CAUSES AND CONSEQUENCES OF THE TRIASSIC-JURASSIC MASS EXTINCTION AS SEEN FROM THE HARTFORD BASIN

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INTRODUCTION

One of the most severe mass extinctions of the Phanerozoic, the Triassic-Jurassic event is greater or equal in magnitude to that at the more famous K-T boundary (Benton, 1994) (Fig. 1). Such severity, at least for marine families is also supported by Foote's (2003) statistical revaluations, although there remain dissenters (e.g. Hallam, 2002; Lucas et al., 2002). The cause of this mass-extinction remains hotly debated; explanations include sea-level change (Hallam, 1990), a methane- and CO_2- generated super-greenhouse triggered by flood basalt eruptions (McElwain et al., 1999; Hesselbo et al., 2002), and bolide impacts (Olsen et al., 1987). During the Triassic, all major extant groups of terrestrial vertebrates evolved, including dinosaurs (whose descendants survive as birds) and mammals. The Triassic-Jurassic mass extinction may have cleared ecological space for the rise of dinosaur dominance much as the K-T mass extinction prepared the way for mammalian ecological ascent (Olsen et al., 2003a).

In this guidebook, we will examine outcrops, exposures, cores, and fossils that provide important new clues about the major features of the Triassic-Jurassic boundary and subsequent events in the Hartford basin, a rich source for data on continental ecosystems during this evolutionary transition. We will focus not just on the physical and biological record of the boundary, but on the post-boundary events, especially those recorded within and above the basin's extrusive zone which may have been characterized by a super-greenhouse environment. We will see spectacular exposures of volcanic structures, including a giant eruptive fissure complex, strata containing fauna and flora documenting the extraordinarily stressed post-boundary biota and its recovery from the suggested greenhouse world.

BIOLOGICAL ANG GEOLOGICAL CONTEXT

The Triassic and Early Jurassic—with its nearly symmetrical meridional supercontinent, Pangaea, stretching about the equator from pole to pole—represents an extreme end member of Earth's geography and climate. A "hot house" world, with no evidence of polar ice (Frakes, 1979), it is marked by deposition of coals in polar and equatorial regions and plausibly extremely high pCO_2 (Berner, 1999). Soil carbonates from the Hartford basin and elsewhere (e.g., Wang et al., 1998) suggest CO_2 levels were between 2000 and 3000 ppm (Ekart et al., 1999; Tanner et al, 2001). Fossil stomatal indices offer lower but still extreme concentrations close to 1000 ppm (McElwain et al., 1999; Retallack, 2001). Despite vast climate differences from the present, a humid equatorial zone of modern dimensions (Kent and Olsen, 2000) existed. Within the transition zone between this humid region and the arid tropics to the north, the Hartford rift basin developed, recording during its long history the Triassic-Jurassic boundary and adjacent events (Fig. 2).

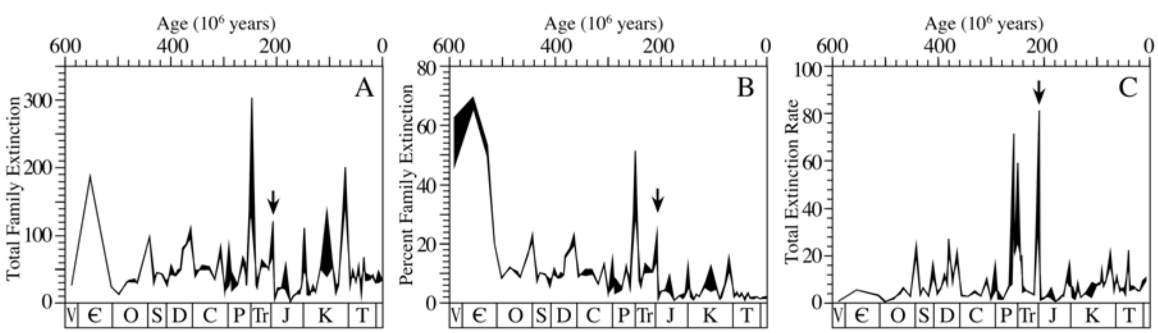


Figure 1. Extinction rate of marine and continental organisms through the last 600 million years (from Benton, 1994) with arrow at Triassic-Jurassic boundary. The upper and lower bounds represent the maximum and minimum curves. A, Extinction rate expressed as the numbers of families that died out in each stratigraphic stage. B, Extinction rate expressed as a percent of families that died out in relation to contemporaneous diversity in each stratigraphic stage. C, Extinction rate expressed as the number of families that died out in relation to the duration of each stratigraphic stage.

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

Pangean rift basins developed largely in a continental milieu during the Middle to Late Triassic along a huge rift zone from Greenland through the Gulf of Mexico in the ~40 m.y. preceding the Jurassic opening of the central Atlantic Ocean (Figs. 2, 3). The Hartford basin is one of the largest segments of the outcropping, deeply eroded North American contingent of these rifts. The basin fill, collectively termed the Newark Supergroup (Fig. 2), apparently formed in entirely non-marine settings. Continental rifting initiated in eastern North America sometime in the median Permian (Olsen et al., 2002d) and finished in the Early Jurassic, although the exact timing of the termination of rifting is poorly constrained. These rifts - in particular the Hartford basin - also record a major tectonic paroxysm that punctuated the beginning of the Jurassic: the emplacement of basaltic intrusions and extrusions of the Central Atlantic Magmatic Province (CAMP) (Marzoli, 1999; Olsen, 1999) - the largest known igneous province (see Philpotts and McHone, this volume).

Both dinosaurs and mammals evolved during the Triassic, along with all of the other major groups of extant terrestrial vertebrates. At the close of the Triassic there was also a large number of other groups of diverse body plans that are now extinct. These included the top predators of the Late Triassic, the fully terrestrial rauisuchians and crocodile-like phytosaurs as well as many strange small arboreal and aquatic forms. Seed plants such as conifers and cycadophytes were abundant along with various ferns and fern allies, including many extant families. Although there is some evidence that angiosperms (flowering plants) may have evolved by the Late Triassic (Cornet, 1989a,b; Wolfe et al., 1989), they were certainly not abundant. Through the Late Triassic, dinosaurs and an extinct conifer group the Cheirolepidiaceae or cheiroleps became relatively abundant. Not until after the Triassic-Jurassic mass extinction did both groups become the most conspicuous element of terrestrial communities. After their ascent however, dinosaurs remained the dominant large land animals for the next 135 million years until the mass-extinction at the K-T boundary, while cheiroleps remained the most abundant tropical trees for the next 80 million years until the proliferation of the angiosperms. The sudden dominance of the dinosaurs after the boundary is one of the main features of the biological record in the Hartford basin as is the extraordinary preponderance of cheirolepidiaceous conifers.

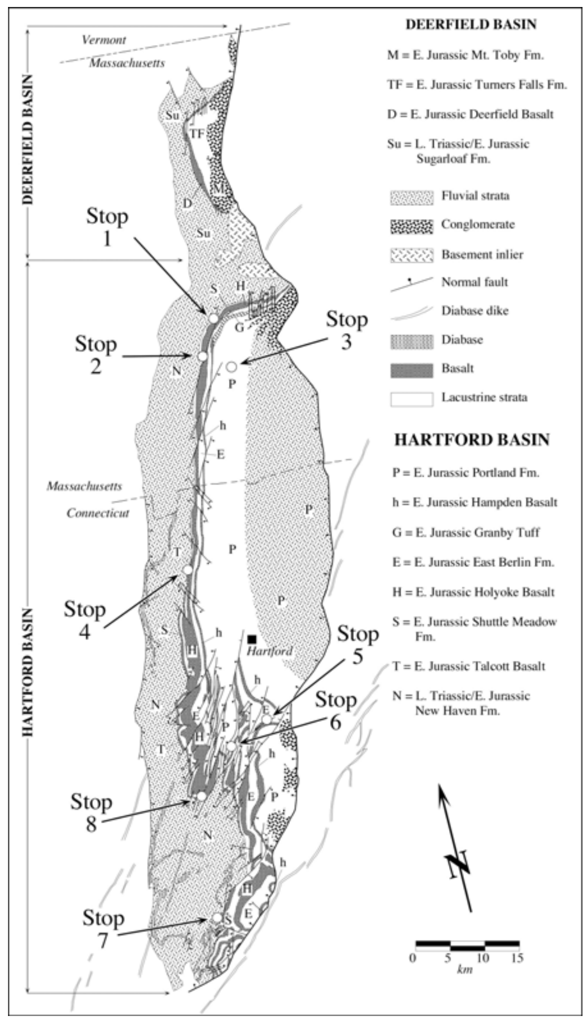


Figure 3. Bedrock geological map of the Hartford basin showing fieldtrip stops. Modified from Olsen and Rainforth (2003).

Robust and well-tested Milankovitch stratigraphy in the Newark basin provides a high-resolution framework for the transition that makes the Newark Supergroup, including the Hartford basin, uniquely suited to document the rates of environmental change through the boundary and the recovery. This cyclicity was first described in detail, and ascribed to astronomical control of climate by Van Houten (1962, 1964, 1969, 1980). All subsequent studies have confirmed and elaborated on these seminal works. An astronomically calibrated time scale has been developed based on more recent outcrop and core work (Kent et al., 1995; Olsen and Kent, 1996; Olsen et al., 1996a, b; Fedosh and Smoot, 1988).

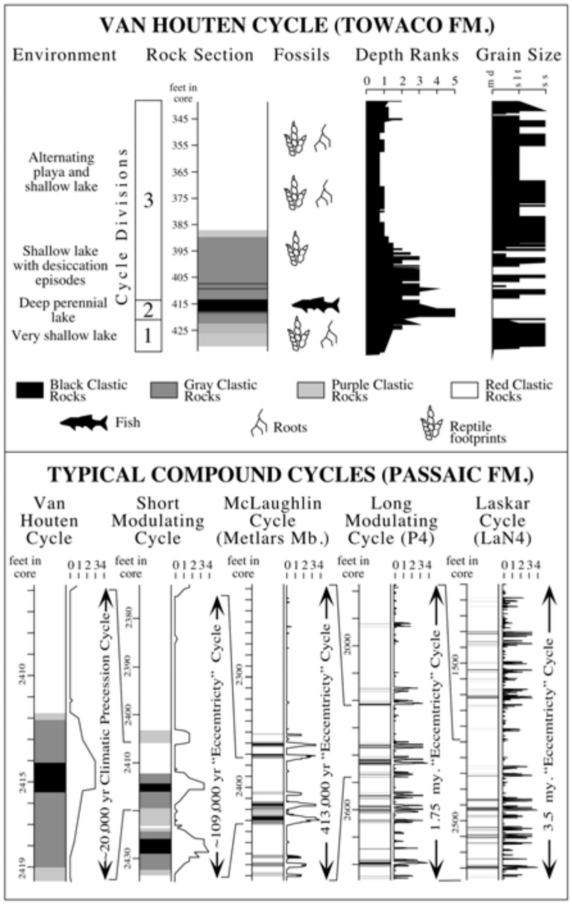


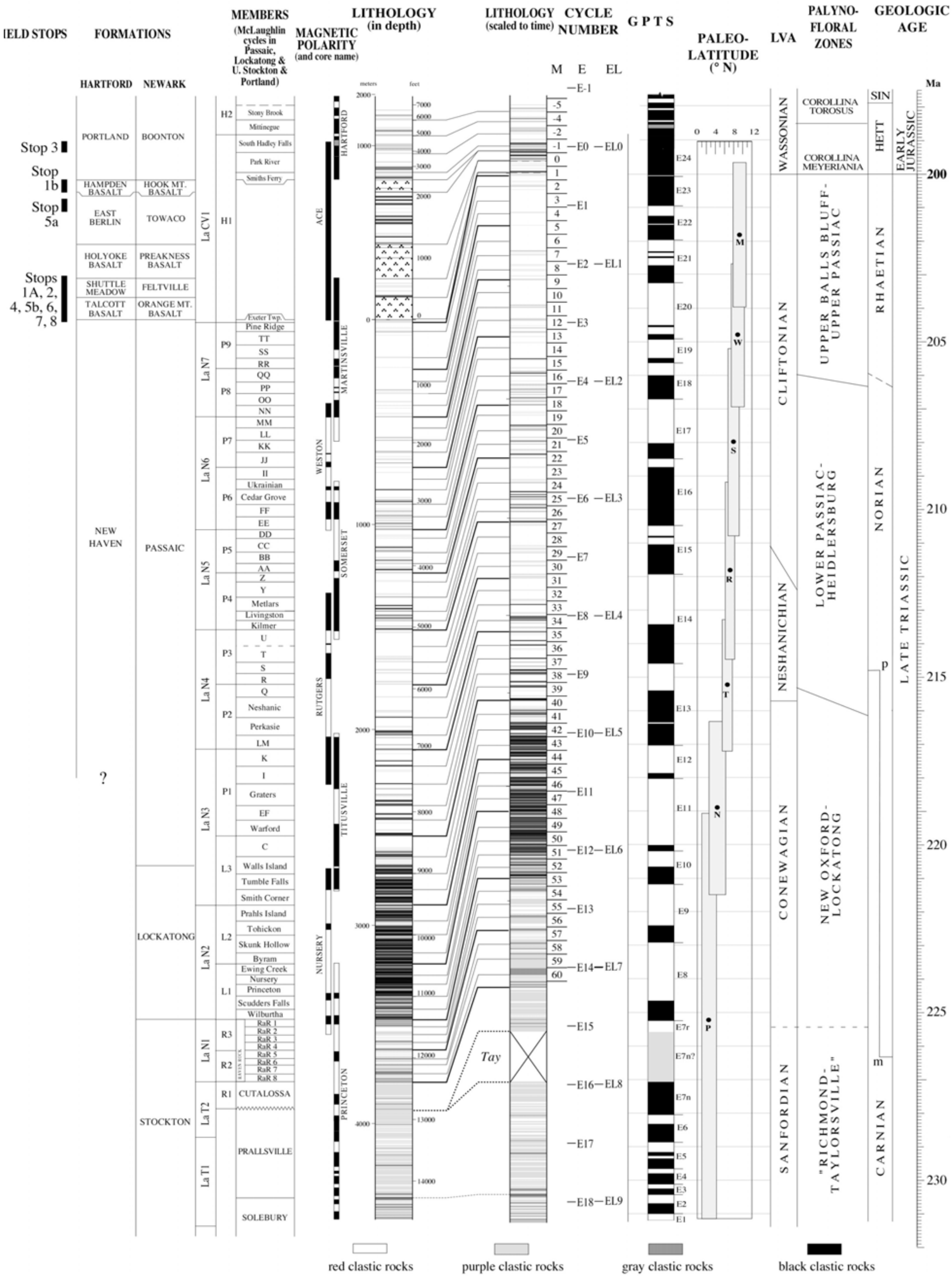
Figure 4. Van Houten and compound cycles. Modified from Olsen and Kent, 1999)

The fundamental sedimentary cycle seen in these sequences, caused by the ~20 ky cycles of climate precession, has consequently been named the Van Houten cycle (Olsen, 1986) (Fig. 4). The Van Houten cycle consists of three lithologically distinct divisions that represent lacustrine transgressive (division 1), high stand (division 2), and regressive followed by lowstand deposits (division 3) (Fig. 4) attributed to climatic variations affecting the rate of inflow and evaporation. Van Houten cycles are modulated in vertical succession by a hierarchy of four orders of cycles (Fig. 4). These are ascribed to modulation of precession by "eccentricity" cycles, which average approximately 100 ky, a 404 ky cycle of eccentricity called a McLaughlin cycle (after an astronomer at the University of Michigan, who mapped 404 ky cycles over much of the Newark basin; Olsen and Kent, 1996), and 1.75 and 3.5 m.y. modulating cycles (Olsen, 2001; Olsen and Rainforth, 2002b). The 404 ky McLaughlin cycle in the Newark basin serves as a basis for an astronomically-calibrated geomagnetic polarity time scale for the Late Triassic (Olsen and Kent, 1996, 1999), and earliest Jurassic (with data added from the Hartford basin), which is pinned in absolute time by radiometric dates from CAMP igneous rocks (Fig. 5). Employing the 404 ky cycle for time scale calibration for an interval hundreds of millions of years ago is justified because this eccentricity cycle is caused by the gravitational interaction of Jupiter and Venus, a cycle which should be stable on the scale of billions of years.

The lacustrine cyclicity pervades the lower three quarters of the Jurassic age section in the Hartford basin. While understood in broad outline for decades (Hubert et al., 1976), the detailed pattern of this cyclicity has been worked out only in recent years as a result of detailed fieldwork and study of industry and Army Corps of Engineers cores (Kent and Olsen, 1999a;b Olsen et al., 2002c, d). Van Houten cycles in the Hartford basin range from 10 and 30 m in thickness, depending on stratigraphic and geographic position. Within single formations in specific areas of the basin, the thickness tends to vary only about 25%.

TECTONOSTRATIGRAPHIC SEQUENCES AND THEIR BIOTA

Four tectonostratigraphic sequences are present in the central Pangean rifts (Olsen, 1997; Fig. 6). Tectonostratigraphic sequences (TS) are similar in concept to marine sequence stratigraphic units in that they are largely unconformity-bound genetically-related packages, but are controlled largely by tectonic events. These are directly relevant to our focus on the Triassic-Jurassic boundary because one tectonostratigraphic sequence boundary within the Hartford basin may cut out the detailed record of the Triassic-Jurassic boundary over most if not all the exposed basin. Tectonostratigraphic sequence I (TS I) is median Permian in age and while present in the Fundy basin of maritime Canada and various Moroccan basins, could exist in the subsurface in other basins. Tectonostratigraphic sequence II (TS II) is of ?Middle (Anisian-Ladinian) Triassic to early Late Triassic (Early to early Late Carnian) age and is present in most Newark Supergroup basins, although again there is no evidence for its


IELD STOPS
FORMATIONS
HARTFORD
NEWARK
MEMBERS (McLaughlin cycles in Passaic, Lockatong & U. Stockton & Portland)
MAGNETIC POLARITY (and core name)
LITHOLOGY (in depth)
LITHOLOGY (scaled to time)
CYCLE NUMBER
M E EL
G P T S
PALEO-LATITUDE (° N)
LVA
PALYNO-FLORAL ZONES
GEOLOGIC AGE
Ma
Stop 3
Stop 1b
Stop 5a
Stops 1A, 2, 4, 5b, 6, 7, 8
PORTLAND
BOONTON
HAMPDEN BASALT
HOOK MT. BASALT
EAST BERLIN
TOWACO
HOLYOKE BASALT
PREAKNESS BASALT
SHUTTLE MEADOW
FELTVILLE
TALCOTT BASALT
ORANGE MT. BASALT
NEW HAVEN
PASSAIC
LOCKATONG
STOCKTON
?
HARTFORD
ACE
MARTINSVILLE
WESTON
SOMERSET
RUTGERS
TITUSVILLE
NURSERY
PRINCETON
Tay
WASSONIAN
CLIFTONIAN
NESHANICHIAN
CONEWAGIAN
SANFORDIAN
COROLLINA TOROSUS
COROLLINA MEYERIANIA
UPPER BALLS BLUFF-UPPER PASSIAC
LOWER PASSIAC-HEIDLERSBURG
NEW OXFORD-LOCKATONG
"RICHMOND-TAYLORSVILLE"
SIN
HETT
RHAETIAN
NORIAN
CARNIAN
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LATE TRIASSIC
red clastic rocks
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gray clastic rocks
black clastic rocks

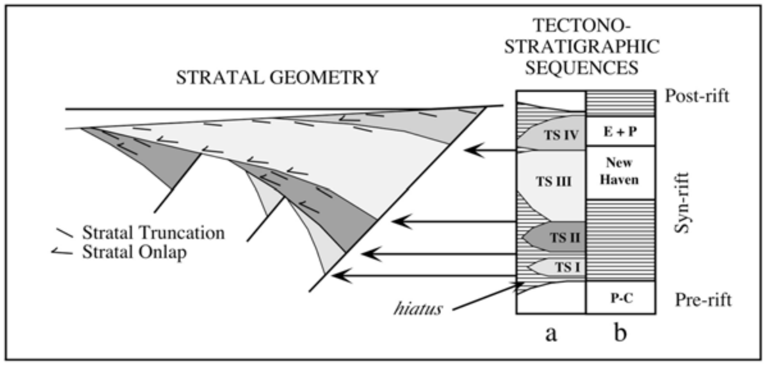

Figure 6. Tectonostratigraphic model of Newark Supergroup basins (after Olsen (1997): a, Tectonostratigraphic sequences; b, sequences known to be present in the Hartford basin.

presence in the Hartford basin. Tectonostratigraphic sequence III (TS III), of early Late Triassic (Late Carnian through early Late Rhaetian) age, is the most widespread of the sequences and dominates nearly all Newark Supergroup basins.

TS III is widespread in the Hartford basin, where it consists completely of the New Haven Formation. It differs dramatically from the Lockatong and Passaic formations of the Newark basin by the lack of lacustrine strata. Instead, TS III consists of red and tan fluvial strata; the stratigraphy and age is relatively poorly known. The basal New Haven formation locally has beds of gray sandstone that at one locality (Forestville, CT: Krynine, 1950; Cornet, 1977) produced a palynoflora closely comparable to that of the lower Passaic Formation. Hence the basal New Haven formation is conventionally assigned a basal Norian age (Cornet, 1977). The rest of the lower New Haven Formation consists of cyclical fluvial strata that have been interpreted as meandering river sequences (McInerney, 1993; Horne et al., 1993, McInenery and Hubert, 2002). These have common and locally well-developed pedogenic soil carbonates. Pure pedogenic micritic calcite from one such carbonate provided a U-Pb date of 211.9 ± 2.1 Ma (Wang et al., 1998), a Norian age on most time scales, including that from the Newark (Gradstein et al., 1995; Kent and Olsen, 1999b). The same exposure has produced a partial skull of the crocodylomorph *Erpetosuchus*, otherwise known from the Lossiemouth Sandstone of Scotland, conventionally given a Carnian age (Olsen et al., 2000b). Previously described reptilian skeletal material from the New Haven Formation in the southern Hartford basin comprises the holotype of the stagonolepidid *Stegomus arcuatus* Marsh, 1896. Lucas et al. (1997) considered *Stegomus* to be a subjective junior synonym of *Aetosaurus* and use *Aetosaurus* as an index fossil for continental strata and the guide fossil for the Neschanichian Faunachron of Huber and Lucas (1993) and Lucas (1997). This again suggests an early to middle Norian age, although this has been questioned (Sues et al., 1999. The middle New Haven Formation in the central Hartford basin consists of mostly red massive sandstone with much less well-developed pedogenic carbonates (Krynine, 1950). There have been virtually no studies of this part of the formation. Apart from abundant *Scoyenia* burrows and root casts, the only fossil from this part of the formation is a scapula of an indeterminate phytosaur ("*Belodon validus*" Marsh, 1893), indicating a probable Late Triassic age. Much more varied lithologies categorize the upper part of TS III and upper New Haven Formation (Hubert, et al., 1978), including meandering and braided river deposits and minor eolian sandstones (Smoot, 1991). Vertebrates from these strata include an indeterminate sphenodontian (Sues and Baird, 1993) and the procolophonid *Hypsognathus fenneri* (Sues et al., 2000). The presence of *Hypsognathus* indicates correlation to the upper Passaic Formation of the Newark basin and thus a later Norian or Rhaetian age (Cliftonian Land Vertebrate Faunachron of Huber and Lucas, 1994.)

Tectonostratigraphic sequence IV (TS IV) is of latest Triassic (Late Rhaetian) to Early Jurassic (Hettangian and Sinemurian) age. It contains the Triassic-Jurassic boundary, extrusive tholeiitic basalts of the CAMP, and occasionally extensive post-CAMP sedimentary strata. TS IV is very well represented in the Hartford basin where more Jurassic strata are preserved than elsewhere in eastern North America. The uppermost New Haven Formation makes up the lowest portions of TS IV. Markedly cyclical lacustrine strata appear to be lacking and there is some

Figure 5: Opposite. Newark and Hartford basin section and combined time scale showing distribution of field stops (adapted from Olsen and Kent, 1999; Olsen et al., 2001a; Olsen et al., 2002a). Abbreviations are: M, 404 ky cycle; E. 1.75 m.y. cycles; EL, 3.5 m.y. cycles; H1-2, Hartford long modulating cycles; P1-9, Passaic long modulating cycles; L1-3, Lockatong long modulating cycles; R1-3, Raven Rock long modulating cycles; LaCV1, Connecticut Valley Laskar cycle; LaN1-7. Newark Laskar cycles; LaT1-2, Taylorsville Laskar cycles; Tay, TS II-III hiatus in Newark basin recognized as strata in the Taylorsville basin; E1-24, magnetostratigraphic polarity chrons; P, N, T, R, S, W, M, paleolatitudes of the Newark basin based on core holes Princeton, Nursery, Titusville, Rutgers, Somerset, Weston, Martinsville, respectively; p, conventional placement of Carnian-Norian boundary based on pollen and spores; m, placement of Carnian-Norian boundary based on magnetostratigraphic polarity correlation to European marine sections (e.g. Channel et al., 2003), LVA, Land Vertebrate Ages of Lucas and Huber (2003).

evidence of a TS III-TS IV hiatus in many outcropping areas (Stops 4b, 5). At least locally there are gray plant and pollen-bearing ?marginal lacustrine strata just below the Talcott Formation (Heilman, 1987). The uppermost few centimeters of gray mudstone and sandstone preserve abundant *Brachyphyllum* shoots and cones and a palynoflorule of typical Early Jurassic aspect, dominated by *Corollina* (Robbins, quoted in Heilman, 1987).

At the hingeward edges of the rift basins, the unconformities between the tectonostratigraphic sequences can represent large hiatuses, but may pass into correlative conformities at depth within the basins without a hiatus. As far as can be seen in outcrop or available shallow cores, the Triassic-Jurassic boundary in the Hartford basin may be represented by a small hiatus at the TS III – TS IV tectonostratigraphic boundary. Because most of the New Haven Formation is undisputedly of Late Triassic age, the position of the boundary probably lies in the red beds closely underlying the gray conifer-bearing sequence.

Above the uppermost New Haven Formation, generally fossiliferous cyclical lacustrine sequences dominate sedimentary sections of TS IV until the middle Portland Formation. Microfloral assemblages present in most gray claystones and siltstones are dominated by the cheirolepidiaceous pollen genus *Corollina* spp., which generally comprise at least 90% of Jurassic palynoflorules (Cornet, 1977). Floral macrofossils are often present in the same units. Assemblages bearing *Clathropteris* and *Equisetites* are common in the Shuttle Meadow Formation and equivalents and the upper portion of the limestone bearing cycles in the lower part of the formation has produced a relatively diverse macroflora of ferns, cycadeoides, ginkophytes, and cheirolepidiaceous conifers (Newberry, 1888). Floral assemblages from the East Berlin and Portland Formations tend to be much more dominated by cheirolepidiaceous conifers, notably *Brachyphyllum* and *Pagiophyllum* and their reproductive structures (Cornet, 1977).

Invertebrates are represented in TS IV by burrows, and also locally by abundant clams, ostracodes, conchostracans, and insects (McDonald, 1992; Huber, et al., 2002) Ostracodes and conchostrachans are common at certain horizons of the Shuttle Meadow, East Berlin and Portland Formations, while unionoid clams occur at several localities within the Shuttle Meadow and Portland Formations (McDonald, 1992; McDonald and LeTourneau, 1989.) A low diversity insect fauna was described from the Hartford basin and correlative rocks of the Deerfield basin by Huber et a. (200) to consist of coleopteran elytra and possible larvae, a blattoid, orthopteran and several generically-indeterminate larval forms. At least four morphotypes of insect walking traces are abundant at certain horizons of the Shuttle Meadow and East Berlin formation.

Articulated fossil fish, often beautifully preserved and very abundant, occur in microlaminated portions of specific Van Houten cycles in TS IV of the Hartford basin (Olsen et al, 1982). The Shuttle Meadow, East Berlin, and lower Portland formations have several species of the “sub holostean” *Redfieldius* and many species of the holostean *Semionotus*. The Shuttle Meadow and East Berlin formations also produce the “sub holostean” *Ptycholepis* and the large coelacanth *Diplurus*. The youngest fish-bearing sequence in the Portland Formation (Chicopee Fish Bed of the “Mittinegue” member) is dominated by *Acentrophorus chicopensis* (Newberry, 1888), a form unknown elsewhere in the Newark Supergroup that, although abundant, is unfortunately poorly preserved and generically indeterminate, but might be a pholidophoridiform.

TS IV of the Hartford basin is the type area of the famous Connecticut Valley footprint assemblage (e.g. Hitchcock, 1836, 1848, 1858, 1865; Lull, 1904, 1915, 1953; Olsen et al., 1998; Olsen and Rainforth, 2002a). The taxonomy of the footprint assemblage is massively over-split and confused. These assemblages, however, like those of TS IV in the Newark basin (above the Triassic-Jurassic boundary) are dominated by dinosaur tracks, particularly grallatorids (theropod dinosaur tracks including *Grallator*, *Anchisauripus*, and *Eubrontes*). Other dinosaurian forms present include *Anomoepus* (Lull, 1953; Olsen and Rainforth, 2002a) and *Otozoum* (Lull, 1953; Rainforth, 2003). There is no obvious difference from the oldest to youngest assemblages.

Osteological remains from TS IV are almost completely limited to the upper, fluvial part of the Portland Formation. Several localities have produced fragmentary to nearly complete skeletons of the prosauropod genera *Anchisaurus* and *Ammosaurus* and the crocodylomorph genus *Stegomosuchus* (Lull, 1953). Marginal lacustrine or fluvial intervals within the cyclical lower Portland have produced a single small theropod skeleton, *Podokosaurus holyokensis*. found in a glacial boulder (Lull, 1953) and a natural cast of an impression of a fragmentary small theropod skeleton (Colbert and Baird, 1958). Additionally, two isolated possible theropod teeth have been found in the lower Shuttle Meadow Formation (McDonald, 1992).

The tectonostratigraphic sequences were probably initiated by major pulses of regional extension that subsequently declined in amplitude, as hypothesized by the basin-filling model (outlined by Schlische and Olsen, 1990, and elaborated on by Contreras et al., 1997) (Fig. 6). As a consequence of growth of the accommodation space during the extensional pulse, the basin depositional environments should follow a tripartite development at their depocenters. Disregarding climate changes, this consists of a basal fluvial sequence, succeeded by a rapidly deepening lacustrine sequence, and culminating in slow upward shallowing. The slowing or cessation of creation of new accommodation space would cause additional shallowing and thus a return to fluvial conditions; eventually erosion would ensue if creation of accommodation space stopped or nearly stopped. Each new pulse of extension would be expected to produce a shift of the depocenter towards the boundary fault system, accompanied by erosion of the hanging wall deposits; this would continue until the basin fill onlapped those areas of the hanging wall. Whether or not the full basin filling sequence - termed a Schlische cycle by LeTourneau (2002) - is actually observed in outcrop, depends on the depth of erosion relative to the basin depocenter and the boundary conditions of the basin geometry and sediment input. In the case of the Hartford basin TS III is entirely fluvial and only TS IV displays a full Schlische cycle, - albeit an excellent one.

We hypothesize that hanging wall unconformities between TS II, III, and IV were each caused by a renewal of extension. This certainly is true of the TS III-IV boundary in the Newark basin, since it is actually a correlative conformity in most presently outcropping areas. On the other hand, there is substantial evidence of a composite unconformity at the TS III-IV boundary in the Hartford basin, which at least locally cuts out the Triassic-Jurassic boundary, although the correlative conformity is probably preserved over much of the basin. The differences may be due to greater depth of erosion within the Newark relative to the Hartford basin.

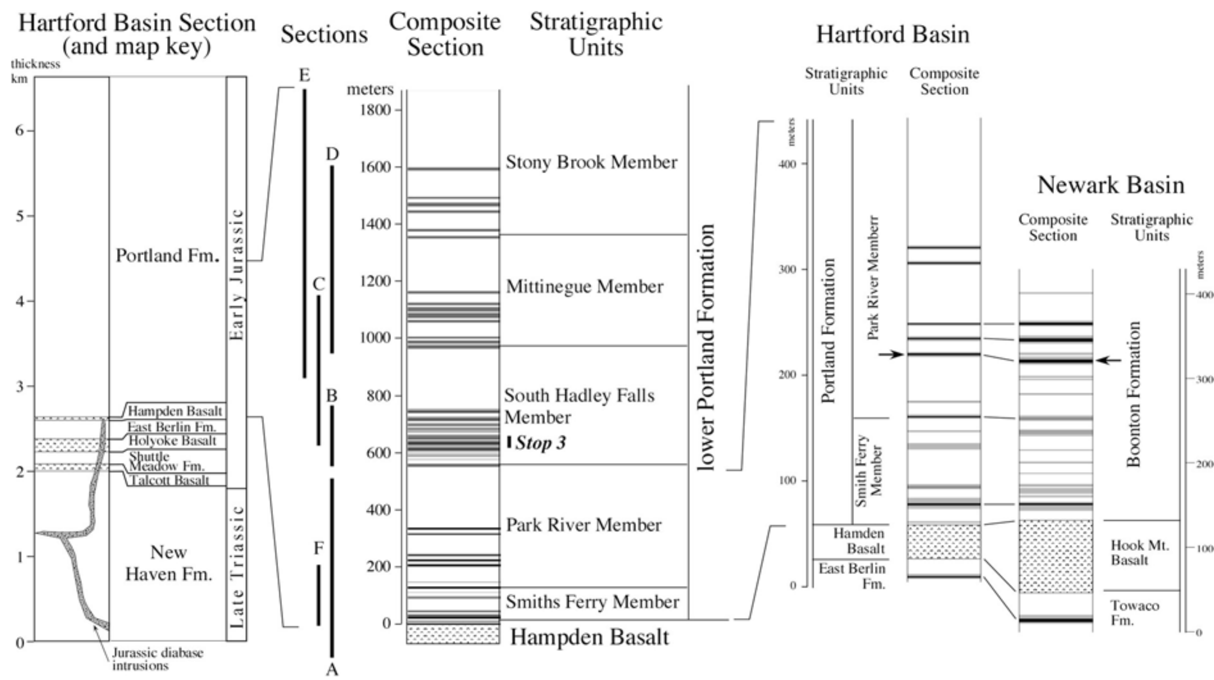

Figure 7. Stratigraphy of the Hartford basin, informal members of the Portland formation, and correlation to the Newark basin. Colors as in Fig. 5.

STRATIGRAPHIC NOMENCLATURE

Traditionally, the Hartford basin section is divided into seven formal lithostratigraphic mappable formations (Krynine, 1950; Lehman, 1959) that generally do not correspond to the tectonostratigraphic divisions (Fig. 7). The cyclostratigraphy of the upper six formations are virtually identical to the temporally overlapping portions of the Newark basin. The formations are, in ascending stratigraphic order: the New Haven Formation, comprised nearly entirely of fluvial tan, and red sandstone and conglomerate, and red mudstone, with very minor gray clastic rocks red eolian sandstone (maximum thickness ~2250 m); the Talcott Formation, which consists of a complex of high titanium quartz normative (HTQ) tholeiitic basalt flows and associated volcanoclastic beds (maximum thickness > 100 m) (Emerson, 1891; 1898a;b; Rogers et al., 1959; Lehman, 1959; Sanders, 1970; Puffer et al., 1981; Philpotts and Martello, 1986); the Shuttle Meadow Formation, composed of a mostly lacustrine and marginal fluvio-lacustrine clastic red, gray and black sequence very closely comparable to the Feltville Formation, although with better developed cyclicity (maximum thickness > 100 m) (Krynine, 1950; Lehman, 1959; Gierlowski-Kordesch and Huber, 1995; Olsen et al., 1996b); the Holyoke Basalt, made up of two major flows of high iron quartz normative (HFQ) basalt (maximum thickness > 100 m) (Emerson, 1898; Rogers et al., 1959; Lehman, 1959; Sanders, 1970; Puffer et al., 1981; Philpotts, 1992); the East Berlin Formation, comprised entirely of cyclical lacustrine and marginal fluvio-lacustrine red, gray, and black mudstone, sandstone and conglomerate, and minor limestone (maximum thickness >150 m); the Hampden Basalt, made up of two major flows of high titanium, high iron, quartz normative (HFTQ) basalt and its volcanoclastic equivalent in the northern Hartford basin, the Granby Tuff (maximum thickness >60 m) (Emerson, 1898; Rogers et al., 1959; Lehman, 1959; Sanders, 1970; Puffer et al., 1981; Philpotts, 1992); and finally the Portland Formation, composed of a lower half consisting of lacustrine and marginal fluvio-lacustrine red, gray and black clastic rocks closely comparable to the Easter Berlin Formation, and an upper half made up almost entirely of fluvial red mudstone sandstone and conglomerate and minor red eolian strata (total maximum thickness ~5000 m).

Olsen et al. (2002c) propose to divide the lower Portland Formation into members in a parallel manner to the Passaic Formation of the Newark basin. They recognize four full McLaughlin cycles in the lower Portland, and one continuing from the underlying East Berlin Formation. These mappable units are proposed as members as follows (from the bottom up): "Smiths Ferry," "Park River," "South Hadley Falls," "Mittinegue," and "Stony Brook" members (Fig. 7). These units are critical to establishing the cyclostratigraphy and time scale for the Hettangian and Sinemurian sites and hence reconstructing the sedimentologic and structural history from the Triassic-Jurassic boundary.

TRIASSIC AND JURASSIC CONTINENTAL COMMUNITIES AND THE TRIASSIC-JURASSIC BOUNDARY IN THE HARTFORD BASIN

The superb time control and resolution provided by the astronomically-calibrated paleomagnetic polarity timescale makes the Newark Supergroup, particularly the Newark and Hartford basins, one of the best venues for examining tropical continental floral and faunal change across the Triassic-Jurassic boundary (Kent and Olsen, 1999b; Olsen and Kent, 1999) and the subsequent recovery. Its one deficit, as cited by Benton (1994), has been a lack of osteological remains of tetrapods, but this is rapidly being remedied (Carter et al., 2001; Olsen et al., 2000b, 2001b; Sues et al, 2000). The changes around the Triassic-Jurassic boundary are seen in most detail in the Newark basin (e.g. Olsen et al, 2002a) where the section is not compromised by a hiatus, but the Jurassic recovery is best seen in the Hartford basin where the section is thicker and much better exposed. Based on the Newark timescale and paleontological correlations with areas outside the central Pangean rift zone, a consistent picture emerges of the profound changes that occurred around the boundary, with some indications of causation.

During the Late Triassic, there were several floral provinces that closely paralleled the geographic distribution of the Permian provinces, apparently following largely-zonal climate belts. There was a vast Gondwanan province in the Pangean southern hemisphere dominated by the pteridosperms *Dicroidium* and *Thinnfeldia* (Anderson and Anderson, 1970; Olsen and Galton, 1984), approximating the distribution of the Ipswich-Onslow microfloral province (Olsen and Galton, 1984). North of this was a tropical zone dominated by cycadophytes such as *Zamites*, and conifers such as *Pagiophyllum* (Ash, 1986; Axesmith and Kroehler, 1989). There was also a northern boreal province dominated by the pteridosperm *Lepidopteris*, dipteridaceous ferns, and tree ferns (Dobruskina, 1988, 1993; Harris, 1931). Both the southern Gondwanan assemblage and the northern boreal province were associated with extensive coal-forming environments. A band of coal-forming environments was also associated with the tropical province, albeit tightly restricted to within a few degrees of the Pangean equator.

Terrestrial tetrapod communities seem, at least in part, to have followed the plant communities. Southern higher-latitude communities, associated with drab-colored sediments, were dominated by synapsids, at least in the early Late Triassic, and at the southern polar regions, amphibians were dominant. A similar synapsid-rich community also existed in proximity to the equator, but otherwise the tropical regions had, by the Late Triassic, become strikingly archosaur-dominated, with large amphibians represented almost exclusively by metoposaurs that occur with a moderate diversity of phytosaurs. This tropical tetrapod province overlaps the Gondwanan *Dicroidium*-dominated province on the Indian plate; hence the tetrapod and plant communities were not completely parallel. Triassic southern boreal tetrapod assemblages again seem to have been dominated by some of the same archosaurs as in the tropical regions; however, amphibians, which included the bizarre plagiosaurs, were far more diverse. No faunas are known from the Late Triassic northern boreal and polar regions.

To some extent the faunas and floras tracked climate as central and southern Pangea drifted north. In most areas dinosaurs became more abundant, more diverse and much larger through the Triassic. The moderate- to large-sized herbivorous prosauropod dinosaurs became common in the later Triassic (Norian and Rhaetian) at the boundaries between the tropical and boreal regions, and perhaps at higher latitudes, but virtually absent from the lower latitudes. The provinciality and within-habitat diversity led to a very high-diversity global terrestrial biota, only now being appreciated (Anderson et al., 1996; see also Lucas and Huber, 2002 for review of tetrapod diversity and distribution).

The Early Jurassic global biota was much more stereotyped. Most floral provinciality was gone, evidenced by the elimination of the *Dicroidium-Thinfieldia* complex complex. Conifers, especially the now-extinct Cheirolepidiaceae (*Corollina*- producers) were extraordinarily dominant in the tropics, a pattern that would continue until the mid-Cretaceous (Watson, 1988). These conifer dominated plant communities of the earliest Jurassic are very well seen in the Hartford basin (Stops 4b, 5). A northern boreal province persisted, with infrequent cheirolepidiaceous conifers, while different groups (e.g. *Thaumatopteris*) (Harris, 1931) dominated. The boreal southern areas had less abundant cheirolepidiaceous conifers.

Cornet (1977 and in Olsen et al., 1989) documented an interesting pattern in conifer leaf physiognomy that is plausibly related to the early establishment of the cheirolep regime (Fig. 8). In the Newark, Hartford and Deerfield basins, post-boundary cheirolepidiaceous conifers tend to have small, stubby leaves with thickened cuticles and sunken and papillate stomata – adaptations usually associated with extreme heat and water stress. In contrast, Late Triassic cheirolepidiaceous conifers tend to have larger, thinner leaves with smooth cuticles. Based on recently collected conifer material from the Newark and Hartford basins, this transition occurred just after the Triassic-Jurassic boundary. Not until well after the youngest known basaltic eruptions of the CAMP (the Hampden Basalt and correlatives), in the Portland Formation, did large-leafed cheirolepidiaceous conifers with smooth cuticle again become dominant (Stop 8). Based on Milankovitch cyclostratigraphy this earliest Jurassic interval of small-leaved cheirolepidiaceous conifers lasted over a million years and plausibly representsplant communities responding to a

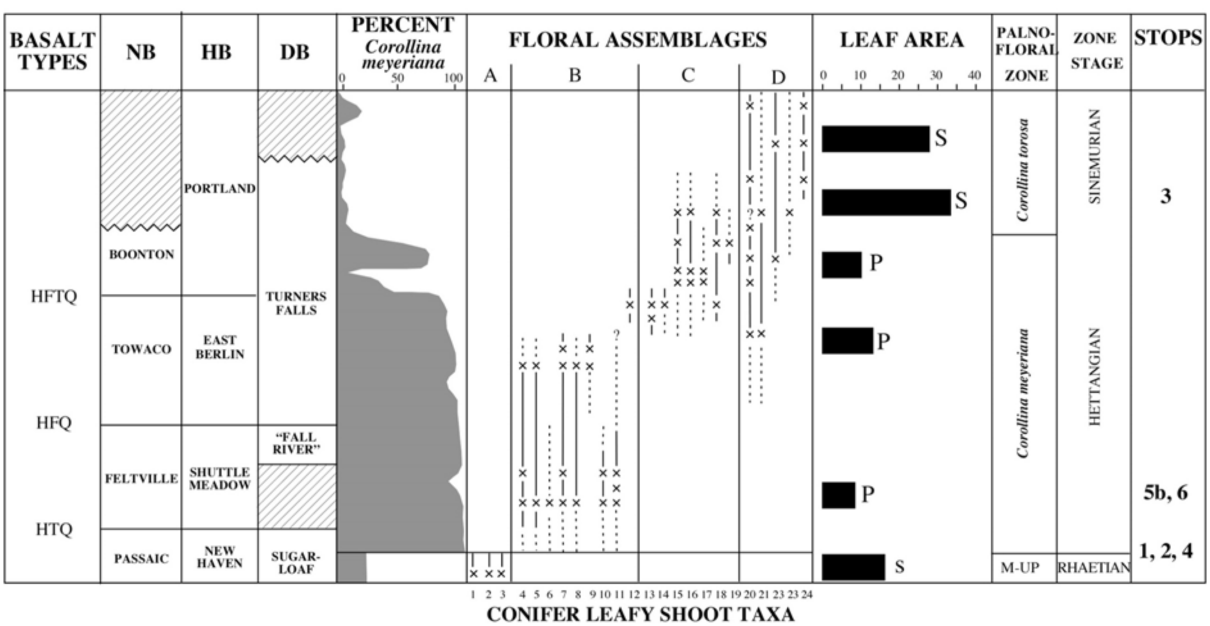

Figure 8. Distribution of different conifer morphologies through the latest Triassic and Early Jurassic in the Newark, Hartford, and Deerfield basins, based on Cornet (1977 and in Olsen et al., 1989). We have calculated average surface area of leafy shoots from Cornet's drawings: s, smooth cuticle; p, cuticles with papillate stomata. M-UP represents "Middle to Upper Passaic Palynoflora". Conifer leafy shoot taxa given in Cornet (1977).

super-greenhouse climate [i.e., McElwain's (McElwain et al., 1999) thermal damage hypothesis (which suggests high species level macrofloral turnover from leaf-temperatures raised above a highly conserved lethal threshold)], in some way triggered by the events around the Triassic-Jurassic boundary.

The tetrapod communities, at least at the beginning of the Early Jurassic, appear to have been virtually cosmopolitan, even at very low taxonomic levels (Shubin and Sues, 1991). Prosauropods and large theropods (larger than any in the Triassic) achieved nearly global distribution, along with crocodylomorphs and several other diapsid groups, with the same genera being reported from Arizona, southern Africa, Nova Scotia and China. Global and within-habitat diversity appears to have been much lower. No longer were there any synapsid-dominated communities; the only surviving members of this group were the tritylodonts, trithelodonts, and mammals, although again with nearly global distributions for several genera. Large amphibians were completely restricted to higher latitudes and had very low diversity. Most critically, non-ornithodiran (i.e. non-dinosaurs and pterosaurs) and non-crocodylomorph archosaurs were gone: these had been the most common large tetrapods of the of the Late Triassic tropics. Roughly 50% of all tetrapod families became extinct at or near the Triassic-Jurassic boundary (Olsen et al., 1987), making this mass extinction, at least for tetrapods, considerably larger than that at the K-T boundary.

The rate at which this change occurred can presently be assessed only in the Newark Supergroup. At this time most of the evidence comes from the Newark basin. In Newark Supergroup basins the Triassic-Jurassic boundary has been identified principally by a microfloral transition characterized by the disappearance of many typically Triassic taxa (Cornet, 1977). The floral change was evidently very abrupt, estimated in the Newark, Fundy, and Argana basins to have occurred over less than 20 ky, and probably much less (Fowell, 1993; Fowell and Olsen, 1993; Fowell et al., 1994; Fowell and Traverse, 1995; Olsen et al., 2000a, 2002a,b), because it occurs within a single Van Houten cycle. Tetrapod data, based mainly on Newark basin tetrapod footprint assemblages, although augmented with data from other Newark Supergroup basins show a similar rate of change. This change is consistent with the much less intensely sampled skeletal data.

In the Newark basin, the floral and faunal changes are directly associated with a trilete fern spore abundance anomaly ("fern spike") and iridium anomaly (Olsen et al., 2001b). The total area covered predominately by ferns, directly after the boundary has not yet been determined, but the Hartford basin provides a candidate for another location for a fern spike interval (Stop 2) and the Fundy basin an additional (Olsen et al., 2002b). The floral and faunal pattern, with the exception of the survival of the non-avian ornithodirans, and the associated iridium excursion is remarkably similar to the pattern seen at the K-T boundary in the North American Western Interior (e.g. Tschudy et al., 1984). This suggests a similar cause for both extinctions - one or more bolide impacts - a suggestion repeatedly made long before the new biotic and Ir data were available (Badjukov, et al., 1987; Bice et al., 1992; Dietz, 1986; Olsen et al., 1987, 1990; Rampino and Caldeira, 1993).

One of the most striking aspects of the Triassic-Jurassic boundary in the central Atlantic margin rifts is the superposition of the oldest CAMP basalts on the boundary, with stratigraphic evidence always exhibiting an intervening small thickness of Jurassic strata. A possible causal link is difficult to ignore, given a similar (although less precisely timed) coincidence between the Deccan Traps and the K-T boundary and the Siberian Traps and the Permo-Triassic boundary (Rampino and Caldeira, 1993). The three largest Phanerozoic mass-extinctions are penecontemporaneous with the three largest Phanerozoic flood basalt provinces. For each of these three flood basalt occurrences, there is some evidence of an asteroid or comet impact. Boslough et al., (1996) proposed a mechanism linking bolides with flood basalts, but the energetics are yet to be reconciled with the observations and the models (Melosh, 2000). Nonetheless, it seems plausible that a massive impact could initiate volcanic eruptions by concentrating the effusive rate of a distant flood basalt province. This topic has yet to be explored quantitatively at the appropriate scale.

The Hartford basin provides perhaps one of the best venues for exploring the role of the CAMP in the Triassic-Jurassic transition (Stops 1b, 3-7). The dynamics of the earliest documented eruptions can be best studied here because it is the only place where eruptive centers themselves have been unambiguously identified (e.g. Stops 1b, 6). Detailed analysis of these eruptive centers as well as the mechanisms of flow emplacement (Stops 3, 4b, 7) should give us some idea of what the eruption of the CAMP ejected into the atmosphere. Although models on atmospheric effects of the Deccan Trap eruptions suggest that volcanic outgassing of mantle volatiles (especially CO_2) was not a significant contributor to greenhouse warming (Caldeira and Rampino, 1990), the short term effects of gigantic, single flow events emplaced over hundreds to thousands of years have yet to be modeled. Such shortlived, large magnitude eruptions could have triggered even larger scale phenomena such as dissociation of methane clathrates (e.g., Hesselbo et al., 2002).

We can paint a speculative picture of the Triassic-Jurassic transition. Current data can be explained by the impact of one or more asteroids or comets (e.g. Spray et al., 1998) that terminated biotic diversity, which otherwise was rising through the Late Triassic. As with the K-T scenario, reduced sunlight and lower temperatures plagued continental biotas for months, and the global carbon cycle, as we can see in the abrupt and prolonged negative excursion in carbon isotope composition of marine organic matter ($\delta^{13}C_{org}$) (Ward et al., 2001; Pálfy et al., 2001; Hesselbo, et al., 2002) was massively perturbed. Similar to the K-T boundary (Beerling, 2002), the Triassic-Jurassic boundary was followed by a significant time of elevated CO_2 (McElwain, et al., 1999; Beerling, 2002), culminating in a super-greenhouse. This resulted in a shift to plants with smaller leaves cuticularly adapted to extreme heat and water-loss stress (e.g., Stops 4b, 5). High temperatures would likely have increased tropical convection resulting in increased regional precipitation and widespread lightning-induced fires leading to conditions of continually arrested ecological succession. To some degree the CAMP flood basalt episode must have contributed (or even caused) these disruptions (e.g., Hesselbo et al., 2002; Cohen and Coe, 2002). For a few thousand years after the disruption only rapidly dispersed spore-bearing plants—largely ferns—populated the tropical regions (Stop 2). Surviving dinosaurs were initially small, but in the next 10 ky theropod dinosaurs would dwarf their Triassic predecessors. The massive and sustained ecological disruption at the Triassic-Jurassic led to the extinction of many tetrapod families presumably dinosaurian competitors, and only afterward did the familiar dinosaur-dominated communities arise, a reign that would last for the next 135 million years.

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ROAD LOG

We have arranged the field trip to begin in the northern Hartford basin at possible Triassic-Jurassic boundary sections (Stops 1, 2) where the first basalt (Talcott) of the extrusive zone is absent and then look at later Jurassic strata that seem to record the recovery from that mass-extinction (Stop 3). We then proceed south to a section where all of the basalts are present (Stop 4) and have lunch. At the next stop (5) we go to Dinosaur State Park where we look at aspects of the continental ecosystem indicative of events within a half million years of the boundary as well as cores of the boundary and overlying strata. Next we continue further south look at the boundary and adjacent deposits focusing on the dynamics of the extrusion of the initial CAMP lavas (Stops 6 and 7) before turning north again to look at how the initial CAMP lavas progressed away from their feeders and influenced local syn-extrusion sedimentation.

Mileage: Time to first stop 25 min.

0.0 Trip begins from the University of Massachusetts Parking Lot #62 off N. Pleasant Street near the Geosciences building (Morrill Hall); leave via Stockbridge Road and turn right onto Infirmary Way.
0.2 Turn left onto N. Pleasant.
0.3 Turn right onto Massachusetts Ave.
1.3 Turn right onto ramp for Route 116 South.
3.1 Turn right onto Route 9 westbound.
8.0 Turn left onto entrance ramp for Route I 91 South.
12.0 Pull off on right shoulder onto dirt.

STOP 1a. POSSIBLE TR-J BOUNDARY SECTION, NORTHAMPTON, MA. (30 MINUTES) Be cautious of traffic; this stop is on an interstate; and beware of poison ivy. SW Mt. Holyoke Quadrangle (approx.) 42° 17.03' N, 072° 36.88' W; Tectonostratigraphic Sequence TS IV; Upper New Haven Fm., lower Shuttle Meadow Fm., Holyoke Basalt; ?Rhaetian-Hettangian age ~200 Ma. Main points are: exposures of basal TS IV with a possible boundary section; absence of Talcott Formation; presence of lower limestone sequence of Shuttle Meadow Formation.

A large cut on the southwest side of the south bound lanes of Interstate 91 on the north side of Mount Tom exposes over 100 m of upper New Haven Formation and lower Shuttle Meadow Formation (Figs. 9, 10). While the upper Shuttle Meadow Formation is not exposed in this cut, the lower flow of the Holyoke Basalt is. This section was described by Brophy et al. (1967) and in part by Cornet (1977) and McDonald (1982), but has never been measured in detail.

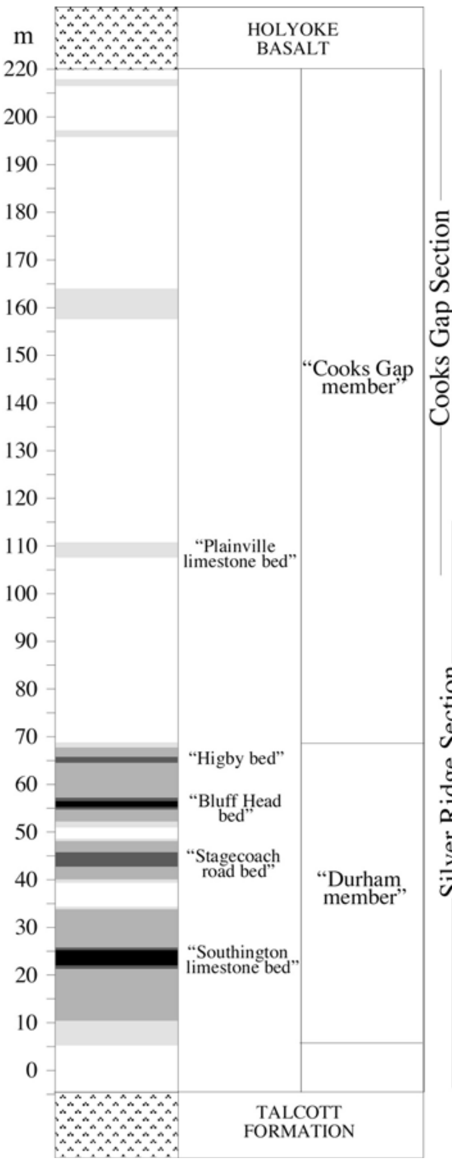


Figure 9. Generalized stratigraphy of the Shuttle Meadow Formation based on the sections at Cooks Gap (exposures) and Silver Ridge (cores and exposures). The scaling between the two sections is uncertain, with the depth scale shown being from Silver Ridge. Rock color as in Fig. 5.

The base of the section consists of conglomeratic sandstone in excess of 15 meters thick, possibly a basal conglomerate of TS IV, lithostratigraphically belonging to the uppermost New Haven Formation. According to Hubert (pers. com., 2002) these coarse units are predominately comprised of debris flows. The Talcott Formation is absent in Massachusetts, but where present in Connecticut, the upper New Haven Formation tends to be finer grained and more distinctly red than lower in the formation, as at Stop 3. This is followed by a 13 m thick sequence of predominately fluvial or marginal lacustrine red sandstone and mudstone that lithologically marks the base of the Shuttle Meadow Formation, and which probably contains the Triassic-Jurassic boundary. The succeeding 3 m is a lacustrine transgressive sequence (division 1) culminating in rooted gray silt and sandstone. Division 2 consists of about 2 m of platy dark gray laminated limestone and calcareous siltstone grading up into crudely laminated black siltstone. The platy laminated limestone contains limestone nodules with well-preserved examples of the fish *Semionotus*, *Redfieldius* and *Ptycholepis* (Cornet, 1977; Cornet and Traverse, 1975; Olsen et al, 1982) as well as abundant coprolites probably of the large coelacanth *Diplurus* (Gilfillian and Olsen, 2000) known from other Shuttle Meadow localities. This is certainly the lower limestone unit of the Shuttle Meadow Formation (Fig. 9), that we term the "Southington limestone bed," deposited in a very large lake covering the Hartford basin (see Stop 5b). A 4 m thick sequence of Gilbert-type delta forests of sandstone overlies the lower limestone and siltstone and tongues downward into it. A gray (2 m), then predominately red (16 m) mostly fluvial clastic sequence follows, overlain by a long expanse of no exposure below the splintery columns of the Holyoke Basalt.

No one has yet attempted to search for the boundary in detail in this section. If it is preserved as an event horizon, it is probably within the lower red clastic sequence, where one of the red or purplish clay rich layers might be prime suspects. Certainly, the magnetic stratigraphy of this section would prove enlightening.

Return to vehicles.

Mileage: Time to next stop 3 min
12.0 Resume driving north on Route I 91 North.
12.4 Holyoke basalt on right with characteristic splintery jointing.
13.4 Diabase plug intruded into and apparently feeding the Granby Tuff.
13.7 Diabase plug in Granby Tuff.

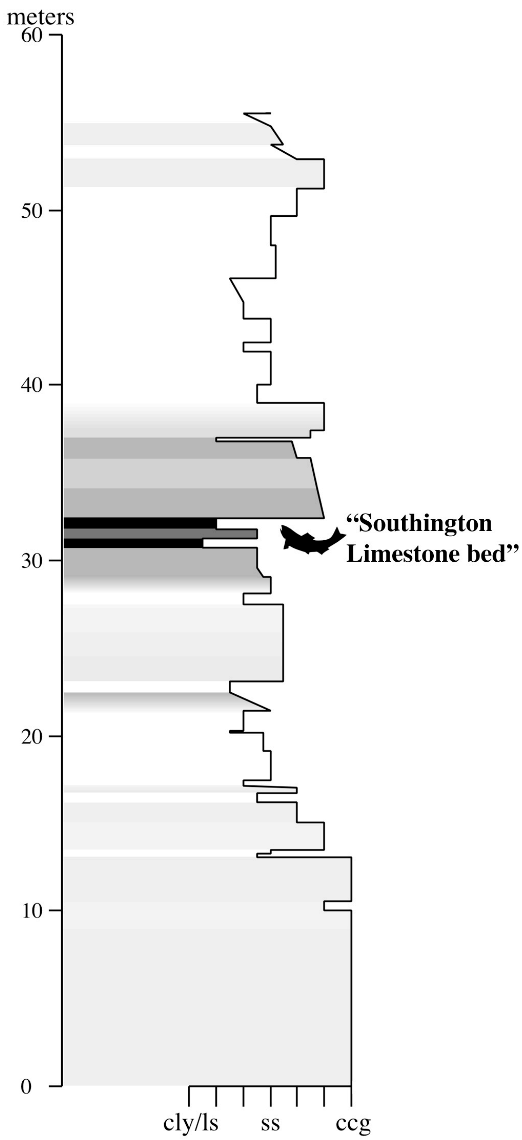


Figure 10. Measured section of upper New Haven and lower Shuttle Meadow formation, I 91 road cut at Mt. Tom, Stop 1. Only the top of the section of conglomeritic New Haven Fomation is shown.

Note: This figure was accidentally omitted by the authors

14.2 Long outcrop of Granby Tuff.
14.5 Pull off Interstate 91 on right into scenic area and park.

STOP 1b. GRANBY TUFF. (10 MINUTES) SW Mt. Holyoke Quadrangle, (approx.) 42°15.28' N, 072° 37.25' W; Tectonostratigraphic Sequence TS IV; Granby Tuff; Hettangian age, ~200 Ma. Main points are: local feeder dikes, eruptive locus of volcanoclastic unit equivalent to Hampden Basalt.

Cuts at and adjacent to a rest stop on the southbound side of Interstate Route 91 expose a few meters of the Granby Tuff. The Granby Tuff was named and described by Emerson (1891, 1898a;b; 1917) and described in more detail by Balk (1937). It is a black to maroon compact and stratified basaltic tuff or tuffaceous sandstone. It locally includes basaltic breccia and is intruded by linear dikes and sills of Blackrock diabase and “Mount Tom plug” (probably the equivalents of the Bridgeport dike). It overlies or is interbedded with flows of the Hampden basalt. The area was the site of fissure eruption for the Hampden basalt (Foose et al., 1968). There are no modern studies on this tuff, but clearly much of it is water reworked.

Adjacent to the feeder dike systems to the flows, such as the Hampden or the Talcott (Stop 6), basalt flows are relatively thin and restricted while volcanoclastics are abundant. This may reflect inflation by sills and local elevation of the land surface in a relatively broad area adjacent to the feeder system during eruption, an idea that will be discussed in more detail at Stop 6. A major difference between the eruptive styles of the feeder systems to the Hampden versus the Talcott is that pillow lavas are absent in the former but prominent in the latter. This is consistent with the emplacement of the Hampden during the drier phases of a 400 ky cycle and eruption of the Talcott during a wet phase of a 400 ky cycle.

Mileage: Time to next stop 7 min.
14.5 Resume driving north on route I 91 north.
14.7 Granby Tuff on right and left.
15.1 Diabase plug in Granby Tuff on right.
15.6 East Berlin and overlying Hampden Basalt on left described by Hubert et al. (1976). Overpass for road to Mt. Tom Ski area. This road has fine outcrops of East Berlin Formation described by Olsen et al. (1989) to the west.
17.2 Take Exit 17 on right for Route 141.
17.5 Turn right onto Route 141 West.
18.3 Turn left onto Southampton Road.
19.4 Passing Stop 2, *Clathropteris* locality.
19.5 Turn right onto Line Road.
19.6 Park on right.

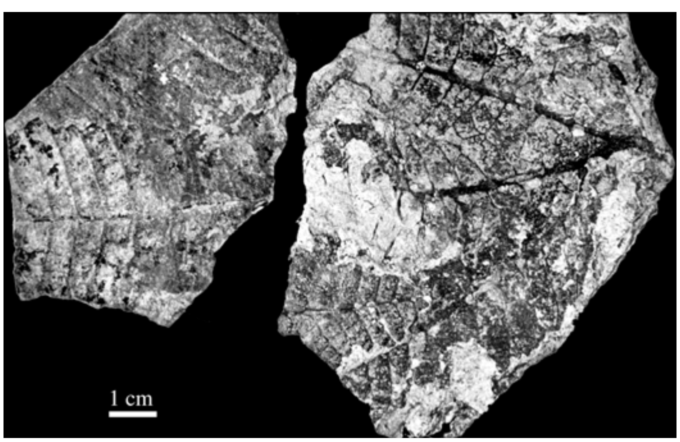


Figure 11. Examples of the fern Clathropteris from Stop 2. Specimens in the collection of the Virgina Museum of Natural History. Photo by Bruce Cornet.

STOP 2. "*CLATHROPTERIS* LOCALITY," HOLYOKE, MA. (45 MINUTES) North central Mount Tom Quadrangle, (approx.) 42° 12.98' N, 72° 39.50' W; Tectonostratigraphic Sequence TS IV; Upper New Haven Fm. or lower Shuttle Meadow Fm.; ?Rhaetian-Hettangian age, 200 Ma. Main points are: physical similarity to Triassic-Jurassic boundary in Newark basin; important fern (*Clathropteris*) macrofossil locality; very high spore counts – "fern spike"?; meaning for biostratonomy of fern spikes..

A small, stratigraphically isolated set of exposures along Southampton Road in Holyoke, MA (Fig. 11, 12) was discovered in the 1970s and has produced the best-preserved examples of the dipteridaceous fern *Clathropteris meniscoides* in

Figure 12. Southhampton Road (south side) *Clathropteris* locality, ca. 1974, Bruce Cornet for scale. Bed with macrofossils and spores at Cornet's feet.

the Newark Supergroup (Cornet, 1977; Cornet and Traverse, 1975) (Fig. 12). This locality bears at least a superficial physical similarity to the Triassic-Jurassic boundary in the Jacksonwald syncline of the Newark Basin.

The Jacksonwald syncline sequence consists of gray clastic rocks with an interbedded carbonaceous layer (comprised of *Clathropteris* and the horsetail *Equisetites*) overlying a white to gray claystone, containing predominately fern spores and *Corollina*. The Holyoke outcrop (Figs. 12, 13) consists of (from the bottom up) a few meters of red and brown sandstone and mudstone, followed by ~5 meters of gray and tan and yellow-weathering pebbly sandstone and siltstone. These coarser units are draped by a few cm of gray rooted clay, covered by a mat of *Clathropteris* and *Equisetities* in growth position (albeit compressed). The *Clathropteris* mat is overlain by tan and yellow-weathering fine conglomerate, pebbly sandstone, and siltstone.

It is clear from surrounding outcrops that this exposure lies below the Holyoke Basalt and is within the upper New Haven Formation or lower Shuttle Meadow Formation. As in Stop 1, the Talcott Formation is absent from this part of the northern Hartford basin. Lithologically, the sequence resembles the Shuttle Meadow although such lithologies begin in TS IV of the New Haven Formation (Stops 4b and 5). While this unit is conventionally mapped as Shuttle Meadow, the chronostratigraphic meaning is unclear. Cornet (1977) positions this section 148 m below the Holyoke basalt, however, the very irregular topography suggests to us the presence of a series of faults, making speculative any thickness estimate outside the existing exposure.

This section may represent another high-resolution Triassic-Jurassic boundary section. Supporting evidence from palynology and reptile footprint turnover is lacking due to stratigraphic isolation. Cornet (1977) retrieved large numbers of *Clathropteris* from this locality, and recovered specimens of the spore *Granulatisporites infirmus* from sporangia of fertile pinnae (Cornet and Traverse, 1975). The mat itself and the underlying claystone comprise a "fern spike" dominated by *Granulatisporites infirmus*, *Converrucosisporites cameronii* (which grades morphologically into the former), and *Corollina meyeriana* with subordinate amounts of *Dictyophyllidites paramuensteri*, *Dictyotriletes* sp., *Pilasporites allenii*, *Podocarpidites* sp., *Corollina torosus*, *Corollina murphyi*, *Corollina simplex*, *Circulina simplex*, *Cycadopites andrewsii*, and *Cycadopites* sp. (Cornet and Traverse, 1975). *Granulatisporites infirmus* is the dominant spore in the Jacksonwald syncline fern spike as well, increasing the likeness. There are no other similar sequences known from the entire Hartford basin.

Two hypotheses present themselves for the similarity between this exposure and the Triassic-Jurassic boundary in the Newark basin: 1) both sites record *Clathropteris* living within overall TS IV, syn-CAMP zone, with the fern spike at the Newark basin boundary section being just a fortuitous occurrence of *Clathropteris* in a boundary transition; 2) the distinctive suite of lithologies and the fern spike represent the boundary phenomena (Whiteside and Olsen, 2003).

The stratigraphic range of *C. meniscoides* stretches from the Late Triassic to the Middle Jurassic (Harris, 1931; Vakhrameev, 1991) with abundant samples from Shuttle Meadow Formation correlative strata. Although in the Newark basin, *C. meniscoides* occurs as the single pre-boundary plant macrofossil in the Perkasie member of the uppermost Passaic Formation (215 Ma), it is rare throughout the rest of the Jurassic Newark Supergroup. It is characteristic of the Rhaeto-Liassic in Greenland, Northern Europe, China and parts of Southeast Asia. It seems plausible that the abundance of *Clathropteris* and its spores at the Triassic-Jurassic boundary and in the lower 400 ky of the Jurassic reflects the earliest and most intense part of the Triassic-Jurassic super-greenhouse. While there is evidence for a Rhaeto-Liassic maximum, there is insufficient data for a global *C. meniscoides* increase specifically at the boundary.

Other data are clearly needed to corroborate this section as a Triassic-Jurassic boundary. Identifying magnetic polarity chron E23r below the *Clathropteris* unit as seen just below the Triassic-Jurassic boundary fern spike in the Newark basin would provide confirmation. We have processed one sample that, while normal, does indicate well-behaved magnetic behavior and thus an obtainable polarity stratigraphy. (D. V. Kent, pers. comm., 2002). Additionally, pollen-bearing horizons are needed below and above the putative "fern spike" to see if the spike lies at the palynological extinction level. Finally there should be an examination of the concentrations of platinum group elements and organic $\delta^{13}C$ for correlation with the marine sections (e.g. Ward et al., 2001). Such investigations may require modest coring at this locality.

Return to vehicles.

Mileage: Time to next stop 15 min.
19.6 Turn around and proceed south on Line Road.
19.7 Turn left on Southampton Road.
20.9 Turn right onto Route 141 East.
21.7 Left is entrance ramp for I91 South (leave out).
22.0 Keep right on Route 141 East.
22.2 Keep left on Route 141 East (Dwight Street).
23.2 Keep right on Route 141 East (Dwight Street).
24.0 Turn left onto Route 116 North (Main Street).
24.1 Main Street merges with Canal Street.
24.4 Turn left on North Bridge Street.
24.6 Turn left onto HWP Co., Private Way.
24.7 Park.

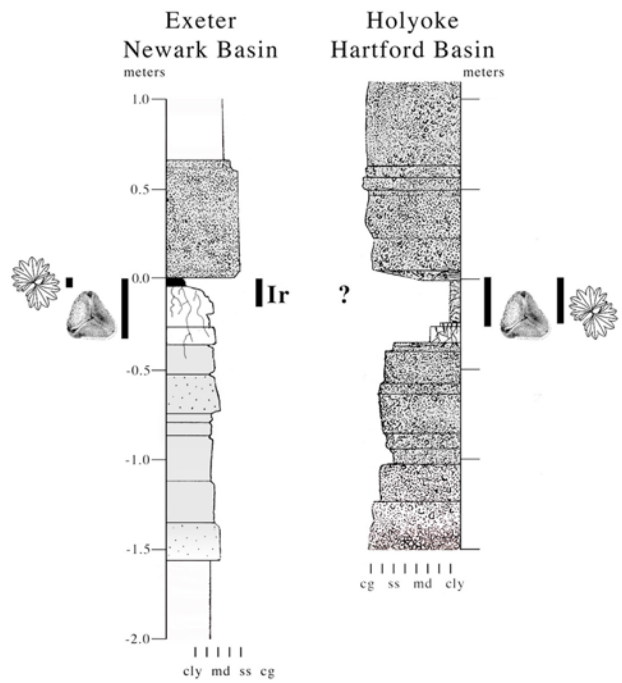


Figure 13. Comparison of the Triassic-Jurassic boundary in the Newark basin of southeastern Pennsylvania with the *Clathropteris* locality in the Hartford basin, Holyoke, Massachusetts. Vertical bars show distribution of: c, macrofossils of *Clathropteris*; g, the spore *Granulatisporites* that was produced by *Clathropteris*, Ir, iridium.

STOP 3. PORTLAND FORMATION AT HOLYOKE DAM, HOLYOKE, MA. (1 HOUR) Central Holyoke Quadrangle, 42° 12.678' N, 072° 36.070' W; Tectonostratigraphic Sequence TS IV; Middle part of the lacustrine part of the Portland Fm.; Late Hettangian age, ~199 Ma. Main points are: apparent recovery from super-greenhouse seen by appearance of large-leaved conifers; significant gray and black lacustrine sequence of "South Hadley Falls member" with abundant and well preserved conifer remains; abundant evaporite pseudomorphs. Note: do not get close to the water in the canal; it is swift and dangerous, and always seek permission from the Holyoke Gas and Electric Department prior to visiting.

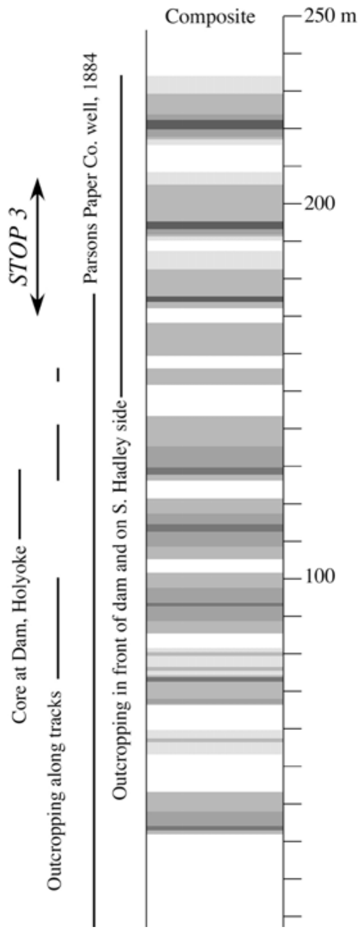
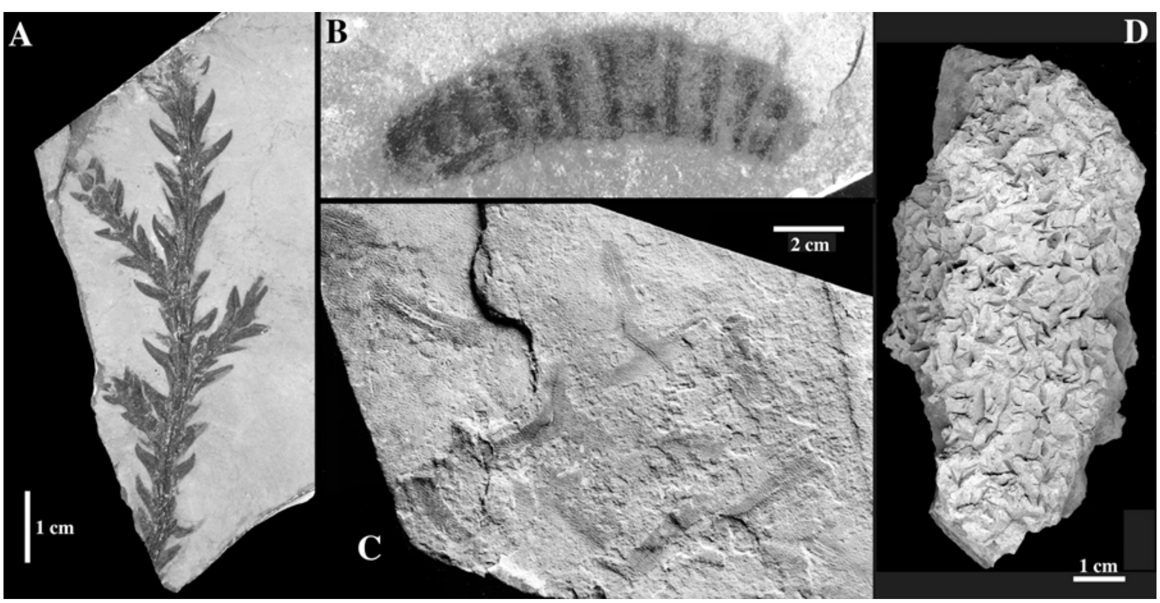


Figure 14. Composite section in vicinity of the Holyoke dam. Parsons Paper Co. well data from Emerson (1898). Rock color as in Fig. 5.

Sedimentary strata in the lacustrine sections of the Hartford basin contain abundant cheirolepidiaceous conifer remains. As outlined above, a very interesting trend in leaf size and cuticle morphology from the Triassic-Jurassic boundary exists through the Shuttle Meadow, East Berlin, and Portland Formations. Small-leafed conifers with thickened cuticle and papillate stomata are overwhelmingly dominant from the boundary itself (see Stop 6) through the "Park River member" of the Portland Formation. The "South Hadley Falls member" overlies the "Park River member" and contains abundant much larger-leaved conifers, a pattern that persists through the rest of the lacustrine Portland Formation. These outcrops and exposures have been described by Emerson (1898), Farquhar (1967), McDonald (1982), Parnell (1983), and Olsen et al., (1989) among others.

The "South Hadley Falls member" is the second complete McLaughlin cycle in the Portland Formation (Fig. 7). Its type section consists of the exposures and outcrops in the river bed and banks of the Connecticut River at South Hadley Falls in front of the Holyoke Dam this stop; now operated by the Holyoke Gas and Electric Department), exposures along the Conrail railroad cut in Holyoke, Massachusetts, augmented by a short core (DH-1, Holyoke Power and Light, Appendix 1), and records of a water well drilled for the Parsons Paper Company (Emerson, 1898b) (Fig. 14). Based on Milankovitch cyclostratigraphy (Fig. 7), this section is about 1.2 m.y. younger than the Triassic-Jurassic boundary. This McLaughlin cycle is unusual in TS IV of the Hartford basin because while it contains the highest proportion of gray and black strata, it lacks fish-bearing well-developed microlaminated units, and has the highest proportion of evaporite pseudomorph-bearing strata (Fig. 15d). As seen here, most Van Houten cycles have an unusually thick, gray division 2 but generally lack a microlaminated portion. Instead, most of division 2 tends to be dominated by thin-bedded mudstones with abundant conifer shoots, seed cones, and fragments. Fragmentary

Figure 15. (below) A, large-leafed conifer, south side of canal, Holyoke dam (N.G. McDonald collection); B, unidentified larva, collected by B. K. Emerson in 1901 near Holyoke, Massachusetts; C, Unidentified larvae, downstream from the Holyoke dam, Portland Formation, Holyoke, Massachusetts; D, evaporite pseudomorphs after a ?sulfate, railroad cut near dam, Holyoke. B and C from Huber et al., 2003.



fish and rare insects are also present (Fig. 15b, c). Divisions 1 and 3 of these cycles tend to be thin and silty with relatively uncommon desiccation cracks and footprints. Evaporite pseudomorphs, (vanished halite and lenticular gypsum crystals or glauberite and halite) are abundant in many layers (Parnell, 1983), especially in the transition between division 2 and 3, sometimes constituting more than 50% of a bed by volume (Fig. 15d). Because of the relatively thick, gray division 2 in many Van Houten cycles, the short modulating cycles are relatively subdued compared to other parts of the Hartford Jurassic section. In general, the lacustrine section appears to have been deposited under a cyclical climate regime with muted precessional fluctuations, but in an overall more arid milieu. This is consistent with deposition of the McLaughlin cycle during a period of low eccentricity in the 1.75 m.y. modulating cycle (Fig. 5).

While sedimentology of the "South Hadley Falls member" gray beds suggest relatively arid conditions, the morphology of the cheirolepidiaceous conifer assemblages suggest more humid conditions. Cornet (1977 and in Olsen et al., 1989) assigned conifer assemblages from this locality to assemblages C and D; the conifers are characterized by relatively large leaves and mostly thin smooth cuticle (Fig. 8). These large leaves contrast dramatically with the underlying Jurassic assemblages B and C typified by predominately small leaves often with papillate sunken stomata (Fig.8). Relatively large leaved conifers with smooth cuticles and normal unsunken stomata in Cornet's assemblage A are found in Late Triassic strata in both the Newark and Hartford basin. Very small leaves and cuticle with papillate stomata are usually considered morphological adaptations to high insolation and dry air, implying that by the time of deposition of the lacustrine strata at Stop 4, the environmental stresses had lessened and returned to, if not surpassed, Triassic levels. Very similar leaf adaptations are seen in conifers from younger strata of the Portland as well. If the thermal damage hypothesis (McElwain et al. 1999) is correct that the small leaves seen in Greenland are a response to the stresses associated with an elevated CO_2-triggered super-greenhouse beginning at the Triassic-Jurassic boundary, then the leaf adaptations seen at this stop should reflect the recovery from that super-greenhouse. Interestingly, the morphological adaptations seen in these younger Portland strata conflict with the sedimentological evidence as well as the increasing aridity caused by the northward drift of tropical Pangea (Kent et al, 1995) throughout the Late Triassic and Early Jurassic (e.g. Olsen and Kent, 1996; 2000).

Return to vehicles.

Mileage: Time to next stop 40 min.
24.7 Turn vehicles around and proceed east along HWP Co., Private Way.
24.8 Turn right onto Bridge Street.
24.9 Turn Right onto MA Route 116 South (Canal Street).
24.9 Turn Right onto Lyman Street.
25.4 St. Colby Drive. Access to Conrail RR cut with excellent exposures of the gray and black parts of the "South Hadley Falls member" is immediately to north.
25.6 Merge right on to US Route 202 North, get left and follow circle around to left to Route 202 South.
25.8 Turn right onto Lincoln Street.
26.1 Turn left on Hampden Street (MA route 141 west).
27.0 Turn left onto ramp for I 91 South.
52.3 Take Exit 40 for Bradley International Airport for CT Route 20 West
55.4 Take right exit for CT Route 20 West
58.1 Turn left onto CT Route 187 South (South Main Street)
60.1 Turn right onto CT Route 540 (Hatchet Hill Road)
60.8 Hill is underlain by Holyoke Basalt.
61.6 Turn left onto CT Route 189 South.
62.7 Cross Farmington River.
57.8 Pass through light and park on shoulder on right.

STOP 4. TARIFFVILLE GORGE: NEW HAVEN FORMATION, TALCOTT FORMATION, AND LUNCH. SE Tariffville Quadrangle, (approx.) 41° 54.41' N, 072° 45.57' W; Tectonostratigraphic Sequence ?TS III and TS IV; New Haven Fm., Talcott Formation, Shuttle Meadow Fm., East Berlin Fm., Holyoke Basalt; ?Rhaetian-Hettangian age, ~200 Ma. Main points are: Triassic-Jurassic boundary section; complete sequence of basalt formations; hanging wall onlap of abbreviated section of Shuttle Meadow Formation.

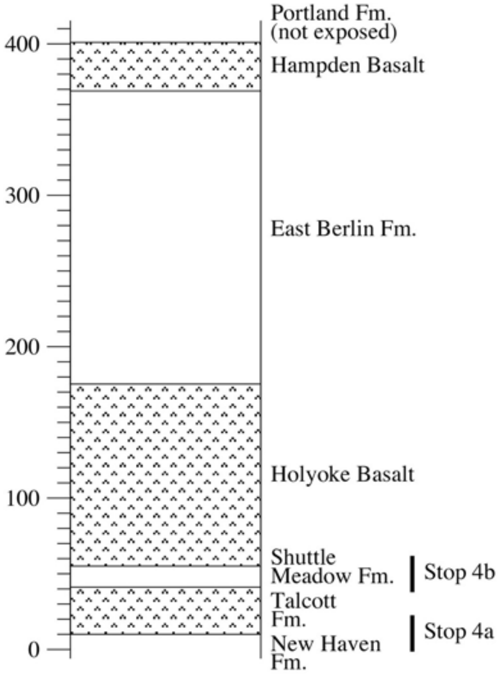


Figure 16. Section along Route 189 and Farmington River in the vicinity of Tariffville, Connecticut.

Outcrops and exposures along the gorge of the Farmington River at Tariffville, CT reveal a nearly complete section of the upper New Haven Formation through Hampden Basalt (Figs. 16, 17). This section is of considerable historical interest because Rice (1886) and Davis (1898) used observations from these outcrops to argue for an extrusive origin of the basalt flows rather than the generally assumed intrusive origin (e.g. Dana, 1874). This section has also been described in some detail by Gray (1982, 1987), Philpotts and Asher (1992), and Philpotts and McHone (this volume).

We have divided this stop into two segments (Stops 4a and 4b) on opposite sides of the river because of accessibility. At Stop 4a on the south side of the river we will look at exposures of the New Haven Formation and the Talcott Formation, and at Stop 4b we will look mostly at the Talcott Formation, Shuttle Meadow Formation, and Holyoke Basalt.

STOP 4a. CUT ON ROUTE 189, SOUTH SIDE OF THE TARIFFVILLE GORGE (20 minutes).

About 8 m of New Haven Formation and nearly all of the Talcott Formation are exposed in a cut for CT Route 189. We will begin at the north end of the exposure on the west side of the road in the New Haven Formation and walk south through most of the Talcott Formation. At this locality the uppermost New Haven Formation (Fig. 17) is entirely red. The lower 6.5 m of the section consists of heavily bioturbated red sandy mudstone and sandstone entirely consistent with the bulk of the New Haven Formation belonging to TS III. Above that, there are numerous sandstone beds with clay drapes with much less bioturbation typical of footprint-bearing facies in the Shuttle Meadow through Portland formations (although we have yet to find any footprints here). On this lithological criterion we hypothesize that the upper 1.5 m of the New Haven Formation belongs to TS IV. The Triassic-Jurassic boundary either lies within this TS IV section or is within a hiatus at the TS III – TS IV contact that should be at about 6.5 m in the section. It should be possible to discriminate between these hypotheses with magnetostratigraphy.

Nearly the entire thickness (about 30 m) of the Talcott Formation is exposed at this locality (Fig. 16) where it appears to consist of two flows. The Talcott Formation is a HTQ-type basalt (high titanium, quartz normative basalt) identical in composition to the oldest flows in the Newark, Gettysburg, and Culpeper basins (Puffer et al., 1981) and apparently derived from magma supplied by the Fairhaven dike system (Philpotts and Martello, 1986). As described by Gray (1982, 1987) and Philpotts and Asher (1992) these exposures show the lowest 1.5 m of Talcott lying on unmetamorphosed New Haven Formation, with some pillows having sunk into the underlying sediment. This is followed by about 6 m of columnar basalt and then massive basalt near the top of the section where the basalt becomes vesicular passing upward into a flow-top breccia. An additional 15 m of vesicular basalt flow lobe units is exposed discontinuously along the road

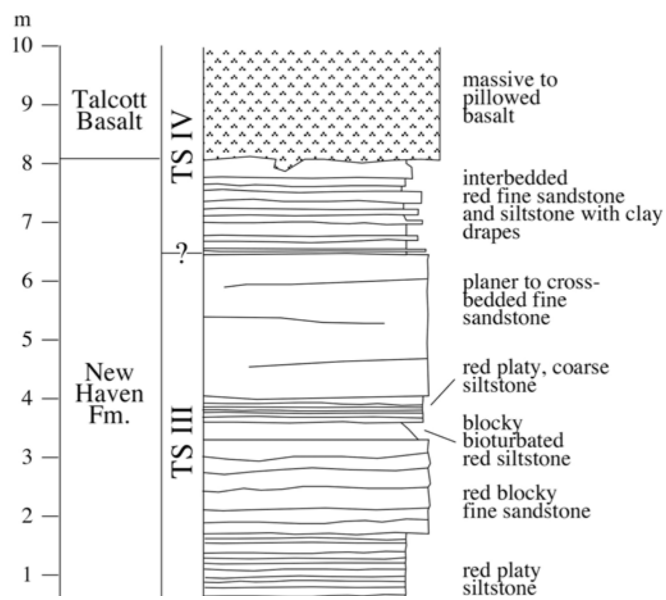


Figure 17. Section of New Haven and Talcott formations on route 189, Tariffville, Connecticut, Stop 4a.

cut and well exposed along the river (Philpotts and Asher, 1992). The presence of the pillows indicates some standing water at the time the flow system transgressed the sediment surface.

There is a 0.5 m basalt breccia-filled fissure with an arkosic matrix in the middle of the road cut (most obvious on the north side) (Gray, 1987), originally observed by Rice (1886) and Davis (1898). As noted by all three authors, this fissure probably filled with sedimentary material from above. Such fissures are classically called Neptunian or clastic dikes and they are locally quite common in CAMP flows. Neptunian dikes were described in most detail by Schlische and Ackerman (1995) from the North Mountain Basalt of the Fundy basin in Nova Scotia, a correlative of the Talcott. They can be very useful for determining the local syndepositional state of stress and occasionally have fossils within them.

Return to vehicles.

Mileage: Time to next stop 9 min.
57.8 Proceed SE on Route 189.
58.4 Passing long cut of Holyoke Basalt with characteristic splintery fracture.
58.8 Limited outcrops of cyclical East Berlin Formation.
59.2 Keep right on 189.
59.3 Exit right for Tariffville Road.
59.5 Turn left onto Tariffville Road.
59.8 Turn right onto entrance ramp for Route 189 North.
60.1 Keep left onto entrance ramp for Route 187 North.
60.7 Cross Farmington River.
61.0 Turn right onto access road for Spoonville Road.
61.0 Turn right onto Spoonville Road.
61.3 Turn right onto Tunxis Avenue.
62.1 Park.

STOP 4B. TARRIFVILLE GORGE: TALCOTT FORMATION, AND REST OF EXTRUSIVE ZONE AND LUNCH (1 HOUR).

Proceed northwest from the parking area and follow the path down to the river at the base of the old bridge abutment to the base of a section exposing the upper surface of the Talcott Formation and nearly the entire Shuttle Meadow Formation (Fig. 16). Rice (1886) described the vesicular basalt and its contact with the overlying sedimentary rock of the Shuttle Meadow Formation noting the conglomerate at the base of the formation that contains abundant well-rounded basalt clasts, followed upward by unmetamorphosed mudstones. Rice stated, “These phenomena seem to lead irresistibly to the conclusion that the lower sheet of trap is contemporaneous.” In other words, the igneous unit is a surface flow, not an intrusion, a very important conclusion for the time.

Follow the path just along the water’s edge about 70 m northwest to an outcrop of the Shuttle Meadow Formation and the overlying Holyoke Basalt. A continuous section of the Shuttle Meadow cannot be seen at this locale, but one is visible on the opposite side of the river. The most striking aspect of this section is the lack of any indication of deeper water lacustrine units (Durham Member, Fig. 9) and the small thickness (about 14 m) of the formation as a whole. This section probably represents only the uppermost part of the formation, thinning by progressive hanging wall onlap because the formation is in excess of 100 m in most of the rest of the basin (Stops 1, 2, 5b, 6) and the sedimentary cyclicity is otherwise laterally persistent. This implies a considerable hiatus (~200 ky) and unconformity between the Talcott Formation and the Shuttle Meadow Formation at this locality. The observations suggest that significant tilting of the basalt occurred after eruption of the basalt, during (or less plausibly, before) deposition of the Shuttle Meadow. The lakes depositing the fish-bearing Van Houten cycles of the Durham Member could very well have extended over this area, but the sedimentary record of them was eroded during lake low stands. This model is entirely consistent with the overall model for the production of the major tectonostratigraphic sequence boundaries and the Schlische cycle represented by TS IV. Thus, we envision accelerated tilting occurring during the TS III – TS IV transition over a few hundred thousand years associated with the eruption of the CAMP basalts, producing a series of unconformities (such as at this stop) in addition to the major one at the TS III – TS IV boundary itself. However, a correlative conformity is present at other places deeper into the basin, such as at Silver Ridge (Stops 5b and 6). We will discuss more evidence for this pattern at Stops 4b, 5, and 7.

Much of the upper part of this section consists of a laminated purplish siltstone with mudcracks. Judging from its lateral continuity along this shore and the opposite, this interval is lacustrine in origin and probably represents an expression of some of the wetter depositional environments at the top of the Shuttle Meadow Formation. Gray (1987) describes these units as being mostly microcrystaline albite and chlorite with the sodium coming from waters percolating downward from the cooling overlying Holyoke Basalt.

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The Holyoke Basalt is an HFQ-type (high-iron quartz normative) basalt, essentially identical in chemistry to the lower two flows of the Preakness Basalt of the Newark basin (Puffer et al., 1981). The massive basal few meters of the flow can be seen above the sedimentary section, but higher, especially along Tunxis Avenue (see below) the flow has a characteristic splintery jointing. In southern Connecticut there are two flows in the Holyoke Basalt. The lower pinches out just north of Farmington (Gray, 1987); the flow outcropping here is the upper flow, which always has a splintery fracture. Interestingly this jointing style is also characteristic of the lower flow of the Preakness Basalt. Prevot and McWilliams (1989) identified a magnetic excursion in the lower part of the Holyoke Basalt here at Tariffville (on Route 187). The same excursion is recorded in the lowest flow of the Preakness Basalt of the Newark basin, characterized by the same kind of splintery fracture, and in the single flow of the Deerfield Basalt. The presence of this excursion virtually assures that these flows were erupted simultaneously. It also implies that the lower flow of the Holyoke Basalt does not have a representative in the Newark or Deerfield Basin.

Return to vehicles.

Mileage: Time to next stop 33 min.

62.1 Turn around and head back SE on Tunxis Avenue.

62.2 Passing exposures of splintery Holyoke Basalt on northwest. These exposures are the subject of Stop 5 of Philpotts and McHone (this volume).

62.4 Passing more massive Holyoke Basalt on northwest. Coarse-grained segregation veins present (Philpotts and McHone,this volume).

62.5 Upper vesicular zone of Holyoke basalt outcrops at river level (Gray, 1987).

62.7 Exposures of middle East Berlin Formation on northwest consisting of red mudstones and sandstone grading up to gray. This marks the base of one of the well-developed black-shale-bearing Van Houten cycles that characterizes the formation (Hubert et al., 1976, 1978; Gray, 1987). The cyclostratigraphy of the East Berlin is nearly identical to the correlative Towaco Formation of the Newark basin (Olsen et al., 1996b).

62.9 Uppermost East Berlin Formation on northwest and Hampden Basalt, a HFTQ-type (high-iron high-titanium quartz normative basalt) (Puffer et al., 1981) virtually chemically the same as the Hook Mountain Basalt of the Newark basin. Only a single flow is evident here, but two flows may be present further south (Gray, 1987).

63.0 Turn left on Spoonville Road.

63.3 Turn left on access road to Route 187.

63.3 Turn left onto Route 187 South.

63.6 Crossing Farmington River.

63.8 Take ramp onto Route 187 South.

64.8 Stay on Route 187 East.

67.6 Turn left onto CT Route 305 (Old Windsor Road/Bloomfield Road).

70.0 Take entrance ramp for Interstate 91 South.

85.2 Take ramp for Exit 23 (West Street, Rocky Hill).

85.5 Turn left onto West Street.

86.4 Turn right into entrance road for Dinosaur State Park.

86.5 Park.

STOP 5a. DINOSAUR STATE PARK, ROCKY HILL, CT. (30 MINUTES) SE Hartford South Quadrangle, (approx.) 41°39.03' N, 72°36.48' W; Tectonostratigraphic sequence TS IV; East Berlin Fm.; Hettangian age, 200 Ma. Main points are: Abundant large theropod dinosaur tracks (*Eubrontes*) in regressive portion of upper gray and black Van Houten cycle of East Berlin; extreme rarity of herbivores; post boundary ecological release; no evidence for herding in theropod dinosaurs.

This site was discovered in 1966 during excavation for the foundation of a state building. Exposures here have revealed nearly 2000 reptile tracks, most of which have been buried for preservation and future exhibition. The present geodesic building at the park houses approximately 500 tracks (Fig.18). The tracks are found in the gray arkoses, siltstones, and mudstones of the East Berlin Formation, about 20 m below the contact with the Hampden Basalt. Ripple marks, raindrop impressions, mudcracks, and the footprints indicate shallow-water conditions and some subareal exposure. The track-bearing beds grade upward into gray mudstone and then red sandstone and mudstone. These strata are part of the uppermost gray cycle in the East Berlin Formation and they are clearly part of the regressive phase of that cycle.

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

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

STOP 5B. HARTFORD BASIN CORES, ON DISPLAY IN THE DINOSAUR STATE PARK AUDITORIUM (30 MINUTES). Cores from NE Meriden Quadrangle, (approx.) 41°35.1' N, 072°45.5’W (see Stop 6); Tectonostratigraphic sequence TS IV; New Haven Fm., Talcott Formation, and Shuttle Meadow Fm.; Hettangian age, 200 Ma. Main points are: cores with continuous sections; TS III – TS IV minor unconformity; gray earliest Jurassic section in top of New Haven Formation; pillowed Talcott Fm. and volcanoclastic member with spherules; cyclical lower Shuttle Meadow Fm. with fish-bearing limestones.

Cores B-2 and B-3 (41°35.053’N, 072°45.647’W; 41°35.010’N, 072°45.655’W). These two cores sample the lower Talcott Formation and the upper New Haven Formation and hence the Triassic Jurassic and TS III – TS IV boundaries (Fig. 19). Core B-2 is the more extensive sampling 44.5 m (146 ft) of Talcott Formation and 58.8 m (193 ft) of New Haven Formation, while Core B-3 has only 1 m (3.3 ft) of Talcott and 3.6 m (11.7 ft) of New Haven.

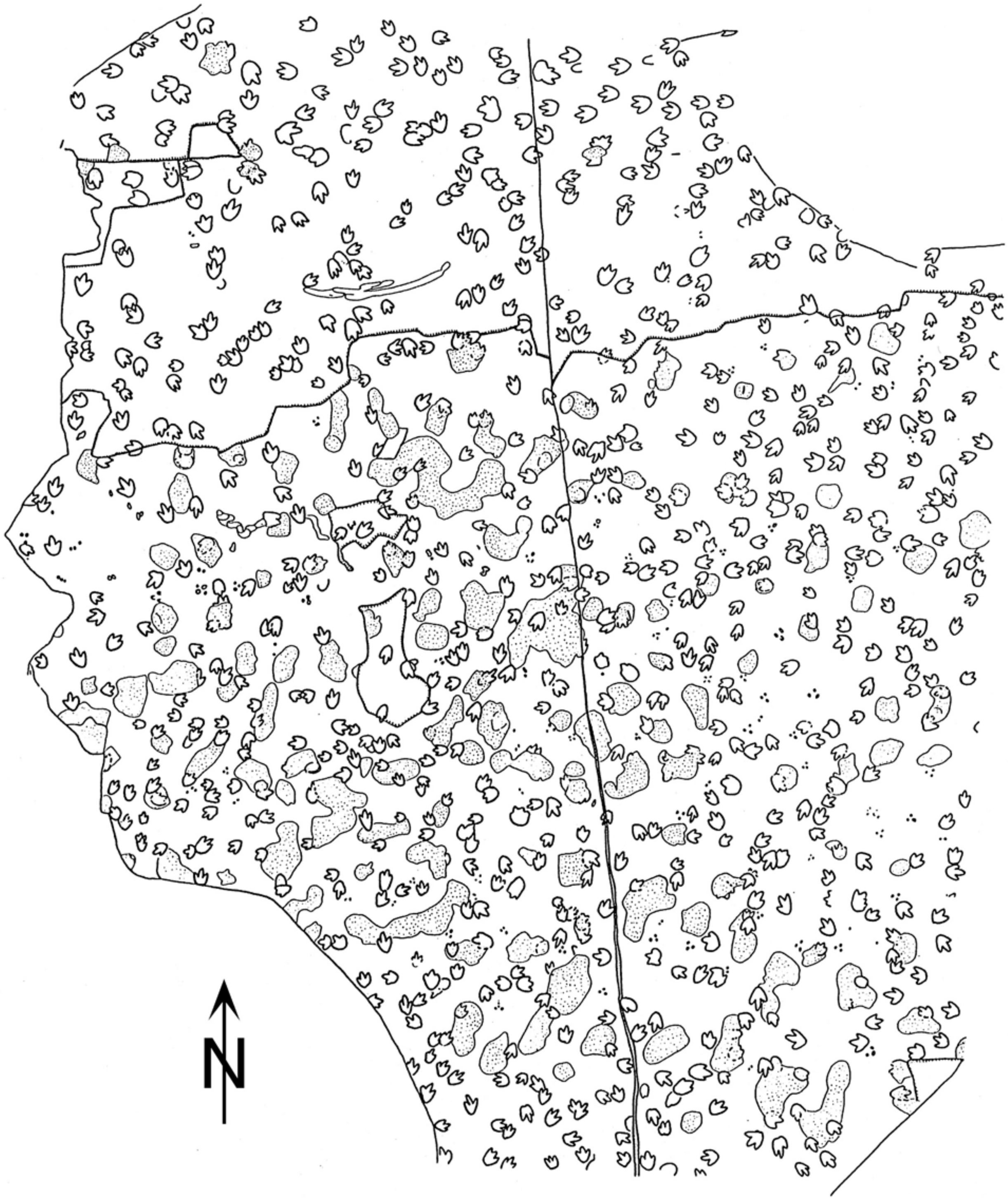

Figure 18. Dinosaur footprints on display *in situ* at Dinosaur State Park (from Farlow and Galton, 2003). Most footprints are 30-40 cm long. Stipple illustrates where overlying rock did not separate cleanly from the upper track-bearing layer, cross-hatched lines indicate boundary between the upper and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

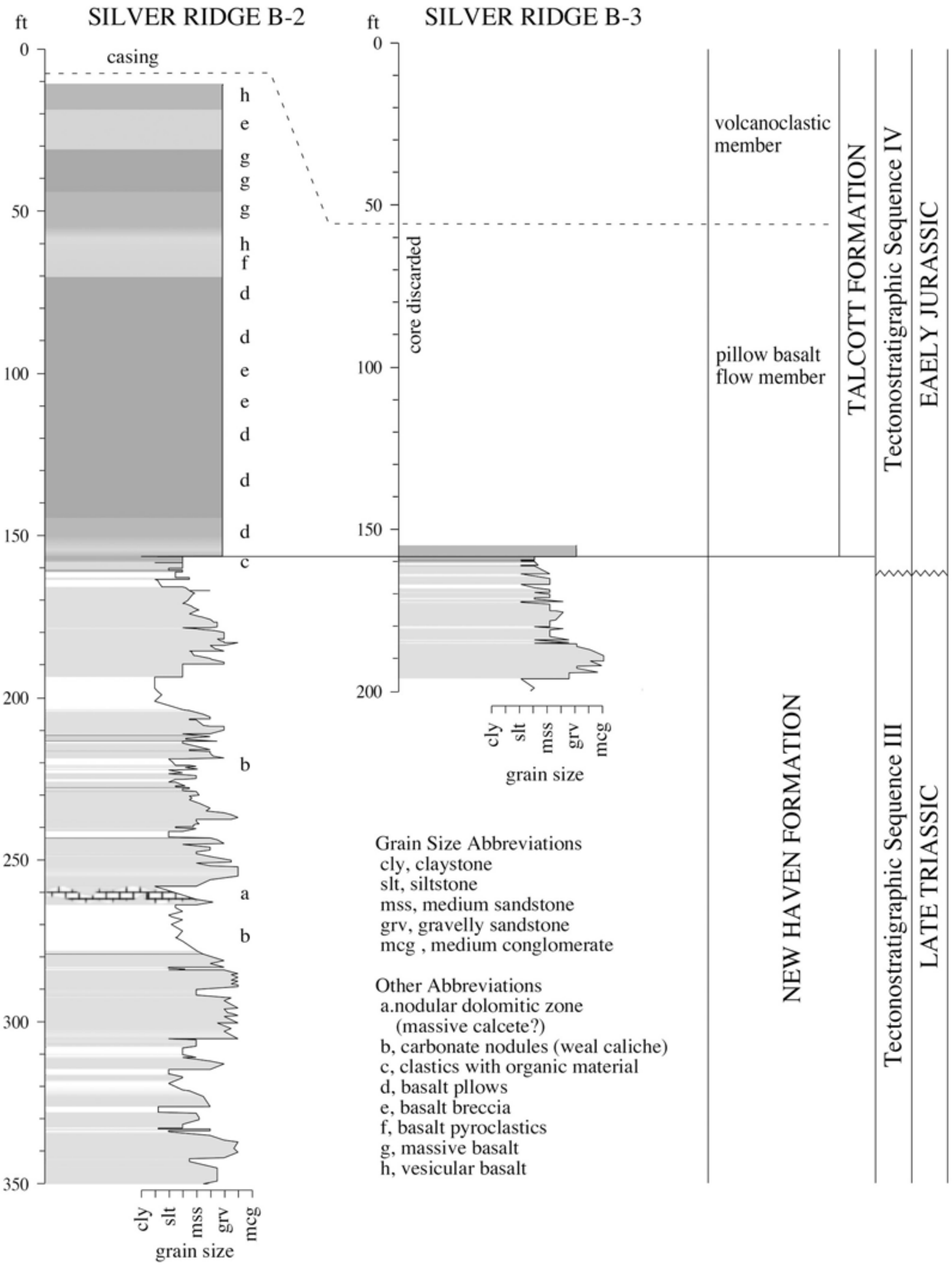

Figure 19. Cores B-2 and B-3 from Silver Ridge (Stop 6) on display at Stop 5b.

In core B-2 the lower 56.5 m (below 164.5 ft) of typical New Haven Formation is characteristic of the alternating "Red stone" and Lamentation facies of Krynine (1950) typical of the Meriden area (Fig. 19). The "Redstone" facies consists of a red micaceous feldspathic sandstone and interbedded bioturbated mudstones while the Lamentation facies is comprised of often conglomeritic coarse gray or purplish white arkose. To the east, the Lamentation facies predominates; to the west the "Redstone" facies is dominant (Krynine, 1950). Both facies tend to be very heavily bioturbated with roots and the burrow *Scoyenia,* and consequently there is very little preservation of ripples and other fine sedimentary structures.

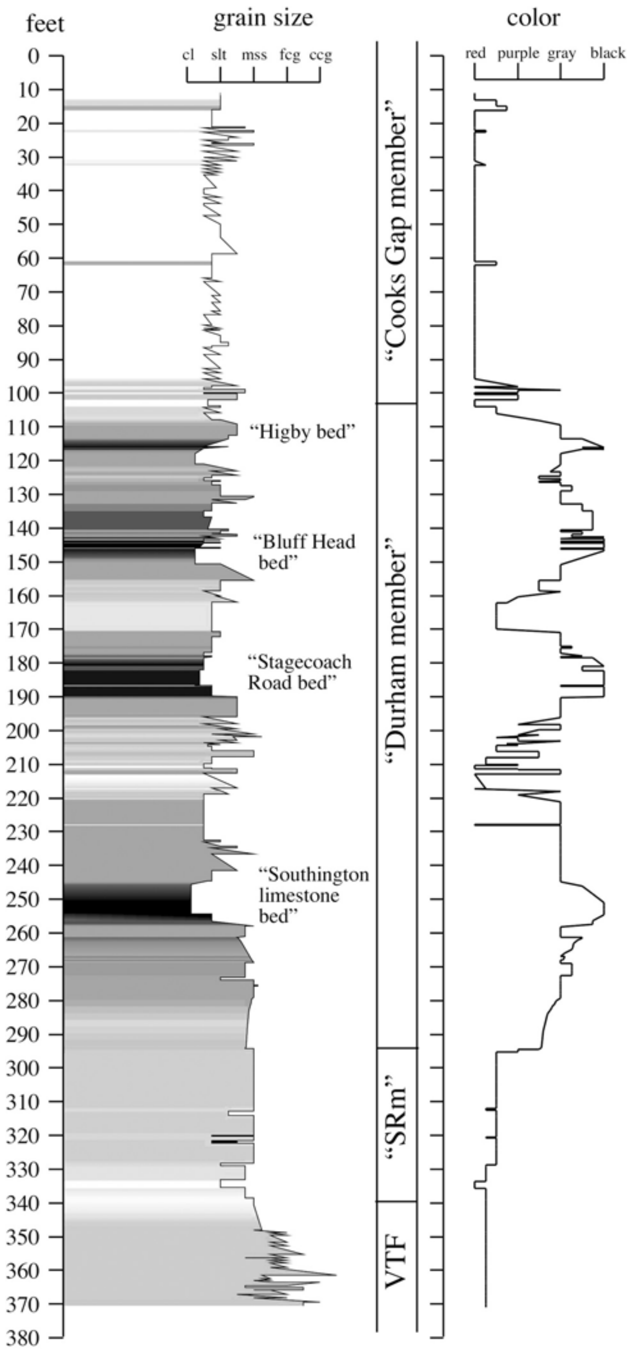

Figure 20. Silver Ridge Core B-1 (Stops 5b and 6)

Above the depth of 164.5 ft in core B-2 and 166.9 ft in core B-3, the New Haven Formation changes abruptly in facies to finely interbedded, red and tan fine sandstone that grades upward into gray fine sandstone and minor mudstone with abundant plant fragments, including conifer shoots. This facies is much more similar to the overlying Shuttle Meadow Formation in the preservation of small-scale sedimentary structures, the reduction of bioturbation, and the preservation of organic matter. At the time of writing we have not extracted pollen from these gray beds, however, a similar sequence of conifer-bearing gray beds occurs just below the Talcott Formation and these have produced a palynoflora of earliest Jurassic aspect, very strongly dominated by *Corollina* (Robbins, quoted in Heilman, 1987), and therefore we believe the gray beds to be earliest Jurassic in age. Kent (pers. comm., 2002) found the New Haven Formation in the B-2 core to be entirely of normal polarity. If the gray beds are indeed of Jurassic age, then the thin reversed polarity zone E23r that lies just below the Triassic-Jurassic boundary in the Newark basin is likely missing. The abrupt facies change at 164.5 ft and 166.9 ft in cores B-2 and B-3, respectively, is thus most easily accounted for by a minor hiatus at the TS III – TS IV boundary. In this interpretation, the Triassic-Jurassic boundary itself is missing as well.

The contact with the overlying Talcott Formation at 156 ft in core B-2 is sharp, but there is some deformation of the uppermost New Haven Formation. From 156 ft to 70 ft, the Talcott is pillowed and brecciated, resembling the exposures at Stops 6b and 8. Radial pipe vesicles are evident as are the chilled margins of pillows. From 70 ft to 32 ft there is massive locally vesicular basalt. This is overlain by a breccia unit to 19 ft, which itself is overlain by vesicular basalt to the top of the core (11 ft). It is impossible to tell whether the contacts at 19 ft and 70 ft are contacts between different major flows or contacts between lobes of one eruption. We favor the latter interpretation based on the geometry visible at Stop 8. Although only slightly more than a meter of the basal Talcott (and the contact with the New Haven Formation) was recovered in Core B-3, the contact is less distorted and the basalt does not appear to be part of a pillow.

Core B-1 (41°35.102'N, 072°45.388'W. This core penetrates nearly the entire lower half of the Shuttle Meadow Formation as well as most of the volcanoclastic equivalent of the Talcott Formation (Fig. 20). Core B-1 reveals for the first time the stratigraphy of the lower, mostly gray and black, highly fossiliferous portion of the Shuttle Meadow Formation that we call the "Durham member" after the famous outcrops that have produced fossil fish for over a century. The "Durham member" is exposed at many locations in the Hartford basin, but no section reveals the complete succession of beds. We also provide informal names to the distinctive beds that comprise the division 2 of each of the three pronounced Van Houten cycles in this member.

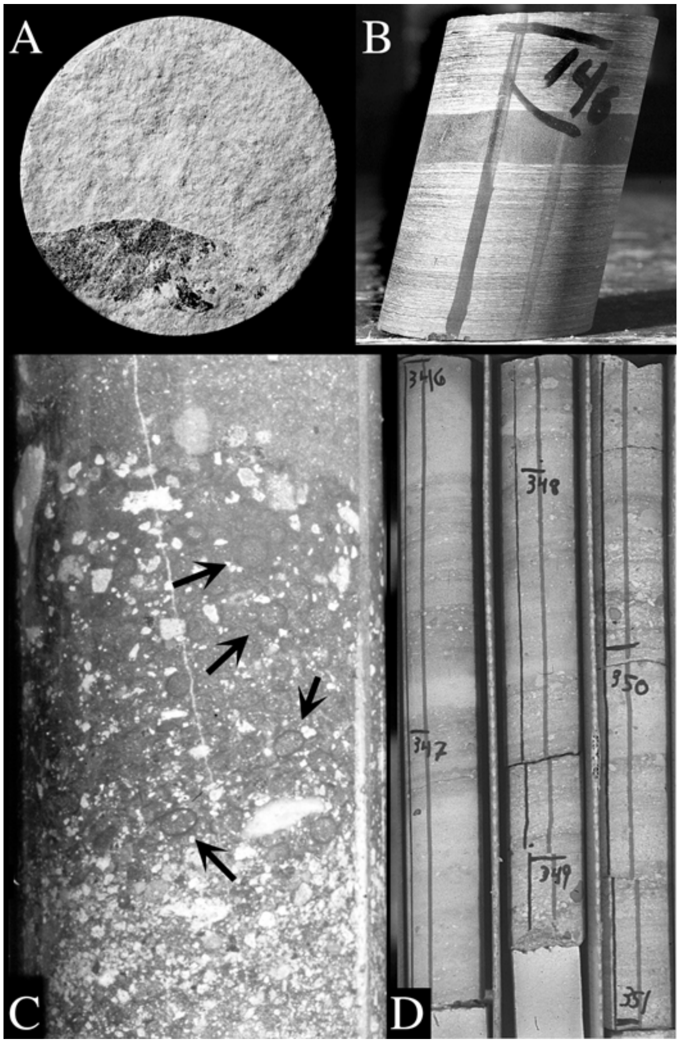

Figure 21. Silver Ridge B-1 core: A, Fish (*Redfieldius*) at 146 ft; B, lateral view of microlaminated calcareous mudstone with fish ("Bluff Head bed") at 146 ft – note small turbidite; C, spherules from 349 ft. (examples at arrows); D, typical Talcott volcanoclastics (footage on core).

The lowest Van Houten cycle has a distinctive black and dark gray division 2 bearing a limestone with rather thick and distinctive laminae with a pelleted or flocculant appearance (254.6-251.0 ft.). This bed, which we call the "Southington limestone bed," outcrops at a variety of places, most importantly at small long abandoned quarries to the south in Southington, Connecticut (Davis, 1898) and the section described by Krynine (1950, p. 61) near the Shuttle Meadow Reservoir. It is also presumably the limestone bed exposed at Stop 1. Rather poorly preserved (partly dephosphatized), but articulated fish are present in this core (e.g., at 252.7 ft, Fig. 21) and adjacent temporary exposures (Stop 6). We believe that the "Southington limestone bed" is the lateral equivalent of the Coe Quarry limestone near Northford Connecticut described by Krynine (1950), Mooney (1979), Steinen et al. (1987) and DeWet et al. (2002).

The second Van Houten cycle lacks a well-developed limestone bed but does have a division 2 with significant black shale (186.7-175.0 ft, Fig. 9). This shale is distinctive in that there are many silty interbeds with some of them bearing reptile footprints (as seen in adjacent exposures, Stop 6). We correlate this interval with the lower fish-bearing black shale (the "crinkle bed") outcropping at the Durham fish locality and name it the "Stagecoach Road bed" after the road immediately east of the stream on which the latter locality is located.

The third Van Houten cycle has two distinctive dark gray to black intervals (Fig. 9). The lower of these two beds (146.5 – 132.8 ft) is unique in that it is highly calcareous, finely microlaminated and contains abundant fish and coprolites (Fig. 21). This bed is clearly the same as the main fish-bearing unit outcropping at the famous Durham and Bluff Head fish localities (McDonald, 1992), and we thus call it the "Bluff Head bed," *Semionotus*, *Redfieldius*, *Ptycholepis* and *Diplurus longicaudatus* have been recovered from this unit (McDonald, 1992) and include some of the best preserved fish in the Newark Supergroup (Fig. 23). The "Bluff Head bed" also provides the best example of a *Semionotus* species flock in the Hartford basin. This species flock and those of the East Berlin, and lower Portland Formation ("Park River member"), akin to their correlates in other Newark basins, are characterized by morphologically similar species of *Semionotus* (also Stops 1 and 6), analogous to species flocks of cichlid fishes of East African great lakes (Olsen, 1980; McCune et al., 1984; Olsen and McCune, 1991; McCune, 1996). Perhaps the evolution of the species flocks was fostered by the extremely low post-Triassic-Jurassic boundary fish diversity in the watersheds of the rift basins, because such species flocks do not seem to occur in Triassic age lacustrine deposits in the same rifts.

The upper of the two dark gray to black beds (118 – 113.6 ft) is not clearly microlaminated, nor as calcareous. It apparently correlates with a black shale outcropping along Highland Brook on the East face of Higby Mountain, and we term this bed the "Higby bed". At the later locality the "Higby bed" produces articulated fish and conchostracans in association with abundant *Anchisauropus*, *Eubrontes*, and *Batrachopus* that occur in bracketing beds of gray siltstone.

The stratigraphy of the Van Houten cycles of the lower Shuttle Meadow Formation ("Durham member") is surprisingly complex; it is doubtful that we could have ever figured out the stratigraphic relationships of the numerous dark shale localities in this formation without this core. Equally surprising, however, was the realization of the profound similarity between the stratigraphy of the Silver Ridge B-1 core and the stratigraphy of the homotaxial Feltville Formation of the Newark basin, known from multiple cores (Olsen et al., 1996a) (Fig.22). This similarity has forced us to reconsider the cyclostratigraphic conclusions of Olsen et al. (1996b, 2003b) that there are only two black shale- and carbonate-bearing Van Houten cycles in the lower Feltville Formation. There are instead three Van Houten cycles and 20 ky must be added to the total duration of the CAMP basalt flows, thus totaling an even 600 ky.

The carbonate rich nature of the "Durham member" is distinctive compared to most of the rest of the Jurassic in the Newark Supergroup, except for other units directly overlying the initial (HTQ) flows of the CAMP, and in the case of only the Hartford basin, directly overlying the Holyoke basalt. Surprisingly there is a strong cyclostratigraphic similarity between the sections above these initial flows over not just eastern North America, but also in Morocco (Olsen et al., 2003b). In central Morocco, laminated, often black limestones still characterize division 2 of the cycles, but bedded evaporites dominate division 3 of the cycles and a second sequence of HTQ flows overly the cyclical sequence, which, because of the cyclostratigraphic homotaxiality we correlate to the "Durham member." A similar stratigraphy of two HTQ flow sequences with two sedimentary cycles is also present in eastern Morocco; however, there the entire interbed is limestone-dominated. Based on homotaxiality we believe that the dry phases of the cycles are represented by bedded to massive limestones and the wetter phases by finely laminated black limestones. No determinable animal fossils have been found in the finely-laminated units, but abundant and diverse serpulid-bearing bivalves are present in the more massive beds. Although the bivalves have not yet been systematically studied, Olsen et al., (2002d) propose that these are consistent with a marine environment, possible part of the pre-planorbis zone of the European early Mesozoic.

The apparent relationship between the initial HTQ CAMP flows, the underlying Triassic-Jurassic boundary and the overlying carbonate rich sequence suggest to us two possible interpretations (Olsen et al., 2003c): 1) The limestone deposition is a response to a super-greenhouse effect caused by a asteroid or comet impact at the Triassic Jurassic boundary, or the massive basaltic eruptions themselves, with effects tapering off over some hundreds of thousands of years; 2) the limestones are weathering products of vast drainage areas newly floored by relatively Ca-rich basalt, and the ferns are a consequence of unusually heterogeneous depositional environments caused by the accelerated tilting and subsidence associated with the eruption of the basalts (and the initiation of TS IV. These scenarios are not mutually incompatible, but they do predict completely different far field effects that can be looked for in other regions far from the CAMP.

The relationship between the accelerated tilting in the lower part of TS IV can be seen in a comparison of the stratigraphy in the lower Shuttle Meadow Formation across the various fault blocks in the Meriden area. At Higby Mountain gray siltstones and sandstones of the "Durham member" rest directly on the Talcott Formation. In the next fault block east with Lamentation Mountain, there are 10 m

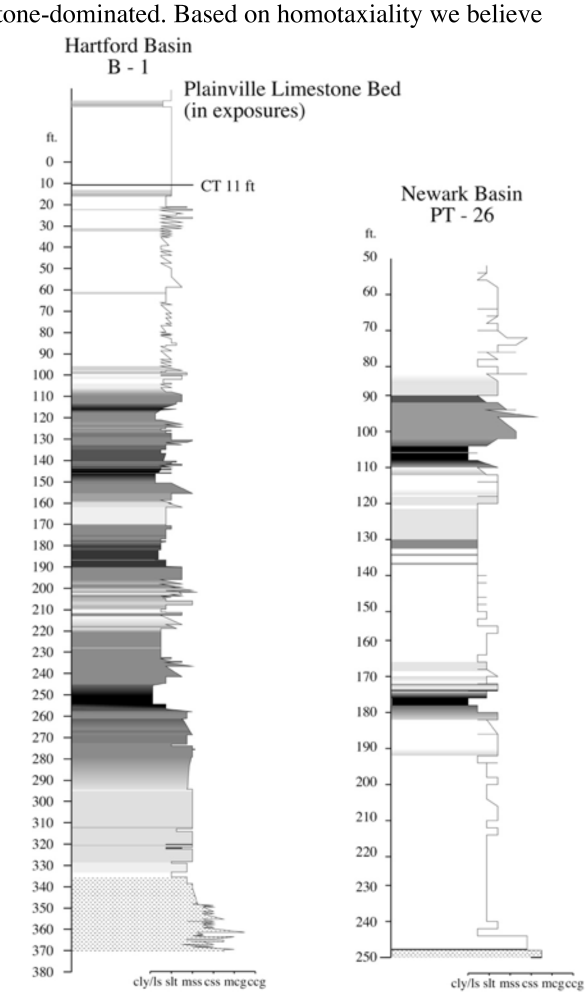

Figure 22. Comparison of the Silver Ridge B-1 core and core PT-26 from the lower Feltville Formation of the Newark basin. Note the homotaxiality of the black units.

of intervening red beds, suggesting additional accommodation due to syndepositional subsidence along the intervening fault. On the next major fault block to the west with Cathole Mountain and Stop 8, the Durham member is missing entirely and the entire formation is much thinner (mileage 148.8, below). Based on outcrop width, this thinning continues further to the west. In contrast there appears to be much less variability of the Talcott Formation. The simplest interpretation of these observations is that considerable structurally controlled depositional relief developed between the emplacement of the Talcott Formation and the deposition of the upper Shuttle Meadow Formation, an interval of time probably no more that a couple of hundred thousand years. These arguments parallel those discussed at Stop 4b.

There are 13.5 m (339.0 to 294.6 ft) of largely red beds between the largely gray “Durham member” and underlying volcanoclastics. A sandstone with a few clasts of basalt at 339 ft marks the top of the volcanoclastics sequence, which otherwise consists of a conglomerate of basalt cobbles and basaltic sandstone that continues to the base of the core at 371 ft. At 362.5 ft and 367 ft there are well-developed layers of reddish spherule layers (Fig. 23) that might represent basaltic lapilli. These are well displayed at the adjacent outcrops (see Stop 6c).

Return to vehicles.

Mileage: Time to next stop 15 minutes.
86.5 Leave parking lot.
86.6 Turn left onto West Street.
89.3 Turn left onto ramp for I 91 South.
89.7 Take Exit 22 on right for CT Route 9 North.
91.9 Pass road cut in upper East Berlin Formation and Hampden Basalt (Hubert et al., 1976, 1978; McDonald, 1982). Abundant small-leaved conifers of assemblage B of Cornet (1977) in fine gray mudstones in uppermost cycle .
92.3 Pass spectacular road cuts of Hamden Basalt and middle to upper East Berlin Formation in a separate fault block. These cuts have been described in numerous papers and guidebooks notably by Lehman (1959), Klein (1968), Hubert et al. (1976, 1978), Olsen et al. (1989), Gierlowski-Kordesch and Rust (1994), and Gierlowski-Kordesch and Huber (1995). These road cuts demonstrate the cyclicity of the Jurassic strata better than anywhere else in the Connecticut Valley. Abundant small-leaved conifers of assemblage B of Cornet (1977) in fine gray mudstones in these black-shale-bearing cycles.
92.3 Take Exit 22 on right for CT Route 15.
92.5 Turn right onto Frontage Road.
92.6 Turn right onto Worthington Road.
92.8 Turn left onto ramp extending east from CT Route 372.
93.1 Keep right toward CT Route 15 South.
95.9 U-turn onto CT Route 15 North.
96.0 Turn right into Silver Ridge Development on Hawthorne Drive and park.

STOP 6. SILVER RIDGE, BERLIN, CONNECTICUT. (optional, 1 HOUR) NE Meriden Quadrangle, (approx.) 41°35.1' N, 072°45.5’W; Tectonostratigraphic sequence TS IV; New Haven Fm., Talcott Formation, and Shuttle Meadow Fm.; Hettangian age, 200 Ma. Main points are: outcrops that partially show what is in the adjacent cores; gray earliest Jurassic section in top of New Haven Formation with small-leaved conifers; pillowed Talcott Fm. and volcanoclastic member with spherules; cyclical lower Shuttle Meadow Fm. with fish-bearing limestones and more small-leaved conifers; “Plainville limestone bed” in middle Shuttle Meadow Formation.

Silver Ridge is the name given to a gated development to the east of Silver Lake and CT Route 15. Roughly 230 m of New Haven Fm., Talcott Formation, and lower to Shuttle Meadow Fm. are patchily exposed in a series of exposures created in the course of construction for highway development, attendant highway improvements and unrelated light industry. We will visit a variety of locations in this general area that illustrate some of the main features.

Begin by walking to CT Route 15 and head south along the west curb. Stop about 30 m south at the wall of basalt and contact with underlying New Haven Formation.

STOP 6a. UPPERMOST NEW HAVEN FORMATION AND LOWER, PILLOWED TALCOTT FORMATION AT CUT ON EAST SIDE OF CT RT. 15.

This cut reveals the uppermost meter of the New Haven Formation of TS IV and 10 meters of the basal Talcott Formation. Here the uppermost New Haven Formation is predominately gray sandstone with subordinate amounts of siltstone. The sandstone contains plant stem and branch compressions and the fine thin bedded sandstones and siltstones contain abundant plant foliage and reproductive structures, so far predominantly small-leaved conifer shoots and cone fragments. This is a weathered version of the gray, uppermost New Haven Formation seen in cores Silver Ridge B-2 and B-3 taken to the immediate southeast and seen at Stop 5b. The exposed New Haven Formation is presumed to be of very earliest Jurassic age, although we await confirmation of this though palynological analysis.

The basal Talcott Formation is heavily pillowed here, but this will be better seen at the next location (see below). There are no visible signs of thermal metamorphism at this site, which is typical of contacts between pillowed basalt flows and sedimentary units. Thus, the pollen preservation should be good.

Walk south and up access road to gate. Enter and go in back of the light industrial building.

STOP 6b. LOWER, PILLOWED TALCOTT FORMATION.

About 20 m of the spectacularly pillowed lower Talcott Formation is exposed at this locality. To our knowledge there are no formal studies of this exposure.

Return to Hawthorne Drive and walk about 300 m to the east and south to Rosewood Lane. Turn left. Look at excavated blocks and "turtle back" exposures between the opposing houses on Hawthorne Drive and Rosewood Lane.

STOP 6c. VOLCANOCLASTIC MEMBER OF THE TALCOTT FORMATION.

The Talcott Formation is a complex of pillowed to massive flows interbedded with volcanoclastic beds and subject to abrupt lateral changes in facies. Davis (1889) described what he termed ashes in the Talcott Formation from this area; one exposure of these can be seen here. This volcanoclastic sequence occurs near the top of the

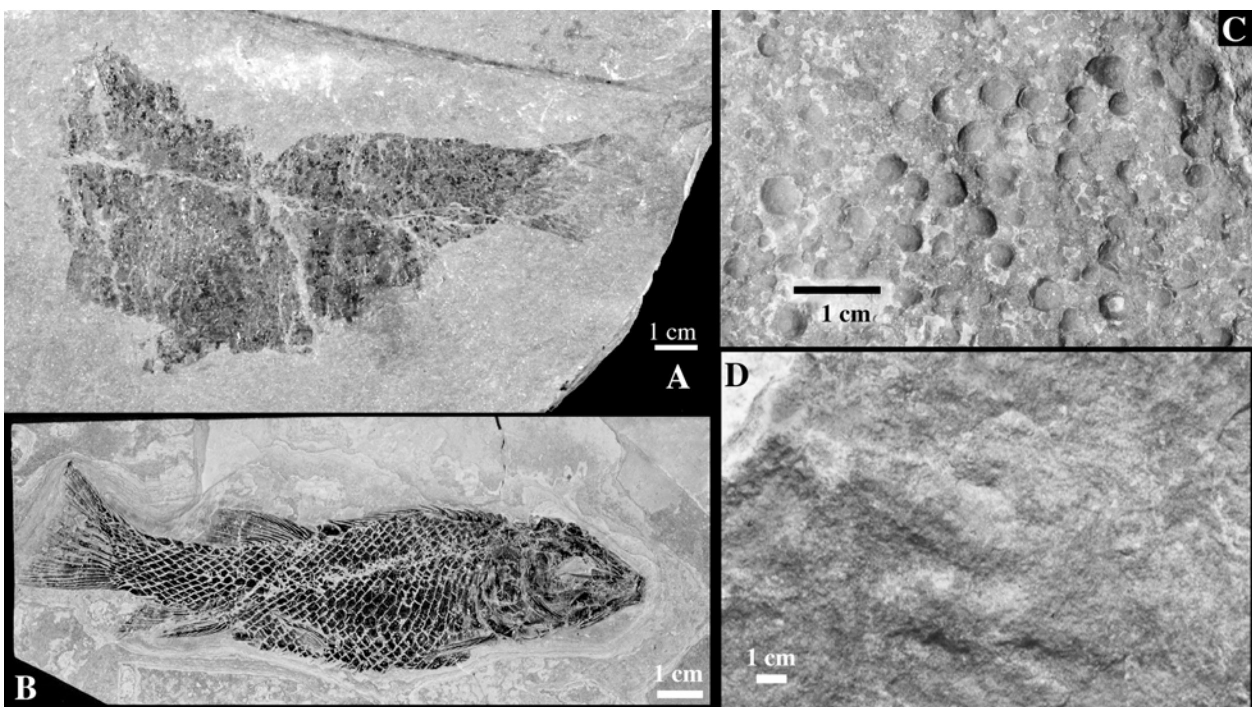

Figure 23. A, *Semionotus* from the "Southington limestone bed", Silver Ridge (Stop 6); B, *Semiontous* from the "Bluff Head bed", Bluff Head, North Guilford, CT N.G. McDonald Collection; C, Spherules from Talcott vocanoclastics, Silver Ridge, (Stop 6) (N.G. McDonald collection); D, uncollected Anchisauripus, Silver Ridge (Stop 6).

Talcott Formation and contains some sedimentary admixture. It is apparently equivalent to the volcanoclastic beds seen in the Silver Ridge B-1 core (Stop 5b).

In addition to the polymict basaltic clasts and palagonitic matrix, numerous spherules are visible on several bedding planes (Fig. 23). These have been called lapilli, but their true nature is unknown. To the south, the Talcott cuesta rises to a much higher elevation. Excavation for houses reveals an upper massive basaltic flow where the ridge is higher, underlain by volcanoclastic beds and then more massive to pillowed basalt. This upper flow must disappear rapidly to the north and northeast by erosion or non-deposition. There are no modern studies of these volcanoclastic beds, and we presume that they consist of some mixture of ash falls, eruptive spatter and lappilli, and reworked eroded basalt.

Here and in cores B-1 and B-2, the volcanoclastic member directly overlies more typical pillowed Talcott Formation. Outcrops at the north end of Hawthorn Drive and to the east and northeast reveal that flows of Talcott Formation overlie and interfinger with volcanoclastics, and thus the latter are intimately related to the Talcott and should not be classified as part of the Shuttle Meadow Formation.

STOP 6d. LOWER SHUTTLE MEADOW FORMATION.

This exposure of predominately gray mudstones and sandstone, along the east side of the road, is probably the regressive part of the Van Houten cycle bearing the “Southington limestone bed” of the “Durham member” of the lower Shuttle Meadow Formation. Even though we have fairly good exposures in this area (by Hartford basin standards,) we could not have worked out the basic cyclostratigraphy of the lower Shuttle Meadow Formation without a continuous core.

Temporary exposures of this same unit to the south revealed the limestone bed itself and it produced a few articulated fish (Fig. 23). Sandstone beds in adjacent units also produced a number of large *Anchisauripus* footprints and laminated gray mudstones produce fairly abundant small-leaved conifers (*Brachyphyllum*).

The “Bluff Head bed” was excavated within a couple of hundred meters to the northwest of this site by Davis and Loper (1861), who predicted its position based on stratigraphy they had seen elsewhere in the Shuttle Meadow Formation.

STOP 6e. MIDDLE SHUTTLE MEADOW FORMATION.

These disturbed exposures show the middle Shuttle Meadow Formation. Silver Ridge B-1 was spudded just to the north. Peter LeTourneau and PEO had hoped we would intersect a distinctive nodular limestone sequence that was supposedly in the middle of the formation but had never actually been seen in unambiguous superposition with the cyclical lower Shuttle Meadow. Silver Ridge B-1 was apparently spudded below this limestone, which much to our pleasure was subsequently excavated at this site. The limestone level apparently lies a few meters above the top of the Silver Ridge B-1 core (Fig. 20).

The limestone beds consist of decimeter-scale nodular gray to purple micrite and gray to purple mudstone and siltstone. Semionotid fish scales, bones, and coprolites are fairly common. Some of the flaggier associated siltstone beds and ripple-cross laminated siltstones have reptile footprints. What is probably the same beds on Higby Mountain to the east have produced numerous unionid bivalves, scales and bones of *Semionotus* and a partial skull and associated bones of a large coelacanth (*Diplurus* cf. *D. longicaudatus*) (McDonald, 1992). Limestone beds, certainly representing the same level are seen at Plainville, CT in Cooks Gap to the north and to the south along US Rt. 1 in Branford, CT a distance of over 100 km (Olsen et al., 1996b). Evidently this limestone was deposited by a relatively shallow lake large in area. These limestone beds are called the Plainville limestone and represent the wet phase (division 2) of one or two Van Houten cycles in the wet phase of a short modulating cycle within the drying phase of a McLaughlin cycle.

Return to vehicles.

Mileage: Time to next stop 50 minutes.
96.2 Turn around and head out Hawthorne Drive.
96.4 Turn right onto CT Route 15 North.
96.8 Take U-turn onto CT Route 15 South.
101.7 Stay on CT Route 15 South.
112.7 Take Exit 63 for CT Route 22.
112.9 Turn left following signs to CT Route 22.
113.0 Turn left onto CT Route 22 East.
117.7 Take right veer onto CT Routes17 South / 22 East.
117.8 keep right on CT 17 South.
121.6 Turn left onto Half Mille Road.
122.4 Entering East Haven.
122.8 Turn right onto Warner Road heading south.
123.0 Park.

STOP 7. ERUPTIVE FISSURE SYSTEM, EAST HAVEN, CONNECTICUT. North-central Branford Quadrangle, (approx.) 41°35' N, 072°82'W; Tectonostratigraphic sequence TS IV and intrusives; New Haven Fm., Fairhaven dike, Talcott Formation, and Shuttle Meadow Fm.; Hettangian age, 200 Ma. Main points are: feeder dike and eruptive fissure complex; volcanoclastics more common than true lava flows.

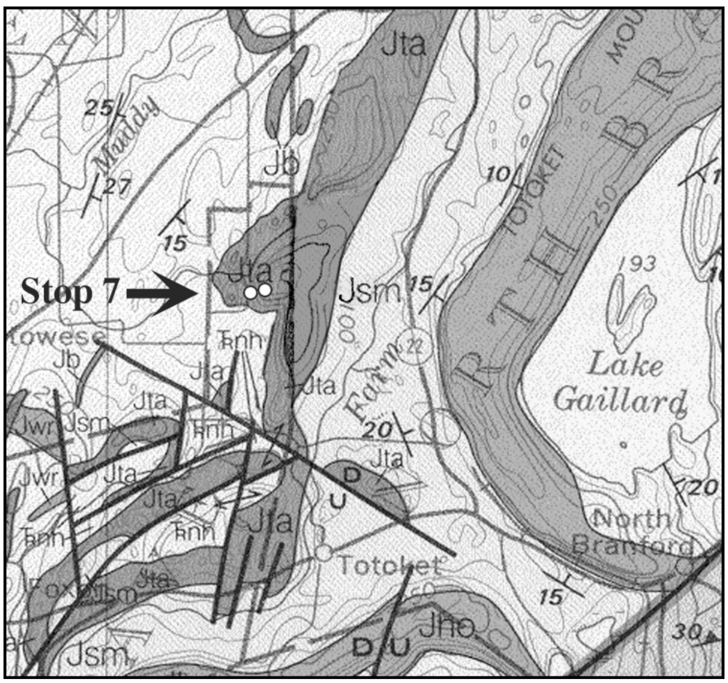

Figure 24. Portion of the Branford Geological Quadrangle (from: http://g3.tmsc.org/geology/bedrock/index.html) but modified in the vicinity of Stop 7. White dots show position of Stops 7a (left) and 7b (right).

We have divided this stop into two sub-stops (Fig. 24). The first (7a) is the locality described by Philpotts and Martello (1986) and Philpotts and Asher (1992) where the Fairhaven dike intersects and feeds the Talcott Formation and related volcanoclastics. The second (7b) is on the top of the eruptive complex comprised of volcanoclastics.

The area around Bayberry Lane and Warner Road in East Haven has played a critical role in our understanding of the Talcott system and the CAMP in general. The Talcott Formation was originally called the "Anterior Trap" by Percival (1842); this name persisted in Connecticut until Lehman (1959) formalized Emerson's (1891; 1898) nomenclature for the extrusive basalts of the Hartford Basin. Emerson called this unit the Talcott Diabase and/or Basalt. He erroneously referred its type section to be located near Talcottville, Connecticut (a small town in the northeastern part of the State, located on crystalline rocks of the Eastern Highlands), but it is reasonable to assume Emerson actually meant Tarrifville, at the northern end of Talcott Mountain, which is upheld in part, by the Talcott Formation. Based largely on field studies of this part of East Haven, Sanders (1968, 1970) attempted to redefine the Talcott in southern Connecticut to consist of basalt and basalt breccia intercalated with sedimentary rocks, with an aggregate thickness of ~300 m. It was this interpretation that drove him to rename the unit Talcott Formation as opposed to Talcott Basalt.

Sanders's interpretation of these outcrops rested on the quite reasonable working hypothesis that all the major units visible in vertical succession could be traced laterally, and where they could not be, they were either covered or faulted out. However, we, following Philpotts and Martello (1986) and Philpotts and Asher (1992), interpret much of the stratigraphy as being related to growth structures, in this case a graben system around the Fairhaven dike fissure system that fed the Talcott Formation basalts and volcanoclastics.

We have examined the most important of Sanders's outcrops, as well as other sections of the Talcott Formation throughout the Hartford basin. Based on our observations and interpretation, the only place the Talcott Formation approaches 300 m thick is within the graben system around the feeder dike, which was subsiding during the eruption of the flows and volcanoclastics. Aside from thin (m-scale or less) clastic interbeds near the top of the formation, no sedimentary members of any appreciable thickness exist as intercalated "members" within the formation, as Sanders (1968, 1970) claimed. The outcrops here, representing the field area on which Sanders based his interpretation, actually consist of a kilometer-scale eruptive complex developed on a relatively straightforward stratigraphic succession. Hence, the stratigraphic units involved consist of the upper New Haven Formation, basalt and breccia of the Talcott Formation, associated volcanoclastics, and the diabase of the feeder dike. The field relationships of these units will be demonstrated at the field trip stop.

The Talcott Formation displays a complex internal stratigraphy, especially at this stop that partially resulted from its extrusion within the basin from multiple feeder dikes located between East Haven and Wallingford, Connecticut. At least two separate dikes intersected the basin surface and generated fissure eruptions from within the basin that are responsible for the association of pillowed and massive basalt and volcanoclastic breccia that comprise the Talcott Formation.

Our interpretation of the geometry of this volcanic complex is in part based on a down-the-plunge view of the structure. Its tight seemingly synclinal shape is not in accord with the bedding in the overlying Shuttle Meadow formation and is more consistent with a graben structure striking roughly N 50° E. The bedding strikes more northernly so the cross-section cannot be properly orthogonal, but it does nonetheless give us some idea of the geometry (Fig. 24). As discovered by Philpotts and Martello (1986) a segment of the Fairhaven dike strikes north directly into this structure. Stop 7a examines this intersection.

There is very little typical Talcott basalt in a broad swath from the region around this stop northeast to north of Reeds Gap along the trend of the Fairhaven dike system. Instead the stratigraphic position of the flows is occupied mostly by volcanoclastics, which in some cases, such as to the east and north of Totoket Mountain the volcanoclastic are white to yellow and highly calcareous. In much of this region there are no flows at all. We intepret these relations to indicate that the land surface may have been elevated adjacent to the fissure system during eruption so that most lava flowed away from the Fair Haven dike without ponding, perhaps flowing through gaps in the volcanoclastics. At the closing phases of the eruption, the land surface may have subsided producing a relative low where there was a relative high. In any case, the zone around the Fairhaven system implies an inflated region from 5 to 10 km wide during eruption. We stress that our rather speculative interpretation requires significant fieldwork before it can be taken seriously.

In many ways, the feeder system for at least part of the Hampden (Black rock dike system) seems to have produced a similar but smaller structure, and one formed during a much drier portion of the Milankovitch cycles.

Walk southeast to opposite 74 Warner Road (labeled 274) to large exposure of igneous rock.

STOP 7a. FAIRHAVEN DIKE AND ASSOCIATED TALCOTT FORMATION AGGLOMERATE AND PILLOW BASALT. (30 minutes). 41°21.277'N, 072°49.385.

Philpotts and Asher (1992) observed that the small cliff facing and paralleling Warner Road is the western contact of the Fairhaven dike with small amounts of sedimentary rock still attached to the dike at the northern end of the exposure. There are several generations of vertical sheets of fine-grained diabase visible that represent successive re-intrusions of magma into the dike. At least one of these sheets has remarkably small, nearly horizontal columns while others have larger columnar joints intruded by additional magma that chilled to what still remains a well-preserved glass containing unaltered olivine phenocrysts. Above this small cliff the diabase passes into agglomerate and then pillowed basalt. Near the center of the dike is a breccia that Philpotts and Asher (1992) interpret to have flowed back into the dike after a surge. The imbrication of the pillows suggests flow to the south.

We interpret the outcrops of this dike and the surrounding volcanoclastics and pillow basalt as part of a 700 m wide graben system formed by collapse of the New Haven Formation during the eruption, that ultimately was capped by volcanoclastics (Stop 7b).

Outcrops on the hill on the east side and paralleling Barberry Road (where it trends nearly north-south) is a ridge of typical red and purplish conglomeritic New Haven Formation. This is overlain by gray siltstone that is in turn overlain by gray siltstone and volcanoclastic units interbedded with pillow basalt making up the crest of the ridge. This is a more normal set of contact relationships suggesting that in this area we are outside of the eruptive fissure system.

Return to vehicles.

Mileage: Time to next stop 3 minutes.
123.0 Turn around, heading north.
123.3 Turn right onto Augur Road.
123.5 Turn right onto Colonial Heights Road.
123.8 Park at crest of hill.

STOP 7b. AGGLOMERATE AND TUFF CAPPING FISSURE COMPLEX. (15 minutes). 41°21.301'N, 072°49.177'W.

Exposures formed by excavations for this new development along Colonial Heights Road reveal the volcanoclastics above Stop 7a. Here stratigraphically high volcanoclastics contain some Paleozoic clasts, but is still mostly basaltic, resembling some of the volcanoclastics at Stop 6 and in the Silver Ridge core B-1 (Stop 5b). Stratigraphically lower units seem in some cases to be nearly pure basaltic tuff (although petrographic work has yet to be carried out).

Return to vehicles.

Mileage: Time to next stop 45 minutes.
123.8 Turn around heading north on Colonial Heights Road.
124.1 Turn Left onto Augur Road.
124.3 Dogleg to continuation of Auger, becomes Half Mile Road.
125.5 Turn right on CT Route 17 North (Middletown Road)
129.2 Merge with CT 22 West.
129.4 Turn left on CT 22 West.
133.3 Turn right onto CT 5 North.
134.0 Take right hand ramp onto I 91 North.
144.6 Take Exit 17 for I 691 West – CT 15 north,
145.6 Take Exit 68W for I 691 West.
148.0 Take ramp for Exit 6 for CT Route 71.
148.2 Turn left at end of ramp onto Lewis Avenue.
148.4 Turn left onto Kensington Avenue, Holyoke Basalt visible ahead.
148.8 Contact between Talcott Formation and Shuttle Meadow on right; no gray beds of the "Durham member" are present (see Stop 5b).
149.0 Turn left on CT Route 71 South (Chamberlain Highway).
149.1 Turn right onto Target shopping center (Cold Spring Avenue).
149.2 Turn left in front of Target.
149.3 Drive to southwest corner of parking lot.

STOP 8. TALCOTT PILLOW BASALT SEQUENCE ON NEW HAVEN FORMATION. (30 MINUTES). Central Meriden Quadrangle, (approx.) 41°33.12'N, 072°48.91'W; Tectonostratigraphic sequence TS IV; New Haven and Talcott formations; Hettangian age, 200 Ma. Main points are: northeast prograding pillow lava forests and flow lobes; onlapping red sandstones with graded beds of altered basaltic gravel and sand.

This spectacular exposure reveals nearly the entire thickness of the Talcott Formation as well as the uppermost few meters of the New Haven Formation. This new exposure is also described in Stop 4 in Philpotts and McHone (this volume). Most of the lower half of the Talcott here consists of forsets and tongues of pillow lava, while most of

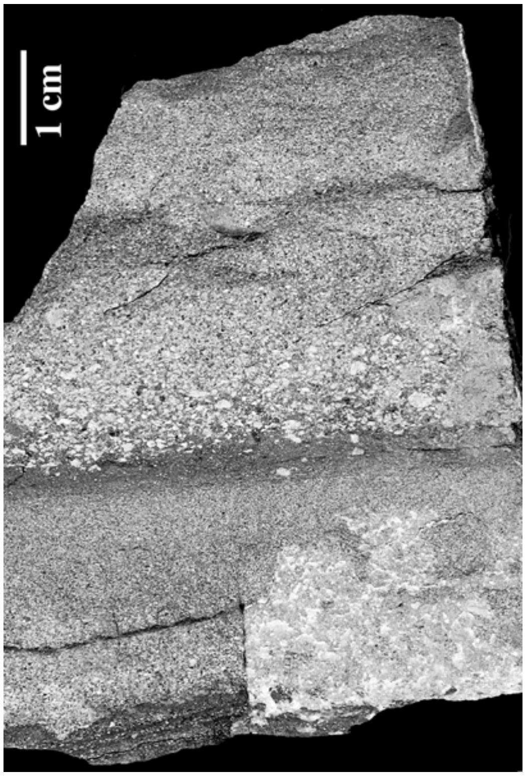

Figure 25. Graded volcanoclastic beds in uppermost New Haven Formation at Stop 8 at the Target.

the upper part consists of non-pillowed flow lobes and vesicular basalt. It is unclear to us how many eruptive events are represented here, but we see no compelling reasons to believe there is more than one.

The base of the Talcott displays an extremely informative relationship to the New Haven Formation. Northeast tapering wedges of pillowed basalt onlap each other in the lower 10 m of the flow complex. These are easily traced out by following the red sandstone and siltstone beds that extend upward from the underlying New Haven Formation into the basalt. In several places these beds are internally stratified and contain numerous large to small clasts of basalt and highly altered basalt. Tracing the red beds downward , they merge with the underlying New Haven Formation, becoming a bed that underlies the next, more north or eastern basaltic wedge. Most of the exposed New Haven Formation here consists of such beds. Internally, they show beautiful graded beds comprised of basaltic gravel and sand (Fig. 25). The basalt and what presumably was basaltic glass is generally altered to a yellow or tan material, superficially resembling a carbonate. As yet no analyses have been done, but all stages of alteration from unquestionable basaltic material to the tan to yellow clasts can be seen. Only the lowest New Haven Formation seems to lack this basaltic material.

Amongst the wedges of basalt pillows are larger lobes of massive basalt without pillows. Higher in the Talcott some of these are over 6 m thick and scores of meters long. These are almost certainly flow lobes that were the sources of the pillowed wedges.

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

If we assume that the various pillow wedges and flow lobes represent one major eruptive event, some constraints are placed on the accumulation rate and water depth of the uppermost New Haven Formation. The couple of meters or so of basalt-bearing sedimentary strata obviously took no more time to accumulate than the flow complex took to advance over the site, which would seem to be on a time scale of years to at most hundreds of years. The lake into which the lava poured had to be at least the depth of the high points of the individual wedges of pillowed basalt (i.e., 5 m or so), but in our view probably not the depth of the thickness of the entire Talcott. This is because the lava displaces the water. As yet we have not been able to find desiccation cracks in the basalt-bearing red beds, so the lake did not seem to dry up, at least locally, during the advance of the Talcott. There are no sedimentological criteria that would indicate that these red beds were deposited under a significant body of water, or even that they are lacustrine strata and this should be kept in mind looking at other Triassic-Jurassic red beds. The obvious lesson, unappreciated by many sedimentologists is that red beds, even sandstones, do not equal fluvial environments.

End of field trip. The local major interstate highways can be accessed by heading out the south exit of the Target and following the signs to I 691 West, towards I 84, or east, towards, I 91 and CT 15.

REFERENCES

Anderson, H.M. and Anderson, J.M., 1970, A preliminary review of the biostratigraphy of the uppermost Permian, Triassic and lowermost Jurassic of Gondwanaland: Palaeontologia Africana, v. 13 (suppl.), p. 1-22.

Anderson, J., Anderson, H., Fatti, P. and Sichel, H., 1996, The Triassic explosion(?): a statistical model for extrapolating biodiversity based on the terrestrial Molteno Formation: Paleobiology, v. 22, p. 318-328.

Ash, S.R., 1986, Fossil plants and the Triassic-Jurassic boundary, *in* Padian, K., ed., The Beginning of the Age of Dinosaurs: Faunal Change Across the Triassic-Jurassic Boundary: Cambridge University Press, p. 21-30.

Axsmith, B.J. and Kroehler, P.A., 1989, Upper Triassic Dinophyton Zone plant fossils from the Stockton Formation in southeastern Pennsylvania: The Mosasaur, v. 4, p. 45-47.

Badjukov, D.D., Lobitzer, H. and Nazarov, M.A., 1987, Quartz grains with planar features in the Triassic-Jurassic boundary sediments from northern Limestone Alps, Austria: Lunar and Planetary Science Conference Abstracts, v. 28, p. 38-39.

Balk, Robert, 1937, Structural behavior of igneous rock: Geological Society of America Memoir 5, 177 p.

Beerling, D. J., 2002, Palaeoclimatology (Communication arising): CO_2 and the end-Triassic mass extinction: Nature, v. 415, p. 386-387.

Beerling, D. J., Lomax, B. H., Royer, D. L., Upchurch, G. R., Jr., Kump, L. R., 2002, An atmospheric pCO_2 reconstruction across the Cretaceous-Tertiary boundary from leaf megafossils: Proceedings of the National Academy of Sciences, U.S.A., v. 99, p. 7836-7840.

Benton, M.J., 1994, Late Triassic to Middle Jurassic extinctions among continental tetrapods: testing the pattern, *in* Fraser, N.C. and Sues, H.D., eds., In the Shadow of the Dinosaurs: Cambridge University Press, p. 366-397.

Berner, R.A., 1991, A model for atmospheric CO_2 over Phanerozoic time: American Journal of Science, v. 291, p. 339-376,

Bice, D.M., Newton, C.R., McCauley, S.E., Reiners, P.W. and McRoberts, C.A., 1992, Shocked quartz at the Triassic-Jurassic boundary in Italy: Science, v. 255, p. 443-446.

Boslough, M.B, Chael, E.P., Trucano, T.G., Crawford, D.A. and Campbell, D.L., 1996, Axial focusing of impact energy in the Earth's interior; a possible link to flood basalts and hotspots, *in* Ryder, G., Fastovsky D. and Gartner, S., eds., The Cretaceous-Tertiary event and other catastrophes in Earth history: Geological Society of America Special Paper 307, p.541-550.

Boyer, P.S., 1979, Trace fossils *Biformites* and *Fustiglyphus* from the Jurassic of New Jersey: Bulletin of the New Jersey Academy of Science, v. 24, no. 2, p.73-77.

Brophy, G.P., Foose, R.M., Shaw, F.C., Szekely, T.S., 1967, Triassic geologic features in the Connecticut Valley near Amherst, Massachusetts, Trip D, Guidebook for field trips in the Connecticut Valley of Massachusetts: New England Intercollegiate Geologic Conference 59th Annual Meeting, Amherst, Massachusetts, p. 61-72.

Carter, J.G., Olsen, P.E., Sues, H.-D., Peyer, K., Campbell, D.C., Campbell, M.R., Coffey, B., and Harlow, S.A., 2001, Life and death in the Triassic: an extraordinary tetrapod assemblage from the Late Triassic of North Carolina: Canadian Journal of Science (submitted).

Cohen, A.S., and Coe, A.L., 2002, New geochemical evidence for the onset of volcanism in the Central Atlantic magmatic province and environmental change at the Triassic-Jurassic boundary: Geology, v. 30, p. 267-270.

Colbert, E.H. and Baird, D., 1958, Coelurosaur bone casts from the Connecticut Valley: American Museum Novitates no. 1901, 11 p.

Contreras, J., Scholz, C. and King G.C.P., 1997, A general model for rift basin evolution: constraints of first order stratigraphic observations: Journal of Geophysical Research, v. 102B, p. 7673-7690.

Coombs, J.R. W.P., 1980, Swimming Ability of Carnivorous Dinosaurs: Science, v. 207, p. 1198-1200.

Cornet, B., 1977, The palynostratigraphy and age of the Newark Supergroup, Ph.D. thesis: University of Pennsylvania, 505 p.

Cornet, B., 1989a, Late Triasic angiosperm-like pollen from the Richmond rift basin of Virginia, USA: Palaeontographica. Abteilung B: Paläophytologie, v. 213, p. 37-87.

Cornet, B., 1989b, The reproductive morphology and biology of *Sanmiguelia lewisii*, and its bearing on angiosperm evolution in the Late Triassic: Evolutionary Trends in Plants, v. 3, p. 25-51.

Cornet, B. and Traverse, A., 1975, Palynological contributions to the chronology and stratigraphy of the Hartford basin in Connecticut and Massachusetts: Geoscience and Man, v. 11, p. 1-33.

Dana, E.S., 1874, Abstract of a paper on the trap rocks of the Connecticut Valley: American Journal of Science, ser. 3, v. 8, p. 390-392.

Davis, W.M., 1889, The ash bed at Meriden and its structural relations: Meriden Science Association: Proceedings and Transactions, v. 3, p. 23-30.

Davis, W.M., 1898, The Triassic formation of Connecticut: U.S. Geological Survey Annual Report, no. 18, part 2, p. 1-192.

Davis, W.M. and Loper, S.W., 1891, Two belts of fossiliferous black shale in the Triassic formation of Connecticut: Geological Society of America, Bulletin, v. 2, p. 415-430.

DeWet, C.B., Mora, C.I., Gore, P.J. W., Gierlowshi-Kordesch, E., and Cucolo, S.J., 2002, Deposition and geochemistry of lacustrine and spring carbonates in Mesozoic rift basins, eastern North America, *in* Renaut, R.W. and Ashley, G.M., eds., Sedimentation in Continental Rifts: SEPM Special Publication No. 73, p. 309-325.

Dietz, R.S., 1986, Triassic-Jurassic extinction event, Newark basalts and impact-generated Bahama nexus: Meteoritical Society 49th Annual Meeting, Abstracts, LPI Contribution 600, p. 1-10.

Dobruskina, I.A., 1988, The history of land plants in the Northern Hemisphere during the Triassic with special reference to the floras of Eurasia: Geologisch - Paläontologische Mitteilungen Innsbruck, v. 15, p. 1-12.

Dobruskina, I.A., 1993, Relationships in floral and faunal evolution during the transition from the Paleozoic to the Mesozoic, *in* Lucas, S.G. and Morales, M., eds., The Nonmarine Triassic: Bulletin of the New Mexico Museum of Natural History and Science 3, p. 107-112.

Ekart, D.D, Cerling, T.E., Montanez, I. and Tabor, N.J., 1999, A 400 million year carbon isotope record of pedogenic carbonate: implications for paleoatmospheric carbon dioxide: American Journal of Science, v. 299, p. 805-827.

Ellenberger, P., 1976, Une piste avec traces de soies épaisses dans le Trias inférieur a moyen de Lodève (Hérault, France): *Cynodontipus polythrix* nov. gen., nov. sp. les Cynodontes en France: Géobios, v. 9, p. 769-787.

Emerson, B.K., 1898a, Holyoke Folio, Massachusetts - Connecticut. U.S. Geological Survey Geological Atlas of the United States, no. 50, 5 maps, 7 p.

Emerson, B.K., 1898b, Geology of old Hampshire county, Massachusetts, comprising Franklin, Hampshire and Hampdon counties: US Geological Survey Monograph 29, 790 p.

Emerson, B.K. 1917, Geology of Massachusetts and Rhode Island: U.S. Geological Survey Bulletin 597, 289 p.

Farquhar, O. C., 1967, Tectonic and neotectonic structures in Triassic sedimentary rocks at Holyoke, Massachusetts: *in* Farquhar, O. C., ed., Economic Geology in Massachusetts, Amherst: Massachusetts Graduate School, University of Massachusetts, p. 317-336.

Farlow, J.O. and Galton, P.M., 2003, Dinosaur Trackways of Dinosaur State Park, Rocky Hill, Connecticut, *in* LeTourneau, P.M. and Olsen, P.E., eds., The Great Rift Valleys of Pangea in Eastern North America, vol. 2: Sedimentology, Stratigraphy, and Paleontology: Columbia University Press, p. 248-263.

Fedosh, M.S. and Smoot, J.P., 1988, A cored stratigraphic section through the northern Newark basin, New Jersey: U.S. Geological Survey Bulletin 1776, p. 19-24.

Foose, R.M., Rytuba, J.J., and Sheridan, M.F., 1968, Volcanic plugs in the Connecticut Valley Triassic near Mount Tom, Massachusetts: Geological Society of America Bulletin, v. 79, p. 1655-1661.

Fowell, S.J. 1993, Palynology of Triassic/Jurassic boundary sections from the Newark Supergroup of Eastern North America: Implications for catastrophic extinction scenarios, Ph.D. thesis: Columbia University, 133 p.

Fowell, S.J., Cornet, B. and Olsen, P.E., 1994, Geologically rapid Late Triassic extinctions: Palynological evidence from the Newark Supergroup, *in* Klein, G.D., ed., Pangaea: Paleoclimate, Tectonics and Sedimentation During Accretion, Zenith and Break-up of a Supercontinent: Geological Society of America Special Paper 288, p. 197-206.

Fowell, S.J. and P.E. Olsen, 1993, Time-calibration of Triassic/Jurassic microfloral turnover, eastern North America: Tectonophysics, v. 222, p. 361-369.

Fowell, S.J. and P.E. Olsen. 1995, Time calibration of Triassic/Jurassic microfloral turnover, eastern North America: reply to van Veen: Tectonophysics, v. 245, p. 96-99.

Fowell, S.J. and Traverse, A., 1995, Palynology and age of the upper Blomidon Formation, Fundy Basin, Nova Scotia: Review of Palaeobotany and Palynology, v. 86, p. 211-233.

Frakes, L.A., 1979, Climates Throughout Geologic Time: Elsevier Scientific Publishing Company, New York, 310 p.

Gierlowski-Kordesch, E. and Huber, P., 1995, Lake sequences of the central Hartford Basin, Newark Supergroup: State Geological and Natural History Survey of Connecticut Guidebook No. 7, p. B1-B39.

Gierlowski-Kordesch, E. and Rust, B.R., 1994, The Jurassic East Berlin Formation Hartford Basin, Newark Supergroup (Connecticut and Massachusetts): a saline lake/playa/alluvial plain system, *in* Renaut, R.W. and Last, W.M., eds., Sedimentology and Geochemistry of Modern and Ancient Saline Lakes: SEPM Special Publication No. 50, p. 249-265.

Gilfillian, A. M. and Olsen, P. E., 2000, The coelacanth *Diplurus longicaudatus* as the origin of the large coprolites occurring in the Triassic-Jurassic lacustrine strata of Eastern North America: Geological Society of America, Abstracts with Programs, v.32, no. 1, p. A-20.

Gradstein, F.M., Agterberg, F.P., Ogg, J.G., Hardenbol, J., Van Veen, P., Thierry, J., and Huang Zehui, 1995, A Triassic, Jurassic and Cretaceous time scale, *in* Berggren, W.A., Kent, D.V., Aubrey, M.-P., and Hardenbol, J., eds., Geochronology, Time Scales and Global Stratigraphic Correlation: SEPM (Society for Sedimentary Geology) Special Publication No. 54, p. 95-126.

Gray, N.H., 1982, Mesozoic volcanism in north-central Connecticut, *in* Joesten, R. and Quarrier, S.S., eds., Guidebook for fieldtrips in Connecticut and south-central Massachusetts, New England Intercollegiate Geological Conference, 74th Annual Meeting, Storrs, Connecticut: Connecticut Geological and Natural History Survey Guidebook, no. 5, trip M-3, p. 173-193.

Gray, N.H., 1987, Mesozoic sedimentary and volcanic rocks in the Farmington River Gorge, Tariffville, Connecticut: Geological Society of America Centennial Field Guide – Northeastern Section, p. 165-168.

Gray, N.H., and Steinen, R.P., 2000, Monoclinic hopper pseudomorphs in Shuttle Meadow mudstones, Hartford Basin, Connecticut: Geological Society of America, Abstracts with Programs, v. 32, no. 1, p. A21.

Hallam, 1990, The end-Triassic extinction event: Geological Society of America Special Paper 247, p. 577-583.

Hallam, A., 2002, How catastrophic was the end-Triassic mass extinction event?: Lethaia, v. 35, p. 147-157.

Harris, T.M., 1931, The fossil flora of Scoresby Sound, east Greenland: Meddelelser om Grønland, Part 2, Part 3, Part 4, 104 p.

Heilman, J.J., 1987, That catastrophic day in the Early Jurassic: Connecticut Journal of Science Education, v. 25, p. 8-25.

Hesselbo, S.P., Robinson, S.A., Surlyk, F., and Piasecki, S. 2002, Terrestrial and marine extinction at the Triassic-Jurassic boundary synchronized with major carbon-cycle perturbation: A link to initiation of massive volcanism?: Geology, v. 30, p. 251-254.

Hitchcock, E., 1836, Description of the footmarks of birds (*Ornithichnites*) on new red sandstone in Massachusetts: American Journal of Science, v. 29, p. 307-340.

Hitchcock, E., 1848, An attempt to discriminate and describe the animals that made the fossil footmarks of the United States, and especially of New England: American Academy of Arts and Sciences Memoir (n.s.), v. 3, p. 129-256.

Hitchcock, E., 1858, Ichnology of New England: a report on the sandstone of the Connecticut Valley, especially its fossil footmarks: Commonwealth of Massachusetts, Boston, 220 p.

Hitchcock, E., 1865, Supplement to the Ichnology of New England: Commonwealth of Massachusetts, Boston, 96 p.

Horne, G. S., N. G. McDonald, P. M. LeTourneau, and J. Z. deBoer. 1993, Paleoenvironmental traverse across the early Mesozoic Hartford Rift Basin, Connecticut, *in* Cheney, J.T. and Hepburn, J.C., eds., Field Trip Guidebook for the Northeastern United States: Geological Society of America Annual Meeting, Boston, Massachusetts, October 25–28, 1993, v.2, p. 1-26.

Huber, P., McDonald, Olsen, P.E., 2002, Early Jurassic insects from the Newark Supergroup, northeastern North United States, *in* LeTourneau, P.M. and Olsen, P.E., eds., The Great Rift Valleys of Pangea in Eastern North America, vol. 2: Sedimentology, Stratigraphy, and Paleontology: Columbia University Press, p. 206-223.

Huber, P. and Lucas, S.G., 1993, Vertebrate biochronology of the Newark Supergroup Triassic, eastern North America, *in* Lucas, S.G. and Morales, M., eds., The Nonmarine Triassic: Bulletin of the New Mexico Museum of Natural History and Science 3, p. 179-186.

Huber P., Lucas, S.G. and Hunt, A.P., 1993, Revised age and correlation of the Upper Triassic Chatham Group (Deep River Basin, Newark Supergroup), North Carolina: Southeastern Geology v. 33, p. 171-193.

Hubert, J. F., Reed, A. A., and Carey, P. J., 1976, Paleogeography of the East Berlin Formation, Newark Group, Connecticut Valley: *American Journal of Science*, v. 276, p. 1183-1207.

Hubert, J.F., Reed, A.A., Dowdall, W.L., Gilchrist, J.M., 1978, Guide to the red beds of Central Connecticut: 1978 Field Trip, Eastern Section, Society of Economic Paleontologists and Mineralogists: University of Massachusetts, Amherst, Department of Geology and Geography Contribution No. 32, 129 p.

Kent, D.V. and Olsen, P.E., 1999a, Search for the Triassic/Jurassic long normal and the J1 cusp: Eos, Transactions, American Geophysical Union, Supplement, v. 80, no. 46, p. F306.

Kent, D.V. and Olsen, P.E., 1999b, Astronomically tuned geomagnetic polarity time scale for the Late Triassic. Journal of Geophysical Research, v. 104, p. 12,831-12,841.

Kent, D.V., Olsen, P.E., and Witte, W.K., 1995, Late Triassic-Early Jurassic geomagnetic polarity and paleolatitudes from drill cores in the Newark rift basin (eastern North America): Journal of Geophysical Research, v. 100B, no. 8, p. 14,965-14,998.

Klein, G. deV., 1968, Sedimentology of Triassic rocks in the lower Connecticut Valley, *in* Orville, P.M., ed., Guidebook for fieldtrips in Connecticut: New England Intercollegiate Geology Conference 60th Annual Meeting, New Haven, Connecticut: Connecticut Geological Natural History Survey Guidebooks No.2, trip C-l, p. l-l9.

Krynine, P.D., 1950, Petrology, stratigraphy, and the origin of the Triassic sedimentary rocks of Connecticut. Connecticut Geological and Natural History Bulletin 73, 247 p.

Lehman, E.P., 1959, The bedrock geology of the Middletown quadrangle: State of Connecticut, Geological and Natural History Survey Quadrangle Report 8.

LeTourneau, P.M., 1999, Depositional history and tectonic evolution of Late Triassic age rifts of the U.S. central Atlantic margin: results of an integrated stratigraphic, structural, and paleomagnetic analysis of the Taylorsville and Richmond basins: Ph.D. thesis, Columbia University, 294 p.

LeTourneau, P.M., 2002, Tectonic and climatic controls on the stratigraphic architecture and paleomagnetic reversal stratigraphy of the Late Triassic Taylorsville basin, Virginia and Maryland, U.S.A., *in* LeTourneau, P.M. and Olsen, P.E., eds., The Great Rift Valleys of Pangea in Eastern North America, vol. 2: Sedimentology, Stratigraphy, and Paleontology, Columbia University Press, p. 12-58.

Lucas, S.G., 2002, End-Triassic Mass extinction or the compiled correlation effect?: Geological Society of America Annual Meeting, Denver, Colorado, October 27-30, 2002, no. 235-7.

Lucas, S.G., 1997, Biostratigraphy, *in* Currie, J. and Padian, K., eds., Encyclopedia of Dinosaurs: Academic Press, San Diego, California/London, UK, p. 65-68.

Lucas, S.G. and Huber, P., 2002, Vertebrate biostratigraphy and biochronology of the nonmarine Late Triassic, *in* LeTourneau, P.M. and Olsen, P.E., eds., The Great Rift Valleys of Pangea in Eastern North America, vol. 2: Sedimentology, Stratigraphy, and Paleontology: Columbia University Press, p. 143-191.

Lucas, S.G., Tanner, L.H., and Chapman, M.G., 2002, End-Triassic mass extinction or the combined correlation effect?: Geological Society of America, Abstracts with Programs, v. 34, no. 6, p. 536.

Lull, R.S., 1904, Fossil footprints of the Jura-Trias of North America: Memoirs of the Boston Society of Natural History, v. 5, p. 461-557.

Lull, R.S., 1915, Triassic Life of the Connecticut Valley: Connecticut Geological and Natural History Survey Bulletin 24, 285 p.

Lull, R.S., 1953, Triassic life of the Connecticut Valley: Connecticut Geological and Natural History Survey Bulletin 81, 336 p.

Malinconico, M.L., 2001, Organic petrographic studies of the Early Mesozoic Newark rift basin: 1) thermal history using borehole and surface vitrinite reflectance and 2) organic sedimentation patterns in orbitally-forced underfilled lake cycles: Society for Organic Petrology, Abstracts with Program v. 18, p. 73-77.

Malinconico, M.L., 2002, Estimates of eroded strata using borehole vitrinite reflectance data, Triassic Taylorsville rift basin, Virginia: Implications for duration of syn-rift sedimentation and evidence of structural inversion, *in* LeTourneau, P.M. and Olsen, P.E., eds., The Great Rift Valleys of Pangea in Eastern North America, vol. 1: Tectonics, Structure, and Volcanism of Supercontinent Breakup: Columbia University Press, p. 80-103.

Marzoli, A., Renne, P.R., Piccirillo, E.M., Ernesto, M., Bellieni G. and De-Min, A., 1999, Extensive 200-million-year-old continental flood basalts of the Central Atlantic Magmatic Province: Science, v. 284, p. 616-618.

McCune, A.R., 1996, Biogeographic and stratigraphic evidence for rapid speciation in semionotid fishes: Paleobiology 22, p. 34-48.

McCune, A.R., Thomson, K.S. and Olsen, P.E., 1984, Semionotid fishes from the Mesozoic Great Lakes of North America, *in* Echelle, A.A. and Kornfield, I., eds., Evolution of fish species flocks: University of Maine at Orono Press, p. 27-44.

McDonald, N. G., 1982, Paleontology of the Mesozoic rocks of the Connecticut Valley; *in* Joesten, R. and Quarrier, S.S., (eds.), Guidebook for fieldtrips in Connecticut and south-central Massachusetts, New England: Intercollegiate Geological Conference, 74th Annual Meeting, Storrs, Connecticut, Connecticut Geological and Natural History Survey Guidebook No. 5, trip M-2, p. 143-172.

McDonald, N. G., 1992, Paleontology of the early Mesozoic (Newark Supergroup) rocks of the Connecticut Valley: Northeastern Geology, v. 14, no. 2-3, p. 185-200.

McElwain, J.C., Beerling, D.J. and Woodward, F.I., 1999, Fossil plants and global warming at the Triassic-Jurassic boundary: Science v. 285, p. 1386-1390.

McInerney, D. P. and Hubert, J. F., 2002, Meandering-river facies in the upper Triassic New Haven Arkose, south central Connecticut: early evolution of the Hartford rift basin. *in* LeTourneau, P.M. and Olsen, P.E., eds., The Great Rift Valleys of Pangea in Eastern North America, vol. 1: Sedimentology, Stratigraphy, and Paleontology, Columbia University Press, p. 88-107.

Melosh, H.J., 2000, Can impacts induce volcanic eruptions?: Catastrophic Events and Mass Extinctions: Impacts and Beyond, LPI Contribution 1053.

Mooney, J., 1979, The origin and diagenetic history of a Mesozoic carbonate hot spring deposit, Coe's Quarry, North Branford, Connecticut: M.S. Thesis, Department of Geology, University of Connecticut, Storrs, Connecticut, 100 p.

Newberry, J. S., 1888, Fossil fishes and fossil plants of the Triassic rocks of New Jersey and the Connecticut valley: U.S. Geological Survey Monographs 14, p. 1-152.

Olsen, P.E., 1980. Fossil great lakes of the Newark Supergroup in New Jersey, *in* Manspeizer, W., ed., Field Studies of New Jersey Geology and Guide to Field Trips, 52nd Annual Meeting:New York State Geological Association, Rutgers University, Newark, p. 352-398.

Olsen, P.E., 1986, A 40-million-year lake record of early Mesozoic climatic forcing. Science 234, p. 842-848.Olsen, P.E., 1997, Stratigraphic record of the early Mesozoic breakup of Pangea in the Laurasia-Gondwana rift system: Annual Reviews of Earth and Planetary Science, v. 25, p. 337-401.

Olsen, P.E., 1999, Giant lava flows, mass extinctions, and mantle plumes: Science v. 284, p. 604-605.

Olsen, P.E., 2001, Grand cycles of the Milankovitch band: Eos Transactions, American Geophysical Union, Supplement, v. 82, no. 47, Abstract U11A-11, p. F2.

Olsen, P.E., Fowell, S.J. and Cornet, B., 1990, The Triassic-Jurassic boundary in continental rocks of eastern North America: a progress report, *in* Sharpton, V.L. and Ward, P.D., eds., Global Catastrophes in Earth History: an Interdisciplinary Conference on Impacts, Volcanism, and Mass Mortality: Geological Society of America Special Paper 247, p. 585-593.

Olsen, P.E. and Galton, P.M., 1984, A review of the reptile and amphibian assemblages from the Stormberg of southern Africa, with special emphasis on the footprints and the age of the Stormberg: Palaeontologia Africana, v. 25, p. 87-110.

Olsen, P.E. and Kent, D.V., 1996, Milankovitch climate forcing in the tropics of Pangea during the Late Triassic: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 122, p. 1-26.

Olsen, P.E. and Kent, D.V., 1999, Long-period Milankovitch cycles from the Late Triassic and Early Jurassic of eastern North America and their implications for the calibration of the early Mesozoic time scale and the long-term behavior of the planets: Philosophical Transactions of the Royal Society of London (series A) v. 357, p. 1761-1787.

Olsen, P.E. and Kent, D.V., 2000, High resolution early Mesozoic Pangean climatic transect in lacustrine environments: Zentralblatt für Geologie und Paläontologie for 1998, no. 11-12, p. 1475-1496.

Olsen, P.E., Kent, D.V., Fowell, S.J., Schlische, R.W., Withjack, M.O. and LeTourneau, P.M., 2000, Implications of a comparison of the stratigraphy and depositional environments of the Argana (Morocco) and Fundy (Nova Scotia, Canada) Permian-Jurassic basins, *in* Oujidi, M. and Et-Touhami, M., eds., Le Permien et le Trias du Maroc, Actes de la Premièr Réunion du Groupe Marocain du Permien et du Trias: Hilal Impression, Oujda, p. 165-183.

Olsen, P.E., Kent, D. V., Et-Touhami, M., and Puffer, J. H., 2003b, Cyclo-, magneto-, and bio-stratigraphic constraints on the duration of the CAMP event and its relationship to the Triassic-Jurassic boundary , *in* Hames, W.E., McHone, J.G., Renne, P.R, Ruppel, C., eds., The Central Atlantic Magmatic Province: Insights From Fragments of Pangea: Geophysical Monograph Series, v. 136, p. 7-32.

Olsen, P.E., Kent, D.V., and LeTourneau,P.M., 2002c, Stratigraphy and age of the Early Jurassic Portland Formation of Connecticut and Massachusetts: a contribution to the time scale of the Early Jurassic: Geological Society of America, Abstracts with Programs, v. 34, no. 2, p. A-61.

Olsen, P.E., Et-Touhami, M., and Puffer, J., 2002d, Comparison of the continental syn-basalt (CAMP) Earliest Jurassic age strata of Eastern North American rift basins with coeval continental to marine strata of Morocco: Geological Society of America, Abstracts with Programs, v. 34, no. 2, p. A-31.

Olsen, P.E., Et-Touhami, M., and Whiteside, J.H., 2003c, The initial CAMP flood basalt event and its relationship to overlying carbonate-rich sequences (North America and Northwest Africa): two options related to the Triassic-Jurassic boundary: Geological Society of America, Abstracts with Programs, v. 35, no. 3, p. 89.

Olsen, P.E., Kent, D.V., Cornet, B., Witte, W.K., and Schlische, R.W., 1996a, High-resolution stratigraphy of the Newark rift basin (Early Mesozoic, Eastern North America): Geological Society of America, v. 108, 40-77.

Olsen, P.E., Kent, D.V., Sues, H.-D., Koeberl, C., Huber, H., Montanari, A., Rainforth, E.C., Fowell, S.J., Szajna, M.J., Hartline, B.W.' 2002a, Ascent of dinosaurs linked to Ir anomaly at Triassic-Jurassic boundary: Science, vol. 296, p. 1305-1307.

Olsen, P.E., Kent, D.V., Sues, H.-D., Koeberl, C., Huber, H., Montanari, A., Rainforth, E. C., Fowell, S.J., Szajna, M.J., and Hartline, B.W., 2003a, Ascent of dinosaurs linked to an iridium anomaly at the Triassic-Jurassic boundary (Reply to Thulborn): Science, v. 301, p. 169.

Olsen, P.E., Koeberl, C., Huber, H., Montanari, A., Fowell, S.J., Et Touhami, M. and Kent, D.V., 2002b, The continental Triassic-Jurassic boundary in central Pangea: recent progress and discussion of Ir anomaly: Geological Society of America, Special Paper 356, p. 505-522.

Olsen, P.E. and McCune, A.R., 1991, Morphology of the *Semionotus elegans* species group from the Early Jurassic part of the Newark Supergroup of eastern North America with comments on the family Semionotidae (Neopterygii): Journal of Vertebrate Paleontology v. 11, p. 269-292.

Olsen, P.E., McCune, A.R. and Thomson, K.S., 1982, Correlation of the early Mesozoic Newark Supergroup by Vertebrates, principally fishes: American Journal of Science, v. 282, p. 1-44.

Olsen, P.E. and Rainforth, E.C., 2002a, The Early Jurassic ornithischian dinosaurian ichnogenus *Anomoepus*, *in* LeTourneau, P.M. and Olsen, P.E., eds., The Great Rift Valleys of Pangea in Eastern North America, vol .2: Sedimentology, Stratigraphy, and Paleontology: Columbia University Press, p. 137-140.

Olsen, P.E. and Rainforth, E.C., 2002b, The "Age of Dinosaurs" in the Newark Basin, with Special Reference to the Lower Hudson Valley. *in* Gates, A.E., ed., Geology of the Lower Hudson Valley, 2001: New York State Geological Association Field Trip Guidebook New York State Geological Association, New York State Museum, Albany, v. 73, p. 59-176,

Olsen P.E, Schlische, R.W. and Fedosh, M.S., 1996b, 580 ky duration of the Early Jurassic flood basalt event in eastern North America estimated using Milankovitch cyclostratigraphy, *in* Morales, M., ed., The Continental Jurassic: Museum of Northern Arizona Bulletin 60, p. 11-22.

Olsen, P.E., Schlische, R.W. and Gore, P.J.W., 1989, Field Guide to the Tectonics, stratigraphy, sedimentology, and paleontology of the Newark Supergroup, eastern North America: International Geological Congress Guidebooks for Field Trips T351, 174 p.

Olsen, P.E., Schneider, V., Sues, H.-D., Peyer, K.M. and Carter, J.G., 2001b, Biotic provinciality of the Late Triassic equatorial humid zone: Geological Society of America Abstracts with Programs, v. 33, no. 2, p. 27.

Olsen, P.E., Shubin, N.H. and Anders, M.A., 1987, New Early Jurassic tetrapod assemblages constrain Triassic-Jurassic tetrapod extinction event: Science, v. 237, p. 1025-1029.

Olsen, P.E., Smith, J.B. and McDonald, N.G., 1998, Type material of the type species of the classic theropod footprint genera *Eubrontes*, *Anchisauripus*, and *Grallator* (Early Jurassic, Hartford and Deerfield basins, Connecticut and Massachusetts, USA): Journal of Vertebrate Paleontology, v. 18, p. 586-601

Olsen, P.E., Sues, H.-D. and Norell, M.A., 2000b. First record of *Erpetosuchus* (Reptilia: Archosauria) from the Late Triassic of North America: Journal of Vertebrate Paleontology, v. 20, p. 633-636.

Ostrom, J.H., and Quarrier, 1968, The Rocky Hill dinosaurs, *in* Orville, P. M., ed., Guidebook for fieldtrips in Connecticut. New England Intercollegiate Geological Conference 60th Annual Meeting, New Haven, Conn. State of Connecticut, State Geological and Natural History

Pálfy, J. Demény, A., Hass., J., Hetényi, M., Orchard, M.J., and Veto, I., 2001, Carbon isotope anomaly and other geochemical changes at the Triassic-Jurassic boundary from a marine section in Hungary: Geology, v. 29, p. 1047-1050.

Parnell, J., 1983, Skeletal halites from the Jurassic of Massachusetts, and their significance: Sedimentology, v. 30, no. 5, p. 711-715.

Percival, J.G. 1842, Report on the geology of the state of Connecticut: New Haven, Connecticut, Osborn and Baldwin pub., 495 p.

Philpotts, A.R., 1992, A model for emplacement of magma in the Mesozoic Hartford basin; *in* J. H. Puffer and P. C. Ragland, Geological Society of America Special Paper 268, p. 137-148.Philpotts, A.R. and Asher, P.M., 1992, The first Mesozoic magma of the Hartford Basin: examination of the feeder dike, a laccolith, and the Talcott lava flow; *in* Robinson, Peter and Brady, J.B., eds., Guidebook for field trips in the Connecticut Valley region of Massachusetts and adjacent states: New England Intercollegiate Geolo 84th Annual Meeting, Amherst, Massachusetts, vol. 1, trip A-6, p. 67-85.

Philpotts, A.R. and Martello, A., 1986, Diabase feeder dikes for the Mesozoic basalts in southern New England: American Journal of Science, v. 286, p. 105-126.

Prevot, M. and McWilliams, M., 1989, Paleomagnetic correlation of Newark Supergroup volcanics: Geology, v. 17, p. 1007-1010.

Puffer, J.H., Hurtubise, D.O., Geiger, F.J., and Lechler, P., 1981, Chemical composition and stratigraphic correlation of the Mesozoic basalt units of the Newark basin, New Jersey, and the Hartford basin, Connecticut. Geological Society of America Bulletin, v. 92, part 1, p. 155-159, part 2, p. 515-553.

Puffer, J.H. and Student, J.J., 1992, Volcanic structures, eruptive styles, and post eruptive deformation and chemical alteration of the Watchung flood basalts, New Jersey, *in* Puffer, J.H. and Ragland, P.C., eds., Eastern North American Mesozoic Magmatism: Geochemistry, Structure, and Stratigraphy. Geological Society of America Special Paper 268, p. 261-277.

Rainforth, E.C., 2003. Revision and reevaluation of the early Jurassic dinosaurian ichnogenus Otozoum: Palaeontology, v. 46, p. 803-838.

Rampino, M.R. and Caldeira, K., 1993, Major episodes of geologic change; correlations, time structure and possible causes: Earth and Planetary Science Letters, v. 114, p. 215-227.

Retallack, G.J., 2001, A 300-million-year record of atmospheric carbon dioxide from fossil plant cuticles: Nature, v. 411, p. 287-290.

Rice, W.N., 1886, On the trap and sandstone in the gorge of the Farmington River at Tariffville, Connecticut: American Journal of Science, v. 32, ser. 3, p. 94-112.

Rizzo, C. A., 1999a, A large coelacanth c.f. *Diplurus* (*Pariostegus*) *longicaudatus*, from the Late Triassic Lockatong Formation, Granton Quarry, North Bergen, New Jersey: The Mosasaur v. 6, p. 85-90.

Rizzo, C.A., 1999b, Evidence for live birth in the Triassic coelacanth *Diplurus* (*Osteopleurus*) *newarki*: The Mosasaur, v. 6, p. 91-95.

Rogers, J., Gates, R. M., and Rosenfield, J.L., 1959, Explanatory text for the preliminary geological map of Connecticut, 1956. Connecticut Geological and Natural History Survey Bulletin 84, 54 p.

Sanders, J.E., 1963, Late Triassic tectonic history of northeastern United States, American Journal of Science v. 261, p. 501-524.

Sanders, J.E.1968, Stratigraphy and primary sedimentary structures of fine-grained, well-bedded strata, inferred lake deposits, Upper Triassic, central and southern Connecticut. Geological Society of America Special Paper 106, p. 265-305.

Sanders, J.E., 1970, Stratigraphy and structure of the Triassic strata in the Gaillard graben, south-central Connecticut: Connecticut Geological and Natural History Survey, Guidebook, no. 3, 15 p.

Schlische, R. W., and Ackermann, R. V., 1995, Kinematic significance of sediment-filled fissures in the North Mountain basalt, Fundy rift basin, Nova Scotia, Canada: Journal of Structural Geology, v. 17, p. 987-996.

Schlische, R.W. and Olsen, P.E., 1990, Quantitative filling models for continental extensional basins with applications to the early Mesozoic rifts of eastern North America: Journal of Geology, v. 98, p. 135-155.

Shubin, N.H. and Sues, H.-D., 1991, Biogeography of early Mesozoic continental tetrapods: patterns and implications. Paleobiology v. 17, p. 214-230.

Smoot, J.P., 1991, Sedimentary facies and depositional environments of early Mesozoic Newark Supergroup basins, eastern North America: Palaeogeography, Palaeoclimatology, Palaeoecology v. 84, p. 369-423.

Spray, J.G., Kelley, S.P. and Rowley, D.B., 1998, Evidence for a Late Triassic multiple impact event on Earth: Nature, v. 392, p. 171-173.

Steinen, R.P., Gray, N.H., and Mooney, J., 1987, A Mesozoic carbonate hot-spring deposit in the Hartford Basin of Connecticut: Journal of Sedimentary Petrology, v. 57, p. 319-326.

Sues, H.D., and Baird , D., 1993, A skull of a sphenodontian lepidosaur from the New Haven Arkose (Upper Triassic: Norian) of Connecticut: Journal of Vertebrate Paleontology, v. 13, p. 370–372.

Sues, H.-D., Olsen, P.E. and Shubin, N.H., 1987, A diapsid assemblages from the Lower Jurassic of Nova Scotia, Canada, in Currie, P.J. and Koster, E.H., eds., 4th Symposium on Mesozoic Terrestrial Ecosystems. Occasional Papers of the Tyrell Museum of Palaeontology, v. 3, p. 205-207.

Sues, H. D. and Olsen, P. E., 1999, A Late Triassic traversodont cynodont from the Newark Supergroup of North Carolina: Journal of Vertebrate Paleontology v. 19. no. 2, p. 351-354.

Sues, H.-D., Olsen, P.E., Scott, D.M. and Spencer, P.S., 2000, Cranial osteology of H*ypsognathus fenneri*, a latest Triassic procolophonid reptile from the Newark Supergroup of eastern North America: Journal of Vertebrate Paleontology v. 20, p. 275-284.

Tanner, L.H., Hubert, J.F., Coffey, B.P., and McInerney, D.P., 2001, Stability of atmospheric CO_2 levels across the Triassic/Jurassic boundary: Nature. v. 411, p. 675-677.

Tschudy, R.H., Pilmore, C.L., Orth, C.J., Gilmore, J.S., and Knight, J.D., 1984, Disruption of the terrestrial plant ecosystem at the Cretaceous-Tertiary boundary, western interior: Science v. 225, p. 1030-1032.

Vakhrameev, V.A., 1991. Jurassic and Cretaceous floras and climates of Earth: Cambridge University Press, 318 p.

Van Houten, F.B., 1962, Cyclic sedimentation and the origin of analcime-rich Upper Triassic Lockatong Formation, west-central New Jersey and adjacent Pennsylvania: American Journal of Science, v. 260, p. 561-576.

Van Houten, F.B., 1964, Cyclic lacustrine sedimentation, Upper Triassic Lockatong Formation, central New Jersey and adjacent Pennsylvania: Kansas Geological Survey Bulletin 169, p. 497-531.

Van Houten, F.B., 1969, Late Triassic Newark Group, north central New Jersey, and adjacent Pennsylvania and New York, *in* Subitzky, S.S. ed., Geology of Selected Areas in New Jersey and Eastern Pennsylvania: Rutgers University Press, p. 314-347.

Van Houten, F.B., 1980, Late Triassic part of the Newark Supergroup, Delaware River section, west-central New Jersey, *in* Manspeizer, W., ed., Field Studies of New Jersey Geology and Guide to Field Trips: 52nd Annual Meeting, New York State Geological Association, Rutgers University, p. 264-276.

Wang, Z.S., Rasbury, E.T., Hanson, G.N., and Meyers, W.J., 1998, Dating the time of sedimentation of clastic sedimentary rocks using U-Pb system of calcretes: Geochimica et Cosmochimica Acta, v. 62, p. 2823-2835.

Ward, P.D., Haggart, J.W., Carter, E.S., Wilbur, D., Tipper, H.W., and Evans, T. 2001, Sudden productivity collapse associated with the Triassic-Jurassic boundary mass extinction: Science, v. 292, p. 1148-1151.

Watson, J., 1988 The Cheirolepidiaceae, *in* Beck, C.B., ed., Origin and Evolution of Gymnosperms: Columbia University Press, p. 382-447.

Whiteside, J.H., and Olsen, P.E., 2003, Possible Triassic-Jurassic boundary sections, Hartford Basin, Eastern North America, *in* Geological Society of America, Abstracts with Programs, v. 35, p. 84.

Wolfe, K.H., M. Gouy, Y.-W. Yang, P.M. Sharp and W.-H. Li., 1989, Date of the monocot-dicot divergence estimated from chloroplast DNA sequence data: Proceedings of the National Academy of Sciences U.S.A., v. 86. p. 6201-6205.