Gneiss was covered by arkose. This would explain the lack of pegmatitic clasts (compare Stop 7), and the presence of arkose rip-up clasts.

The other point to be made here is that the Hartford basin experienced inversion. The moderate S to SSW dip of the bedding here has been interpreted to reflect folding of these rocks on the south side of the Rocky Hill anticline (Resor and de Boer, 2005). This fold and several other folds in Early Jurassic sediments (unnamed on Fig. 1) make a strong case for N-S shortening of these rocks in the later Jurassic. Such deformation supports the interpretation that the Hartford basin was inverted later in the Early Jurassic. This type of shortening parallels the kinematics of folds seen in Stop 3B, and is consistent with other brittle faults in the hinterland of this basin. Not only are the sediments of the Hartford basin in a foreland setting to the crystalline rocks to the east, but they may share a N-S deformational event caused by sinistral transpression along the Atlantic margin.

61.8 Return to cars and resume road log when passing gate.
62.1 Turn right onto Old Maids Lane
62.5 Turn left onto Rte. 17 North.
68.0 Merge with Rte 2 North
69.4 Exit 5D (right) into Rte. 3 South. Area is developed up to the edge of the flood plain of the Connecticut River. Cross Connecticut River at 71.7.
72.1 Bear right onto I-91 North (toward Hartford).
75.2 Rte. 15 merges on left. Migrate into left lane for next exit.
75.8 Left at exit 29A: Capitol Area
76.3 Right onto Columbus Avenue
76.5. Turn right into Marriott Hotel passenger discharge area.
END OF TRIP

NE GSA Welcome Reception 6:30-8:30pm - Exhibits Open/Cash Bar

UNRAVELING ALLEGHANIAN OROGENESIS IN SOUTHERN CONNECTICUT: THE HISTORY OF THE LYME DOME

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INTRODUCTION

This field trip highlights the geologic history of the Lyme dome, from Neoproterozoic deposition and magmatism to widespread deformation and migmatization in the Carboniferous to Permian Alleghanian orogeny. Recent studies of the Lyme dome (figs. 1 and 2) (Walsh and others, 2007, 2009) constrain the origin of the rocks in the core of the dome, the absolute timing of four principal deformational and thermal events attributed to Alleghanian orogenesis, and the processes that generated the dome. The trip summarizes the recent work and visit some spectacular bedrock exposures along the Connecticut coast of Long Island Sound (figs. 1 and 2). We will focus on the mechanisms and timing of dome formation identified through geologic mapping, structural analysis, and sensitive high-resolution ion microprobe (SHRIMP) U-Pb geochronology. We will also discuss why the potential discovery of Gander zone rocks in the core of the Lyme dome (Wintsch and others, 2005; Aleinikoff and others, 2007) is critical to testing the concept of continuity of terranes and the timing of terrane accretion throughout the northern Appalachians.

Detrital zircon geochronology in combination with ages on intrusive rocks brackets the deposition of quartzite in the core of the dome sometime between about 925 and 620 Ma. Granite and granodiorite intruded the Neoproteorozic metasedimentary rocks in the core of the dome between about 620 to 580 Ma. Four major early Permian events associated with the Alleghanian orogeny affected the rocks in the Lyme dome area. Syn-tectonic migmatization and widespread penetrative deformation (D_1, ≈300 – 290 Ma) included emplacement of alaskite at 290±4 Ma during regional foliation development in the silllimanite-orthoclase zone. This event was associated with the foliation-parallel (S_1) formation of neosome in stromatic migmatites (metatexite migmatite of Sawyer, 2008). Rocks of the Avalon terrane may have wedged between Gander cover rocks and Gander basement in the core of the Lyme dome during D_1. Limited structural evidence for diapiric uplift of the Lyme dome indicates that diapirism started late in D_1 and was completed by D_2 (≈290 – 280 Ma) when horizontal WNW contractional stresses dominated over vertical stresses. Second sillimanite metamorphism continued and syn-tectonic D_2 granitic pegmatite (288±4 Ma) and the Joshua Rock Granite Gneiss (284±3 Ma) intruded at this time. North-northwest extension during D_3 (≈280 – 275 Ma) led to granitic pegmatite intrusion along S_3 cleavage planes and in extensional zones in boudin necks during hydraulic failure and decompression melting. Intrusion of a dike of Westerly Granite at 275±4 Ma suggests that D_3 extension was active, and perhaps concluding, by about 275 Ma. Late randomly oriented but gently dipping pegmatite dikes record a final stage of intrusion during D_4 (≈275 – 260 Ma), and a change from NNW extension to rapid vertical unloading and exhumation. Monazite and metamorphic zircon rim ages record this event at ca. 259 Ma. The evolution of the Lyme dome involved D_1 mylonitization, intrusion, and migmatization during north-directed contraction, limited late D_1 diapirism, D_2 migmatization during WNW contraction with associated flexural flow and fold interference, D_3 NNW horizontal extension and decompression melting, and final D_4 vertical extension and rapid exhumation. Late regional uplift, extension, and normal faulting at higher crustal levels may have been caused by diapiric rise of the underlying lower crust, below the structural level of the Lyme dome. The rocks record no evidence of Acadian metamorphism or deformation, suggesting that the Gander zone here was not tectonically juxtaposed with Avalon until the Alleghanian orogeny.

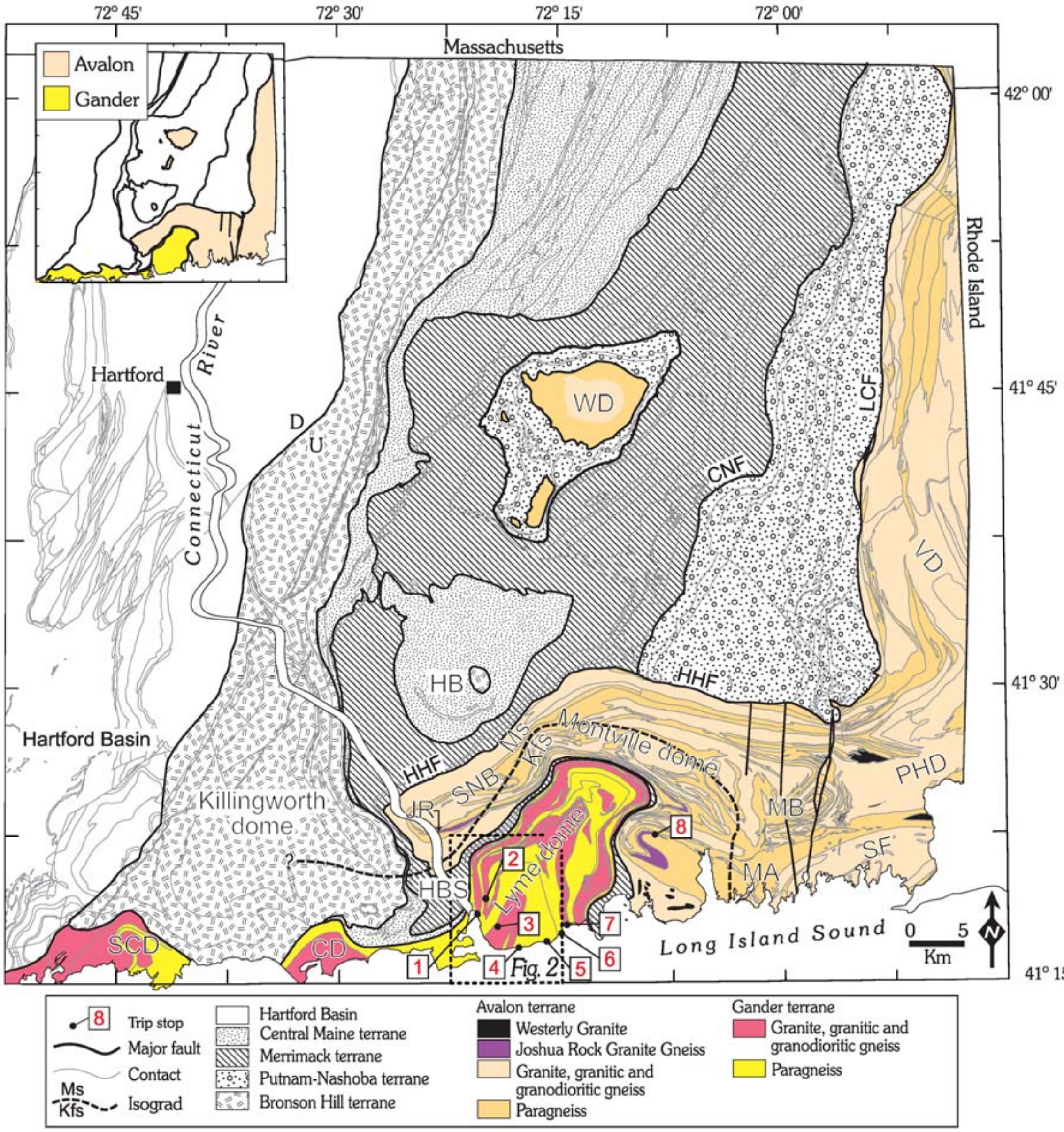

Figure 1. Generalized tectonic map of eastern Connecticut showing the distribution of terranes and major Alleghanian domes and basins (modified after Rodgers, 1985; Goldmith, 1985; Walsh and others, 2007). Upper left inset map shows the distribution of the Gander and Avalon terranes. Abbreviations: LCF – Lake Char fault, CNF – Clinton-Newbury fault, HHF – Honey Hill fault, WD - Willimantic dome, HB – Hopyard basin, HBS – Hunts Brook slice, SCD – Stony Creek dome, CD – Clinton dome, SNB – Selden Neck block, MA – Mystic antiform, MB - Mystic basin, SF – Stonington fold, PHD – Potter Hill dome, VD – Voluntown dome, JR – Joshua Rock, and Ms – muscovite and Kfs – potassium feldspar along the second sillimanite isograd. Reproduced by permission of the American Journal of Science.

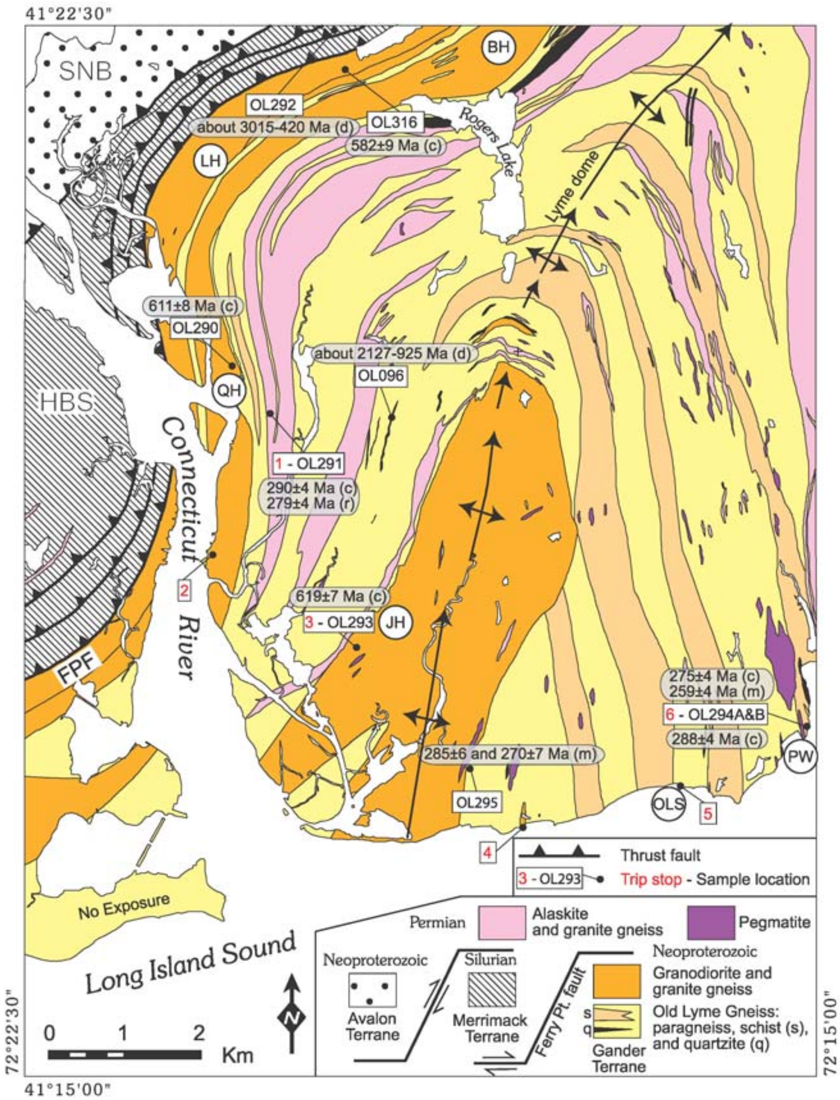

Figure 2. Simplified geologic map of the Old Lyme quadrangle showing field trip stops (labeled in red), sample locations (labeled in black), and associated U-Pb zircon ages (in gray boxes); modified after Walsh and others (2007, 2009). Abbreviations: SNB – Selden Neck block, HBS – Hunts Brook slice, FPF – Ferry Point fault, BH – Becket Hill, JH – Johnnycake Hill, QH – Quarry Hill, LH– Lord Hill, OLS – Old Lyme Shores, PW – Point O'Woods, c – igneous zircon core, d – detrital zircon, m – metamorphic zircon, and r – metamorphic zircon rim. Reproduced by permission of the American Journal of Science.

ROAD LOG

This trip will begin and end in Hartford, CT. We will travel southeast along the Connecticut River to Old Lyme, and then proceed eastward along the shores of Long Island Sound. We will return via routes 11 and 2 to Hartford. The drive to the Old Lyme area and the return trip take about 45 minutes. In this report stop coordinates are given in latitude and longitude (WGS84 datum). Minerals in rock descriptions are listed in order of increasing abundance.

Meeting point: Meet at 8:45 AM on Saturday March 17, 2012 at the Hartford Marriott Downtown, 200 Columbus Boulevard, Hartford, CT 06103 (41°45.833'N, 72°40.183'W). We will leave at 9:00 AM and return to Hartford by 6:00 PM.

- From downtown Hartford get on I-91 South.
- Take exit 22S on the left for Route 9 South. An outline of the geology exposed along Route 9 can be found in Skehan (2008).
- At the end of Route 9, bear left onto I-95 and U.S. Route 1 North.
- Cross the bridge over the Connecticut River and take exit 70 on the right for Route 156 in Old Lyme. Mileage for the road log begins at the traffic light at the bottom of the off ramp.

Mileage

0.0 Turn left onto Route 156 West.
0.1 Continue straight at the traffic light at the junction of Route 156 and U.S. Route 1.
0.8 Turn right onto Talcott Farm Road.
0.9 Turn left staying on Talcott Farm Road.
1.0 Park at roadcut for Stop 1.

STOP 1. Permian alaskite and granite gneiss at Talcott Farm Road (20 minutes)
(41°19.731' N, 72°20.105' W)

This roadcut exposes light pink, locally rusty-weathering, medium- to coarse-grained, well-foliated quartz-plagioclase-K-feldspar alaskite gneiss to biotite-K-feldspar-quartz-plagioclase granite gneiss. Thin screens of rusty weathering biotite-rich schist occur parallel to the foliation, and they are interpreted as xenoliths of the surrounding country rock, the Old Lyme Gneiss. Pegmatite to granitic leucosome occurs parallel to the foliation, in cross-cutting irregular patches, and as dikes.

Alaskite and granite gneiss (fig. 2) occurs as large sills and irregular intrusive bodies into the rocks of the Old Lyme Gneiss in the Lyme dome. The rock generally exhibits a strongly developed foliation, but locally it is coarse grained and poorly foliated. The foliation developed within the gneiss has the same relative age as the dominant foliation within the host rocks of the Old Lyme Gneiss, and it is mapped as S_1. Here the foliation has a strike and dip of 180º, 60º and we are on the western limb of the dome.

Lundgren (1967) mapped similar rocks as units "sga," "sgba," and "sgm" of the Sterling Plutonic Group. On the State map, Rodgers (1985) agreed with Lundgren and further subdivided Lundgren's units as Potter Hill Granite Gneiss and others as Hope Valley Alaskite Gneiss. Our mapping suggests a compositional similarity with some of the rocks mapped as the Neoproterozoic Sterling Plutonic Group (or Suite) (Rodgers, 1985), but a SHRIMP U-Pb zircon age from the alaskite and granite gneiss sampled at this outcrop is 290±4 Ma (sample OL291; Walsh and others, 2007), indicating the rock is Permian. We, therefore, have abandoned the use of the names Sterling Plutonic Group (or Suite), Potter Hill Granite Gneiss, and Hope Valley Alaskite Gneiss for the alaskite and granite gneiss. Instead, we classify this unit as unnamed Permian intrusive rocks. The presence of foliated early Permian intrusive rocks in this part of Connecticut suggests that other rocks mapped as Sterling Plutonic Group in southern Connecticut (Rodgers, 1985) need to be re-evaluated. Because modal abundances of alaskite gneiss overlap with those of the granite gneiss at Becket Hill (Zgb), a degree of uncertainty in identification of granitic units exists. The granite gneiss at Becket Hill generally contains more biotite. We will see this rock again at Stop 7, but there it is undated, and it will highlight the challenge of mapping in areas that contain very similar rocks with quite different ages.

Mileage
0.0 Make a U-turn.
0.1 Bear right staying on Talcott Farm Road.
0.2 Turn left at stop sign onto Route 156 East.
1.0 Junction with I-95 at Exit 70; continue on Route 156 East.
1.5 Turn right on Ferry Road
1.9 Turn left into parking lot for Stop 2.

STOP 2. Granite gneiss at Becket Hill exposed along Ferry Road (20 minutes)
(41°18.792’ N, 72°20.683’ W)

Coarse-grained, moderately to well-foliated, pink magnetite-biotite granite orthogneiss is exposed at a small abandoned quarry on the east bank of the Connecticut River (fig. 2). Magnetite and biotite are more abundant here than at Stop 1 and the relative abundance of these minerals was useful for mapping and for distinguishing this rock from the alaskite and granite gneiss unit in the field.

Locally, but not at the quarry, lenses of Old Lyme Gneiss are enclosed within the map unit, and the units contact are discordant relative to layering in the Old Lyme Gneiss. Here the foliation (S_1) has a strike and dip of about 200º, 60º and we are still on the western limb of the dome. A weak L_2 crenulation lineation plunges gently to the north-northeast, and this is the expression of the dome-stage folding in the orthogneiss at this location.

On the State map (Rodgers, 1985), the granite gneiss at Becket Hill (Zgb) was assigned to the Potter Hill Granite Gneiss of the Sterling Plutonic Group. Hermes and Zartman (1985) later redefined the Sterling Plutonic Group as the Sterling Plutonic Suite. The type locality for the Potter Hill Granite Gneiss is located in the Avalon terrane in Rhode Island (Feininger, 1965). Walsh and others (2007, 2009) used the informal name, granite gneiss at Becket Hill, because the granite gneiss mapped here was interpreted to be part of the Gander terrane. The granite gneiss at Becket Hill cuts the rocks of the Old Lyme Gneiss. A SHRIMP U-Pb zircon age of 611±8 Ma from a sample at Quarry Hill (fig. 2, sample OL290, Walsh and others, 2007) further constrains the Neoproterozoic age of 634±29 Ma obtained by Zartman and others (1988) from a sample on Quarry Hill (their sample no. 2 PEC–707).

Mileage
0.0 Turn right out of the parking lot onto Ferry Road.
0.4 Turn right onto Route 156 East.
1.9 Turn left onto Mile Creek Road, and then make an immediate left onto Johnnycake Hill Road.
2.0 Park at the end of the road for Stop 3.

STOP 3. Granodiorite gneiss at Johnnycake Hill exposed at Black Hall (20 minutes)
(41°18.792' N, 72°20.683' W)

This stop is in the core of the Lyme dome (fig. 2) where light-gray, moderately foliated to massive, medium- to coarse-grained biotite-K-feldspar-quartz-plagioclase granodiorite orthogneiss crops out. Coarse pegmatite is exposed along the east side of the road across from the mailbox. Outcrops in this unit are typically homogeneous and show very little compositional segregation. Here the rock contains accessory amounts of hornblende and rare, small (<10 cm) pods or xenoliths of garnet-sphene-diopside-hornblende-biotite-quartz-plagioclase calc-silicate rock. The rock locally contains anastomosing granitic leucosomes. The massive nature, the presence of amphibolite and calc-silicate xenoliths, and the crosscutting nature of the contact with the Old Lyme Gneiss support the interpretation that the rock is a pluton. A SHRIMP U-Pb zircon age of 619±7 Ma (sample OL293, Walsh and others, 2007) provides the oldest reliable age from an igneous rock in coastal Connecticut. The age places an upper limit on the Old Lyme Gneiss which yields detrital zircons ranging from about 2,127-925 Ma (Walsh and others, 2007). The dated granodiorite gneiss was collected from a low railroad cut on the north side of the tracks, west of the foot bridge. Caution, these train tracks are extremely dangerous!

The granodiorite gneiss here was not mapped by Lundgren (1967) but was briefly described (1967, p. 28) as "quartz-dioritic and granodioritic in composition, whether volcanic or intrusive" in the lower part of his Plainfield Formation (our Old Lyme Gneiss). On the State map, Rodgers (1985) reinterpreted Lundgren's (1967) map and identified the rocks in the core of the Lyme dome as a mixture of Plainfield Formation (our Old Lyme Gneiss), Potter Hill Granite Gneiss, and Narragansett Pier Granite (unit "Zp+Zsph+Pn?"). Our work supports Lundgren's findings, and suggests a lithologic similarity to rocks mapped in the Avalon terrane as either Neoproterozoic Rope Ferry Gneiss or New London Gneiss (Rodgers, 1985; Webster and Wintsch, 1987; Aleinikoff and others, 2007). Further work is needed, however, to more clearly document the ages and isotopic signatures of these rocks to determine the validity of the assignment of these rocks to the Gander or Avalon terranes.

Mileage

0.0 Make a U-turn.
0.1 Turn right onto Mile Creek Road and then make an immediate left onto Route 156 East.
1.0 Turn right onto Old Shore Road.
1.7 Turn right onto Brighton Road
2.2 Enter the Old Lyme Beach Club (or park at the gate if the gate is closed; fig. 3).

STOP 4. Granodiorite gneiss at Johnnycake Hill exposed at the Old Lyme Beach Club
(60 minutes - Lunch)
(41°16.778' N, 72°17.887' W at gate; 41°16.776' N, 72°17.752' W at outcrop)

This outcrop offers an isolated exposure of the biotite granodiorite orthogneiss in the core of the dome (fig. 2). The closest outcrops, some 800 m away along Route 156, are metasedimentary rocks of the Old Lyme Gneiss and Permian pegmatite, and the outcrop at this stop is interpreted as an isolated body of the granodiorite or possibly an apophysis of the main pluton found in the core of the dome. The gneiss is cut by biotite pegmatite dikes and sills, and contains irregular patches of granitic and pegmatitic leucosome. The irregular patches of neosome create a net-structured migmatite. Large pegmatite bodies vary from foliation-parallel sills to irregular discordant masses that generally follow the trend of the S_1 foliation but are locally highly discordant. Here we are on the east limb of the Lyme dome, and the foliation (S_1) strikes north-south and dips moderately at about 40-50º to the east. The dominant foliation (S_1) is deformed by weakly developed, gentle north-northeast plunging F_2 folds with more readily visible L_2 intersection lineations and east-west trending boundinage and open folds (F_3) and shear bands. Late tabular, moderately to gently dipping pegmatite dikes cross-cut all ductile fabrics and are related to fracturing as the result of rapid decompression during D_4 (Walsh and others, 2007).

Figure 3. Google Earth image of the Old Lyme Beach Club showing the outcrop at Stop 4.

Mileage

0.0	Make a U-turn. Mileage reset from entrance gate.
0.5	Turn right onto Old Shore Road.
0.8	Turn right onto Route 156 East.
2.1	Turn right onto Billow Road, entering Old Lyme Shores.
2.5	Park on the right at the end of Billow Road. Walk southeast across the beach to the outcrop (fig. 4).

STOP 5. Migmatitic Old Lyme Gneiss at Saltworks Point, Old Lyme Shores (30 minutes)
(41°17.100' N, 72°16.300' W at parking area; 41°17.063' N, 72°16.272' W at outcrop)

This is an excellent exposure of some of the major rock types found in the Old Lyme Gneiss. The outcrop is dominated by stromatic migmatite (metatexite) consisting of well-layered, K-feldspar-biotite-plagioclase-quartz gneiss (fig. 5). The migmatitic rocks at this location were previously described by McLellan and others (1993) and by Lundgren (1966, 1967). Alternating layers of gray biotite-rich quartz-feldspar gneiss with lesser garnet amphibolite and minor quartzite form the paleosome of the migmatite. The biotite-rich quartz-feldspar gneiss, mapped as the "Biotite gneiss of the Old Lyme Gneiss" (unit "Zo" of Walsh and others, 2009) is the most abundant metasedimentary rock in the area. Thin layers of dark-gray to black garnet-sillimanite-K-feldspar-biotite-plagioclase-quartz schist are characteristic of the "sillimanite schist of the Old Lyme Gneiss" (unit "Zos" of Walsh and others, 2009) mapped elsewhere, but are too thin to map separately here. An area of amphibolite and garnet amphibolite gneiss several meters thick occurs on the northwest end of the outcrop, and this rock is mapped as "amphibolite of the Old Lyme Gneiss" (unit "Zoa" of Walsh and others, 2009). Metamorphic zircon from a similar amphibolite (OL295, fig. 2) yielded ages of 285±6 and 270±7 Ma that are interpreted to reflect two distinct periods of growth of zircon that are correlated with episodic regional D_2 and D_3 activity (Walsh and others, 2007).

Granitic neosome occurs sub-parallel to the dominant foliation (S_1) and as cross-cutting dikes. The S_1 has a strike and dip of about 0º, 50º and we are on the eastern limb of the dome. The dominant foliation (S_1) is locally tightly folded by F_2 folds, but it is the younger D_3 boudinage and F_3 folds that are the more readily apparent younger deformation at this stop (fig. 5). Note how the boudin necks act as dilational sites for emplacement of D_3-stage granitic pegmatite and granite. McLellan and others (Stop 16, 1993) noted that miarolitic cavities in the youngest leucosomes provide evidence for rapid decompression following peak metamorphic conditions of about 735ºC and 7 kbar. Walsh and others (2007) attributed the onset of decompression to the D_3 phase of dome formation.

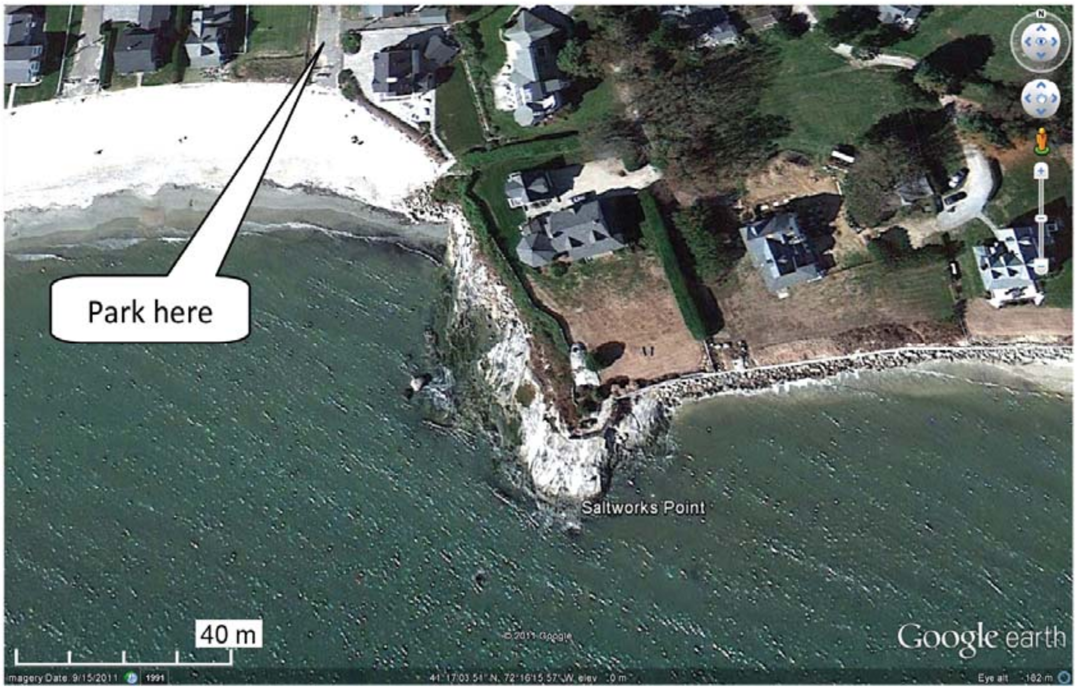

Figure 4. Google Earth image showing the outcrop at Saltworks Point (Stop 5).

Figure 5. Map view photograph of stromatic migmatite (metatexite) in the Old Lyme Gneiss at Saltworks Point (Stop 5). The rock is classified as the biotite gneiss unit (Zo) of Walsh and others (2009). The dominant foliation (S_1) is parallel to the hammer, and is deformed by east-west trending F_3 folds and boudinage that create dilational sites for D_3-stage granite and pegmatite.

Mileage

0.0 Make a U-turn.
0.3 At stop sign, turn right onto Route 156 East.
1.2 Turn right onto Connecticut Road and go under the railroad tracks.
1.3 Immediately after the railroad overpass, turn left onto Hillcrest Road.
1.5 Bear left staying on Hillcrest Road.
1.9 Park on the right at the end of the road, and walk to the outcrop along the shore (fig. 6).

STOP 6. Migmatitic Old Lyme Gneiss at Point O'Woods (45 minutes)
(41°17.450' N, 72°15.110' W at parking area)

This is an excellent exposure of the "sillimanite schist of the Old Lyme Gneiss" (unit "Zos" of Walsh and others, 2009). Here we will also see the dated sample locations of the Permian Westerly Granite dike and Permian granite pegmatite reported in Walsh and others (2007).

The outcrop is dominated by migmatitic well-layered, garnet-sillimanite-K-feldspar-biotite-plagioclase-quartz schist to gneiss. Granitic neosome occurs parallel to the S_1 foliation, similar to the rocks at Stop 5. Note that this unit is more schistose than the "biotite gneiss" (unit Zo) at Stop 5, and the more pelitic character of this rock was the primary criterion for separating these two major rock units (Zo and Zos) in the Old Lyme quadrangle (Walsh and others, 2009). The eastern shore of the Point O'Woods peninsula exposes an area of amphibolite (unit "Zoa" of Walsh and others, 2009), and a belt of interlayered quartzite and schist (unit "Zolq" of Walsh and others, 2009).

Granitic neosome which is sub-parallel to the dominant foliation (S_1) is highly deformed by upright north-striking and east-dipping F_2 folds with moderate to steeply plunging fold axes. The F_2 folds are related to the main stage of dome formation (Walsh and others, 2007; 2009), and locally outcrop-scale analogs to the Lyme dome can be seen (fig. 6). In places, granitic pegmatite cross-cuts folded S_1 and in turn is deformed by the F_2 folds and associated S_2 cleavage. In one such place (fig. 6), the pegmatite yielded a U-Pb zircon age of 288±4 Ma and a monazite age of 287±2 Ma, placing an upper age limit on D_2 deformation (sample OL294B, Walsh and others, 2007). Younger, light-pink dikes of Westerly Granite post-date most of the pegmatite and all ductile structures in the deformed rocks. Sample OL294A (Walsh and others, 2007) was collected from a 0.8-m-thick granite dike here at Point O'Woods, near the gazebo adjacent to the seventh house north of the point (fig. 6). The sample yielded a U-Pb zircon age of 275±4 Ma, which places an upper age limit on the timing of major ductile deformation in the area.

Figure 6. Google Earth image of Point O'Woods (Stop 6) showing the locations and SHRIMP U-Pb zircon ages of two dated samples from Walsh and others (2007). Inset photograph shows the location of an F_2 fold that is an outcrop-scale analog to the Lyme dome.

Mileage

0.0	Make a U-turn.
0.6	Turn right onto Connecticut Road and go under the railroad tracks.
0.7	At stop sign, turn right onto Route 156 East.
1.5	At stop sign bear right, staying on Route 156 East.
2.6	Turn right into Rocky Neck State Park. Follow the signs to the beach.
4.1	Park in the parking lot, and walk under the railroad tracks through the West Beach Access (fig. 7).

STOP 7. Alaskite and biotite granite gneiss at Lands End in Rocky Neck State Park (45 minutes) (41°18.033' N, 72°14.583' W at parking lot)

From the West Beach Access follow the path towards the pavilion to extensive exposures of light gray to pink biotite granite gneiss and lesser pink alaskite gneiss. We are now in the Niantic quadrangle, where Goldsmith (1967a) mapped these rocks as biotite granite gneiss of the Sterling Plutonic Group (or Suite) of "Pre-Pennsylvanian" age. Rodgers (1985) included this belt of rocks in the Neoproterozoic Potter Hill Granite Gneiss. This belt of rock was traced to the north and west around the Lyme dome to Stop 1, where the rock yielded an age of 290±4 Ma. This stop highlights the challenge of mapping rocks like this and assigning ages without the benefit of modern U-Pb geochronology. Is it Neoproterozoic or Permian, or a combination of the two? Clearly there are areas of younger pink to orange granitic leucosome and pegmatite which post-date the foliated granitic gneiss and are easiest to interpret as Alleghanian melts, but what is the age of the granitic host rock?

At the point of the Lands End peninsula (fig. 7, 41°17.833' N, 72°14.792' N) you can see abundant granitic pegmatite cutting the foliated granitic gneiss. An exposure on the western shore of the peninsula (fig. 7, 41°17.900' N, 72°14.854' N) marks one of the few places where you can see the sharp contact with the older paragneissic host rock. The host is a migmatitic sillimanite-biotite-k-feldspar-quartz schist to gneiss that resembles the Old Lyme Gneiss seen in the Old Lyme quadrangle to the west. The contact between the paragneiss and the orthogneiss is sharp, parallel to the dominant foliation (S_1), and cut by pegmatite dikes.

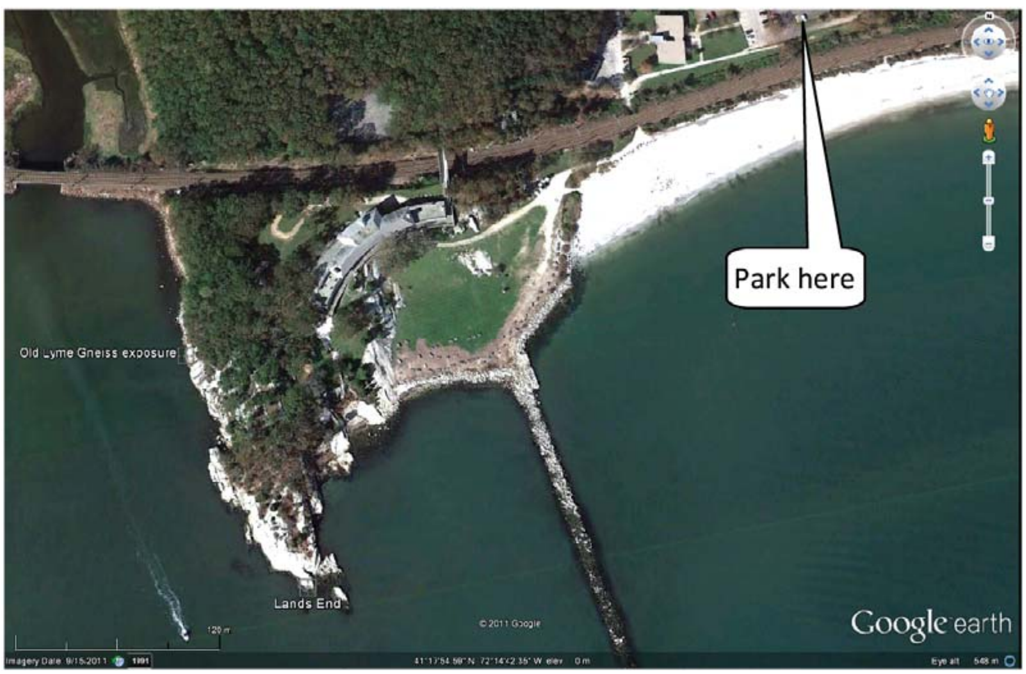

Figure 7. Google Earth image of Lands End at Rocky Neck State Park (Stop 7).

Mileage

0.0 Make a U-turn and return to the entrance of Rocky Neck State Park (RNSP) on Route 156.
1.4 Turn left onto Route 156.
1.6 At the traffic light, turn right onto the access road for I-95.
2.1 Bear right onto the I-95 North.
6.8 Bear right at the junction with I-395, staying on I-95 North.
10.1 Take exit 82 on the right for Route 85.
10.4 At the traffic light at the bottom of the off ramp, turn left onto Route 85 North.
10.8 At the traffic light, turn left into the Crystal Spring Mall.
10.9 Bear right at the fork, keeping the mall to your left (west-southwest).
11.1 Park in the mall parking lot, across from the roadcut on the right (northeast).

STOP 8. Joshua Rock Granite Gneiss at the Crystal Mall in Waterford (30 minutes)
(41°22.500' N, 72°08.520' W)

This roadcut exposes the Joshua Rock Granite Gneiss (JR) where it intruded the New London Gneiss in the Avalon terrane (Zartman and others, 1988). This rock yielded a Permian SHRIMP crystallization age of 284±3 Ma (Walsh and others, 2007), which supersedes a poorly constrained conventional thermal ionization mass spectrometry (TIMS) age (Zartman and others, 1988, sample PEC-702). The long roadcut exposes light-gray to light-pinkish-gray, aegerine-augite granite of the JR below a belt of darker colored, banded amphibolite and granodioritic gneiss of the New London Gneiss (fig. 8). The JR is exposed on the southeast end of the roadcut where it intruded amphibolite gneiss (fig. 8). Well-layered gneiss is exposed at the northwest end of the cut (fig. 8). The JR contains a planar foliation that is "parallel to the structural trend of the Alleghanian orogeny" (Zartman and others, 1988, p. 392). Goldsmith (1961) interpreted this belt of rock as a phacolith, and originally mapped the aegerine-augite granite as the Joshua Rock Gneiss Member of the New London Gneiss (Goldsmith, 1967a, b). At the roadcut, the granite contains a very weak, wispy foliation that is generally parallel to the dominant fabric in the gneissic country rock (S_1) (fig. 8). Local rafts (schollen) of the New London Gneiss occur in the JR (fig. 8). The dominant fabric appears to be partially to totally annealed or resorbed in places, locally leaving remnant schlieren of biotite and hornblende. The amphibolite at the contact with the granite gneiss contains abundant granitic leucosomes and thin, foliation-parallel to slightly discordant intrusive sills suggesting that the host rock was injected with leucosome during, or very late in the development of S_1. Zartman and others (p. 398, 1988) stated that, "it is ambiguous whether the rock exhibits well preserved igneous texture or if the rock has been completely annealed and is nearly strain free due to high temperature recrystallization." The rock could be classified as a schollen diatextite migmatite using the conventions of Sawyer (2008). Second generation folds (F_2) and weak foliation (S_2) deform the Joshua Rock Granite Gneiss, and several S- and Z-shaped folds are outcrop-scale examples of the large "S fold" on the Connecticut State geologic map (Rodgers, 1985) (fig. 1). The local orientations of the F_2 folds and S_2 foliation generally strike southeast-northwest, dip steeply, and differ from the orientations seen in the Old Lyme quadrangle where they strike northeast. Goldsmith (fig. 3, 1985) showed both the NE and SE-NW trending folds as undifferentiated younger folds, and we suggest that that they are generally contemporaneous D_2 structures.

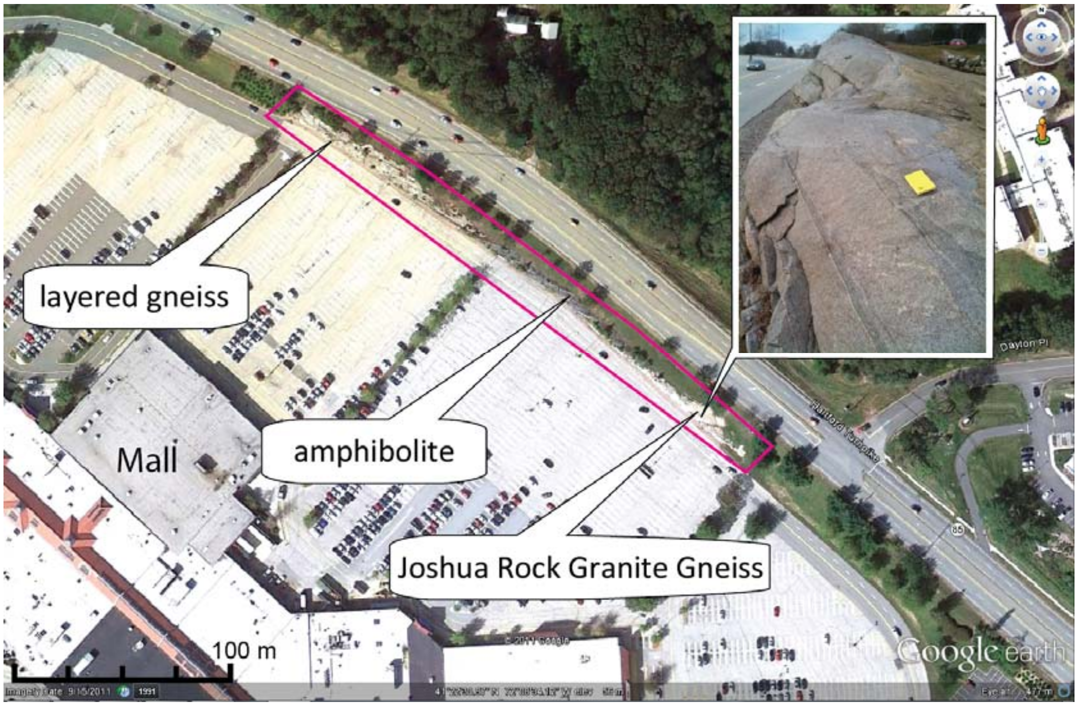

Figure 8. Google Earth image of the roadcut (outlined in magenta) at the Crystal Mall (Stop 8). The inset photograph shows the light-pinkish-gray Joshua Rock Granite Gneiss as viewed looking northwest along the roadcut, with darker-colored schollen of the older gneiss parallel to the S_1 foliation; the yellow book is 20 x 13 cm.

END OF TRIP

Return to Hartford by following the instructions below:

- From the mall turn left onto Route 85 North.
- See statue of Monty the T-rex at The Dinosaur Place on the left in Oakdale.
- In Salem, turn left on Route 82 West following the signs for Route 11 North.
- Turn right onto Route 11 North. From here to Hartford you will be on a limited-access highway.
- In Colchester, Route 11 merges with Route 2 West. Follow Route 2 West to Hartford.
- Take Exit 2W on the left toward Downtown Hartford
- Turn Left onto Columbus Blvd
- Hartford Marriott is on the left after the Connecticut Science Center

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THE GEOLOGY OF WALDEN POND

Robert Thorson, University of Connecticut, Storrs, Connecticut

Full length of Walden Pond from east to west in early September, 2010.
Dark band at bottom is the swimming beach in deep shade shortly after sunrise.

CAVEAT
Though I will lead this trip and have written its guide, I am not an expert on the local geology. I have neither carried out nor published original field research relevant to the field trip stops, and must therefore rely on the technical literature as filtered through my substantial familiarity with the terrain. Happily, there is plenty of excellent work to cite and summarize.

INTRODUCTION

My expertise related to the field trip is not with the geology, *per se*, but with the way in which that geology has become incorporated into local history and culture. I refer to something I informally call "cultural geology," a phrase I first heard from Harry Foster, my former editor at Houghton & Mifflin, who uttered it while trying to make sense of my purpose in writing about the historic stone walls of New England. I consider cultural geology to be our discipline's contribution to American Studies. The obvious part of this effort concerns the environmental determinism of the physical landscape: a tough mountain root dominated by granite and high-grade metamorphic rocks; exhumed by denudation along the Atlantic passive margin; and rough-handled by multiple glaciations within the last million years. The result is a landscape dominated by: clean streams flowing in watersheds cut below ancestral erosion surfaces; a rocky coast exclusive of the sandy moraine archipelago to the south; and interior soils that are stony, sandy, or muddy, depending on whether sub-glacial, glacio-fluvial, or glacio-lacustrine processes stood last in line to bury the now-tranquil, crystalline legacy of the Paleozoic. The geology creates the geography, which influences the economy and history.

The less obvious and more interesting part of cultural geology is more subtle. It concerns the means by which the fairly new subject of geology (> 1807) influenced literate thinking within the fairly new republic of the United States (>1787), especially for the heirs of the American Revolution. More specifically, geology was the rising “science of the day” during the time when our national psyche was being forged. Years ago, I tried to summarize these somewhat inchoate ideas in two dense statements for the inaugural geology entry of the *Encyclopedia of New England*. First, "Geology, the meat and bones of science, gave natural history the depth of time and the universality of process that would help transition European Calvinism into American Transcendentalism." More to the point: “The soul of New England perches on a rock.” (Thorson, 2006)

Continuing with this endeavor for much of the last decade, I’ve devoted much of my scholarly attention targeting an audience dominated by those primarily interested in environmental literature and history. This field trip seeks to do just the opposite: to provide geologists with an opportunity to visit one of the most culturally significant sites in America: arguably the birthplace of its nature writing and the symbolic epicenter of its current environmental consciousness. “No parallel tract of body of water or place,” the historian W. Barksdale Maynard writes, “has so captivated the human imagination.” Walden has become “an international shrine” (Thorson, 2009). My working premise is that, having discovered or re-discovered the geology of Walden Pond, you’ll have something new to talk about with your humanities-based colleagues. My interest in the geology of Walden Pond follows legacy dating back to self-taught glacial geologist, John Muir and his successors (Inners, 2006).

To make a long story short, and amid, more typical academic obligations, I eventually found myself under contract to complete a book manuscript by January 2013 on the physicality of Walden Pond -- its geology, hydrology, limnology, meteorology, optics, acoustics – as related to Thoreau’s masterpiece *Walden,* which – to a large extent – derives from his scientific observations. When my colleague Jean Crespi, (the technical chair of GSA’s Northeastern Section Meeting) caught wind of this, she agreed that a trip to a popular tourist site like Walden Pond would be a nice complement to the other more scientifically rigorous trips. On the downside, being under contract to Harvard University Press, prevents me from giving away the store in writing. So what you get here is the in-stock inventory instead.

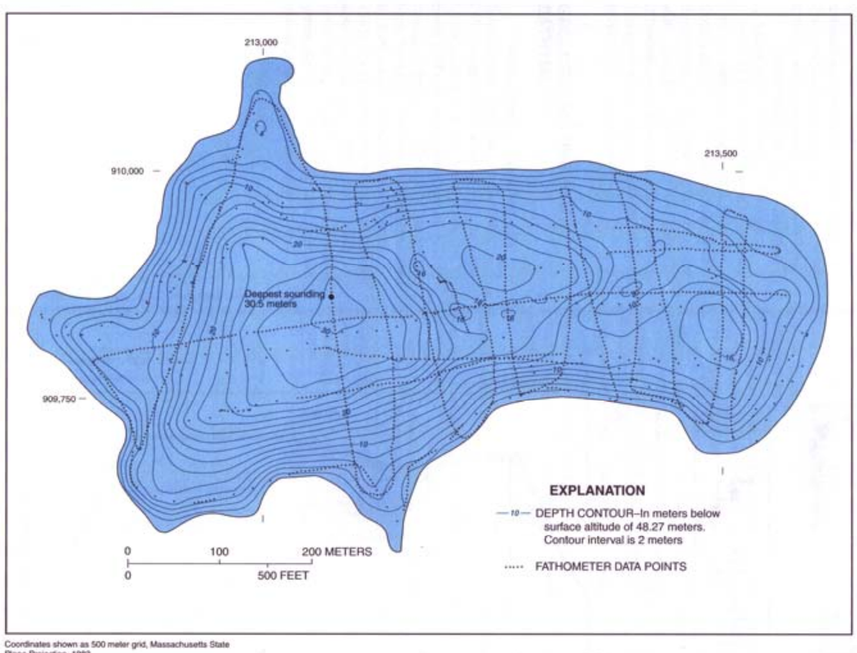
FUNDAMENTALS

To make another long story short, and telling it backwards in time:

- Walden Pond is technically a ***lake*** based on every modern criteria – area (62.5 acres; 25.3 ha), depth (102-107 feet; ca. 31-33 m), and ecological functioning (annual hypolimnion) and was generally considered a lake even in Thoreau’s day. Its bathymetry, geohydrology and limnology are very well constrained by Colman and Friesz (2001). The lake basin doesn’t “hold” water. Rather, it is a deep void that extends approximately 100 feet (31 m) below the average height of the water table, which fluctuates within a historic range of about 12 feet (4 m).

- It's a ***kettle** lake*, meaning one formed by the downward and/or inward melting of one or more blocks of glacial ice. In this case, the coalescing of four separate kettles, basins, one of which (Wyman's Meadow) toggles back and forth between being a bay of the lake into which one may float a boat and a dry marsh. The largest kettle is the main basin to the west, which is surrounded by coves that are aligned with rock structure and represent differential collapse (Thoreau, 1854). The date of deglacation is only broadly constrained between 17,200 and 15,800 calendar years before present (13.3-14.3 ^{14}C kyr B.P.) based on the chronology of the North American Varve Project (Ridge, 2004 and on-line updates).

- The *kettle* formed by collapse of ***ice blocks*** that were completely surrounded and buried beneath glaciodeltaic meltwater sediments forming the proximal facies of what Woodworth (1898) called a glacial wash plain, Goldthwait (1905) a glacial sand plain or kame delta, Koteff (1963) a delta, and more recent work a "morphosequence," (Koteff and Pessl, 1981) built into, and above the stable water-plane surface of the Cherry Brook phase of Glacial Lake Sudbury largely by meltwater streams issuing from the base of the ice sheet (Gustavson and Boothroyd, 1987) and draining its margin in a west-southwesterly direction. Ridges of coarse gravel containing large boulders (Heywood Peak) protrude well above the delta surface, are aligned with the ice margin, and separate the delta plain from the nearly coeval and more highly collapsed and topographically lower terrain immediately to the north: they could be considered moraine ridges. Following drainage of Glacial Lake Sudbury (and the subsequent Glacial Lake Concord) unfilled sections of the lake became extensive poorly drained flats, owing to the combination of low infiltration, and isostatic uplift (Koteff et al., 1993) of resistant reaches to the north. These became the rich meadows responsible for colonial settlement. (Donahue, 2004)



Bathymetry of Walden Pond, Concord, Massachusetts

- The two deepest *ice blocks* lay above a thin cover of "till" of unspecified origin within ***bedrock basins*** (Colman and Friesz, 2001) created by sub-glacial erosion within a pre-glacial bedrock valley aligned in a southeast-northwest direction along the strike of a Paleozoic thrust fault syntectonically intruded by the Andover Granite (Barosh, 1993; Zen et al., 1983).

- The bedrock basins pre-glacial ***bedrock valleys*** formed above less resistant lithologies within the Nashoba thrust belt, a distinct terrane of intensely deformed and highly metamorphosed sediments largely of volcanic origin that was caught in the collision between the Avalon Terrane to the east (separated by the Bloody Bluff Fault Zone) and other rocks to the west (separated by the Newbury Fault Zone). The dominantly granitic intrusive near Walden Pond (Andover Granite) may range from 630-380 million years in age. (Barosh, 1993), but are generally considered Ordovician-Silurian in age (SOagr of Zen et al., 1983). A younger suite of brittle fractures oriented generally to the north-northwest is responsible for the cross grain of the trellis drainage pattern that developed during prolonged denudation. The more resistant rocks of the leading edge of the Avalon Terrane (largely NeoProterozoic and originally of volcanic origin; Zv of Zen et al., 1983) create the generally southwest-northeast drainage divide between the Sudbury-Concord River watershed to the west and north, which drains to the Merrimack River, and the Charles River watershed, which drains to Boston Harbor.

FIELD TRIP STOPS

For this trip, I focus on the geological host of the world's most famous sand sinkhole: the *proglacial* lake delta of Glacial Lake Sudbury that was impounded in a north-draining watershed by ice to the north, forcing flow over the divide into the Charles River drainage. This we will accomplish in three stops; the fourth stop provides tourist opportunities:

- Stop #1 – the spillway responsible for creating that glacial lake
- Stop #2 - the largest "kame" delta built into that lake (with a local and important digression on the subsequent partial collapse and stabilization)
- Stop #3 - the breadth of that large lake and the properties of its distal bottomset beds, with Glacial Lake Concord standing in for Glacial Lake Sudbury
- Stop #4 – Concord Center tourist sites

Given the title of this field trip, most of our time will be spent at Stop #2. And given that all sites are easily located using internet map browsers, I dispense with the familiar road log.

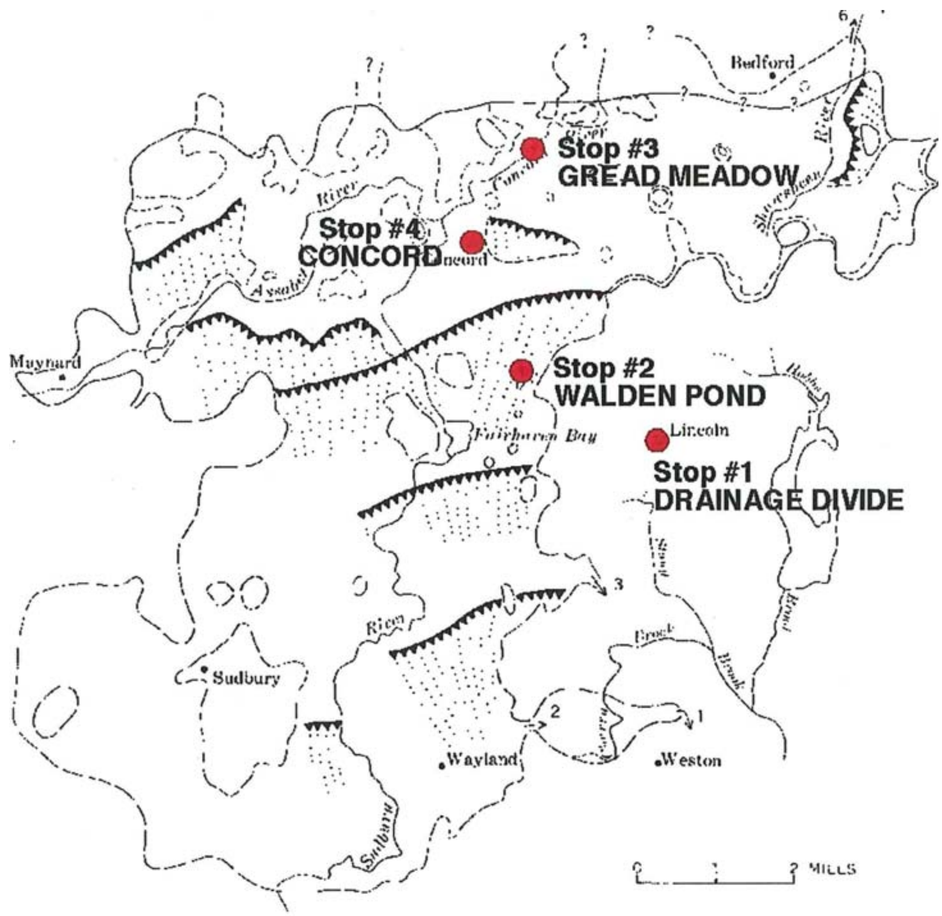


Figure 1. Map showing approximate location of field trip stops in the vicinity of Lincoln and Concord, Massachusetts. Stop #1 will traverse the terrain from Weston to north of the map, crossing the divide at Lincoln, indicated by large red dot. Base is from Koteff (1963) with the explanation removed. The relevant features are: bold lines with triangles with dotted lines, “ice-contact head and generalized flow lines of delta deposits;” long-dashed line, “shoreline of glacial lake Sudbury, Cherry Brook Stage;” and numbered arrows, “Location of glacial-lake outlet or spillway.”

Stop #1 – Reconnaissance of Drainage Divide (30 minutes drive-by, possible brief stop)

While traveling north from Weston, we will view in sequence, the:

a) Main spillway of Glacial lake Sudbury where it overflowed from north to south through the valley of Cherry Brook (Koteff, 1963);
b) Headwaters of the Charles River on resistant Avalonian rocks within the watersheds of Stony Brook and Hobbs Brook;
c) Crest of that drainage divide near historic Lincoln town center; and
d) Bloody Bluff historic site within Minute Man National Historic Park in Lexington, where a skirmish between British redcoats and colonial militia took place, and which made it the namesake for the fault zone.

Stop #2 – Walden Pond State Historic Preserve (120 minutes)

After parking on the extensive surface of the kame delta (the top of its topset facies, paved over), we will:

a) Descend into the kettle, examining its basic form, basic hydrologic setting and the anomalously clean water for such a densely settled region;
b) Climb to the highest point of the ice-contact face of the delta at Heywood's Peak to see the steep collapse slope stabilized near the angle of repose;
c) Pause while crossing a footbridge separating Walden Pond from Wyman's Meadow to visualize how the larger lake resulted from coalesced kettles;
d) Visit Thoreau's house site and the adjacent cairn of stones; and
e) Take an optional 30-minute walk around the pond or return to the headquarters area.

Cairn of stones (mostly locally derived) at Thoreau's house site, looking south to the Pond though the excessively drained sand of the Glacial Lake Sudbury delta and is therefore dominated by pitch pine.

Hot summer day at public swimming beach at Walden Pond in July 2010. Homo sapiens are the dominant geologic agent at Walden Pond today, making this scene appropriate for the Anthropocene.

Stop #3 – Great Meadows National Wildlife Refuge (60 minutes)

During the colonial era, these were lush grassy meadows, which gave the river its native name, "Musketaquid," which means "Grass-ground River or River of Meadows" (Donahue, 2004). The fertile, extensive meadows that nucleated more than a millennia of continuous occupation of natives during the Woodland Period, and which drew English settlers in 1635, reflect the fact that practically all meadows in the regions, and certainly all the large ones, lie on the poorly drained bottoms of glacial lakes dammed by a north-receding ice front, later tilted southward by a dimensionless gradient of 0.00085 (Koteff et al., 1993). We will:

a) Overlook the extent of Glacial Lake Concord
b) Walk across the historic meadow
c) Examine the channel geomorphology of the sluggish Concord River on the far side.

(*Note: Busses are not allowed on the neighborhood access road to the refuge headquarters, requiring that we walk ten minutes each way.)*

Stop #4 – Concord Center (30 minutes)

Time permitting, we will offload participants at Monument Square in Concord village. Within the distance of a city block are the: First Parish Unitarian Church where Thoreau could not prevent himself from attending during his baptism and funeral; the old Mill Dam, where colonists converted an Indian fishing weir into their first source of hydropower and water supply; the site of the old jail where Henry ruminated on civil disobedience; the Concord Lyceum for public speeches and lectures; the old burying ground; and Main Street. Within a mile are dozens of other tourist sites for another day.

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ANCIENT LAKE AND RIVER ENVIRONMENTS: A CORE AND FIELD WORKSHOP ON THE JURASSIC PORTLAND FORMATION

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INTRODUCTION

The lower Jurassic Portland Formation, was originally designated as the *Portland Arkose* for the rocks exposed in the brownstone quarries near Portland, CT (Krynine, 1950). The unit was formally changed to the *Portland Formation* (Leo et al., 1977) in recognition of the variety of lithologies that comprise the formation. The formation is best known as the source of the "brownstone" building stone that was quarried commercially from the mid-1600's until today (Guinness, 2003), and as the host of some of Connecticut's best dinosaur fossil discoveries (McDonald, 1992; Guinness, 2003). However, despite over 200 years of Hartford Basin study (see historical background provided in Lorenz, 1988), the Portland Formation remains an interesting research topic. Since the comprehensive review of the Hartford Basin strata by Krynine (1950), the Portland Formation has been addressed in a handful of manuscripts (for example, Froelich and Olsen, 1984; Lorenz, 1988; Smoot, 1991; Hubert et al., 1992; LeTourneau and Olsen, 2003; Wolela and Gierlowski-Kordesch, 2007), local field guidebooks (for example, Hubert et al., 1978, 1982; Horne et al., 1993; Gierlowski-Kordesch and Huber, 1995; Olsen et al., 2005; LeTourneau and Thomas, 2010), and several Master's theses (LeTourneau, 1985; McInerney, 1993; Zerezghi, 2007).

The Portland Formation is lower Jurassic (Hettangian – Pleinsbachian) in age (Cornet and Traverse, 1975; Horne et al., 1993; Olsen et al., 2002). It is estimated by most researchers to exceed 2 km in thickness (for example, Krynine, 1950; Horne et al., 1993), and may exceed 4 km (Olsen et al., 2005). It overlies the Hampden Basalt and is the youngest unit preserved in the Hartford Basin (Figures 1 and 2). The Portland Formation extends the length of the Hartford Basin, from near New Haven, CT in the south, to near Amherst, MA in the north (Figure 2). It is composed of strata that represent deposition in a variety of continental environments. The lower portion of the formation is poorly exposed in outcrops, but where it does occur, it is composed of alternating reddish-brown mudstone and sandstone deposits representing playa and alluvial plain environments, and dark gray to black mudstone representing a perennial lake environment. A nearly complete section of the lower Portland Formation is preserved in a series of 31 cores that were taken by the Army Corp of Engineers in preparation for the drilling of the Park River Tunnel in Hartford, CT in the 1970's. These cores show a cyclicity of playa and lacustrine facies (c. 10 lake cycles) in the lowest 600 meters of section.

The upper part of the Portland Formation is also poorly exposed, but there are concentrations of outcrops in various locations within the basin (for example, near Manchester, Portland, and Durham, CT, and Wilbraham, MA) associated with relay zones along the Eastern Border fault system. At these locations, the formation is composed primarily of reddish-brown sandstone and conglomerate interpreted to represent braided river or alluvial fan deposits, with thin floodplain mudstone layers.

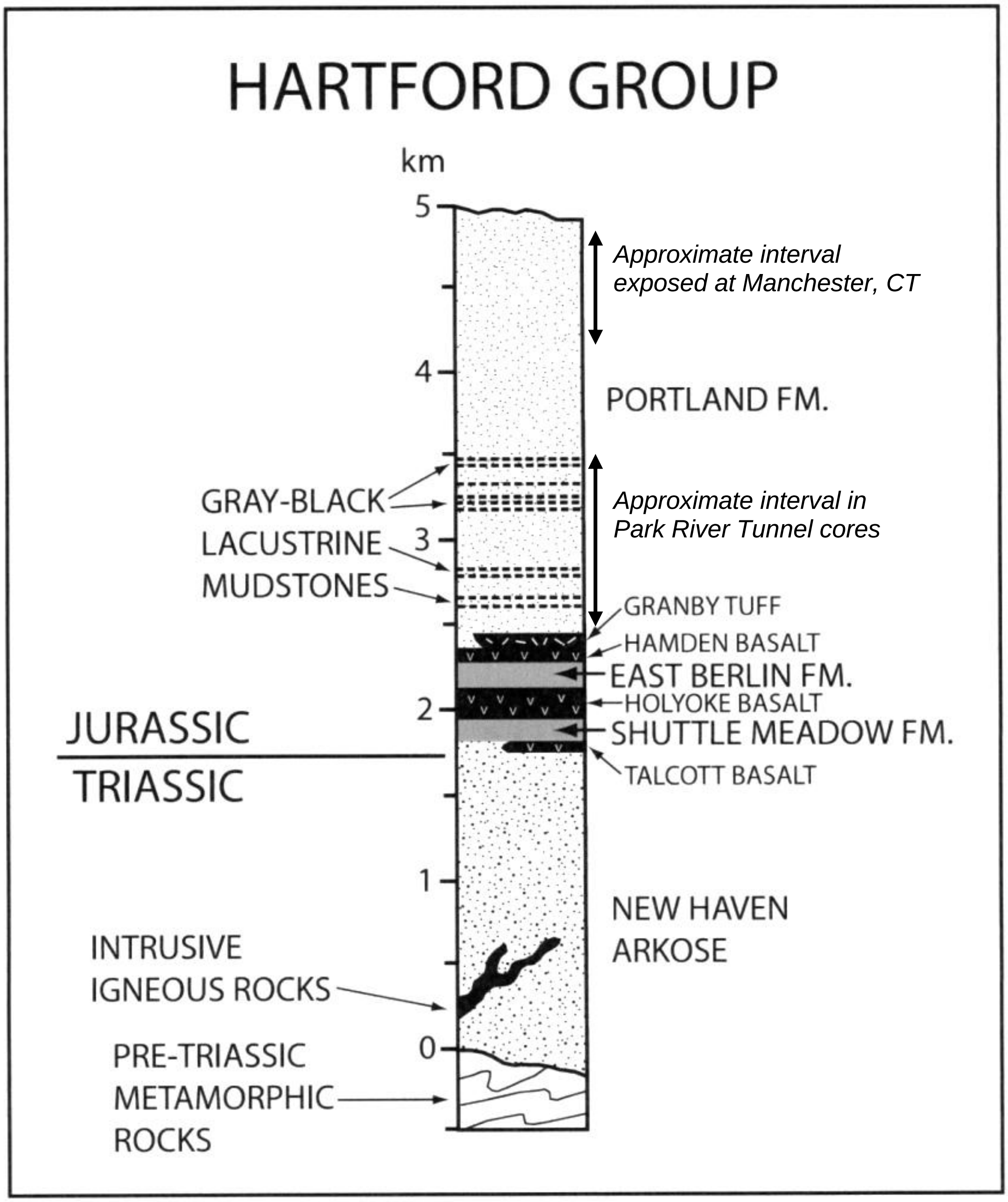

Figure 1. Generalized stratigraphic column of the formations within the Hartford Basin. The cores are taken from the lower, lacustrine-dominated portion of the Portland Formation, and the outcrops are in the upper, fluvial-dominated portion. Modified from: Hubert, Feshbach-Meriney, and Smith, 1992.

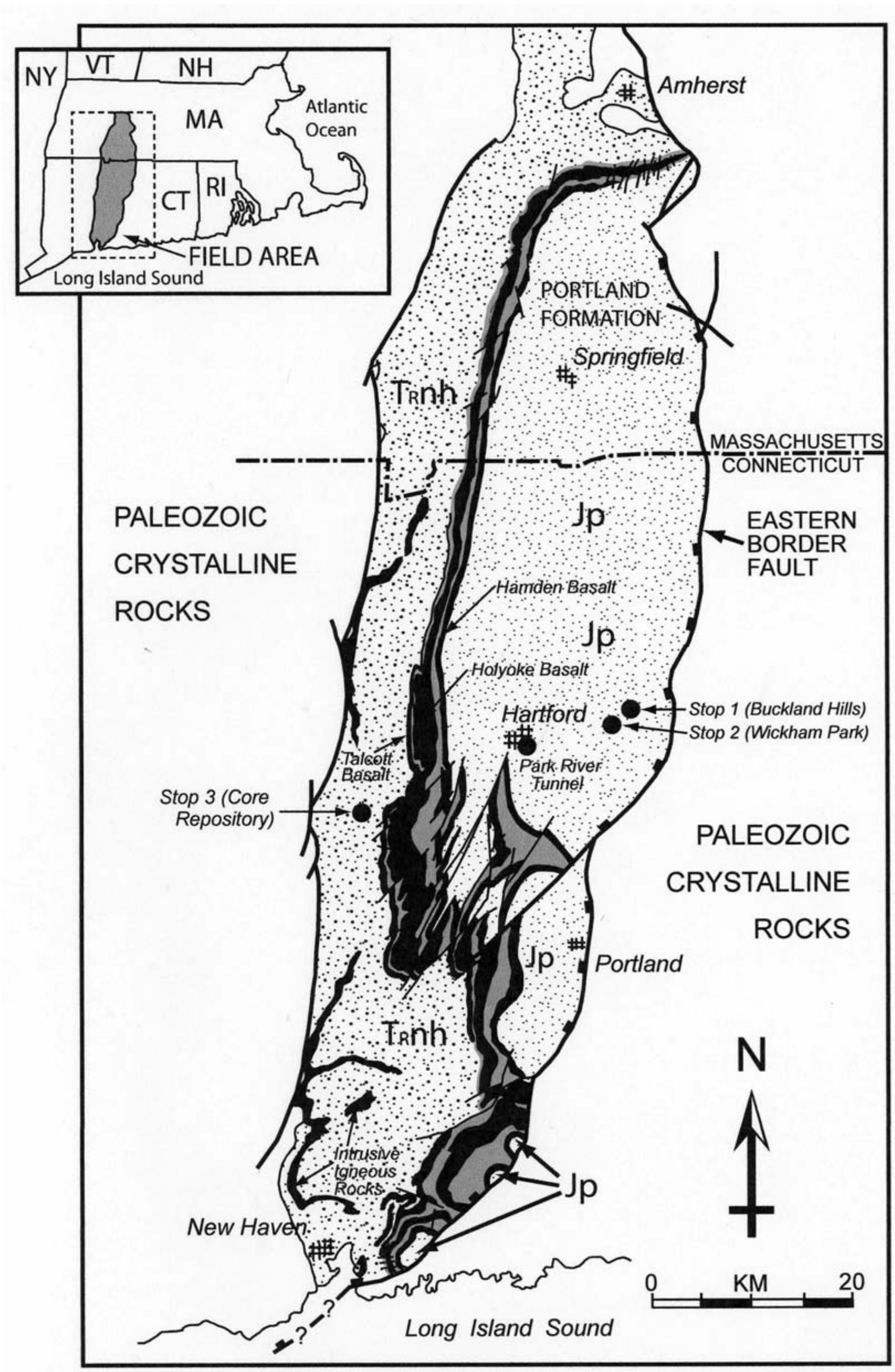

Figure 2. Geological map of the Hartford Basin showing the distribution of formations and the location of the field trip stops as well as the location from which the Park River Tunnel cores were taken. See Figure 1 for a key to formations. TRnh = Triassic New Haven Arkose; Jp = Jurassic Portland Formation. Modified from: Hubert, Feshbach-Meriney, and Smith, 1992.

This workshop will provide a basic description of the facies that comprise both the lower and upper parts of the Portland Formation, as well as an understanding of their paleoenvironmental interpretation. Climatic and tectonic controls on the deposition of these facies will be discussed. The trip will begin with an examination of braided river deposits in the upper Portland Formation at two outcrop locations in Manchester, CT. It will conclude with an examination of lower Portland Formation lake cycles in several of the Park River Tunnel cores currently housed at the Connecticut Geological Survey's Core Repository in Farmington, CT.

GEOLOGIC SETTING

The Portland Formation is the youngest stratigraphic unit preserved within the Hartford Basin. This basin (Figure 2) developed during early extension in the Triassic and Jurassic Periods that eventually formed the North Atlantic Ocean (Manspeizer, 1988). The rifting created a topographic low which was filled with non-marine sediments and basaltic lava flows. The basin is an asymmetric half-graben that is bordered on its eastern margin by a set of faults which are collectively referred to as the Eastern Border Fault (Schlische and Olsen, 1990; Schlische, 2003). Strata within the basin generally dip to the east by about 10°-20° as a result of syndepositional tilting. A number of anticlines and synclines (fault-displacement and fault-deflection folds; Schlische, 2003) are located along the eastern margin of the basin. Normal faults, that are both synthetic and antithetic to the border faults, dissect the strata.

The overall stratigraphy of the Hartford basin is described in Hubert *et al*,1978, 1982, 1992; Froelich and Olsen, 1984; Olsen et al, 1989; Gielowski-Kordsch and Huber, 1995. The strata of the Hartford basin (Figure 1) are divided into four sedimentary formations separated by three lava flows. The oldest formation, the New Haven Arkose, was deposited during the late Triassic to earliest Jurassic Periods and consists of fluvial red beds, alluvial fan conglomerate, and floodplain mudstone containing paleosols. It is bounded above by the Talcott Basalt. The Shuttle Meadow Formation overlies the Talcott and is early Jurassic in age. It is composed of red mudstone and siltstone, interpreted as a playa environment, and gray to black shale with thin carbonate layers representing a perennial lake environment. The Shuttle Meadow is bounded above by the Holyoke Basalt. The East Berlin Formation is located between the Holyoke Basalt (below) and Hampden Basalt (above). It was deposited during the early Jurassic period and is composed of red playa mudstone and fine sandstone, gray and black lacustrine mudstone, and minor fluvial and alluvial sandstone and conglomerate. Overlying the Hampden Basalt is the Portland Formation which was deposited during the early to middle Jurassic Period. The lower portion of the Portland Formation is composed primarily of gray lacustrine strata and red playa mudstone. The upper portion of the Portland contains primarily reddish-brown braided river and alluvial fan sandstone and conglomerate.

Generally, the source of the sediment for the Portland Formation was the highlands that were located to the east of the basin (Wintsch et al., 1992; Kunk et al., 2001; Burton et al., 2005) and the deposits are coarsest along the Eastern Border Fault. They are interpreted as the deposits of alluvial fans and braided rivers along the eastern margin of the basin. In the center of the basin, the lower Portland is exposed. It is composed of lacustrine, playa, and alluvial plain mudstone to fine sandstone. Sediment contribution from the west is difficult to assess, because tilting and erosion has removed all of the Portland Formation that originally occurred in proximity to the western margin of the basin. However, work by Wintsch and Bleveus-Walker has established an eastern source for all basin sediments sampled (Bleveus-Walker 2008, Wintsch et al 2012, this volume).

These strata display the tripartite sediment packaging (Figure 1) common in rift basins that is interpreted to reflect the increase and then decrease in accommodation growth within the basin (Lambaise 1990; Schlische and Olsen, 1990; Gawthorpe and Leeder 2000). When rifting is initiated, accommodation is low and coarse fluvial sediment easily progrades out into the basin. In the Hartford Basin, this is represented by the New Haven Arkose. As subsidence in the rift basin increases, coarse sediment is trapped at the edges of the basin, and only fine sediment is deposited within the basin center. This is represented by the lacustrine strata of the Shuttle Meadow, East Berlin, and lower Portland Formations in the Hartford Basin. Volcanic rocks associated with these strata further suggest maximum crustal thinning and rift-related tectonic subsidence at this time. Finally, as rifting subsides, and basin subsidence decreases, accommodation growth will also decrease and coarse, fluvial sediment can once again prograde into the basin. In the Hartford Basin, this is represented by the upper Portland Formation. This workshop will examine the strata in the Portland Formation that record the waning of tectonic subsidence in the Hartford Basin, and the associated shift from lacustrine to fluvial sedimentation.

The workshop starts at the Hartford Marriott Downtown, and begins with an outcrop component in Manchester, CT. Vans will provide transportation.

FIELD TRIP STOPS

Stop 1. Braided River Deposits (Upper Portland Formation) at the Eastern Edge of the Hartford Basin, Intersection of Buckland Rd. and Buckland Hills Rd., Manchester, Connecticut

Road Log from the Hartford Marriott Downtown to Stop 1, Home Depot at the Buckland Mall (Manchester, CT)

Driving Directions	Mileage for Step	Cumulative Mileage
1. Start out from the Hartford Marriott Downtown going north on Columbus Boulevard toward Grove Street.	0.1	0.1
2. Turn right onto State Street.	0.1	0.2
3. State Street becomes the Founders Bridge.	0.2	0.4
4. The Founders Bridge becomes CT-2 E.	0.2	0.6
5. Merge onto I-84 E via EXIT 2 toward E Hartford/Boston.	5.7	6.3
6. Take the Buckland Street exit, EXIT 62.	1.1	7.4
7. Turn left onto Buckland Street.	0.4	7.8
8. Turn right onto Buckland Hills Drive. You will drive past the outcrop we will be examining on the right (south) side of the road.	0.2	8.0
9. Take the first 1st right into the Home Depot parking lot. We will park here and walk down to the outcrop that we just passed on Buckland Hills Drive.	Total Distance = 8.0 miles	

Setting:

A series of outcrops were exposed in western Manchester (CT) as a result of the construction of retail businesses (primarily the Buckland Mall) and the I-84 interstate. The outcrops occur just 4.5 km west of the Eastern Border Fault, and the edge of the Hartford Basin. Only a few small outcrops are known to occur along I-84 between these outcrops and the Eastern Border Fault, and they are comprised of similar facies. Thus, the outcrops in the western Manchester region represent facies closest in proximity to the source area for the sediment entering this region of the basin. They are also some of the youngest sediments in the basin, but the paucity of fossils within these strata makes an accurate assessment of the age difficult. Stone has been quarried for centuries from several locations within the immediate vicinity, and some of these quarries (including a former quarry at the location of our field stop) have produced fossil remains of prosauropod dinosaurs from the Portland Formation (Hanrahan, 2004).

Description:

The outcrop along Buckland Hills Drive is part of the upper Portland Formation. For logistic reasons, we will examine the outcrop on the southern side of the road, where 33 meters have been measured. About 55 meters of section are exposed on the northern side of the road, but accessibility issues and vegetation make this side harder to examine. Strata at this outcrop trend N25°E on average, and they dip approximately 15°SE. The rocks at this outcrop are primarily coarse-grained in nature, and are comprised of three main lithologies:

1. *Poorly-sorted, coarse reddish-brown arkose* (Figure 3): This is the most abundant facies within this outcrop. Planar bedding and low-angle trough cross-bedding are common, and high-angle trough cross-bedding occurs in some beds. However, the bedding is only obvious in low-angle afternoon sunlight, and the arkose more typically appears to lack internal stratification. Conglomerate is common within this lithofacies, and typically occurs in thin lenses at the base of channel-fill sequences. Some isolated quartz clasts reach cobble size (Figure 3). Quartz, feldspar, mica, and lithic fragments are the main constituents of the sand-sized grains. Gneiss, quartz/quartzite, phyllite, and schist make up the gravel clasts. This facies is interpreted to represent the channel-fill (bars) of a braided river system.

2. *Polymictic conglomerate* (Figure 4): Lens-shaped deposits of conglomerate with a coarse sand matrix are common. In fact, this facies and the arkose discussed above form a continuum of coarse lithologies that vary based on the amount of sandstone vs. conglomerate. These two facies can be distinct in the field, but are commonly intercalated. Sedimentary features are generally absent, but planar bedding and low-angle trough cross-bedding are locally preserved. Imbricated clasts are rare. More commonly, clasts are aligned with their long axis parallel to bedding. The composition of the sand- and gravel-sized clasts is similar to those described in the arkose above. Like the arkose facies, the conglomerate is interpreted to be channel fill in a braided river system, but it likely represents the thalweg lags at the base of channels.

3. *Dark reddish-brown micaceous mudstone* (Figure 4): Thin intervals (less than 20 cm thick) of reddish-brown siltstone and shale occur throughout the outcrop. Usually these weather recessively and are highly friable. The highly weathered nature of the mudstone limits the ability to recognize sedimentary structures, but burrows were obvious when the outcrop was new (McInerney, 1993). Near the top of the outcrop, continental trace fossils (mostly *Scoyenia* burrows) have been observed. Mudcracks occur on the underside of sandstone beds that overlie the mudstones. Mudstone layers are both continuous across the outcrop, and occur in thin, discontinuous lenses. This facies is interpreted to have been deposited by floods at some distance from the braided river axis.

An analysis of 23 paleocurrent indicators, mainly from trough cross-beds preserved within channel fill, reveal a NW (average azimuthal direction of 349°) flow direction (McInerney, 1993).

Figure 5 is a stratigraphic column of the outcrop on the southern side of the road, and Figures 6 and 7 are details of parts of the outcrop that illustrate the channel architecture preserved within these deposits. A hierarchy of channel elements can be recognized based on grain size and the distribution of facies. Individual channel-fill sequences represent the smallest of these elements. They range from less than 0.5 meters to about 2 meters in thickness and are typically composed of conglomerate and sandstone that fine upward. They are bounded below by a surface that records an abrupt grain size increase; many show evidence of erosion. In some places the channel-fill is capped by a thin, discontinuous mudstone layer that represents a time of channel abandonment and deposition of fine sediment as currents waned. These channel-fill elements amalgamate both laterally and vertically to form larger-scale channel belts (Figures 5 and 6). The channel belts range between 1.5 and 11 meters in thickness, and in general either fine-upwards, or they first coarsen, then fine-upwards. Although channel-fill elements thicken and thin irregularly within a channel belt unit, the channel belts maintain a relatively uniform thickness. Each channel belt is bounded below by an erosional surface. Thick (typically >5cm), continuous mudstone layers cap these channel belts. Finally, the entire outcrop is composed of a stack of channel belts that preserve an overall fining upward trend, representing the largest-scale order of cyclicity.

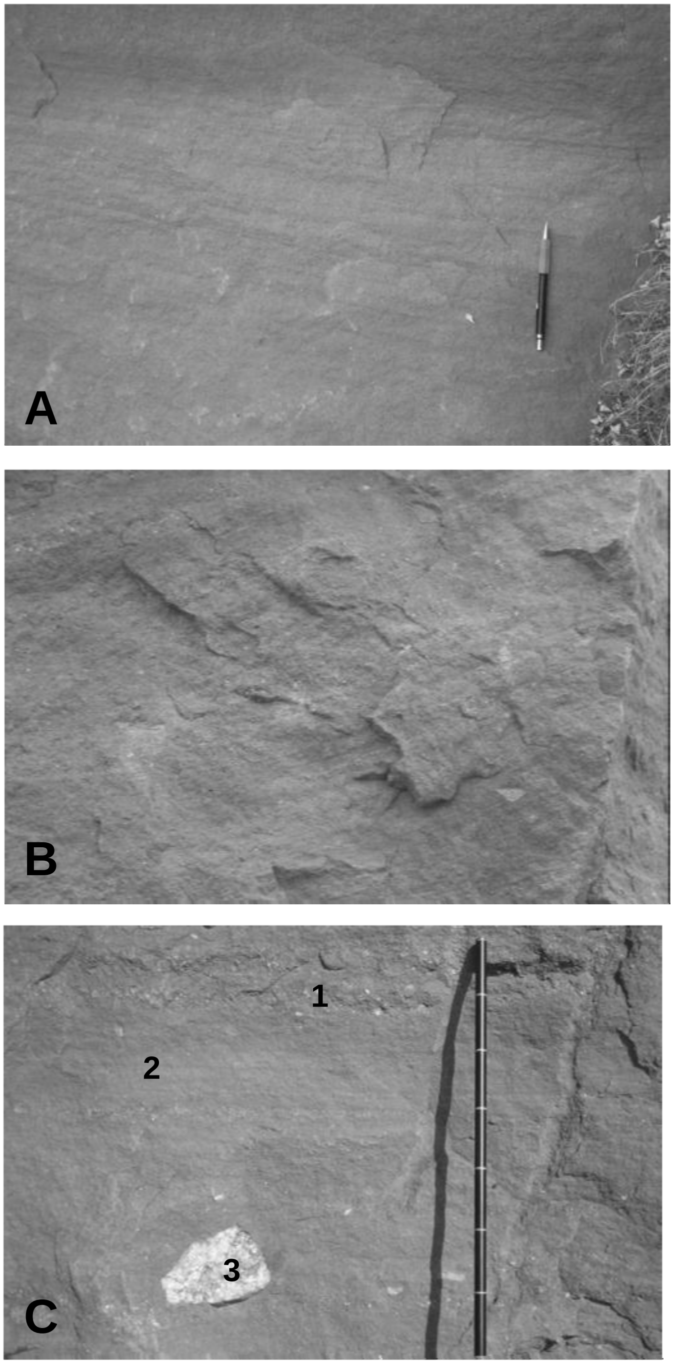

Figure 3. Outcrop photographs of the poorly-sorted, coarse reddish-brown arkose.
A. Typical expressions of this facies show planar or low angle cross-beds (note pencil for scale). Often, internal laminae are obscure within this facies.
B. Reddish-brown sandstone with high-angle trough cross-beds (field of view is about 2m wide).
C. Reddish-brown sandstone lithology with thin conglomerate layers (1), planar bedding (2), and a large, isolated quartzite clast (3). These clasts are scattered about the outcrop, particularly near the top of the section. The Jacob staff is divided into 10 cm intervals.

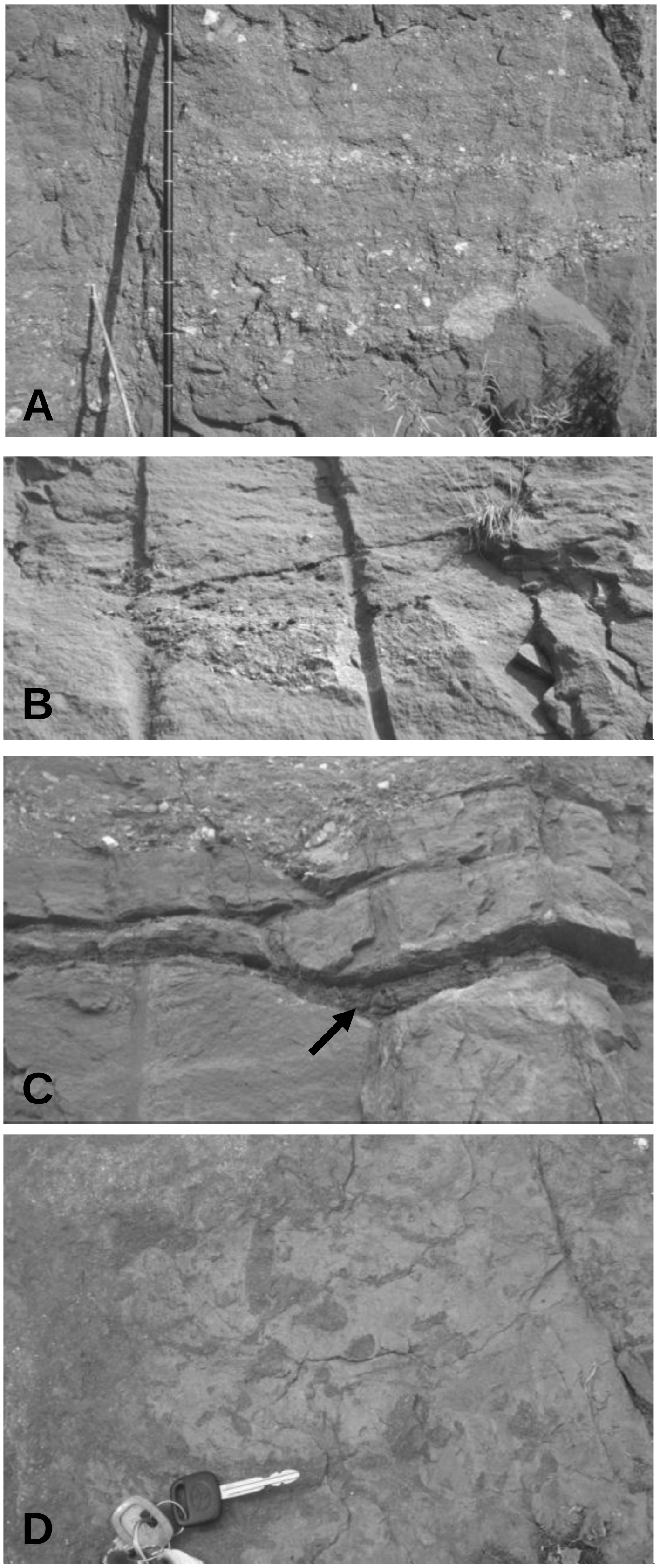

Figure 4. Outcrop photographs of the polymictic conglomerate and the dark reddish-brown micaceous mudstone lithiofacies.
A. Layers of conglomerate with current aligned and some imbricated clasts (the Jacob staff is divided into 10 cm intervals).
B. Lens-shaped deposit of conglomerate representing fill of a small channel (lens is about 2 meters across). The conglomerate layers typically have sharp bases with channel-shaped cross-sectional morphologies.
C. Thin mudstone layers between thicker sandstone and conglomerate beds that are interpreted to mark the boundaries of channel belts within the upper Portland Formation (shale layer at the arrow is about 10 cm thick).
D. Plan view of continental trace fossils (*Scoyenia*) within micaceous siltstone (note keys for scale).

Interpretation:

The facies in these outcrops are interpreted to represent braided river deposits (McInerney, 1993), based on the low abundance of overbank mudstone, the coarse grain size, and the high degree of channel amalgamation. Paleocurrent flows to the NW have been used to suggest these streams formed the distal part of larger alluvial fan system that was distributing sand to the NW at this location, before the rivers entered the valley floor and flowed south along the axis of the basin. The arkose and conglomerate represent the preserved remains of channel-fill sediments, including bar forms as well as the coarser thalweg lags at the bases of channels. The mudstone facies represents periods of waning flow that occur either when the river shifts the channel in which water and sediment is moving or when it avulsed to an entirely new location.

Individual channel fill sequences are thin and laterally continuous, implying the water was flowing in shallow, broad channels. This interpretation is supported by the abundance of planar bedding which is typical of fast, but shallow channels. Large isolated cobbles (Figure 3) indicate that the competence of the river is higher than it appears, supporting the notion of a swift-moving stream. Sands preserved within these channels were deposited in bars that migrated over time. Channels amalgamate into channel belts through the normal processes of channel and bar migration that occur in braided rivers. The relatively constant thickness of individual channel belts indicates that accommodation played a fundamental controlling role in the stratal architecture. Channels migrated laterally and aggraded vertically until they filled available accommodation. The fining nature of the tops of channel belts reflect reduced flow through a channel system that has filled its available accommodation and avulses to another location where accommodation is greater. Thus, the channel belts can be interpreted to represent "accommodation cycles". Finally, the entire outcrop could represent a larger, fining upward stacked channel belt complex that is likely responding to allogenic (base level, tectonic, and/or climatic) controls. The lack of strata above that exposed in the Buckland Hills roadcut indicates that perhaps the fluvial system has migrated from this location, and only finer grained alluvial plain sediments (now covered) were deposited.

Cycles similar to those recorded above have been identified by others (McLaurin and Steel, 2007; LeLeu et al., 2009, 2010). LeLeu et al. (2009, 2010) describe the cyclicity within Triassic rift strata (Wolfville Formation) from the Fundy Basin in Nova Scotia, Canada. Like the Hartford Basin, the Fundy Basin is a failed rift associated with the opening of the Atlantic Ocean. The Wolfville Formation is composed primarily of planar cross-stratified sandstone and granule-rich sandstone, with minor amounts of mudstone, and is interpreted to represent bar elements in a braided river system, and the associated floodplain deposits. Several orders of cyclicity have been reported in these strata (Leleu et al., 2009, 2010). The channel bar elements represent the smallest scale of cyclicity. These smaller scale surfaces and bedforms are interpreted to reflect the autocyclic migration of channels within the braided river system.

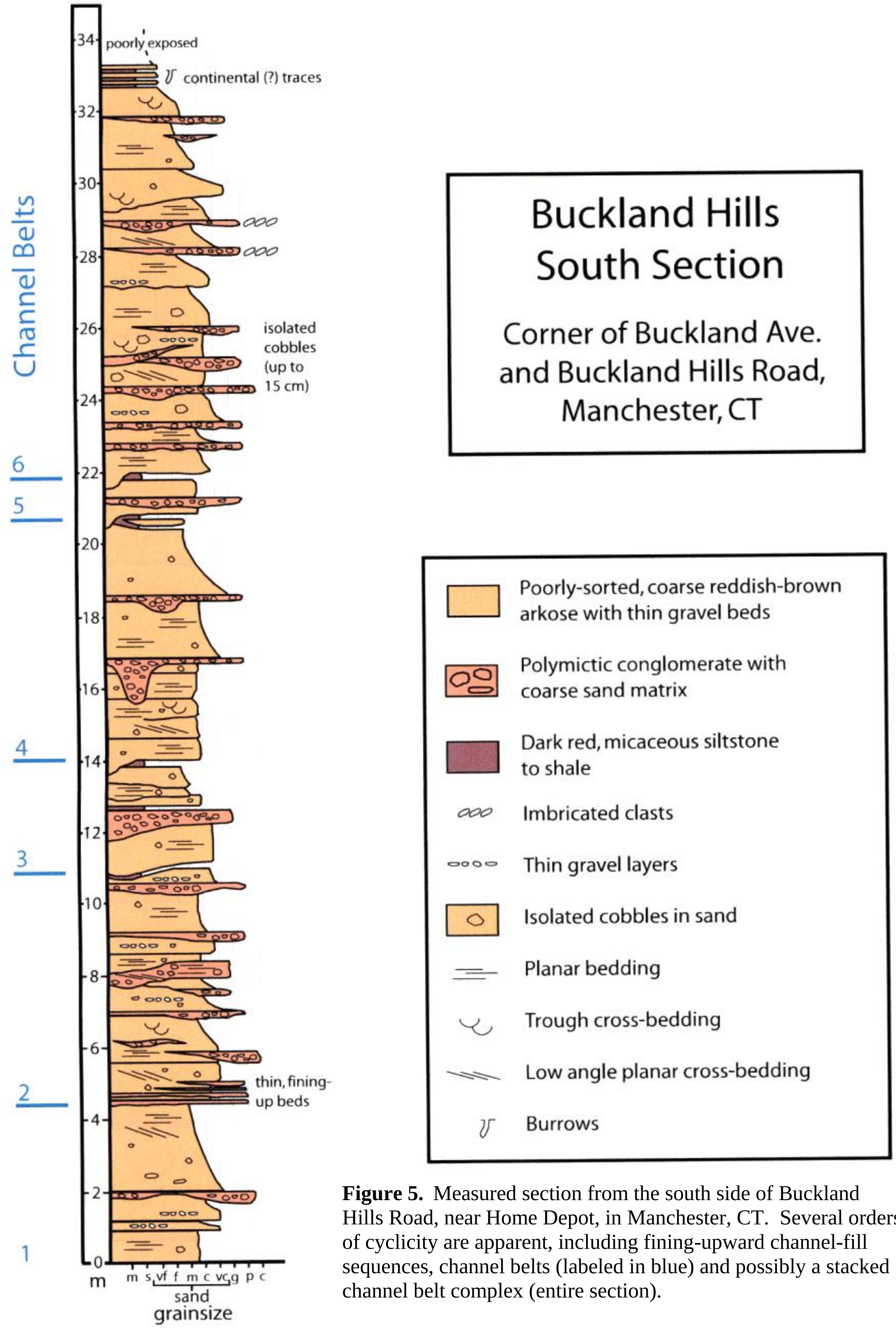

Figure 5. Measured section from the south side of Buckland Hills Road, near Home Depot, in Manchester, CT. Several orders of cyclicity are apparent, including fining-upward channel-fill sequences, channel belts (labeled in blue) and possibly a stacked channel belt complex (entire section).

Figure 6. Outcrop photographs and interpreted drawing of the western end of the outcrop (south side) on Buckland Hills Road). The thicker lines are the boundaries between channel belts, which are defined by continuous mudstone layers. The numbers refer to the tops of the channel belt cycles. The red-colored intervals near the top of the section on the interpretation are thicker (> 5 cm) mudstone layers.

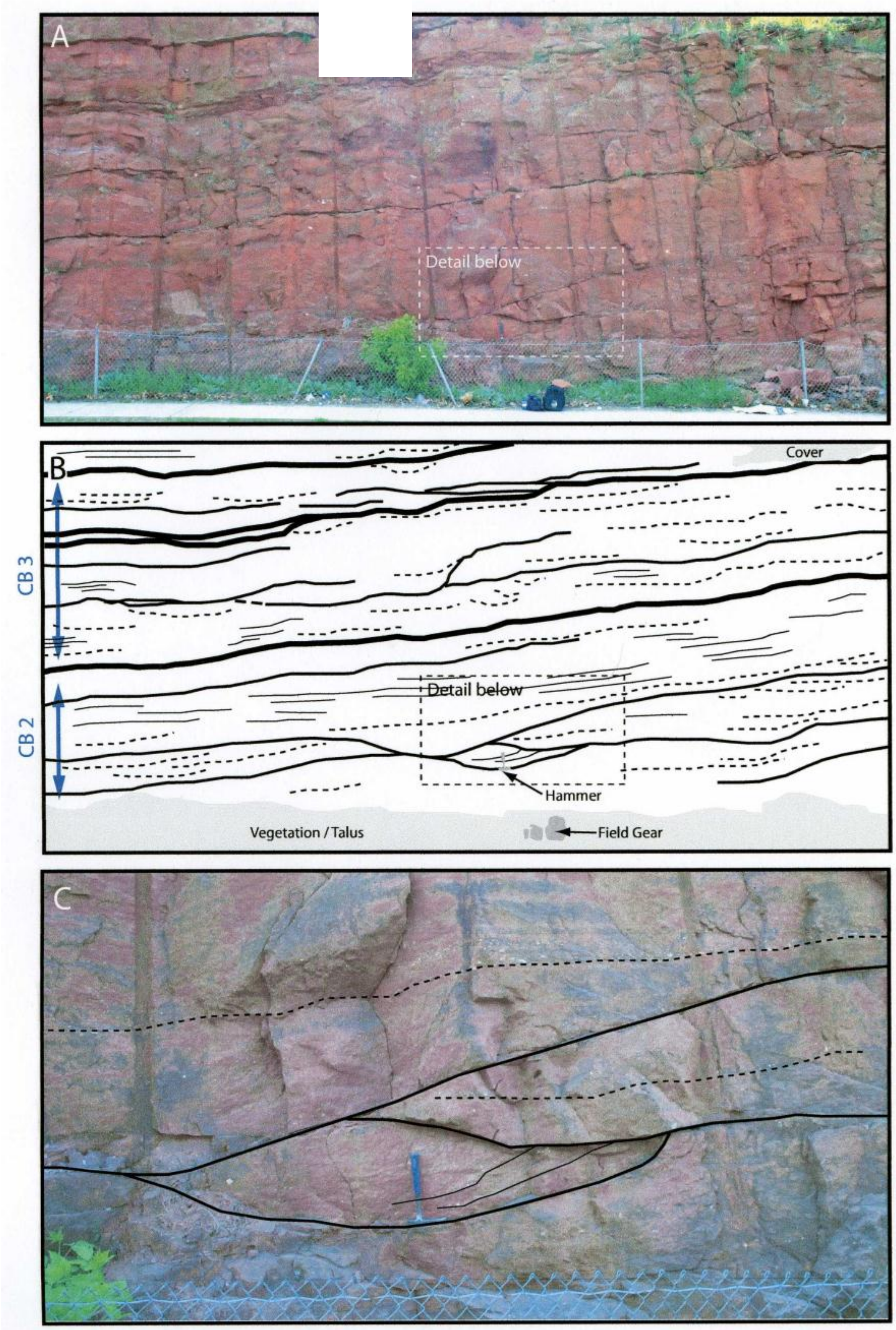

Figure 7. Photographs and interpretations of details of the outcrop at Buckland Hills Road. This shows the bedding styles associated with channels and channel belts. Note field gear (backpack- top photos) and hammer (bottom photo) for scale.

The individual channel-bar systems amalgamate into larger channel belts that represent multistory channel-bar systems created by lateral migration in times of low accommodation growth. The cycles are bounded on the base by a major erosional surface, and fine-upward through a series of channel-bar elements and eventually into finer-grained floodplain deposits. These larger scale bounding surfaces and channel-fill patterns are interpreted to result from an allocyclic climate control influencing the capacity of the rivers to transport bedload (Leleu et al. 2010) or affecting accommodation (McLaurin and Steel, 2007). The fining-upward nature of the channel belts reflects a decrease in generation of accommodation.

These same processes may be responsible for cycle development in the Portland Formation. The smallest scale channel-fill elements are likely caused by migrating channels and bars within a braided river system. The larger channel belts likely represent autogenic avulsions within the overall fluvial or alluvial fan environment, and the outcrop-scale fining upward trend may be controlled by allogenic factors such as climate, tectonics, or changes in base level.

There are a number of interesting questions that can be associated with the strata preserved in this outcrop. These include:

1. What is the overall depositional environment? The current direction data can be best explained if these sediments were deposited on the distal margin on the north side of an alluvial fan. Is there any sedimentologic evidence to support this?

2. What is the significance of the lack of trough cross-beds and the abundance of planar beds (or low angle cross-beds) in regards to understanding how sediment was transported within the braided river system? Upper plane beds typically develop in very swift and/or very shallow currents. At what velocity were these rivers flowing? How deep were the channels?

3. What is the relevance of the larger clasts (Figure 3c) preserved within the sandstones? What do they tell us about bedload transportation in the river? What do they tell us about the competence and capacity of the river?

4. Are the fluvial cycles observed within the upper Portland Formation the result of allogenic forcing, or can they be explained in terms of natural autogenic processes (such as avulsions, or channel migrations)? What does this tell us about the controls on fluvial architecture within these braided systems?

Stop 2. Braided River Deposits (Upper Portland Formation) near the Eastern Edge of the Hartford Basin, Ramp from I-84 onto US-44 (Manchester, CT)

Road Log from the Home Depot Parking Lot at the Buckland Hills Mall (Manchester, CT) to Wickham Park (Manchester, CT)

Driving Directions	Mileage for Step	Cumulative Mileage
1. Leave the Home Depot parking lot by turning left onto Buckland Hills Drive.	0.2	0.2
2. Buckland Hills Drive becomes Pleasant Valley Road.	0.2	0.4
3. Turn left to take the US-44/Middle Turnpike. Follow signs to US-44/Middle Turnpike. The outcrop we will be examining is located on the right side of this ramp, just before we turn onto US-44.	2.0	2.4
4. Turn right onto Middle Turnpike W/US-44.	0.2	2.6
5. Take the first right into the entrance of Wickham Park and follow the road to the first parking lot on the right.	0.1	2.7
6. Park in this lot. We will walk back to the outcrop we just passed outside Wickham Park. The outcrop is located on a Rails-to-Trails bike path, so please be considerate of bikers.	Total Distance = 2.7 miles	

Setting:

A series of outcrops exposed in westernmost Manchester (Connecticut) at the interchange between I-84 and US-44 provide a small, 3-dimensional window into the upper Portland Formation. The outcrops are about 3.2 km SW of the Buckland Hills Road outcrop (Figure 2), and are approximately 6.5 km from the Eastern Border Fault. They represent a depositional environment that is more distal than that at Stop 1. This generalization is complicated by the fact that the strata at this location are likely older than those at Buckland Hills. McInerney (1993) suggests these strata are about 1400m above the base of the Portland. Normal faults probably occur between the two outcrops, resulting in further correlation complications. Even with these uncertainties, it is likely that the environments exposed in the Wickham outcrops are more distal equivalents to those at Buckland.

Description:

The outcrop that will be examined along the bicycle path near Wickham Park is part of the upper Portland Formation, and contains many of the same lithofacies as those exposed at Buckland Hills. Although there are about 300m of lateral exposures at this location, approximately 77 meters of section have been measured. On average, the strata strike N15°E, and dip 20°SE.

The rocks at this outcrop are primarily coarse grained (similar to Buckland Hills), and are comprised of the same three main lithologies. There are some differences, though. Facies include:

1. *Poorly-sorted, coarse reddish-brown arkose* (Figure 8): This is the most abundant facies within this outcrop. Planar bedding and low-angle trough cross-bedding are common, and high-angle trough cross-bedding also occurs. Pebble and granule beds are commonly interlayered with the arkose. Some isolated pebble- and cobble-sized lithic clasts occur. Quartz, various types of feldspar, and lithic fragments (mostly granite and gneiss) are the main constituents of the sand-sized grains. This facies contains channels up to 3m thick, and is interpreted to represent the channel-fill (bars) of a braided river system.

2. *Polymictic conglomerate* (Figure 8): Like at the Buckland outcrop, lens-shaped deposits of conglomerate with a coarse sand matrix are common. Sedimentary structures include planar bedding and low-angle trough cross-bedding. Imbrication of clasts is observed. Gneiss, quartz/quartzite, pegmatite, feldspar, phyllite, and schist make up the gravel clasts. The conglomerate occurs as distinct layers or channel-shaped lenses. It can also be interlayered with the arkose facies above. It is interpreted to be channel fill in a braided river system, in particular the thalweg lags at the base of channels.

3. *Micaceous mudstone* (Figure 9): Intervals (up to 80 cm thick) of micaceous siltstone, shale, and fine sand occur throughout the outcrop. They weather recessively, and most lack distinct bedding features. McInerney (1993) described abundant root traces from the outcrop that were obvious when it was first exposed. What remains now are distinct light greenish-gray mottles (Figure 9) that formed in association with these roots. They originate in mudstone layers but may penetrate down into underlying sandstone layers. Continental trace fossils (mostly *Scoyenia* burrows) have been observed in the mudstones. Mudstone layers are both continuous across the outcrop, and occur in thin, discontinuous lenses. This facies is interpreted to represent pedogenicly altered overbank deposits associated with periods of waning flow and/or river avulsion.

An analysis of 62 paleocurrent indicators, mainly from trough cross-beds preserved within channel fill, reveal a SW (average azimuthal direction of 216°) flow direction (McInerney, 1993).

Interpretation:

The facies exposed at Wickham Park differ from those at Buckland Hills in several ways. First, channelized sand bodies are easily recognized at the Wickham outcrop (Figure 10), particularly if you step back from the outcrop. This is most likely as a result of the thicker and more abundant mudstone layers that enhance the visibility of channels through differential weathering. Second, overbank/floodplain mudstones are thicker and display more evidence of pedogenic alteration (Figure 9) than the mudstones at Buckland Hills. Finally, the central portion of the outcrop at Wickham Park contains channels with a much higher width-to-depth ratio than anything seen at Buckland Hills.

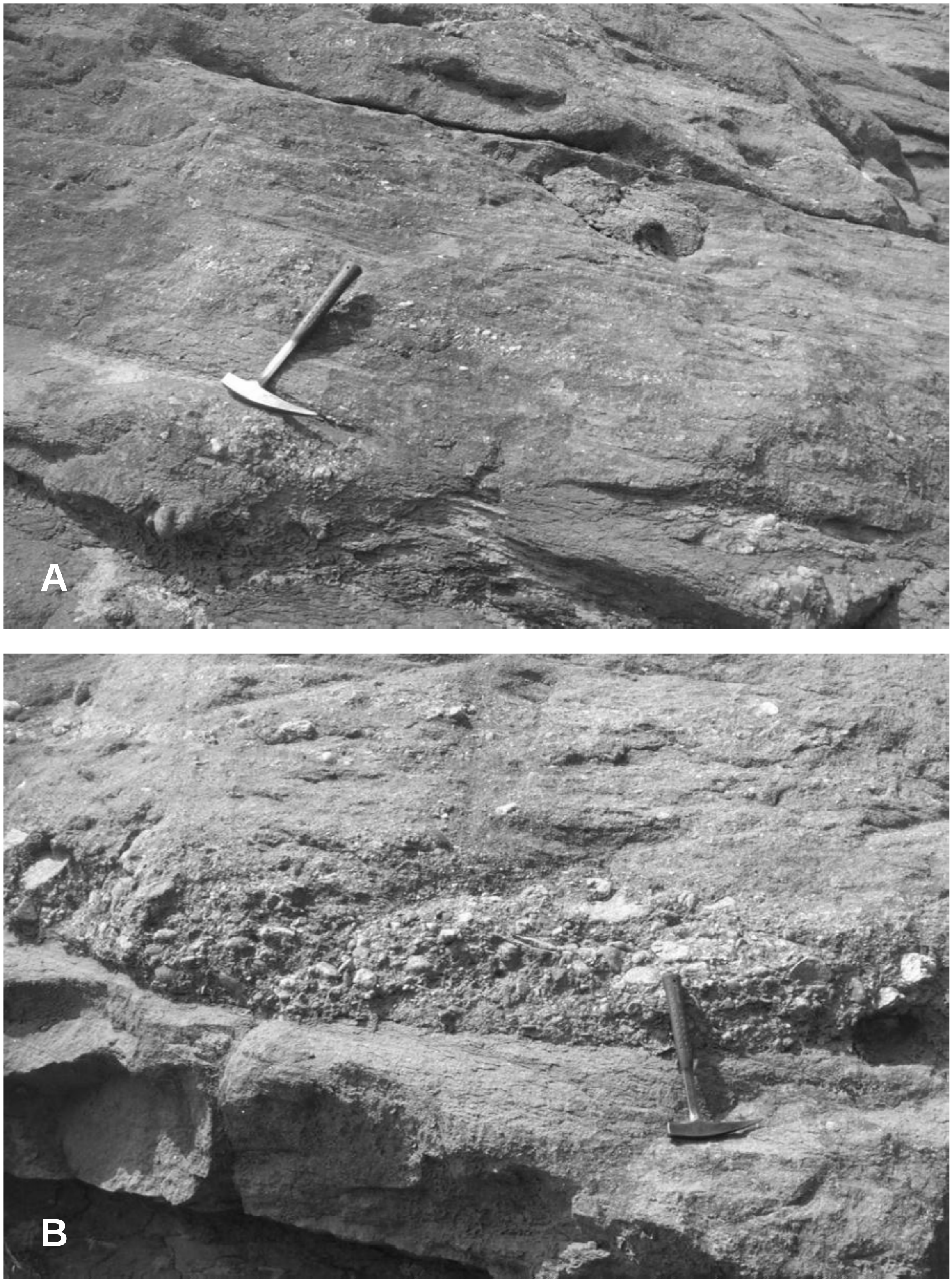

Figure 8. Outcrop photographs of the poorly-sorted, coarse reddish-brown arkose and polymictic conglomerate.
A. Planar bedded arkose within some thin conglomerate layers (note hammer for scale).
B. Conglomeratic lag at the base of a channel associated with an erosional surface (note hammer for scale).

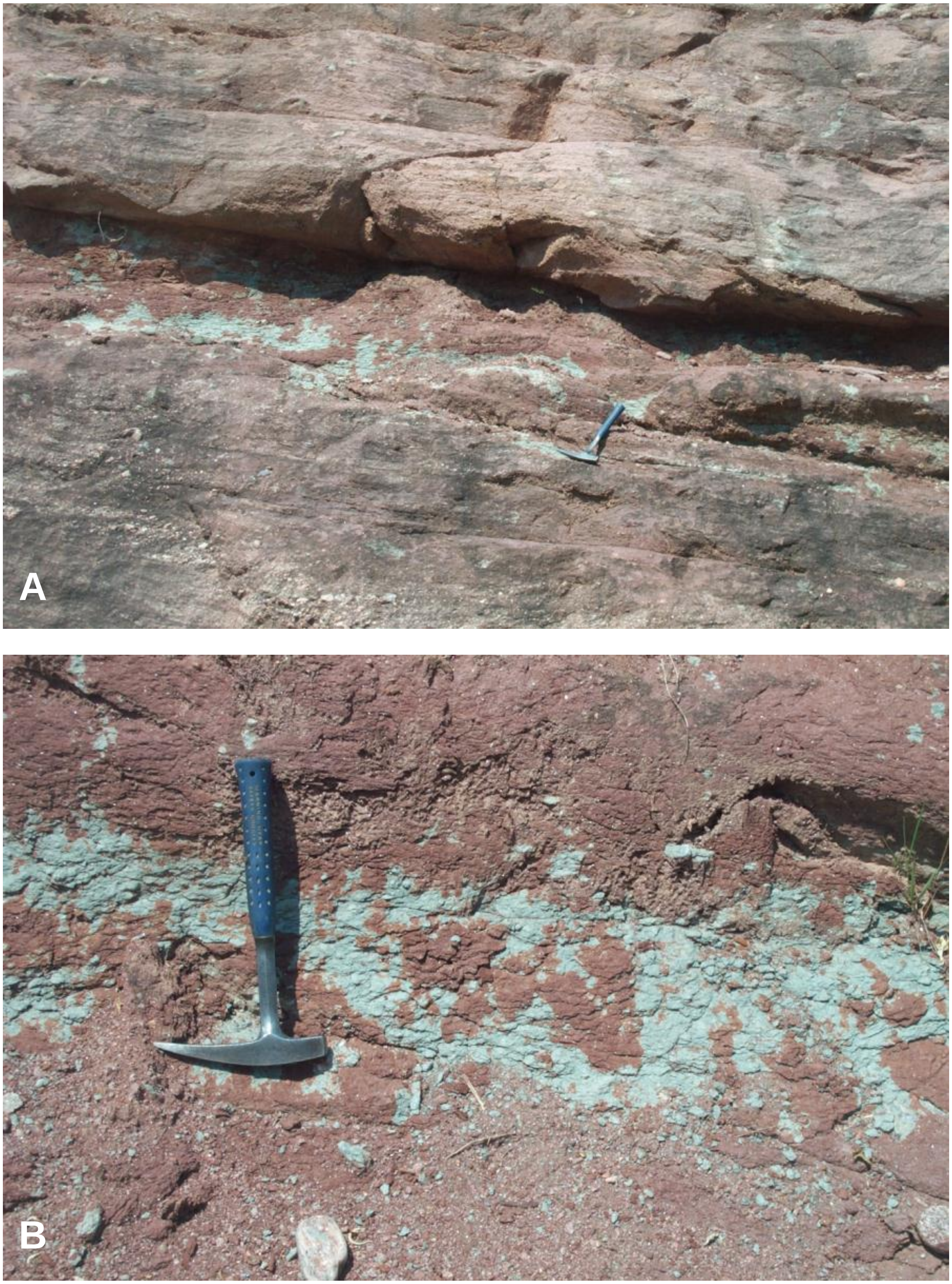

Figure 9. Outcrop photographs of the micaceous mudstone. **A.** Thick mudstone layer between thicker planar-bedded sandstone beds. Note the light greenish-gray mottling associated with the mudstone (hammer for scale). **B.** Close-up of the mudstone showing the mottled texture common in this lithofacies (note hammer for scale).

Like the Buckland Hills outcrop, the braided river strata at Wickham Park display a hierarchy of cycles. The smallest scale cycles are individual channel-fill elements. Channels have erosive bases, commonly cutting into mudstone, and are between 0.5 and about 2 meters thick (Figure 11). The channel-fill typically has conglomerate at the base and fines upward into planar or trough cross-bedded sandstone. Many channels are capped by thin (< 20cm), discontinuous mudstone layers. They are interpreted to represent autogenic processes of channel and bar migration typical of braided river systems.

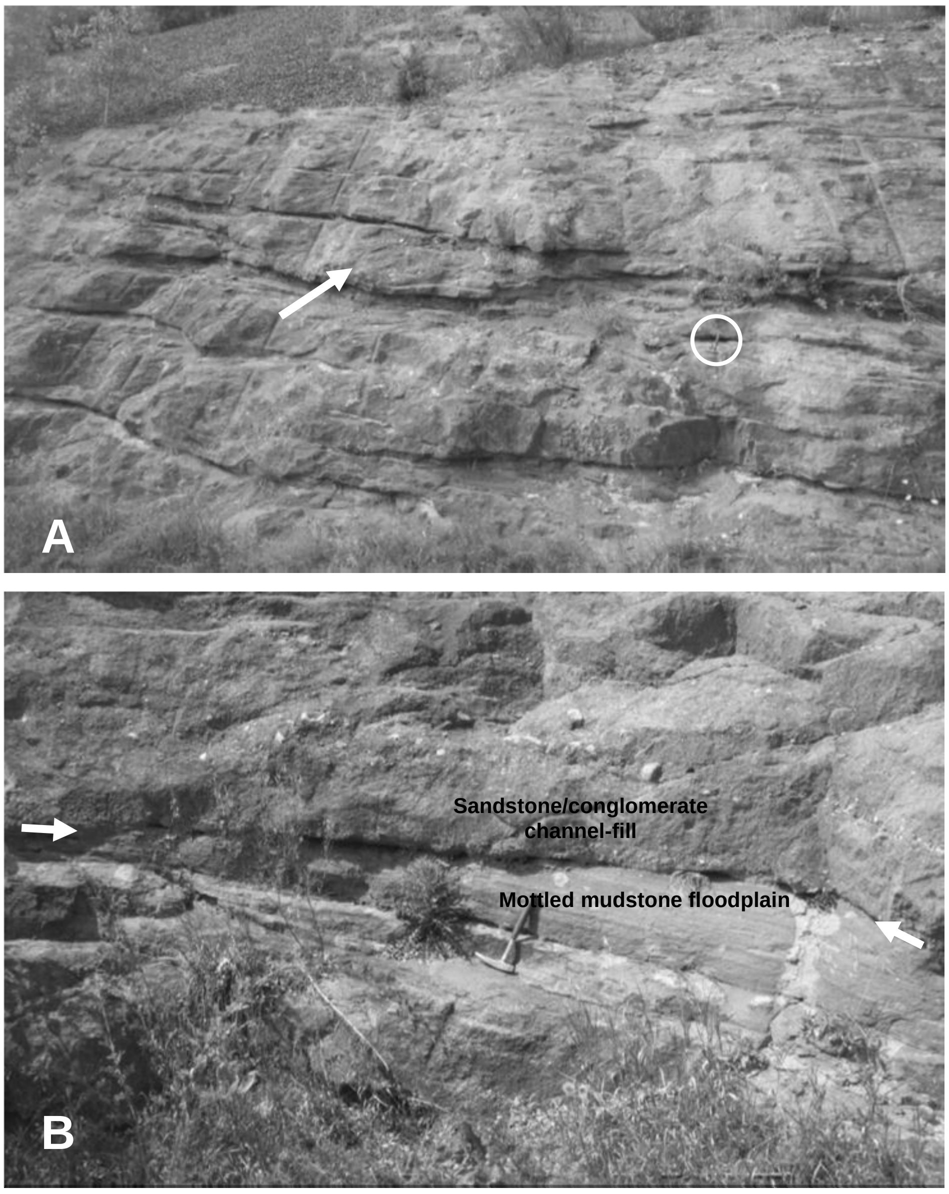

Figure 10. Outcrop photographs of sandstone channels and erosion surfaces at the Wickham Park outcrop.
A. Small sandstone channel (arrow) surrounded by shale layers (note circled hammer for scale).
B. Erosion surface (between arrows) that places sandstone unconformably over mudstone (note hammer for scale).

The channel-fill elements amalgamate laterally and vertically into larger scale channel belts (Figure 12) that are similar to those at Buckland Hills. The base of each channel belt is erosive, but the erosion is typically not more than a few centimeters in depth, resulting either from the cohesion of the mudstones which are being eroded, or from a balance between degradation by erosion and the amount of sediment and water entering the system. Channel belts are typically on the order of 4 to 10 meters thick, greater than those at Buckland Hills. This may indicate higher accommodation near the center of the basin relative to the edge of the basin. Each channel belt is capped by a thick (typically > 50cm) mudstone layer exhibiting pedogenic alteration. These mudstone layers are interpreted to have formed over a considerable time when the channel system avulsed to a location with greater accommodation. As at the Buckland outcrop, these channel belts can be considered "accommodation cycles" that are controlled by autogenic river avulsions or possibly allocyclic climate controls influencing the capacity of the rivers to transport bedload (Leleu et al. 2010) or affecting accommodation (McLaurin and Steel, 2007).

The nature of the channels that comprise the channel belts at Wickham Park changes over time. Figure 12 shows the lower part of the outcrop at Wickham Park. The lowest few channel belts are comprised of relatively small channels that migrate laterally and vertically over time. The middle part of the Wickham Park section (lower picture in Figure 12) shifts to channel belts composed of fewer, more laterally continuous channels with relatively constant thicknesses and higher width-to-depth ratios. They stack vertically within channel belts. The top of the outcrop (not in Figure 12) shows a return to channel belts composed of smaller, highly amalgamated channels similar to the base of the outcrop. Further analyses are required to see if there is larger scale cyclicity, such as the stacked channel belt complexes observed in the Fundy Basin (LeLeu et al. 2009, 2010).

As noted above, the floodplain deposits exposed at Wickham Park are thicker than those at Buckland Hills, and preserve evidence of greater pedogenic alteration. Even so, the alteration is limited to burrows, roots, and gleying (light greenish-gray mottles). No horizonization or significant alteration of parent material has occurred. Thus, these mudstones are interpreted to represent entisols. This is in contrast to the mudstone layers at Buckland Hills, which exhibit only mudcracks and burrows, and are classified as inceptisols (at best). Better soil development at Wickham Park can be attributed to several factors related to a more central location in the basin, such as the finer grained nature of sediment, different paleohydrogeologic conditions, or more time for formation due to increased accommodation. Alternatively, variations in soil development may be related to difference in climate at the times of deposition of the Wickham Park vs. Buckland Hills outcrops.

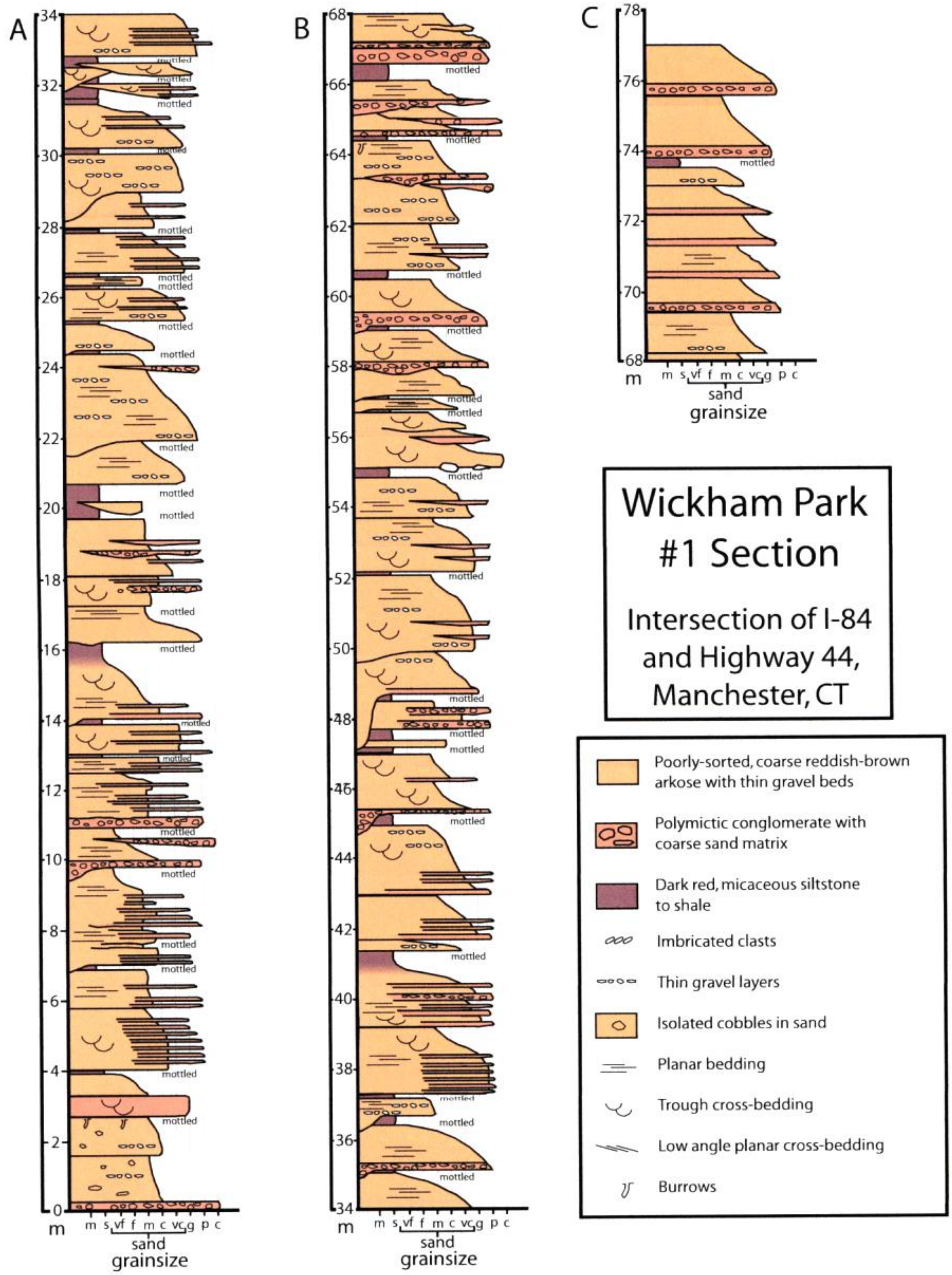

Figure 11. Measured section along the Rails-to-Trails bike path adjacent to Wickham Park, Manchester, CT. The 77m section is composed of meter-scale fining upward conglomerate and sandstone beds punctuated by mottled reddish-brown and light greenish-gray mudstone.

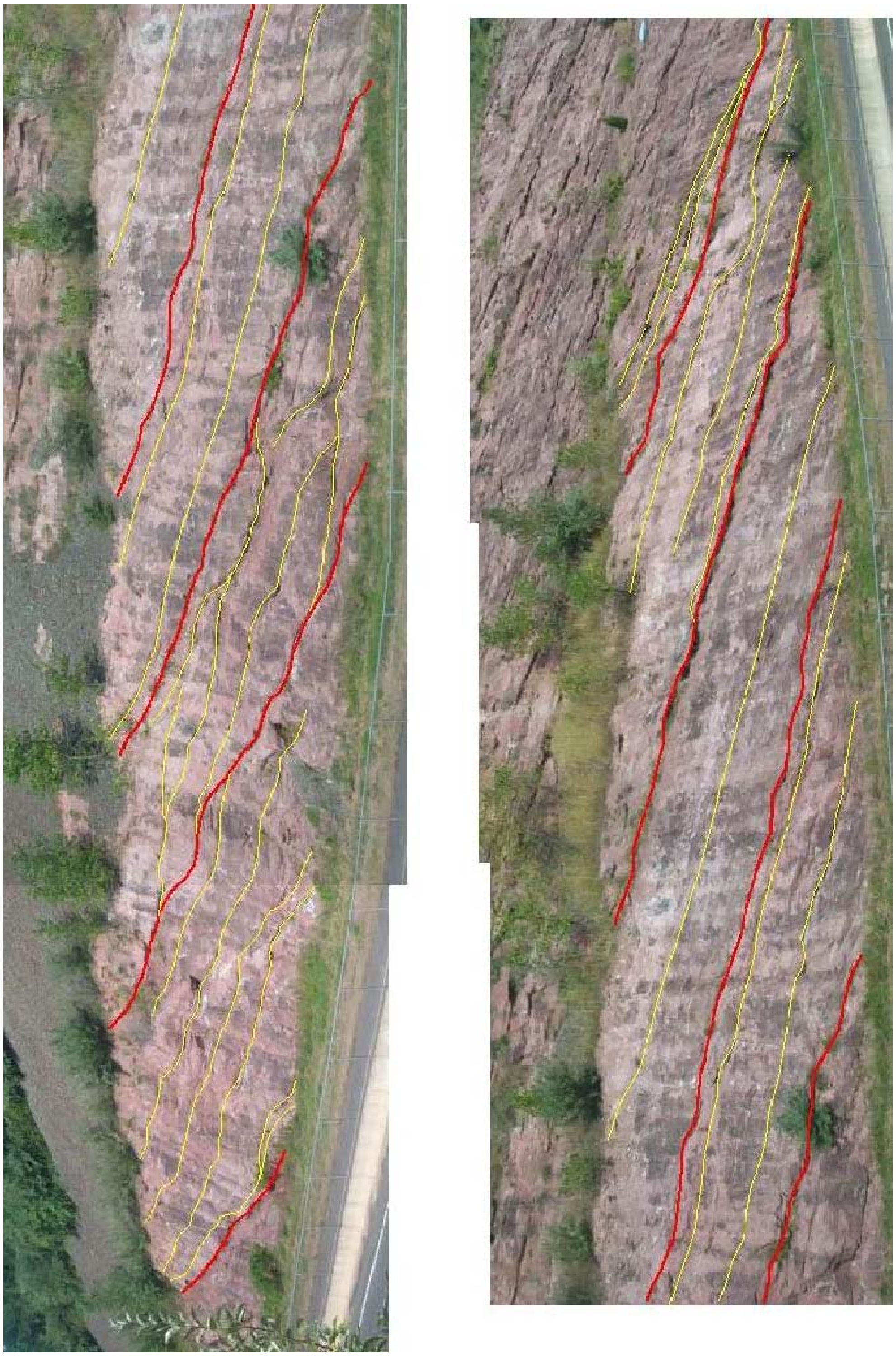

Figure 12. Photograph of the lower part of the Wickham Park outcrop that highlights the channel belts (red lines) recognized by persistent, relatively thick shale layers. The yellow lines represent smaller scale channel-fill elements.

A depositional model for the upper Portland Formation (Fig. 13) was developed primarily from the examination of the Buckland Hills and Wickham Park outcrops. The Buckland Hills outcrop (1) is located closer to the sediment source area than the Wickham Park outcrop (2).

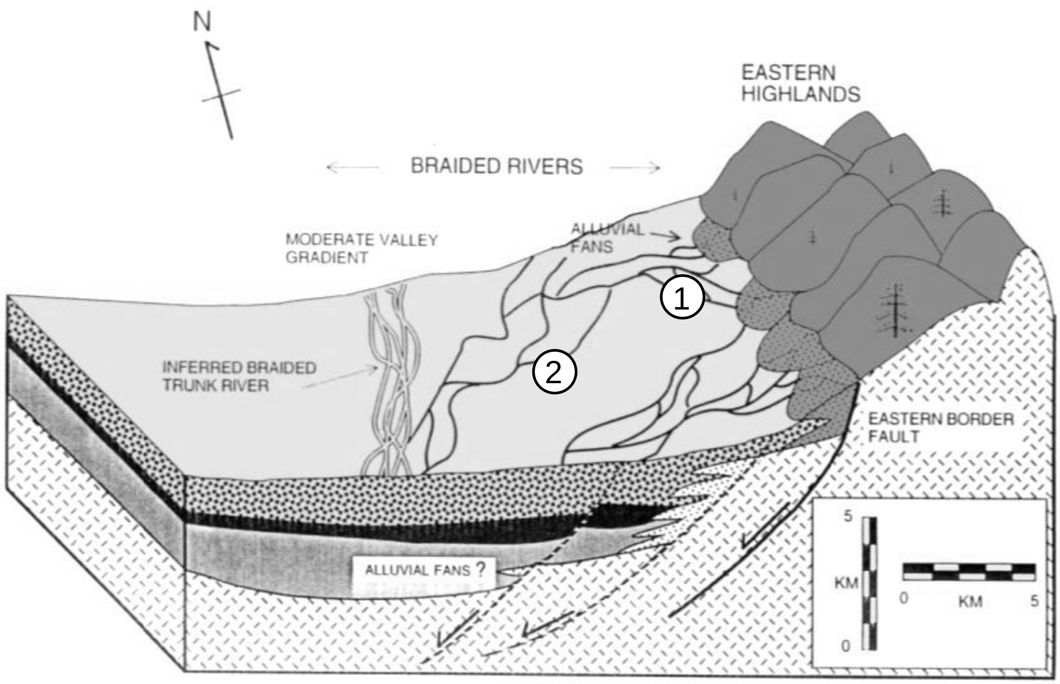

Figure 13. Depositional model for the upper Portland Formation modified from McInerney (1993). Location 1 is the relative position of the Buckland Hills outcrop and location 2 is the relative position of the Wickham Park outcrop.

There are a number of questions to discuss at the Wickham Park outcrop, including:

1. What is responsible for the shift in channel belt style from small, amalgamated channels to larger, laterally continuous channels and back again? Is there any evidence of larger scale stacked channel belt complexes as observed in similar age rift sediments in the Fundy Basin (LeLeu et al. 2009, 2010)?

2. What factors promoted better soil development at Wickham Park as compared to the Buckland Hills outcrops?

3. Are the fluvial cycles observed within the upper Portland Formation the result of allogenic forcing, or can they be explained in terms of nature autogenic processes (such as avulsions, or channel migrations)? What does this tell us about the controls on fluvial architecture within these braided systems?

Stop 3. Lacustrine/Playa Cores (lower Portland Formation) from the Center of the Hartford Basin (South Hartford, CT), stored at the Connecticut Geological Survey Core Repository (Farmington, CT)

Road Log from the Parking Lot of Wickham Park (Manchester, CT) to the Connecticut Geological Survey Core Repository (Farmington, CT)

Driving Directions	Mileage for Step	Cumulative Mileage
1. Turn left out of the parking lot at Wickham Park and head along the road that takes you to the park exit.	0.1	0.1
2. Turn left onto Middle Turnpike W/US-44.	0.2	0.3
3. Merge onto I-84 W/US-6 W via the ramp on the left toward Hartford. We will pass the outcrop that we just examined at Stop 2.	13.9	14.2
4. Merge onto US-6 W via EXIT 38 toward Bristol. The rocks you pass through along this stretch of the trip are Jurassic basalts and some sedimentary rocks from the Hartford Basin	4.6	18.8
5. Turn right into the driveway for the Connecticut Geological Survey Core Repository. There is a small sign at the entrance. This driveway also serves as a parking lot for local hiking trails. The gate to the repository is usually locked, but will be open for our visit.	0.1	18.9
6. Park in the lot at the end of the drive. We will be spending the rest of the field trip at this location.	Total Distance = 118.9 miles	

Introduction to the Cores:

A series of cores through the lower Portland Formation were taken along an east-west transect in southern Hartford in preparation for drilling a tunnel that redirected the course of the Park River. Although collected in the central portion of the basin, the cores now reside in the Connecticut Geological Survey Core Repository in Farmington, CT. These cores collectively provide a record of the lowest c. 600 meters of the formation, and are interpreted to have been deposited in a variety of playa, lacustrine, and alluvial environments (Zerezghi, 2007). Thirty-one cores were taken in all, and we will be examining 2 or 3 that best represent the facies in this part of the formation. The most comprehensive data set collected to date comes from core FD-30T. The information contained within this guidebook is derived from this core, but the observations are representative of the other cores.

Description:

Eight depositional facies were interpreted from the strata preserved in core FD-30T. The only detailed account of the playa/lacustrine facies of the lower Portland Formation comes from a MS thesis (Zerezghi, 2007), although, Olsen et al. (2005) give a brief overview. The facies are similar to those in the underlying East Berlin that has been studied in more detail (Demicco and Gierlowski-Kordesch, 1986; Gierlowski-Kordesch and Rust, 1994; Drzewiecki and Zuidema, 2007). The eight facies include:

1. *Laminated Black Shale*- This facies (Figure 14a) is composed of thinly laminated, organic-rich fissile black shale with rare thin discontinuous carbonate laminae. Framboidal pyrite crystals are common. Some laminae are contorted by post-depositional processes. The depositional environment of this facies is interpreted to be a deep and/or oxygen-poor perennial lake.

2. *Structureless Black Mudstone*- This facies (Figure 14b) is similar to the laminated black shale, but typically lacks laminae. The texture is uniform throughout. It is interpreted to represent shallower perennial lakes that contain enough oxygen to permit bioturbation.

3. *Thin-bedded Gray Shale*- This facies is structureless, very thin bedded, or laminated shale and siltstone. The color is light to medium gray and there are rare disruptions within the laminae caused by mudcracks, burrows, or soft-sediment deformation (Figure 14c). The depositional environment is interpreted to be a shallow perennial lake margin that experienced exposure.

4. *Crinkly-laminated Siltstone*- This facies is made up of thinly interbedded reddish-brown mudstone, siltstone and very fine-grained sandstone (Figure 15a). The laminae are highly convoluted and crinkly. Sand-filled mudcracks are abundant throughout this facies. The depositional environment is interpreted to be shallow ephemeral lakes in a playa environment. The crinkly laminae are interpreted to be microbial in origin.

5. *Current-rippled Siltstone*- This facies is composed of red siltstone, very fine-grained sandstone and/or medium-grained sandstone that contains current ripples (Figure 15b). It is equivalent to the interbedded sandstone and mudstone facies and ripple cross-laminated siltstone facies of Gierlowski-Kordesch and Rust (1994) and Gierlowski-Kordesch (1998). Asymmetric current ripples, and some flaser-like bedding are pervasive throughout this facies, and mudcracks are common. This lithofacies was deposited by sheetfloods (perhaps into shallow playa lakes) on playa mudflats and sandflats.

6. *Mudcracked Mudstone*- This facies is composed of reddish-brown mudstone to very fine-grained sandstone that is typically structureless but can contain obscure ripple cross-laminae (Figure 15c). Sand-filled mudcracks and burrows are abundant. The depositional environment is interpreted to be playa mudflats and/or sandflats.

7. *Cross-bedded Sandstone*- This lithofacies is white to light brown, medium- to coarse-grained arkosic sandstone with trough cross-bedding and some planar bedding (Figure 15d). Individual sandstone beds are between 10 and 50 cm thick with erosional bases overlain by red siltstone and mudstone rip-up clasts. This facies is interpreted to represent deposition from flowing water on playa mudflats and sandflats. It may represent sheetflood deposits, sheet deltas, and/or ephemeral braided streams (Gierlowski-Kordesch and Rust, 1994).

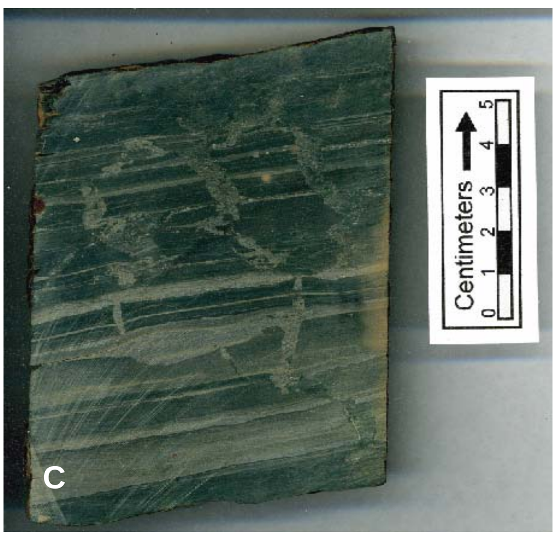

Figure 14. Gray to black mudstone facies.
A. Laminated black shale interpreted to be a deposited in a deep and/or anoxic perennial lake environment. Note framboidal pyrite.
B. Structureless black mudstone interpreted to be deposited in a shallow, oxygenated perennial lake environment.
C. Thin-bedded gray shale containing burrows and possibly mudcracks interpreted to be deposited in a lake margin environment.

8. *Pedogenically-altered Siltstone*- This lithofacies is composed of red siltstone to very fine-grained sandstone with evidence of pedogenic alterations (Figure 16). Some diagnostic characteristics include carbonate nodules and homogenized (bioturbated) sand and mud. There is also a very high concentration of mudcracks throughout the facies. This facies represents exposure of playa mudflats and sandflats. It is interpreted to represent incipient soil development.

Figure 15. Lithofacies interpreted to represent playa mudflats and/or sandflats. **A**. Crinkly-laminated siltstone with mudcracks interpreted to be playa lake environment. **B**. Current-rippled siltstone. **C.** Mudcracked red mudstone. The depositional environment is interpreted to be playa mudflats and/or sandflats. **D**. Cross-bedded sandstone with distinct trough cross-bedding.

Interpretation:

The facies in these cores form two distinct associations. The first three facies (gray and black shale and mudstone, Figure 14) were deposited in a perennial lake environment. The last five facies (reddish-brown mudstone and cross-bedded sandstone, Figures 15 and 16) were deposited in playa and alluvial plain environments. It should be noted that although the gray/black mudstone typically represents perennial lake facies and the reddish-brown mudstone typically represents playa facies, in some instances playa facies are gray in color. These differences in playa and perennial lake facies associations can be related to variations in climate. Although the climate was arid overall, the perennial lake facies likely represent periods that were slightly more humid. Many of the Park River Tunnel cores are characterized by the regular alteration between playa and perennial lake facies (Figure 17) that have been linked to cyclic changes in climate (Olsen and Kent, 1996; Olsen et al., 1996).

Figure 16. Examples of the pedogenically-altered siltstone facies.
A. Highly disrupted (almost homogenized) mixtures of mud and sand with mudcracks common.
B. Carbonate (dolomitic) nodules are also present in the pedogenically-altered facies.

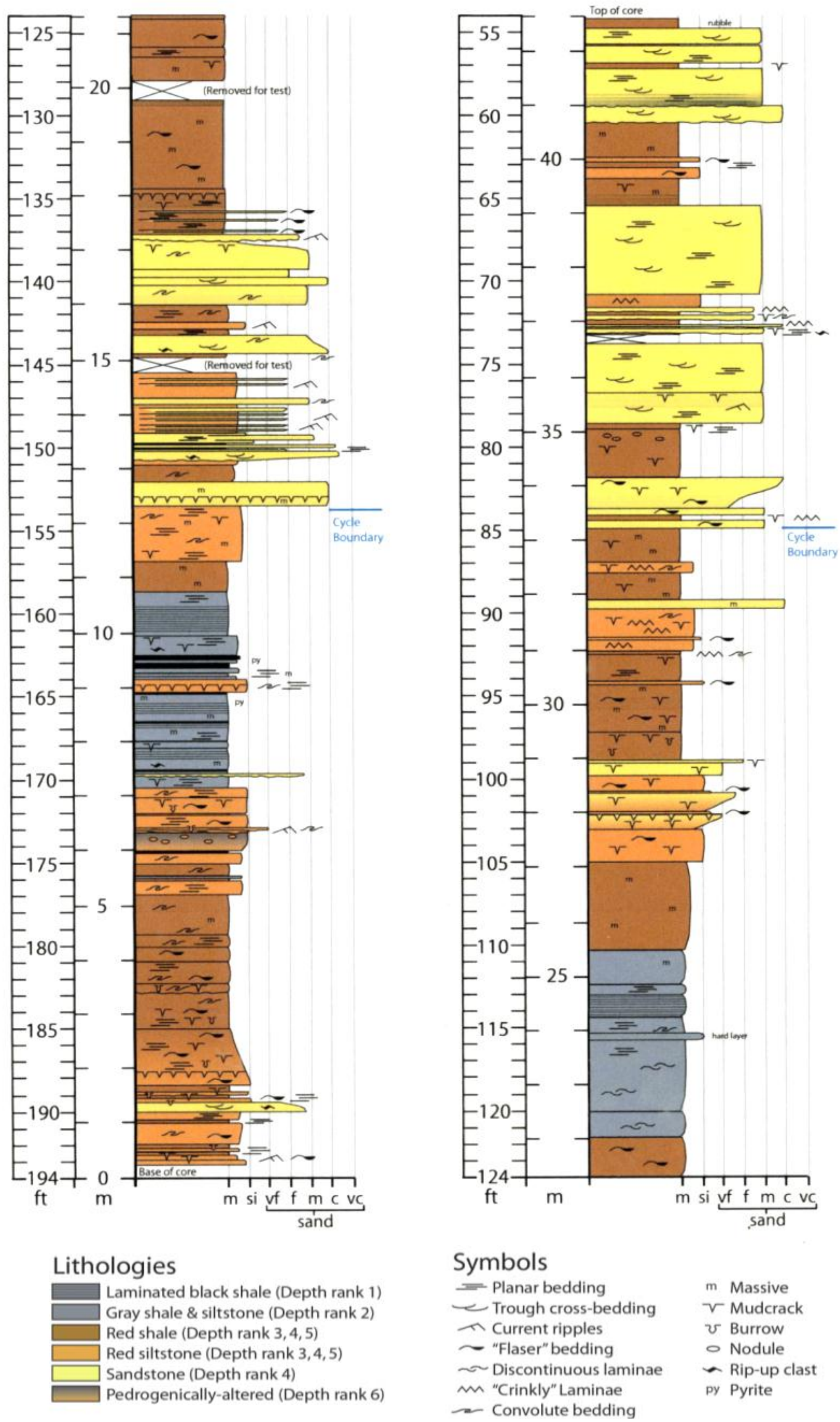

Figure 17. Measured section for core FD-30T.

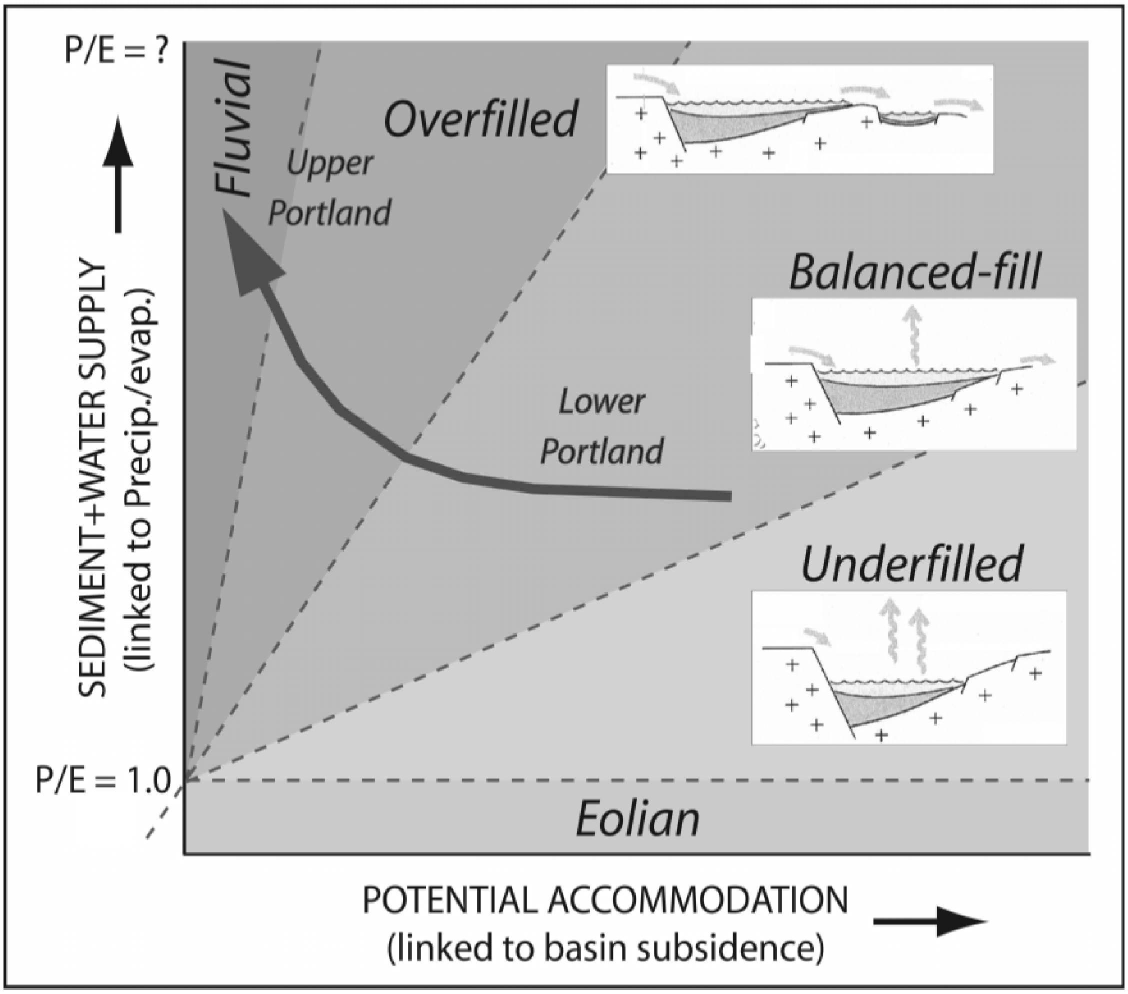

Figure 18. Diagram showing the relationships among accommodation, sediment+water supply, and lake system type. The arrow shows the direction of evolution of lake systems from underfilled to balance-filled in the lower Portland Formation, to fluvial in the upper Portland Formation. P/E stands for the ratio of precipitation to evaporation. Modified from Bohacs, Carroll, Neal, Mankiewicz (2000).

Carroll and Bohacs (1999) and Bohacs *et al.* (2000) proposed two fundamental controls on lake systems (Figure 18), including accommodation (the amount of space available for sediment to be deposited in a basin) and sediment+water supply (the amount of sediment and water flowing into the basin). Playa and perennial lake facies of the lower Portland Formation are consistent with deposition in balanced-fill lake basin types as defined by Carroll and Bohacs (1999). Balanced-fill lakes are defined as those that have rates of sediment+water supply that approximately balances the rates of changes in accommodation. These systems are characterized by better developed freshwater lakes interbedded with fluvial deposits.

In addition to the cyclic nature of the lower Portland Formation, the Park River Tunnel cores also display a gradual change from lacustrine-dominated strata to fluvial-dominated strata. This trend begins in core FD-30T, where an increased abundance of cross-bedded sandstone can be observed within the playa sediments at the top of the core (Figure 17). Perennial lake strata are absent in the few cores located stratigraphically above core FD-30T. The ultimate change from playa facies with increased fluvial input to fully fluvial environments like those seen at the Buckland Hills and Wickham Park has not been directly observed in the field. Olsen et al. (2005) estimate another 1000m or so of lacustrine cycles are stratigraphically above those preserved in the Park River Tunnel cores prior to the transition to a fluvial environment.

The playa facies and associated lake facies in the lower part of core FD-30T are interpreted as balanced-fill lake systems (Figure 18; Carroll and Bohacs, 1999; Bohacs *et al*., 2000). The sandstone in the upper part of the core represents sheetflood and, possibly, fluvial facies. The upper Portland Formation is composed of braided river deposits. The facies present do not necessarily indicate any fundamental change in climate so the shift from playa to fluvial dominated facies is interpreted to represent a decrease in accommodation associated with decreased movement along the Eastern Border Fault. As accommodation growth associated with basin subsidence waned in the middle Jurassic, the space required for lake development decreased and coarse sediment prograded out into the basin through braided rivers. This interpretation is consistent with rift basin stratigraphic models (Lambaise 1990; Schlische and Olsen, 1990; Gawthorpe and Leeder 2000), and is illustrated in Figure 18.

While at the core repository, we will have the opportunity to discuss some of the latest research pertaining to the lower Portland lacustrine strata. Proper correlation of lake cycles among the various cores is necessary for a comprehensive understanding of how the lake system evolved through time. Correlation is complicated by the fact that this region of the basin is cut by numerous normal faults (one is preserved within the cores) with significant vertical displacements. Thus, core correlation based on bedding orientations and horizontal distances between cores is unreliable, and other techniques are required to construct a robust stratigraphic framework. We are currently utilizing handheld X-ray Fluorescence and Gamma-ray Spectrometry, as well as organic and isotopic geochemistry to test and confirm correlations. In addition, the geochemistry will provide a better understanding of the types of lakes that occupied the basin, and how they evolved over time.

SUMMARY

This trip provided opportunities to examine: (1) the sandstone and conglomerate fluvial facies of the upper Portland Formation in outcrop, and (2) the mudstone to fine-grained sandstone facies of the lacustrine-dominated lower Portland Formation in core. The environments in which the sediments of the Portland Formation were deposited evolved gradually from lacustrine to fluvial as a result of a decrease in accommodation associated with the cessation of rift-related tectonics. Higher frequency cyclicity is superimposed on this tectonic trend, as exemplified by the lacustrine cycles in the lower Portland Formation and the channel belts in the upper Portland Formation. Climate was likely a contributing factor in the creation of the lake cycles. Autogenic river avulsions, or perhaps an allogenic controls (climate, tectonics), created the channel belts. Current work on Portland strata is focusing on the application of organic geochemistry, gamma-ray spectrometry, and X-ray fluorescence spectrometry to improve correlations among the cores, and to characterize the nature of the lakes.

Road Log from the Connecticut Geological Survey Core Repository (Farmington, CT) to the Hartford Marriott Downtown

Driving Directions	Mileage for Step	Cumulative Mileage
1. Leave the parking lot of the core repository heading back toward Scott Swamp Road.	0.1	0.1
2. Turn left (east) onto Scott Swamp Road/US-6 toward New Britain Avenue. Continue to follow US-6 E	4.5	4.6
3. Merge onto I-84 E/US-6 E via the ramp on the left toward CT-9 S/Hartford.	7.8	12.4
4. Merge onto I-91 S via EXIT 52 toward New Haven.	1.0	13.4
5. Take EXIT 29A toward Capitol Area.	0.2	13.6
6. Merge onto Whitehead Highway.	0.1	13.7
7. Take the Columbus Boulevard ramp toward Convention Center.	0.1	13.8
8. Turn right onto Columbus Boulevard.	0.1	13.9
9. The Hartford Marriott Downtown (200 Columbus Boulevard) is on the right.	Total Distance = 13.9 miles	

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