Fossil Great Lakes of the Newark Supergroup – 30 Years Later

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ABSTRACT

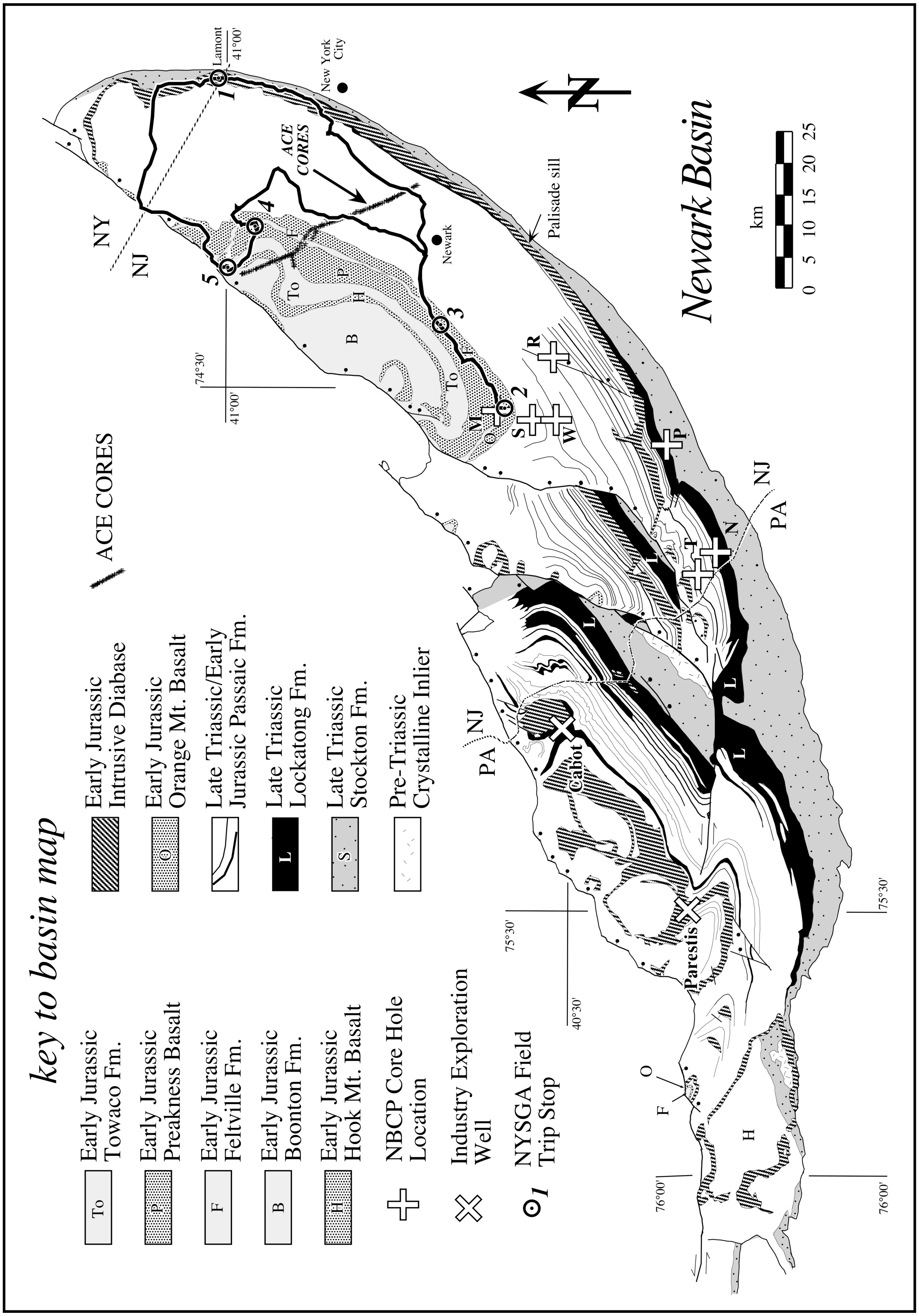
In the 1980 NYSGA meeting (Olsen, 1980b) I led a fieldtrip to look at strata of the Triassic and Jurassic great lakes of the Newark basin which are among the largest lakes known of any time. Our view of these giant lake systems has been much amplified since that time because of the collection of industry seismic lines, the drilling of two deep exploration holes, about 40,000 ft of core by the Army Corps of Engineers, 26,700 ft of core by the Newark Basin Coring Project (NBCP), and quite a few Ph.D. theses (including my own). The purpose of this fieldtrip is to examine outcrops and related cores in the Newark basin, that illustrate some of the key features of these giant lake strata and interbedded flood basalts and how they play key roles in current debates on climate change, mass extinctions, evolution, ascent of the dinosaurs, the rifting process, carbon sequestration, and chaos in the Solar System.

INTRODUCTION: THE TRIASSIC-JURASSIC GREAT LAKES OF CENTRAL PANGEA

This guidebook focuses of the deposits, fossils, and context of huge lakes, related sedimentary deposits, and giant lava flows formed during latest Triassic and Early Jurassic in the Newark basin of New York, New Jersey, and Pennsylvania, one of a series of remnants of a rift valleys system preserved in Eastern North America (Figure 1). I define the term Great Lakes as meaning lakes whose dimensions during high stands were on the scale of over 100 km in longest dimension comparable to the scale of the American Great Lakes or the East African Great Lakes and perhaps approaching the scale of land-licked seas such as the Black or Caspian seas. This introduction discusses in general terms the major themes that guided the selection of the stops and which inform the impotance of observations we can make on the outcrop.

As should be expected ideas have changed considerably in 30 years and new observations, especially from cores, have molded new ones. As much as possible I will try to point out how ideas have changed and also how interpretations of outcrops have changed since 1980 (Olsen, 1980b).

This contribution is dedicated to Warren Manspeizer, editor of the 1980 NYGSA guidebook who encouraged a generation of scientists, including me, intellectually in general and urged me specifically to write two papers for that guidebook (Olsen, 1980a, 1980b).



key to basin map
Early Jurassic Towaco Fm.
Early Jurassic Preakness Basalt
Early Jurassic Feltville Fm.
Early Jurassic Boonton Fm.
Early Jurassic Hook Mt. Basalt
NBCP Core Hole Location
Industry Exploration Well
NYSGA Field Trip Stop
Early Jurassic Intrusive Diabase
Early Jurassic Orange Mt. Basalt
Late Triassic/Early Jurassic Passaic Fm.
Late Triassic Lockatong Fm.
Late Triassic Stockton Fm.
Pre-Triassic Crystalline Inlier
ACE CORES
Newark Basin
km
0 5 10 15 20 25
Palisade sill
New York City
Newark
Lamont
NY
NJ
PA
Cabot
Parestis
41°00'
74°30'
40°30'
75°30'
76°00'

The Rifting Process and Origin of the Atlantic Ocean

The rifting process that eventually resulted in the formation of the central part of the Atlantic Ocean was protracted over a period of at least 40 million years from the Late Permian to the Early Jurassic, with at least one spectacular and probably catastrophic punctuation (the CAMP eruptions). The rifting zone itself[1] is the largest we know of on our planet, arguably stretching roughly 8000 km northeast-southwest from Greenland and Europe to the Gulf of Mexico and the Pacific and north west-southeast about 1000 km (Figure 2). In this rifting zone there were many, perhaps hundreds, large, mostly normal, fault systems along which there was geologically significant hanging wall subsidence leading to the development of large rift valley systems, many of which were individually hundreds of kilometers long, and that filled with thousands of meters of sediments. Over much of this rifting region bisecting central Pangea, the rifts formed far from access to the sea and hence when subsidence exceeded the rate at which sediments could fill them, which apparently was most of the time, lakes, great lakes filled the deficit. Thus, in most cases the sediments filling the individual rift basins were completely of continental origin, and largely lacustrine. As far as we know, only adjacent to the Tethys and along the north-central axis of the rifting zone did marine brines or open marine waters make it into the rift basins, and most of that occurred late in the Triassic-Jurassic part of the rifting process.

As is the case with rift systems today, many of the individual rift basin systems within this huge rifting zone were almost certainly connected with one another not only by river systems, but also by veneers of alluvial strata shed from protruding footwall basement or inselbergs remaining from pre-rift topography. At times of extreme lake level highs many of the otherwise isolated great lakes may have become united into lakes much larger than the depocenters marking the main individual rift basins systems forming lakes the scale of the Caspian Sea as illustrated by the stops on this field trip.

The CAMP LIP

While the rifting process itself was protracted, it was marked by one brief but immense basaltic igneous event involving the emplacement of the Central Atlantic Magmatic Province (CAMP) (Marzoli et al., 1999) that may have lasted only about 600 ky (Olsen et al., 1993, 2003). The intrusive system of the CAMP is spread from France through Iberia, Eastern North America, West Africa, to western Brazil, covering and area

Figure 1: (previous page). Map of the Newark basin showing location of field stops, trip route, and wells and cores.

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[1] It is useful to make the distinction between the broad rifting zone, which is the entire region of thinning crust with dimensions of thousands of kilometers cut by swarms of normal faults, and the many individual rift basins and rift basin systems with dimensions of hundreds of kilometers or smaller that formed along the largest of the normal fault systems. In context of our current understanding of these rifting zones, the concept of "failed rifts" or "aulacogen" as a "failed arm" of a triple junction makes no sense, as is also true of the concept of an "axial dike" in a rift basin.

of about 11 million square kilometers. The dikes form an apparent radial pattern (May, 1971) the locus of which is in southern Florida or the Bahamas (Figure 3). Plausibly this radial swarm reflects the stress pattern along a thermal welt caused by the plume head. However, the existence of the radial pattern itself has been questioned by a number of authors (McHone, 2000; Coltice et al., 2009), but it as close to a radial pattern as seen anywhere in our Solar System, notably Venus (Figure 4) (e.g., Ernst et al., 2001; Hansen & Olive, 2010). This radial pattern may be the consequence of the arrival of a plume head (White & McKenzie, 1989) as is supposed to be the case for most other large igneous provinces (LIPs), although this concept is also strongly questioned (Anderson, 2000)[2].

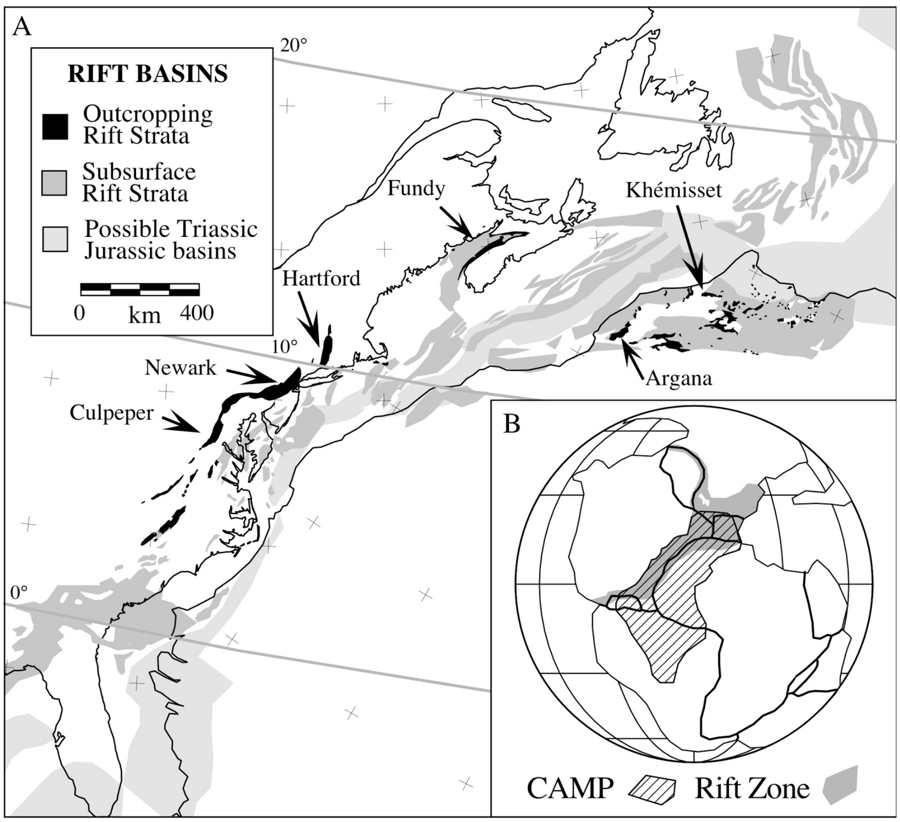

Figure 2: Distribution of rift basins of central Pangea and the distribution of the CAMP. A, Rift basins of the central Atlantic margins (for the Late Triassic); B, Pangea during the earliest Jurassic showing the distribution of the CAMP, overlapping much of the Triassic-Jurassic rift zone.

[2] One of the strongest arguments against the plume model for the CAMP has been the apparent absence of a "hot spot" track. It is clear that there is no such track on the North American plate. However, this is problem is illusionary because the North American Plate drifted almost straight northward after the CAMP and most of the hot spot track would be on the continental part of South American Plate, in northern South America, which did not move much during the Mesozoic, and where there are basaltic eruptives of appropriate age. Indeed modern plate reconstructions (e.g., Schettio et al., 2010) predict the current position of the CAMP hot spot to be at the Cape Verde Islands as suggested by (White and McKenzie, 1989; Oliveira et al., 1990; Ernst et al., 1995), which is of the right age, or to me less plausibly the Fernando de Noronha hot spot (Hill, 1991).

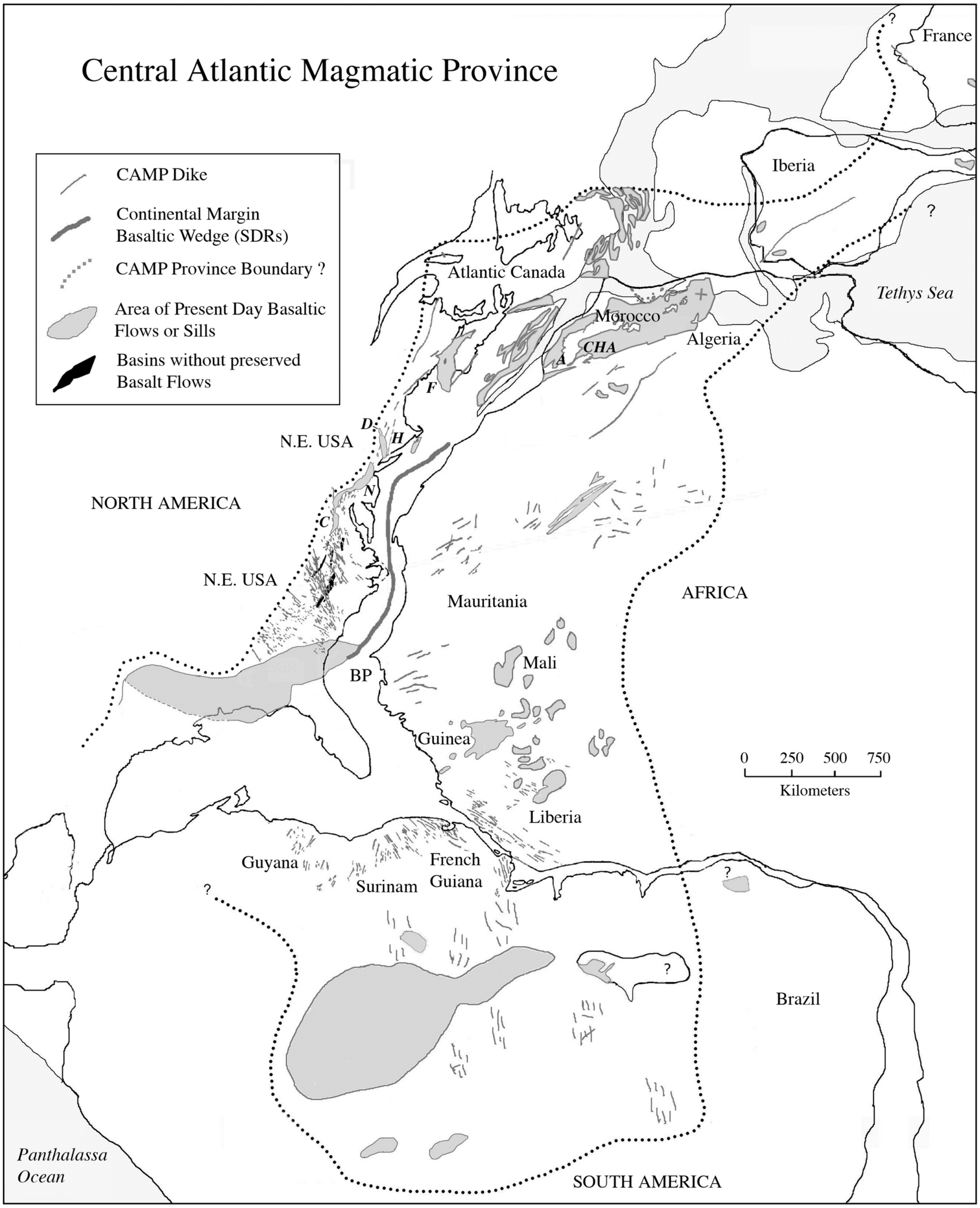

Figure 3: Distribution of the CAMP in central Pangea showing basins described in text. Modified from McHone (2000) and Whiteside et al. (2007).

The CAMP lava flows and other extrusives emanating from the intrusive system are known over a much smaller area than the intrusions, but this is certainly a result of extensive post-rift erosion. If the flows covered most of the area marked by the dike system the CAMP lavas would comprise the largest known continental flood basalt province on Earth, although whether they really covered such a large area remains to be tested (e.g., Olsen et al., 1999).

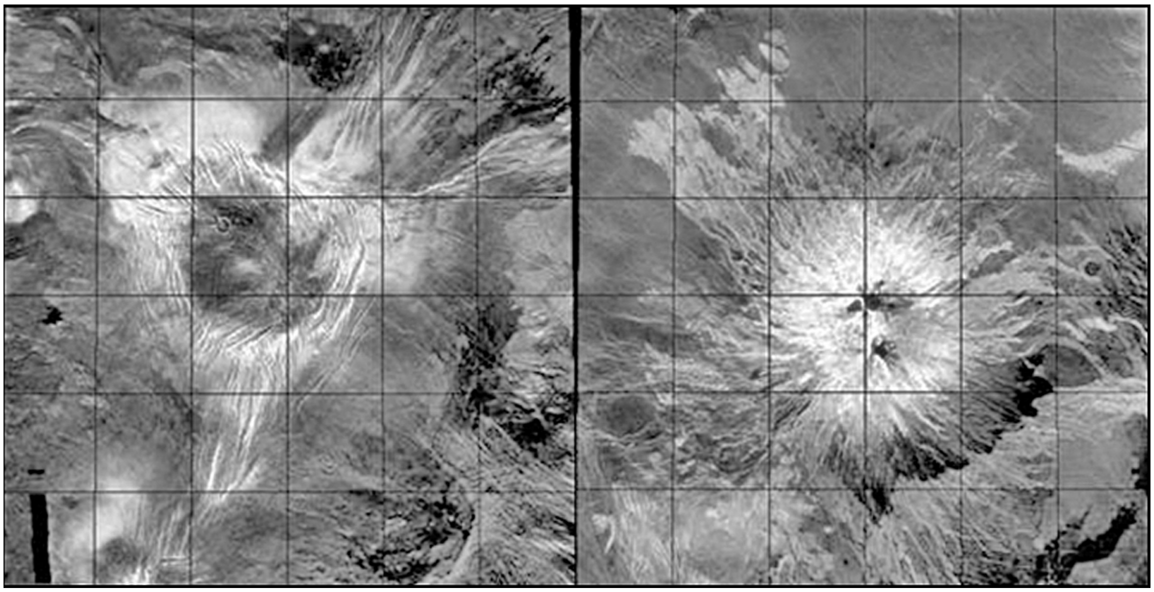

Figure 4: Radial dike swarms on Venus at Irnini Mons (left) and Sapas Mons (right). Note that these swarms are not purely radial, but rather have sets of parallel dikes and intersecting systems with different orientations. NASA Magellan Synthetic Aperture Radar (SAR) mosaic images from http://www.geology.pomona.edu/research/Faculty/ Grosfils/Venus/Volcano/large_ volcanoes.htm thus undermining the argument that such features in the CAMP dikes indicate the *lack* of a radial dike swarm and hence the lack of a plume.

CAMP igneous activity may also mark the initiation of Atlantic sea floor spreading (Schlische et al., 2003; Schlische, 2003; Olsen et al., 2003). What appear to be CAMP lavas (the Clubhouse Crossroads Basalt) are present in the subsurface in South Carolina and Georgia, and based on seismic reflection profiles, these seem to connect to the giant wedge shaped edifices of seaward dipping reflectors below the southeastern continental shelf (Austin et al., 1990; Holbrook et al., 1994; Kelemen & Holbrook, 1995; Oh et al., 1995; Talwani et al., 1995). The latter are taken to be subareally emplaced lavas formed at the onset of seafloor spreading as first identified in offshore Norway (Mutter et al., 1982). An accurate direct date of the Clubhouse Crossroads Basalt is still elusive, however, and the attractive hypothesis that the CAMP lavas mark the onset of seafloor spreading has yet to be critically and directly tested.

However, following the assumption that the seaward dipping reflectors are part of the CAMP to its logical conclusion, it seems plausible that the arrival of the CAMP plume head initiated the spread of magma, not only along the radial dike swarm, but perhaps also along crest of asthenosphere along the thinnest zone of crust approximating the axis of the rifting zone, setting off the extremely concentrated eruption of the seaward dipping reflectors, and initiating seafloor spreading. This axial asthenospheric crest would have been perpendicular to the minimum compressive stress direction on the northeastern side of the thermal welt. The thermal welt should also have produced maximum compressive stresses perpendicular to the normal fault system in the southeastern US, perhaps both terminating extension and causing small-scale tectonic inversion features consistent with the observed patterns seen in that region (e.g., Schlische et al., 2003; Maliconico, 2003), while at the same time accelerating continental extension further to the north along the rifting axis.

Critical to all arguments about the effects of the CAMP eruption are not just the huge volume of igneous rocks involved emplaced in a geologically relatively short period, but specifically the rate and magnitude of individual eruptive events. We will see some evidence at Stop 4 that individual lava flows spanned distances of minimally 800 km along the trend of the rifts and were in places over 200 m thick, making them among the largest if not the largest single lava flows known on Earth.

Cyclical Climate Change Reflected in the Deposits of Great Lakes.

Perhaps the single best-known feature of the Newark basin (and other Newark Supergroup basins) is the nearly pervasive sedimentary cyclicity in the abundant lacustrine deposits. This cyclicity was first identified implicitly by the mapping work of Dean B. McLaughlin of the University of Michigan in the 1930s through 1950s (e.g., McLaughlin, 1933, 1944, 1946a, 1946b, 1959) and explicitly by Franklyn Van Houten of Princeton University in the 1960s (Van Houten, 1962, 1964, 1969) and who's later works included a NYGSA guidebook paper (Van Houten, 1980) in the Manspeizer (1980) volume. Both workers, especially Van Houten, had a profound effect on my thinking and practice of stratigraphy (e.g., Olsen, 1980b).

Subsequent studies have confirmed and elaborated on the seminal works of McLaughlin and Van Houten, using quantitative methods of times series analysis that avoid the need to explicitly identify the cycles in advance (e.g., Olsen, 1996a, 1999). After application of an age model to the thickness periodicities, not only does do these methods recover the ~20 ky cycle, but also the ~100 and 405 ky cycles identified by Van Houten by counting cycles, as well as cycles with period of 1.8 and 3.5 m.y. In 1986 and 1996a, I named two the sedimentary cycles present in the Newark lacustrine strata after Van Houten and McLaughlin (Figure 5). Van Houten cycles are the meter-scale lake level cycle obvious at outcrop-scale paced by the ~20 ky climatic precession cycle and McLaughlin cycles are the map-scale cycles traced out by McLaughlin himself (Olsen, 1986; Olsen et al., 1996a) that were paced by the 405 ky eccentricity cycle. Van Houten cycles are lake level cycles controlled by precipitation and evaporation changes tracking the intensity of tropical insolation[3] (Olsen & Kent, 1996). The overall degree of

[3] The mechanism is thought to be that changes in tropical insolation (solar radiation energy received on a given surface area in a given time – watts/meter/unit time^2), controlled by celestial mechanical control of the Earth's orbital geometry, changes the intensity of upwelling warm air (tropical convergence) and hence precipitation. It rains, more or less, during the time and place of maximum insolation with the magnitude of precipitation being positively related to the magnitude of insolation.

development of wet and dry sedimentary facies (Figure 6) and facies-sensitive fossils in Van Houten cycles vary in their expression and in doing so comprise the four larger cycles. These are: the short modulating cycle, corresponding to ~100 ky eccentricity cycles; the McLaughlin cycle, corresponding to the 405 ky cycle, and the two additional eccentricity cycles of roughly 1.8 and 3.5 m.y. (Olsen & Kent, 1996, 1999).

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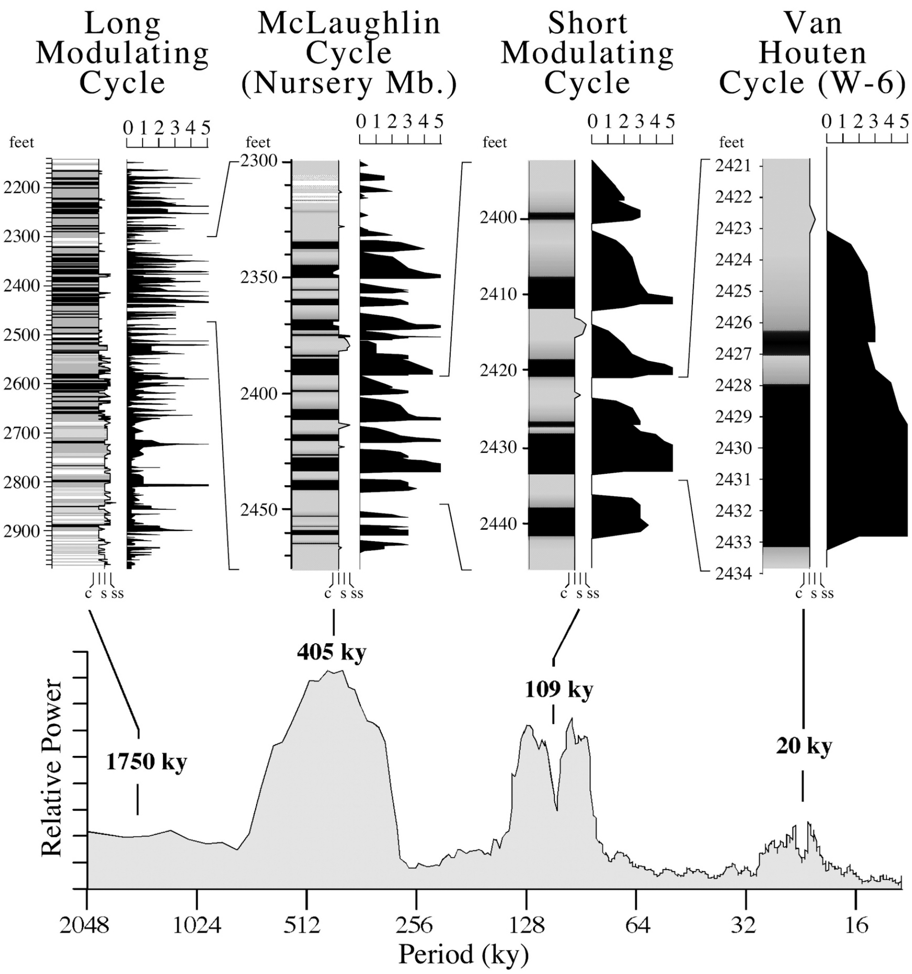

Figure 5: Lacustrine cycles of the Newark basin (above) from the Nursery no. 1 core and power spectrum tuned to the 405 ky cycle. White represents red , very light gray is purple, gray is gray, and black is black.

McLaughin never seemed to state an opinion in print of the origin of the cycles he mapped, although he and Van Houten were contemporaries and Van Houten correctly ascribed them to what we more precisely identify today as a 405 ky cycle caused by the gravitational interaction of Venus and Jupiter. This is somewhat ironic because McLaughlin was a professor of astronomy at the University of Michigan and in those circles is known for his work on volcanism on Mars, novae, and Be class stars.

Fossils found in Newark lacustrine strata occur in predicable facies within the context of the Van Houten cycles tracking the climatic precession cycle and the modulating longer term cycles paced by eccentricity (Figure 6). In particular, articulated fish tend to occur in the microlaminated facies in the deepest-water phases of the Van

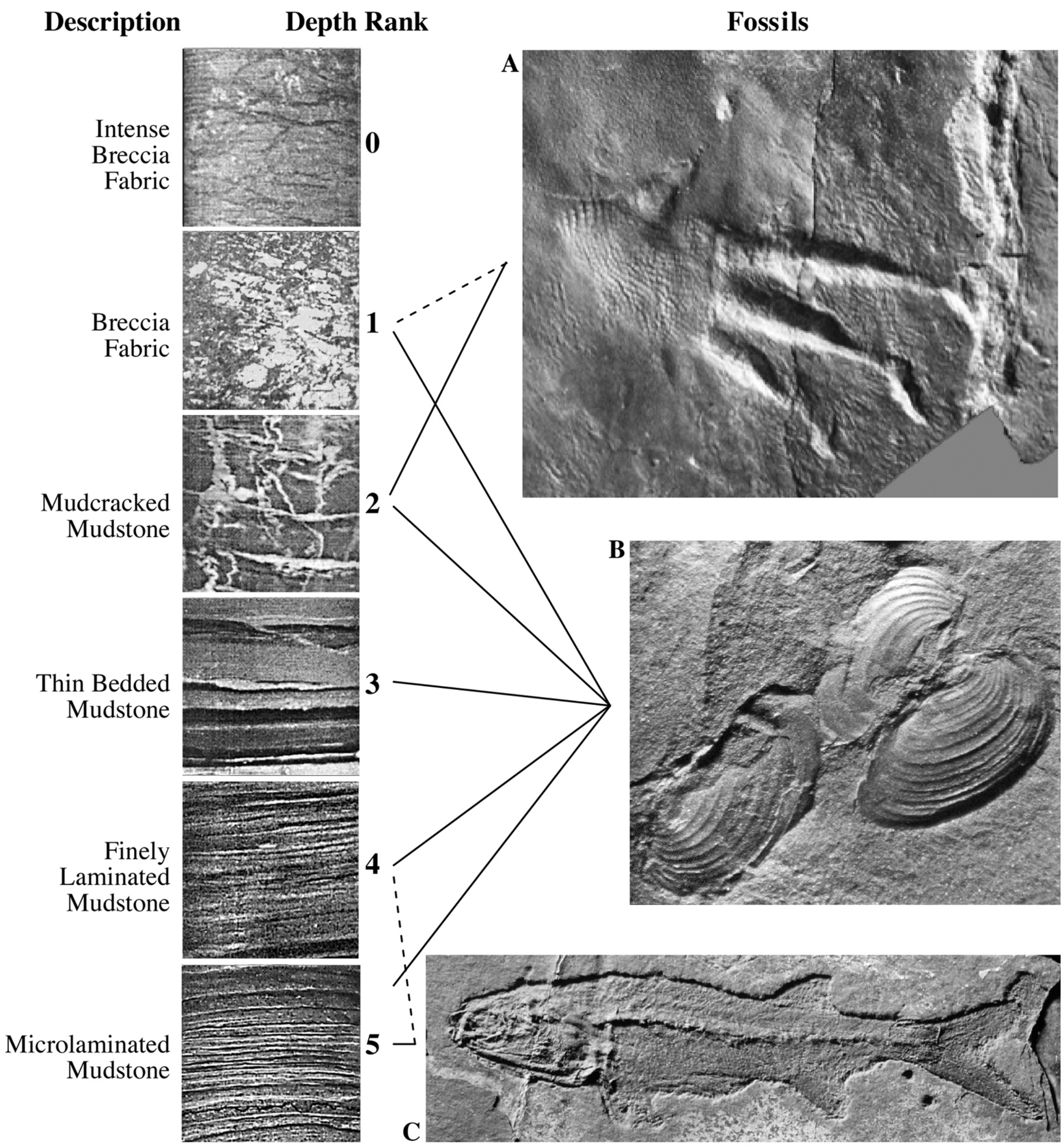

Figure 6: Sedimentary fabrics and fossils associated with various depth ranks (from Olsen & Kent, 1996). A, *Rhynchosauroides* cf. *hyperbates* from the Lockatong Formation; B, conchostracans from T-8 in the Argana basin; C, *Turseodus* sp. From the Lockatong Formation.

Houten cycles occurring during times of maximum precession variance. In the same cycles, footprints tend to occur in the shallowest water facies. In contrast, footprints tend to occur in the "deepest" water facies of those Van Houten cycles formed during times of lower precession variability, while few fossils at all are found times of low precessional variability in the overall 405 ky cycle.

The spectrum of Milankovitch cycles present in the Newark basin has allowed the development of a timescale for the Late Triassic (the Newark Basin Astronomically calibrated Geomagnetic Polarity Time Scale NBAGPTS: Figure 7). The historical development of this times scale is detailed in Olsen et al. (2010). The Milankovitch pattern and the NBAGPTS have withstood several independent quantitative tests

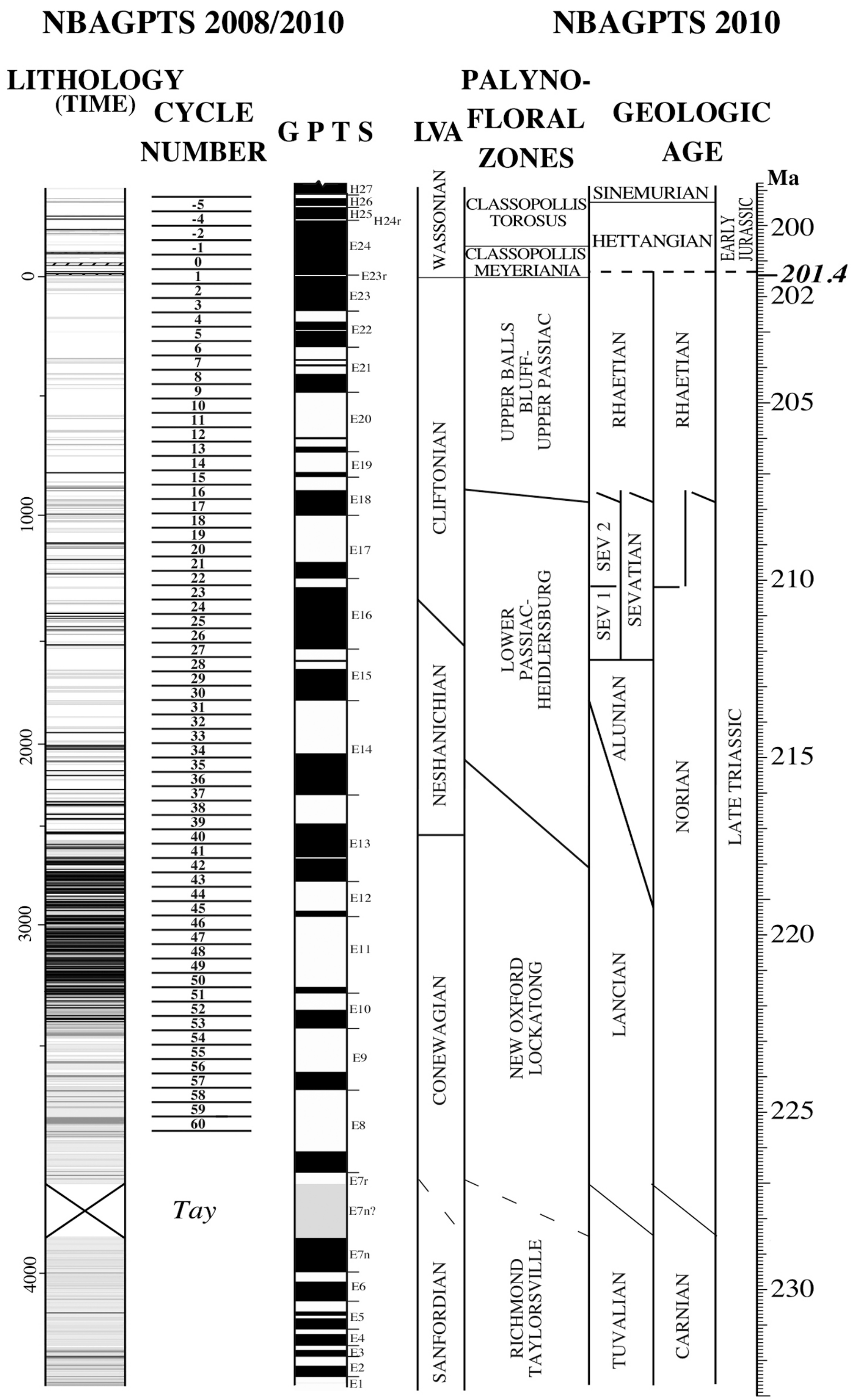

Figure 7: Newark basin astronomically calibrated geomagnetic polarity timescale (NBAGPTS). Modified from Olsen et al.. 2010.

(Kominz & Bond, 1990; Kominz et al., 1991; Baily & Smith, 2008) and independent radioisotopic ages registered to the NBAGPTS (Figure 8) via paleomagnetic polarity correlation (Furin et al., 2006; Olsen et al., 2010) as well as via ages from within the Newark itself (Rasbury et al. 2003; Blackburn et al., 2009). The NBAGPTS has also allowed determination of the duration of the stages of the Late Triassic (Muttoni et al., 2004, 2009; Channell et al., 2003) and earliest Jurassic (Kent & Olsen, 2009), as well as a guide to the duration of events during the Late Triassic and Early Jurassic such as the mass extinction near the close of the Triassic (Olsen et al., 2002a; Whiteside et al., 2007, 2010a), and duration of the CAMP episode (Olsen et al, 1996b, 2003).

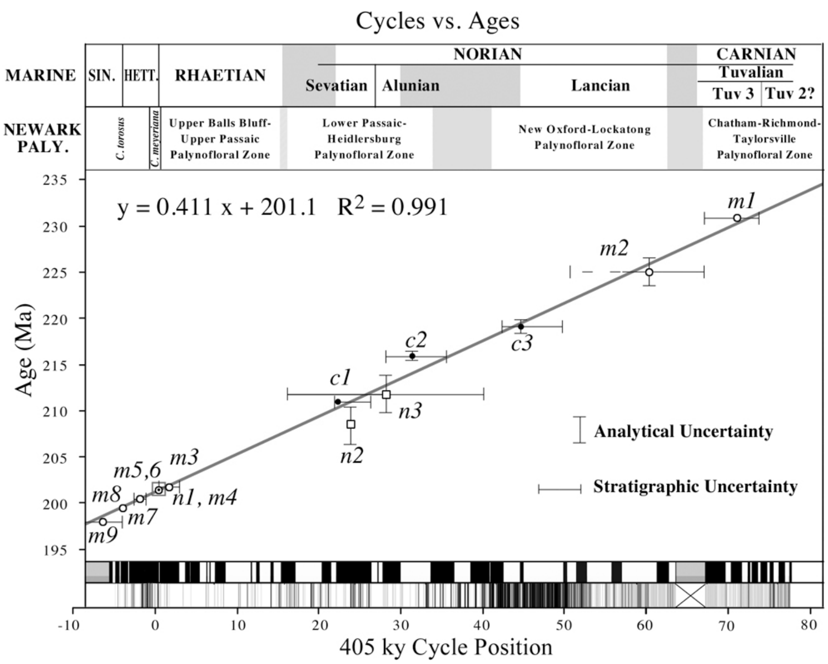

Figure 8: Correlation between number of putative 405 ky cycles in the Newark and Hartford basin sections (from Newark-APTS 2010) and 206Pb/238U ages with linear regression yielding a slope of 411±11 ky / cycle indistinguishable from the hypothesized 405 ky duration of the cycle. Ages are from marine sections (circles, m labels), the Newark and Fundy basins (squares, n labels), and the Chinle Formation (black dots, c labels). See Olsen et al., 2010 for list of ages used and their sources.

Size of the Great Lakes

Core and outcrop analysis of the lacustrine sequences in the Newark basin as well as other Newark Supergroup basins leaves little doubt that the high stands sequences of Van Houten cycles can be traced across the basin (Olsen et al., 1996). For the Newark basin, for Triassic strata this indicates the minimum area of these lakes was in the order of 7400 km^2 assuming the lake extended to the present edges of the basin exclusive of the "narrow neck". However, the Newark basin is deeply eroded (1-6 km: Malinconico, 2009), and its original size must have been considerably larger tan at present, but how much larger is a subject for debate.

The broad terrane hypothesis of Russell (1880) postulates that the Newark and Hartford basins, and other pairs of Newark Supergroup basins with boarder fault systems of opposing polarity were originally the sides of a full graben system that was postdepositionally arched along the graben axis and deeply eroded. In the original version of this hypothesis, the present geometry of the basins as half graben with strata tilting towards the border fault systems would be completely postdepositional. This concept was elaborated in a more modern form by Sanders (1963) who nonetheless retained the basic

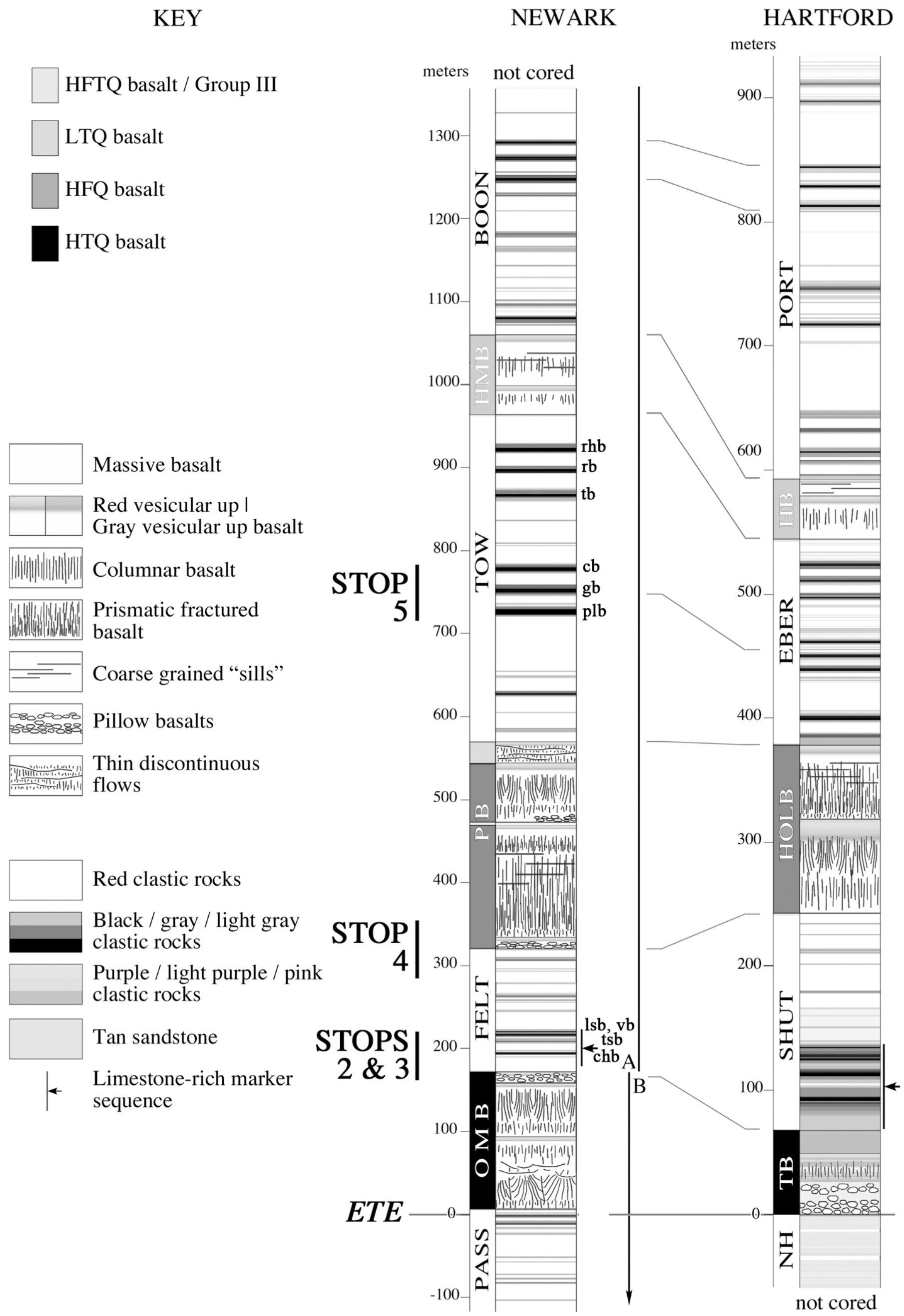

Figure 9: Measured Triassic-Jurassic sections from the Newark and Hartford basins showing the field trip stops and distribution of lacustrine strata of different colors. Darker colors generally conform to units exhibiting deeper water lacustrine facies, while more purple and red colors correspond to shallower and drier lacustrine facies, except in the cases of the New Haven (NH), Sugarloaf formations (SA), and Fall River (FB) beds, which are in part fluvial, and the white limestones that can also be partially fluvial. Abbreviations for lithological units are: BOON, Boonton Formation; EBER, East Berlin Formation; FELT, Feltville Formation; HB. Hampden Basalt; HMB, Hook Mountain Basalt; HOLB, Holyoke Basalt; NH, New Haven Formation; OMB, Orange Mountain Basalt; PASS, Passaic Formation; PB, Preakness Basalt; PORT, Portland Formation; SHUT, Shuttle Meadow Formation; TB, Talcott Formation; TOW, Towaco Formation; Abbreviations for measured sections are: A, Martinsville no. 1 core; B, ACE, Army Corps of Engineers cores (from Olsen et al., 1996b) Arrows with bar at apex are significant carbonate rich intervals. From Whiteside et al. (2007). Bed designations are: cb, Colfax Bed; chb, Cataract Hollow Bed; gb, Glen Bed; lsb, Lake Surprise Bed; plb, Pines Lake Bed; rb, Roseland Bed; rhb, Riker Hill Bed; tb, Totowa Bed; tsb; Trailside Bed; vb, Vosseller Bed.

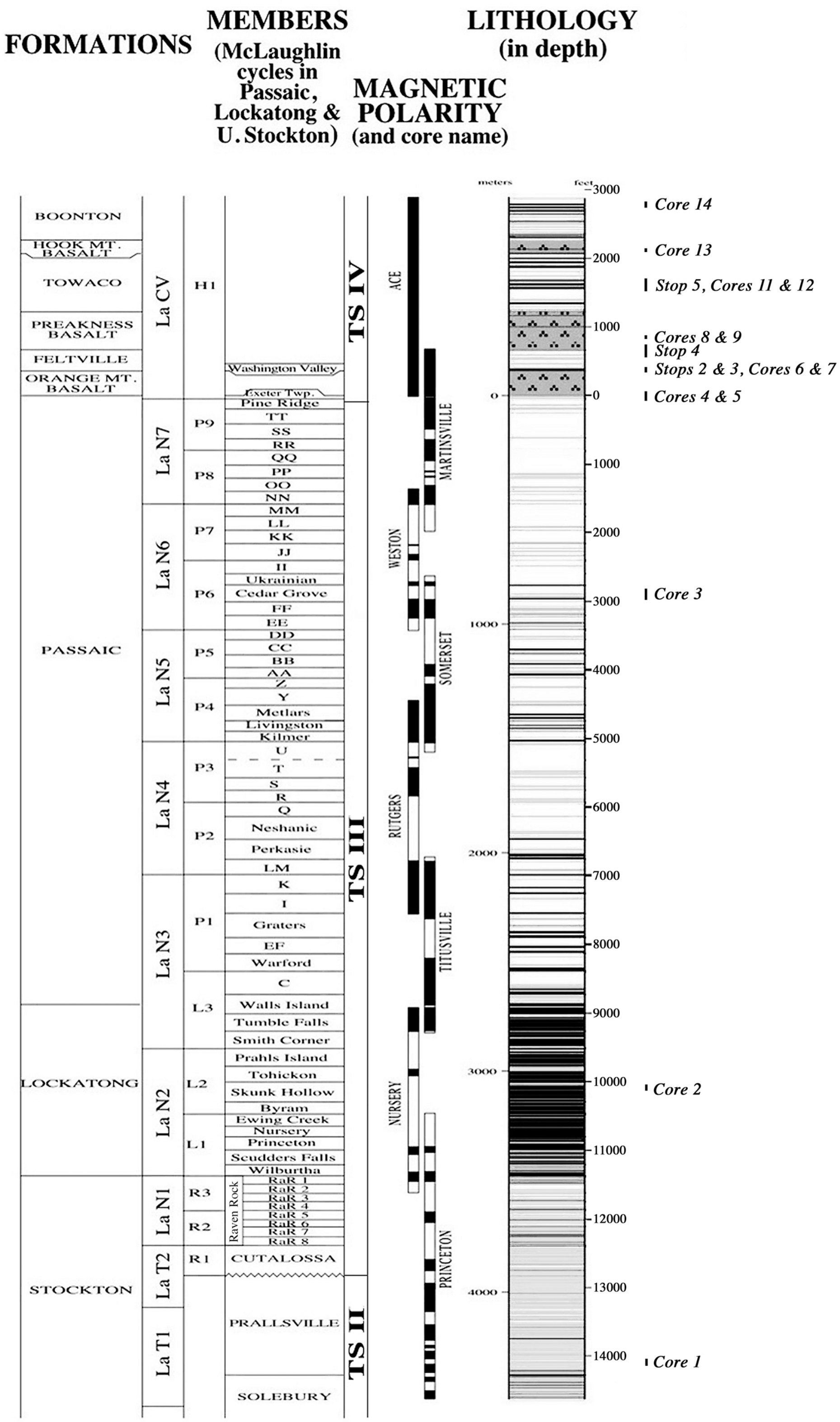

Figure 10: Newark basin sections showing position of stops and cores (Stop 1).

form of the idea. Ample data from industry reflection seismic profiles, correlation between cores, drill holes, and outcrops summarized by Olsen (1997), Schlische (2003), Schlische & Withjack (2005), and Withjack & Schlische (2005) shows that strata in the Newark and Hartford basin exhibit divergence towards the border fault systems and hanging wall onlap, demanding syndeposition assymetrical accommodation space growth (see Stops 1 – 3). In addition, the small Pomperaug basin in Connecticut has a very condensed section of all units relative to the Newark and Hartford basins and has clasts with local crystalline basement provenance (LeTourneau& Huber, 2006), which is inconsistent with its position relative to the original broad terrane model. As originally proposed the broad terrane model cannot be supported.

However, the degree to which the now isolated basin depositional centers were connected by continuous depositional systems during the Triassic and Jurassic remains largely unknown because of deep erosion. The same is true of the interconnectedness of the lakes themselves. McHone (1996) argued for a modified broad terranne hypothesis in which basalt flows extended over basement highs during the Early Jurassic extrusion of the CAMP. Condensed and abbreviated sections of sediment could well have connected the individual depocenters. In regions where post rift erosion has not proceeded as far as in eastern North America, such as Morocco, this is exactly what is seen (Medina et al., 2007). As many have pointed out the eastern North American basins may have looked topographically similar to modern rifting areas such as the Basin and Range or the East African rift system consisting of depocenters interconnected by broader expanses of condensed and abbreviated section.

In fact, as we have learned more about the details of the stratigraphy and geochemistry of the individual basins, some rather startling similarities present themselves, even in the face of clear data supporting the growth of local accommodation space and local provenance of sediment. In specific, there is a striking similarity in the syn-CAMP stratigraphic details in basalt geochemistry, paleomagnetics, and especially cyclostratigraphy (Puffer. 1992, 2003; Prévot & McWilliams, 1989; Olsen et al., 2003; Kent and Olsen, 2009; Whiteside et al., 2007) (see Figure 9). The recent discovery of a thin air fall ash (Pompton Tuff) in the precisely homotaxial cycle in the Hartford and Newark basins has shown that the lacustrine cyclicity of those two basins was linked at the decadal to seasonal level (Stops 1 and 5) and leaves no doubt at all of the synchronicity of the cycles. The remarkably tight environmental linkage this implies seems greater than we might expect of lakes with different watersheds and suggests either that the strength of regional climate and even weather variation overwhelmed local effects or that the lakes were actually the same water body during high stands.

Faill (2003) suggested that the Newark, Gettysburg, and Culpeper basins were part of a much larger more or less continuous basin and depositional system that he termed the Birdsboro basin. The present extent of the Birdsboro basin is nearly as large as Lake Tanganyika, Malawi, or Baikal in length and would certainly be as large as these basins in terms of what would be preserved in the geological record (Whiteside et al., 2010b). The stratigraphy of the Gettysburg and Culpeper basins is not nearly as well known as that of the Newark and Hartford basins, but what is known suggests the lakes of the Birdsboro basin could also have been connected at high stands. If there was one continuous lake during high stands of the syn-CAMP sedimentary units, this lake would be over 800 km in length and be larger than all but the largest of modern inland waters, the Black and Caspian Seas, and would be the largest rift lake known to exist on Earth.

End Triassic Mass Extinction

Whatever the origin of the CAMP, plume or otherwise, associated with the initiation of seafloor spreading or not, it is clear from recent high-precision U-Pb (^{206}Pb/^{238}U) dating of zircons from some of the oldest CAMP basalts and from marine strata bracketing the Triassic-Jurassic boundary (Pálfy & Mundil, 2006; Schoene et al., 2010) that eruption of the lavas was extremely and perhaps synchronous with the initiation of the mass extinction at the close of the Triassic, the end-Triassic extinction event (ETE)[4] and that the marine and terrestrial extinction were synchronous within the ±30 ky level of precision of the method (CA-ID-TIMS; Mattinson, 2005). The oldest basalt flows in the Newark Supergroup include the high titanium quartz normative (HTQ) basalts of Puffer and Lechler (1980) and Puffer (2003) notably the Orange Mountain Basalt of the Newark basin and the North Mountain Basalt of the Fundy basin, both of which lie above the initial part of the ETE (Fowell & Traverse, 1995; Olsen et al., 2002; Whiteside et al., 2010a; Cirrilli et al., 2009), with latter yielding a ^{206}Pb/^{238}U age of 201.38±0.02 Ma, an age indistinguishable from ^{206}Pb/^{238}U ages from ashes just above the last appearance of the Late Rhaetian guide ammonite *Choristoceras* (Schoene et al., 2010) and the latter are about 70 +220/–70 ky older than an ash above the first appearance of the GSSP marker ammonite *P. spelae*. This radioisotopic estimate is close to the more precise astrochrological estimate of 120 ky derived from the Newark and Hartford basins (Whiteside et al., 2010a) and 140-160 ky derived from the marine St. Audries Bay section in Somerset England (Ruhl et al, 2010). In any case, in terms of Newark basin stratigraphy, the position of the Triassic-Jurassic boundary should lie within the middle Feltville Formation (Figure 10). This movement of the boundary to within the extrusive zone certainly makes matters difficult for geo-cartographers but there is no apparent escape.

Proposed mechanisms for the ETE via the CAMP generally involve abrupt injections of gasses into the atmosphere causing catastrophic climate change (see review by Rampino, 2010), and oceanic acidification by CO_2 absorption (van de Schootbrugge et al., 2007), but direct poisoning by CAMP-related gases has also been proposed (Svensen, 2009). The climate-altering gasses could have included just CO_2, or methane-derived

[4] Formerly the ETE was considered synonymous with the Triassic-Jurassic boundary. However, the recent establishment of the GSSP (Global boundary Stratotype Section and Point) of the base Hettangian at the marine section at Kuhjoch, Austria (Northern Calcareous Alps) at the first occurrence of the ammonite *Psiloceras spelae* (Hillebrandt et al., 2007; Morton, 2008a, 2008b; Morton et al. 2008) and its ratification by the IUGS in 2010 (Morton, 2010), defines the ETE as a Late Triassic event. By onset of the ETE, or the initial ETE, I mean the first wave of extinctions, which in eastern North American paleotropics include the vessicate pollen taxon Patinasporites densus and its "relatives", and the footrint taxon *Brachychirotherium*, and in the marine realm includes the last appearance of the ammonite *Choristoceras*.

from methane clathrate dissociation[5] (Hessebo et al., 2002), that would oxidize to CO_2, or thermogenic methane derived from CAMP intrusions into organic-rich sediments[6] (Svensen, 2009), that would oxidize to CO_2, all of which would cause global warming – a super-greenhouse - that could have lasted thousands of years for *each* large eruption in the CAMP. Sulfur aerosols have also been proposed as a CAMP eruptive product that would have caused global cooling (Schoene et al., 2010), but with a much shorter recovery time on a per-eruption basis. Global warming, higher CO_2 and related phenomena may also have caused ocean anoxia (Hallam & Wignall, 1997). Toxic thermogenic chorine or fluorine could have been released by intrusions injected into evaporates (Svensen, 2009). Some scenarios for CAMP-related extinction have become very complicated indeed involving thermal expansion-driven transgressions (Kidder & Worsley, 2010). All these are possible, even, plausible causes, but what is needed is direct evidence of the timing of the CAMP relative to the extinctions and observable effects of specific proposed mechanisms.

There is in fact evidence for a sharp rise in CO_2 at the ETE, based on plant leaf stomatal data from Greenland and Sweden (McEwain et al., 1999) and Germany (Bonis et al., 2010). This rise appears to coincide (in Greenland for sure) with a dramatic drop in plant diversity (McEwain et al. 1999, 2009). In Greenland, the data indicate a fourfold increase in CO_2 from pre-ETE levels of 600 to 2100-2400 ppm during the ETE corresponding to a 3° to 4°C super-greenhouse warming (McEwain et al., 1999). Unfortunately, these data are from beds lacking any CAMP volcanic products and thus the relation to the timing of CAMP are highly inferential involving long distance correlation (e.g., Whiteside et al., 2010a).

It is thus important to be clear about what we actually know of the timing relationship between the CAMP and the onset of the ETE. Presently, in North America there are what have been taken to be Jurassic-aspect pollen and spore assemblages between the ETE and the oldest CAMP lavas in all of the basins producing pollen below the oldest basalt (Fowell, 1994; Fowell & Traverse, 1995; Olsen et al., 2002; Whiteside et al., 2007, 2010a). This has been taken to indicate that the ETE could not be caused by the CAMP eruptions because the extinction *preceded* the proposed cause (Fowell et al., 2007; Whiteside et al., 2007).

However, even the initial ETE could still have been caused by the CAMP despite the eruptions postdating the beginning of the extinction - if there were a lag between the intrusion of CAMP sills and the surface eruption of the lavas, with venting of thermogenic environmentally deleterious gases preceding the eruption of the flows - which is not an unreasonable scenario. Indeed we know that CAMP sills did metamorphose a large body of organic rich strata, notably the Triassic Lockatong Formation[7] in the Newark basin (Van Houten, 1969), organic rich Paleozoic shales and

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[5] Originally proposed for the Paleocene-Eocene thermal maximum (PETM) by Dickens et al. (1997).

[6] Also proposed for the PETM by Svensen et al. (2004).

[7] The entire volume of the metamorphic aureole of the CAMP Palisade sill complex, including its continuation into Pennsylvania (Quakertown and Coffman Hill intrusions etc.) and New York is an order of magnitude smaller than that modeled for the North Atlantic Vøring and Møri basins for the PETM (<u>very</u> roughly 3.6×10^{12} m^3 for the Palisade sill and 3.5×10^{13} m^3 for the PETM). Even if the Palisade complex was intruded entirely into the Lockatong, which it is

evaporites in the Amazonian basin (Milani & Zalán, 1999; Barata & Caputo, 2007) Svensen et al., 2009), and CAMP intrusions also invaded Triassic evaporites in Morocco (M. Et-Touhami, pers. com.). However, proxies for halide or halocarbon gasses are not known.

However, there is some evidence that some CAMP flows were in fact coeval with the initial ETE. Marzoli et al. (2004) described palynoflorules in the Argana and High Atlas basins in Morocco of Triassic aspect directly in contact with the oldest basalts, which if true would argue that those basalt were slightly older than the oldest basalts in Eastern North America and indeed could have been contemporaneous with the onset of the ETE and therefore could have caused it. More recent work by Whiteside et al (2010a) confirms two cases in Morocco (Khemisset and Central High Atlas basins) that conform to Marzoli's observations[8]. In addition Cirrilli et al. (2010) have discovered pollen and spore taxa immediately above the North Mountain Basalt in Nova Scotia typical of the Northern European expression of the latter part of the ETE and by definition of latest Triassic age (in as much as the first appearance of *P. spelae* is above in the same sections). This seems to indicate that extinctions *persisted* after the oldest basalts even in eastern North America. This is also indicated by the stable carbon isotope data of Whiteside et al. (2010a) that suggests that the extinctions of the conodonts in marine strata followed shortly after the extrusion of the earliest Newark basin basalts. Finally the exposed lavas of the CAMP are but a small portion the entire provinces and slightly older CAMP lavas could in fact be widespread.

In 2002 (Olsen et al., 2002) my colleagues and I showed that in eastern North America, tetrapod extinctions, based largely on the spectacular Newarkian track record, was synchronous with the floral extinctions (Figure 11) and a "fern spike". We related this extinction event to a modest iridium anomaly that we reasoned by analogy to the K-T iridium anomaly could have been produced by the impact of a giant bollide and caused the mass extinction (Figure 12). Subsequent work in the Fundy basin by Tanner and Kyte (2005) and Tanner et al. (2008) showed that there were multiple Ir anomalies associated with the initial ETE in that basin and resampling of a much larger stratigraphic section around the ETE in the same section in the Newark basin reported on in 2002 also showed there to be multiple Ir anomalies, each associated with a redox boundary (unpublished data by N. Nair and P. E. Olsen). This argues against a bolide impact for the origin of the ETE, and at the present time there is no convincing evidence of a role for impact at this

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not, and even if the Lockatong Formation were as rich in organic carbon as the North Atlantic Cretaceous and Paleocene mudstones (TOC averaging ~1.4%), which it probably was not, the expected amount of thermogenic methane would still be an order of magnitude less than that postulated for the PETM (following all the same assumptions as Svenson et al. (2004) and thus unlikely to be an important cause for the extinction by itself. No other large volumes of organic-rich strata intruded by giant sills are known to be present in the Newark Supergroup.

[8] That said, Whiteside et al. (2007) presented data from the Argana basin that differed from that of Marzoli et al. (2004) by lacking vessicate pollen in the samples just below the basalt, while still having abundant other pollen, thus conforming to the eastern North American pattern. The same kind of observation was made by Deenen et al. (2010) in a different section in the Argana basin. This may be due to differential thermal degradation of the vessicate forms adjacent to the basalt as suggested by Deenen et al. (2010).

mass extinction event. In my view the role for the CAMP in the ETE is much more compelling, but requires further testing and exploration.

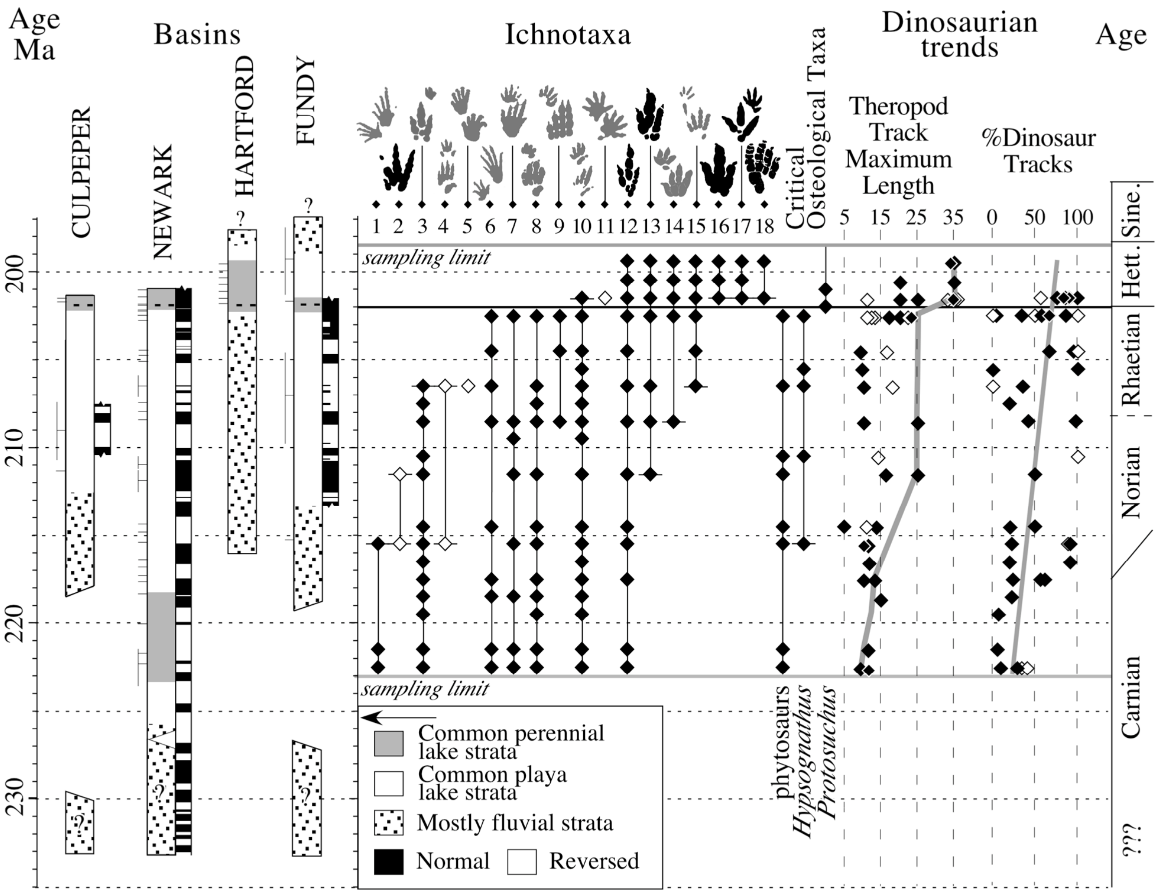

Figure 11: Correlation of four key basins of the Newark Supergroup showing the temporal ranges of footprint ichnogenera and key osteological taxa binned into 1-My intervals showing the change in maximum theropod dinosaur footprint length (line drawn through maximum) and percent at each 1-My level of dinosaur taxa. Short, horizontal lines adjacent to stratigraphic sections show the position of assemblages, and the attached vertical lines indicate the uncertainty in stratigraphic position. Solid diamonds indicate samples of footprints, and open diamonds indicate samples with <10 footprints. Horizontal, dashed gray lines indicate the limits of sampling; thick gray line indicates trend in maximum size of theropod tracks; ?, age uncertain. Ichnotaxa are as follows: 1, *Rhynchosauroides hyperbates*; 2, unnamed dinosaurian genus 1; 3, *Atreipus*; 4, *Chirotherium lulli*; 5, *Procolophonichnium*; 6, *Gwyneddichnium*; 7, *Apatopus*; 8, *Brachychirotherium parvum*; 9, new taxon B (8); 10, *Rhynchosauroides* spp.; 11, *Ameghinichnus*; 12, "*Grallator*"; 13, "*Anchisauripus*"; 14, *Batrachopus deweyii*; 15, "*Batrachopus*" *gracilis*; 16, *Eubrontes giganteus*; 17, *Anomoepus scambus*; and 18, *Otozoum moodii*. Ma, million years ago; Hett., Hettangian; Sine., Sinemurian. Modified from Olsen et al. (2002).

Within the Newark basin are several features of the sedimentary record that appear to be tied to the climatic and tectonic events related to the ETE and CAMP episode and these are illustrated at our field stops and cores (Figure 10). The first feature is readily seen in the cores at Stop 1 and outcrops as Stops 2, 3, and 5. Based on the pervasive cyclicity, the accumulation rate of the interflow units increases from the upper

Passaic Formation by a factor of more than 4 (5 m per cycle to 20 m per cycle). This is despite the fact that accommodation space was being filled very fast by CAMP lava flows. Second, the deeper water units within sedimentary cycles in the Feltville Formation are very carbonate-rich (Stops 1, 2 and 3), in contrast to the succeeding cycles in the overlying Towaco and Boonton Formation or the underlying Passaic or Lockatong formations. A very similar pattern is seen in the Hartford basin of Connecticut and Massacussetts, the Culpeper basin of Virginia, the Fundy basin of Nova Scotia and New Brunswick Canada, and even in the Moroccan basins (Whiteside et al., 2007). It seems plausible that these carbonate beds are related to elevated weathering rates of early CAMP basalts under intense super-greenhouse conditions with resulting heavy loading of bicarbonate and calcium from the basalt in the Feltville and coeval lakes. Also peculiar to the Feltville Formation, but similar to its correlates in other basins is the abundance of the large leathery-leaved dipteridaceous fern *Clathropteris meniscoides* (Figure 13) and elevated levels of fern spores. Both of these features are also characteristic of the strata underlying the Orange Mountain Basalt but above the ETE. Again this is plausibly related to elevated CO_2 and perhaps also to increased frequency of fire which in turn increased the frequency of early successional plant groups. Cornet (1975) showed that *Clathropteris* was associated with cherolepidaceous conifers (*Brachyphyllum* spp.) that had unusually short and small leaves with thickened papillate cuticle. This is consistent with adaptations to aridity, but the associated sediments show no signs of unusual aridity such as evaporites, suggesting that perhaps the adaptations were not to aridity *per se*, but rather to high heat stress and high canopy evaporation levels, inline with the thermal damage hypothesis of McElwain et al. (1999) for the ETE super-greenhouse.

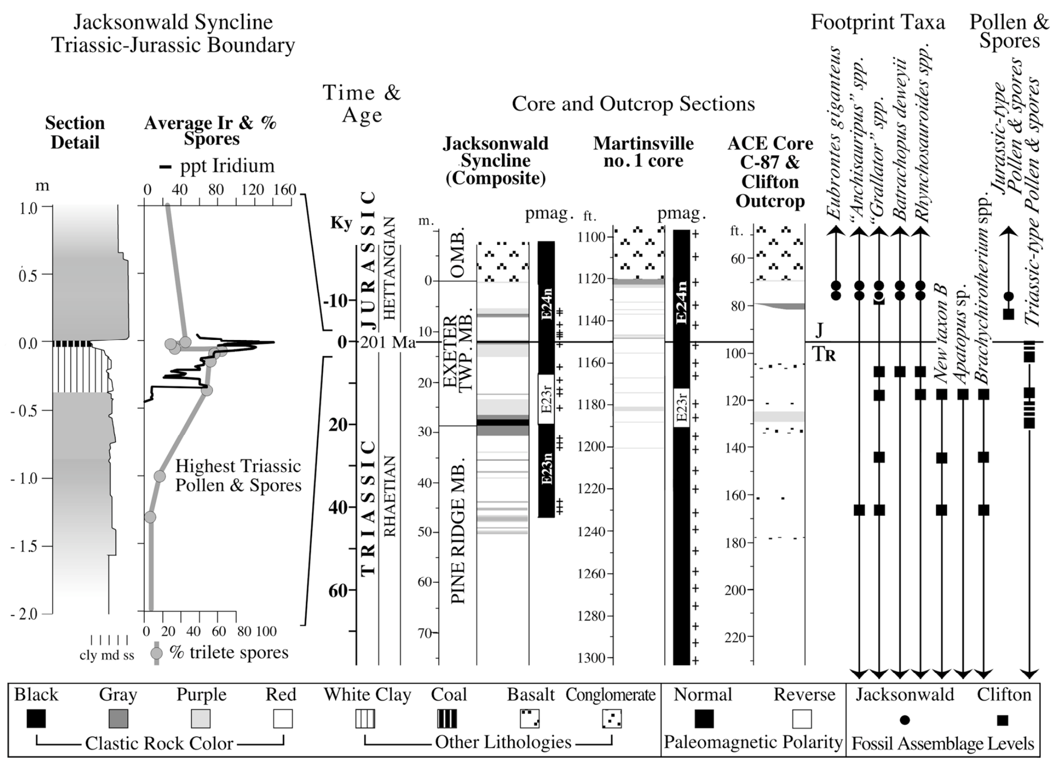

Figure 12: Details of the ETE in the Newark basin (from Olsen et al., 2002).

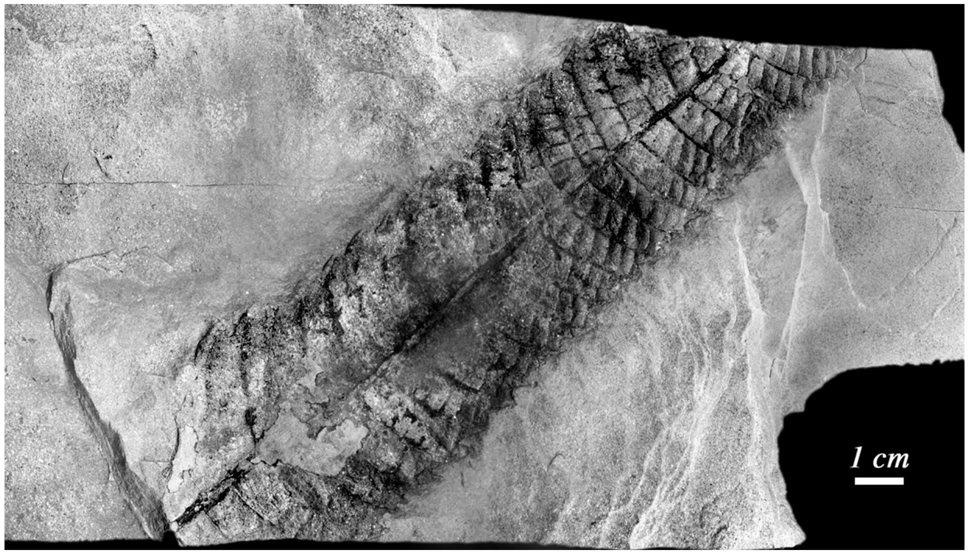

Figure 13: Example of the fern *Clathropteris* from the cycle bearing the Cataract Hollow Bed, lower Feltville Formation, Stop 3, Station 3.

Ascent of the Dinosaurs

Ironically, the ETE decimated the diversity of tetrapods at the end of the Triassic, notably hitting the curotarsians[9] (crocodile relatives) especially hard (Brusatte et al., 2008a, 2008b), but afterward dinosaurs and crocodiliomorph crurotarians became ecologically dominant, and stayed that way until the close of the Cretaceous. The ecological ascent of the dinosaurs is particularly well-displayed, albeit somewhat indirectly, in the strata of the Newark basin, where the transition in tetrapods is seen in particular detail thanks to the extremely abundant footprints around the ETE. Olsen et al. (2002) examined the transition using the NBAGPTS for time calibration (Figure 11). The last appearances of the larger curotarsian tracks such as *Brachychirotherium* and *Apatopus* were within 30 ky of the floral extinction marking the ETE. In the oldest assemblages above the floral transition marking the initiation of the ETE, and below the oldest Newark basalt (Orange Mountain Basalt) the size of the theropod dinosaur tracks is 20% larger than anytime before, corresponding to the oldest *Eubrontes giganteus*. Olsen et al. (2002) argued that this could be explained either by character release

[9] Crurotarsians are members of the Crurotarsi (often called crocodile-line archosaurs), closely related to the concept of the Pseudosuchia, which include modern crocodilians, and are the sister group to the Ornithodira (or bird-line archosaurs), which include the non-avian dinosaurs and their near relatives as well as birds. The Crurotarsi are defined (Sereno, 1991; emended by Sereno, 2005) as all forms closer to *Crocodylus niloticus* (Nile crocodile) than to Passer domesticus (the House Sparrow) and thus includes all forms evolutionarily closer to crocodiles than to birds.

following ecological release[10] after the extinction of dinosaurian competitors, or by emigration of the larger theropods from elsewhere, but favored the former hypothesis. This transition from a cruotarsian-dominated ecosystem to a dinosaur dominated one marks the true ecological ascent of the dinosaurs that had evolved at least 30 million years earlier.

Lucas et al. (2006) and Lucas & Tanner (2007a,b) have argued that "… tracks of large theropod dinosaurs assigned to *Eubrontes* (or its synonym *Kayentapus*) are known from the Triassic of Australia, Africa (Lesotho), Europe (Great Britain, France, Germany, Poland–Slovakia, Scania) and eastern Greenland, invalidating the "ecological release" hypothesis…" (Lucas & Tanner, 2007a,b). However, the concept of *Eubrontes* used by these authors is broad and encompasses basically all of the larger brontozooid (or grallatorid) kinds of footprints, while the ecological release hypothesis is framed using only the largest of these, the ichnospecies *Eubrontes giganteus*, which is the type species (Olsen et al., 1998), the other species within the genus being explicitly excluded. Of the occurrences listed by Lucas of *Eubrontes*, none can be assigned to *E. giganteus* except the Scania (Sweden) occurrence. The latter is part of an assemblage of brontozoid tracks from the roof of the Gustaf Adolf coal mine in Höganäs and was originally described by Bölau (1952). The sandstone natural casts in the roof of the mine are from the Rhaetian Bjuv Member ("lower" or "B" of coal seam) of the Höganäs Formation. This interval in other mines produces the upper occurrences of the seedfern *Lepidopteris*. This taxon has been used as the guide fossil for the upper Rhaetian, and hence it has been assumed that the tracks are Triassic in age. I see no reason to doubt the association of *Lepidopteris* and *Eubrontes* giganteus at this locality. If this occurrence were indeed earlier than the ETE it would in my view indeed falsify the ecological release hypothesis. However, as McElwain et al. (2009) has shown, the ETE begins before the last appearance of *Lepidopteris*, which in fact has its last occurrence at the end of the ETE in Greenland, and therefore this occurrence of *Eubrontes* is probably approximately contemporaneous with the first occurrence in Eastern North America. There is no evidence of *Eubrontes giganteus* below the ETE and the ecological release hypothesis still stands as viable[11].

Interestingly, herbivore tracks, unlike later Jurassic assemblages are very rare, in fact absent from the earliest *Eubrontes giganteus* localities in the Passaic and lower Feltville formations. Skeletal evidence also suggests herbivores were relatively rare. These, just-post-ETE theropods, were probably opportunistic carnivores, and may have been part of a largely aquatic-based ecosystem. There is some skeletal evidence for this in the form of apparent fish-eating adaptations among the post-ETE theropods (Milner &

[10] Ecological release is a term coined by Wilson (1961) for the release of competitive (and selective) pressure that occurs when a species arrives on an island devoid of its competitors from whence it came. While couched originally in terms of geography and island biogeography (MacArthur & Wilson, 1967) it can easily be viewed in terms of an extinction of competitors.

[11] Gigantism is an often-cited consequence of ecological release and a particular good example of which is the recently extinct giant eagle (*Harpagornis moorei*) of New Zealand derived from a much smaller Asian/Australian form 1.8 million years ago (Scofield & Ashwell, 2009). This example provides a potential way to test the ecological release hypothesis in theropod dinosaur skeletons, because Scofield, & Ashwell (2009) show brains size lags behind body size and thus we would predict the encephalization quotient of the early very large theropods would be below expected for all theropods.

Kirkland, 2007). In any case, the post-ETE track assemblages are the oldest in which dinosaurs, particularly large theropod dinosaurs, were unquestionably ecologically dominant.

Evolution of Species Flocks

While the dinosaur tracks of the Passaic Formation and the shallow-water portions of Van Houten cycles in succeeding formations interbedded with the lavas of the CAMP document the one set of the effects of ecological release on the ascent of the dinosaurs, the deep-water portions of the Van Houten cycles formed during times of high precessional variability show the effects of ecological release on fish assemblages, in this resulting in the proliferation of closely related species in the same area in this case a lake, called a species swarm or flock (Mayr, 1963). The genus containing these species flocks is *Semionotus*, a holostean fish related to living gars (Figure 14). The Feltville, Towaco, and Boonton formations all have species flocks of *Semionotus*. McCune (1990, 1996, 2004), and we will visit 3 exemplar localities in the former two formations (Stops 2, 3, 5).

McCune (1996, 2004) has described the species flocks of semionotids, mostly in Towaco and Boonton forms, and to a lesser extent in the Feltville. Morphological variation in *Semionotus* species in a single Van Houten cycle is largely in body shape and details of dorsal ridge scale pattern. Similar body shapes of *Semionotus* occur within different clades in successive cycles, strongly suggesting iterative evolution.

Other taxa than semionotids are less common but include rare *Ptycholepis* cf. *marshii* and *Redfieldius* sp. in the Feltville Formation and one specimen of *Ptycholepis* new sp. and relatively common *Redfieldius* spp. and rare *Diplurus longicaudatus* in the Boonton Formation (Figure 15). In this assemblage, the semionotids probably were in a large variety of niches involving visual selection. *Redfieldius* has a mouth and jaw suspension apparatus suggestive of modern planktivorous taxa. *Ptycholepis* was probably a small-prey piscivore. *Diplurus* was a top predator and its distinctive phosphatic coprolites composed of digested fish bones are often present in large numbers.

It is difficult to assess diversity in the non-semionotid fishes because they are so uncommon, with the exception being *Redfieldius* in the Boonton Formation that does seem to be represented by several species. One of the points made by my 1980 NYSGA paper was that it is very difficult to assess true diversity because even in the face of a very large numbers of specimens, such as the 3000 of so specimens collected by McCume most taxa are rare, so rare they are not sampled at all.

The diversification of *Semionotus* species flocks in the latest Triassic and Early Jurassic Newark lakes reflects another example of ecological release followed by character release and adaptive radiation, perhaps analogous to the macroevolutionary pattern seem among tetrapods, particularly dinosaurs after the ETE. Except in the case of the semionotids it happened again and again as lakes waxed and waned tracking the orbital cycles, all without any perceptible directional change, closely analogous to the cichlid fishes in the East African Great Lakes, which also fluctuate dramatically to orbitally driven climate changes (Johnson et al., 1996; Cohen et al., 2007).

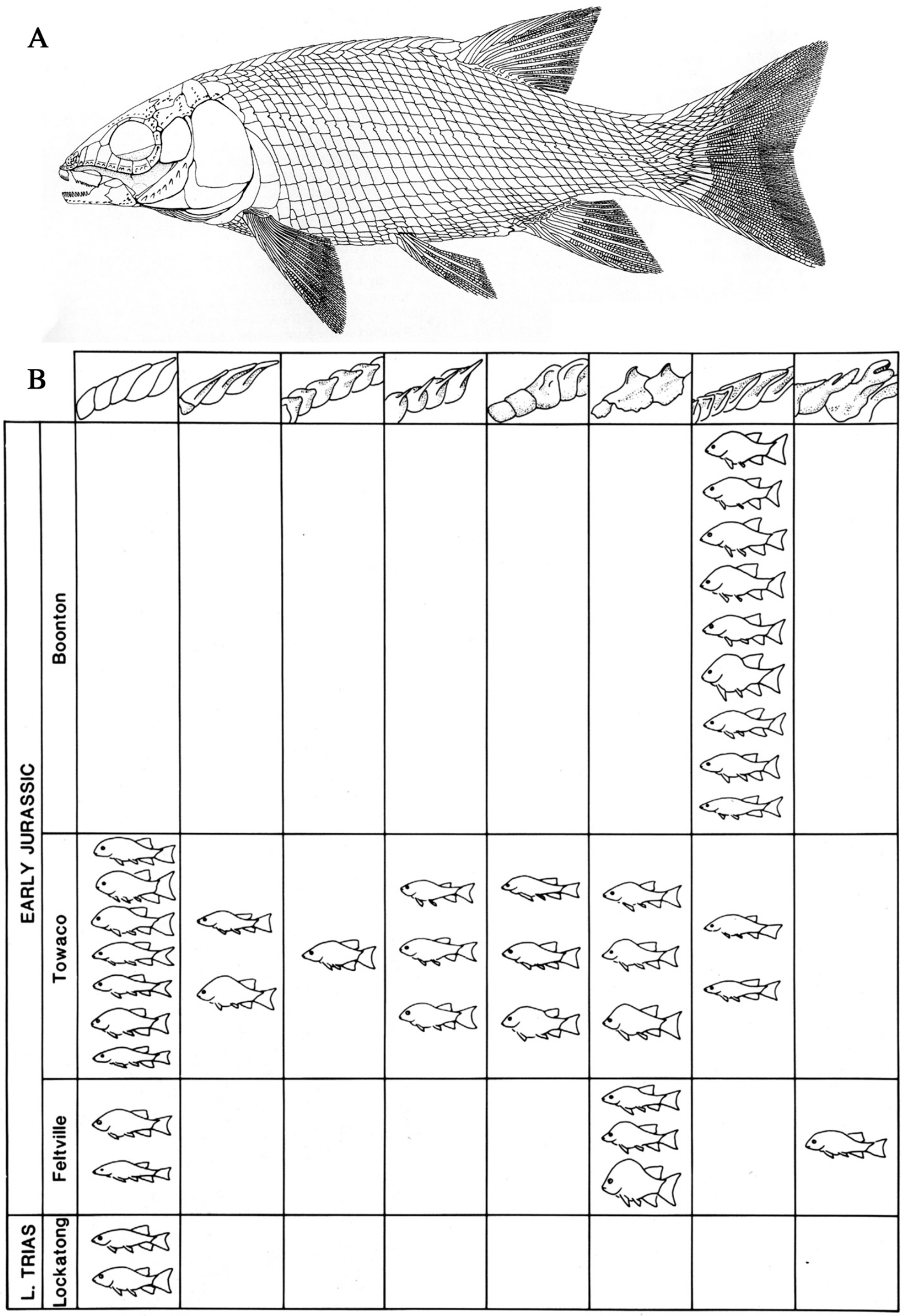

Figure 14: *Semionotus* (A) and species flocks of *Semionotus* (B) of the Newark basin, from Olsen & McCune (1991) and McCune et al. (1984).

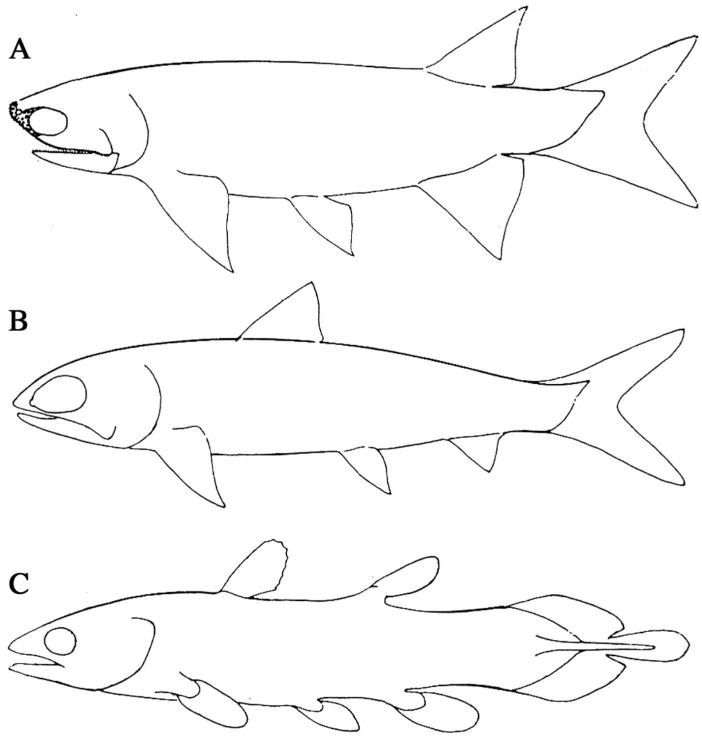

Figure 15: Other non-semionotid fish genera from the syn- and post-CAMP interval. A, *Redfieldius*; B, *Pycholepis*; C, *Diplurus longicaudatus*. From Olsen et al. (1982).

Chaotic Evolution of the Solar System

Milankovitch cyclicity is caused by variations in the orientation of the Earth's spin axis and shape and orientation of the Earth's orbit caused by the gravitational interactions of bodies in the Solar System following the laws of celestial mechanics. These were first formulated in their simplest form by Kepler and then explained within the context of calculus and the law of universal gravitation by Newton and a plethora of subsequent workers. With analytical solutions of the one and two body problem solved in the 18th century it seemed that the motions of the celestial bodies could be projected forward and back infinitely in a clock-work-like solar system with just a little more mathematical elaboration. However the 3-body and n-body problems proved intractable[12] and in the 19th century Poincaré showed that they are fundamentally not soluble using differential equations and in the process laid the foundations for the modern theory of deterministic chaos[13]. The Solar System has proven not in fact to be clock-work-like, but rather is chaotic on timescales of 10s of millions of years (Laskar, 1999). Thus, numerical solutions of the gravitational problem for the Solar System diverge completely after 10s of millions of years with extremely minor input variations which means it is impossible to construct accurate curves of insolation for the Earth for the Triassic or Jurassic or even to predict the periods of the climate cycles of eccentricity and obliquity modulating

[12] The $n=3$ and the $n > 3$ problems were solved in specific cases by Sundman (1912) and Wang (1991), respectively.
[13] A very readable book reviewing the interwoven history of celestial mechanics and chaos theory is by Diacu & Holmes (1996).

cycles greater than 400 ky, particularly the cycles caused by the interaction of Earth and Mars. In addition a significant percentage of the solutions (~1%) exhibit catastrophic singularities for the inner planets after 100s of millions of years including planetary close encounters, collisions, and ejections (Batygin & Laughlin, 2008; Laskar & Gastineau, 2009).

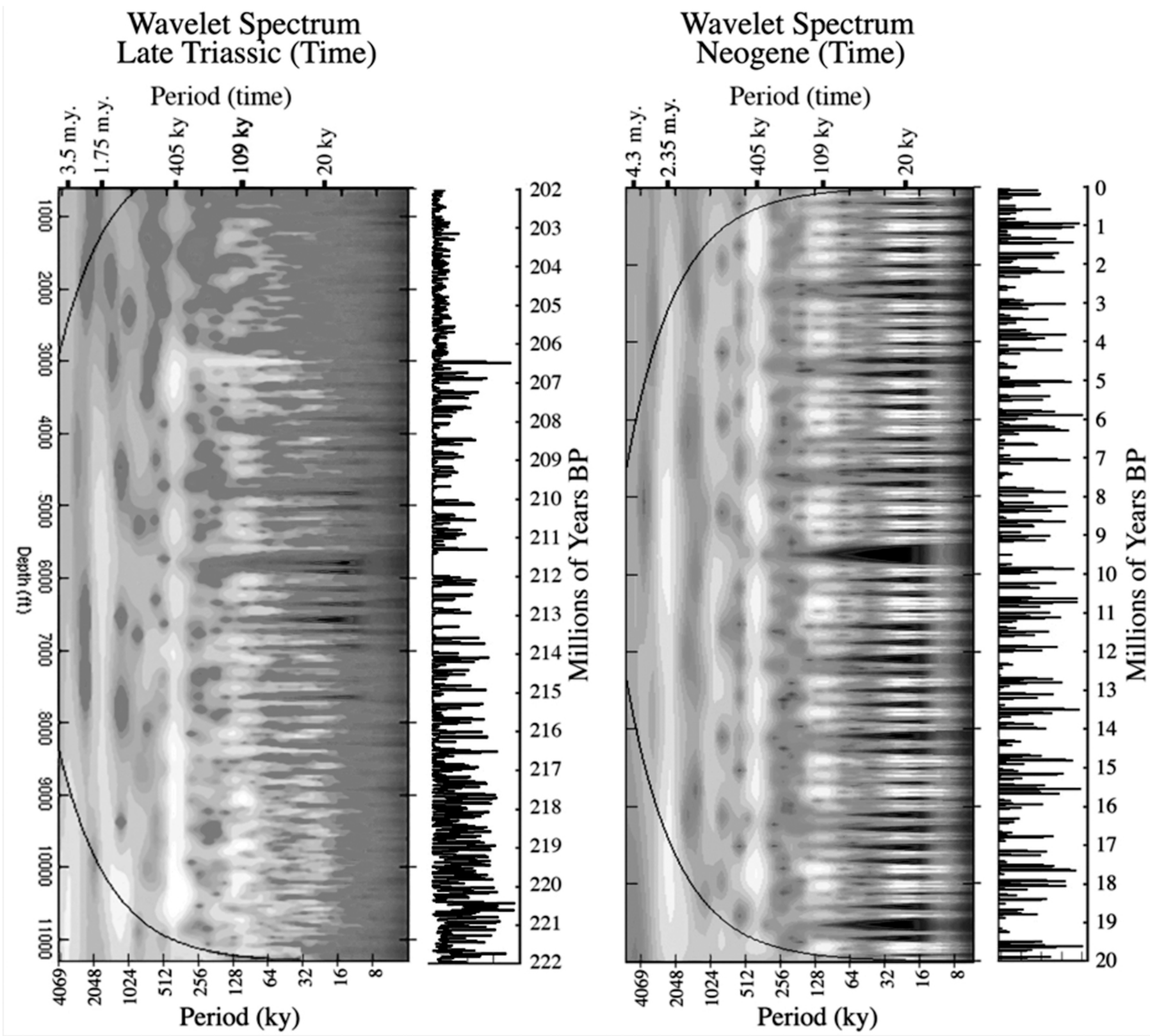

Figure 16: Comparisionn of wavelet spectra of depth rank data from the Passaic and Lockatong formations of the Newark basin and a clipped insolation curve of the last 20 m.y. (based on Laskar et al., 2004). White denotes highest spectral power.

While chaotic diffusion prevents the recognition of accurate solutions of planetary behavior and hence accurate predictions of long-period climatic cycles, those planetary interactions did occur in the past and the geological record of climate change related to those cycles provides a way to limit the range of possible solutions. Presently radioisotopic dates do not provide a direct mechanism of calibrated geological records at the precision necessary to select among of potential solutions. However one celestial mechanical cycle is close to metronomic accuracy on scales of 100s of millions of years, and that is the eccentricvity cycle caused by the interaction of Jupiter and Venus which has a very constant period averaging 405 ky with a maximum error of less than 500 ky projected back to 250 Ma (Laskar et al., 2004). Of course this cycle is the same as what was used to tune the NBAGPTS and which corresponds in lithology to the newarkian McLaughlin cycle.

Tuning the Newark basin record to the 405 ky cycle reveals longer term cycles of 1.8 and 3.5 m.y. (Olsen & Kent, 1999). A comparison between a wavelet spectrum of the tuned Newark lacustrine relative water depth record and 20 m.y. of the Laskar et al. (2004) reveals that the homologous modern cycles have periods of 2.4 and 4.5 m.y. (Figure 16) and are caused largely by the interaction of Earth and Mars. The differences between the current and Triassic values are well within the range of chaotic diffusion. That the Triassic values for these cycles are indeed reasonable and not due to periodic episodes of non-deposition as suggested in a review of my 1999 paper by F. Hilgen (Olsen & Kent, 1999) is shown by the correspondence between the number of McLaughlin cycles present and U-Pb dates from the Newark or correlated to the Newark by paleomagnetic polarity stratigraphy (Figure 8). Thus, the gravitational solutions must conform to this data, while current published versions do not.

In map view, the 1.8 m.y. cycle shows up as clusters of the named members of the Passaic Formation and the ones with the majority of the fossils. These same intervals, which derive from times of maximum precession variability generally support topographic highs. This cycle provides a natural way of segmenting the stratigraphy at a large scale.

Interestingly, one of the times of maximum precessional variability in the 1.8 m.y. cycle begins at the ETE. This is not to infer that the 1.8 m.y. cycle caused the ETE, but it is quite possible that the high CO_2, directly or indirectly, caused by the CAMP enhanced the strength of the hydrological cycle during the time of maximum precession variability. This in turn enhanced both seasonal and precessional contrasts, thus further destabilizing ecosystems. Long period orbital cycles do seem to have paced Neogene mammalian evolutionary turnover (van Dam et al., 2006). The mean lifespan of Neogene mammals is about 2.5 m.y. which appears to be regulated by the present 2.4 m.y. eccentricity cycle. The same was probably true for Late Triassic and Early Jurassic tetrapods, but the Newark record does not provide enough vertebrates to see this level of taxonomic change. Correlative vertebrate records from western North America might have high enough sampling density to see these sorts of patterns once the time scale is worked out in sufficient detail via coring and correlation to the Newark (e.g., Olsen et al., 2008; Geissman et al., 2010).

Carbon Sequestration

Burning of fossil fuels is adding CO_2 to the Earth's atmosphere in ever increasing amounts. The increase in CO_2 is analogous ti what seems to have happened around the ETE because of the emplacement of the CAMP. It is therefore profoundly ironic that one mechanism for sequestering anthropogenic CO_2 is to pump CO_2 back into CAMP basalts and diabase where the carbonation reaction results in the production of carbonate minerals such as magnesite[14]. Goldberg et al. (2010) describes possible CAMP reservoirs

[14] One carbonation reaction for basalt is olivine (Mg_2SiO_4) + $2CO_2$ = magnesite ($2MgCO_3$) + quartz (SiO_2). Other reactions occur with feldspars.

for CO_2 sequestration in CAMP basalts that can have zones with elevated porosity and permeability in the form of vesicular intervals at the top of flows (Stops 1 & 4) or fracture zones such as in the lower Preakness Basalt (stop 4).Measurements on samples from the Martinsville core of the Orange Mountain Basalt (Figure 17) have porosities of ~10% (Goldberg et al. 2010). In these cases rather than a threat for instability, time is an ally in extending the reaction time with the basalt that brings longer stability. Other possible targets for CO_2 sequestration in the Newark basin are sandstones, and that possibility is also being investigated (e.g., http://www.tricarb.org/tricarb/default.aspx).

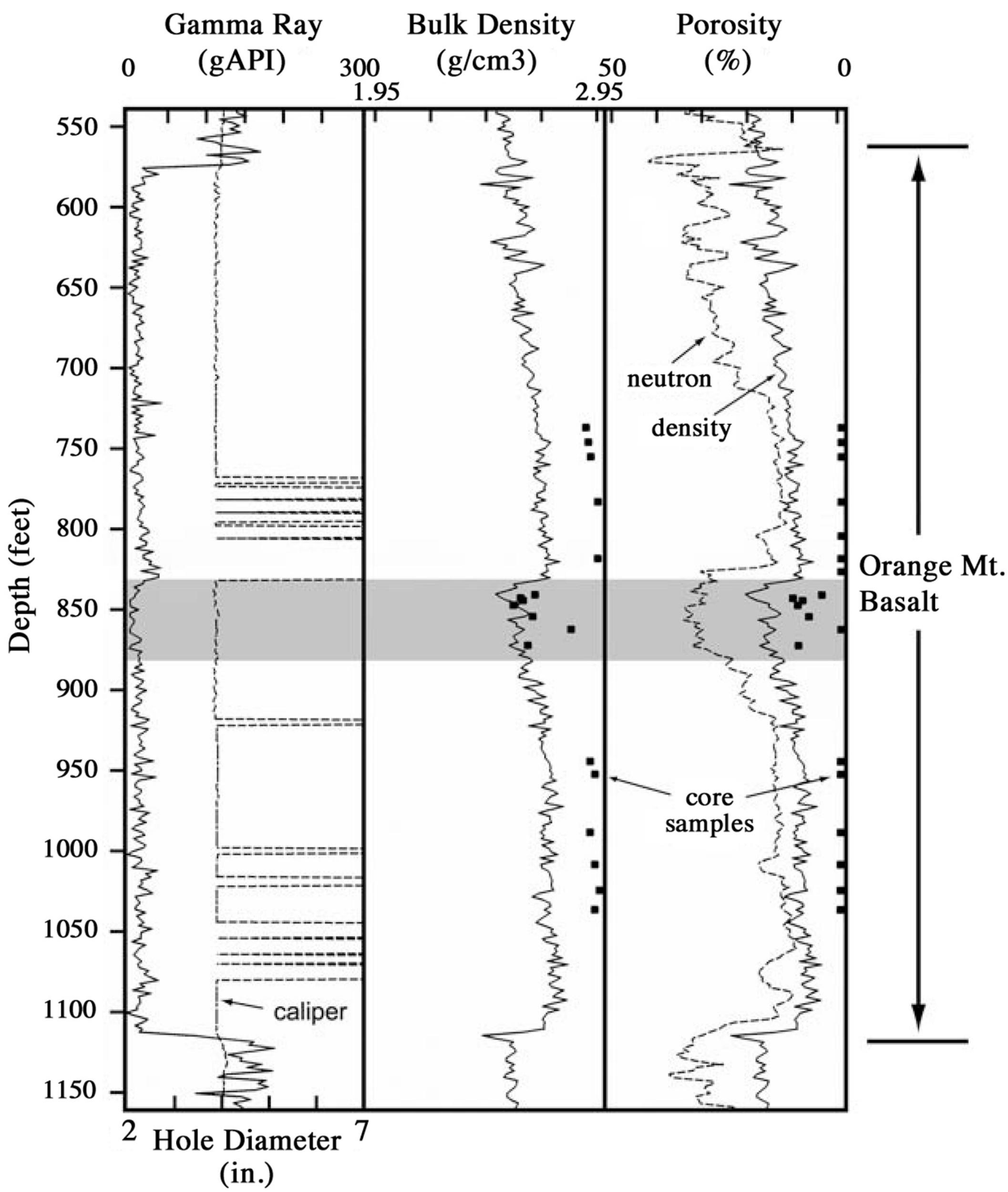

Figure 17: Geophysical log profiles through Orange Mountain basalt. Modified from Goldberg et al. (2009).

FIELD TRIP AND ROAD LOG

The field trip begins and ends at Lamont-Doherty Earth Observatory (LDEO), Palisades, NY in the parking lot closest to 9W. (http://www.ldeo.columbia.edu/about-ldeo/maps-contact). The route (Figure 1) takes us through the Palisades sill, Locakatong and Passaic formations and then into the CAMP lavas and finally back through the Passaic Formation and Palisade sill.

0.0 mi Walk from parking area to the south side of the Machine Shop of LDEO.

Stop 1: Lamont-Doherty Earth Observatory (Machine Shop Area): Newark Basin Cores, Palisade Sill Outcrop, and Borehole Through Sill.

Latitude and Longitude: 41° 0.239'N, 73° 54.760'W.
Stratigraphic Units: Stockton – Boonton Formations and Palisade Sill
Age: ?Late Carnian (early Late Triassic) to Early Jurassic (~230-200 Ma)
Main Points:

1. Representative cores of the Newark basin
2. Lacustrine cyclicity
3. Correlation between Newark and Hartford basins based on core and outcrop
4. Pompton Tuff and the size of the great lakes
5. Carbon sequestration

Cores: Cores will be on display of all of the formations of the Newark basin (Figure 10) as well as a core of the Shuttle Meadow Formation of the Hartford basin (Figure 18) for comparison with the Feltville Formation. Detailed descriptions and interpretations of the sedimentology of these cores and related outcrops are given by Smoot (2010).

Core 1: Stockton Formation, Prallsville Member (Figures 7, 10).

This segment of core is from the Princeton no. 1 core drilled on Princeton University property near the Forrestal Campus (Kent & Olsen, 1995) (Figure 1). It is a fluvial sandstone with some of the highest directly measured porosity and permeability in the basin (3174.50 ft: 131-118 millidarcys permeability, 14.2-14.0% porosity). The Prallsville Member is below the postulated tectonosequence boundary in the Stockton (Figure 7) and is thus of Late Carnian age at about 231 Ma on the NBAGPTS.

Core 2: Lockatong Formation, Tohickon Member (Figures 7, 10).

This segment, from the Nursery no. 1 core (Figure 1) taken off Nursery Road in Ewing Township, is representative of the middle Lockatong Formation and shows nearly one complete Van Houten Cycle. The age of this unit is Early Norian about 219.5 Ma on

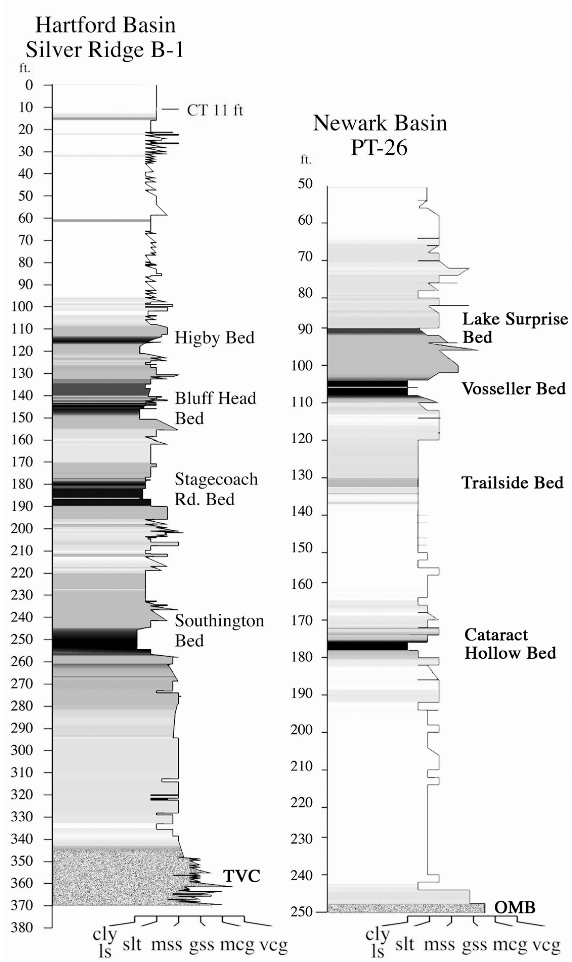

Figure 18: Comparison between the lower Shuttle Meadow Formation of the Hartford basin and the lower Feltville Formation of the Newrak Basin (from Olen et al., 2005).

the NBAGPTS. This represents one of the cycles deposited during times of high precessional variability, but does not represent the deepest of the Lockatong lakes and has not produced any fossil fish in outcrop. This core has well-developed pinch-and-swell lamination and some bioturbation in the deeper water intervals, polygonal desiccation cracks, and brecciated massive mudstone. When the brecciated massive mudstone is very well-developed it is difficult to see anything at all in the core. According to Smoot (2010) the brecciated massive mudstone formed by repeated wetting and drying of mud initially deposited in standing water in a playa with cracks forming over previously filled cracks.

Core 3: Passaic Formation, Cedar Grove Member (Figure 7).

This long segment is from the Weston Canal no. 1 core near Zarephath, NJ. shows much of a 100 ky cycle at around 207.4 Ma. The Cedar Grove Member is one of the most easily mapped units being identified from Newark New Jersey along Route 287 to near Birdsboro, PA. This is a nice example of penecontemporaneous gypsum including abundant nodules and anhedral crystals. According to Smoot (2010), the crystal styles and upward-coarsening successions resemble modern gypsum soil occurrences (Smoot and Lowenstein, 1991) where gypsum dust is blown onto the surface, and then redistributed downward by rainfall rather than gypsum crystallizing out of brines in playas. There are also veins of various orientations filled with fibrous gypsum that formed much later. The muds themselves however formed in very shallow lakes, with the gypsum dust possibly being blown in from very far away (as suggested by Smoot, 2010) along with a considerable amount of lithic dust. The lower part of the member has a Van Houten cycle with one of the best-laminated deep-water intervals in the Passaic Formation. The very well-developed nature of this cycle contrasts with the poorly developed succeeding cycles. In outcrop this well-developed cycle has produced palynomorphs and reptile footprints of characteristic Triassic aspect including the silesaurid dinosauromorph track *Atreipus milfordensis*, the crurotarsian *Brachychirotherium* cf. *parvum*, and the lepidosauromorph *Rhynchosauroides* sp.

Cores 4 & 5: Passaic Formation, Exeter Member and Orange Mountain Basalt (Figure 7, 10).

Two cores are shown here from the Passaic Formation – Orange Mountain Basalt contact. Core 4 is a segment of the Martinsville no. 1 core of the NBCP from Martinsville, NJ from a more central basin location and Core 5 is part of the ACE series, PT-38, spanning the same interval but closer to the northern edge of the basin. Outcrops of this interval in the Jacksonwald syncline have darker shales and much more obvious cyclicity.

In addition to the facies change seen across these three sites, from better cyclicity with dark gray shales in the deeper water units in the southwest, to obscure cyclcity in red beds in the northeast, there is also a parallel change from more or less consistent vertical patterns of cycles to a vertical pattern with fully fluvial conglomeritic facies below to a more marginal lacustrine facies above at the site of PT-38. The Jacksonwald area has been a major source of evidence on the initial ETE, including abundant Triassic-aspect crurotarsian tracks such as *Brachychirotherium* and *Apatopus*, while outcrops adjacent to the location of PT-38 have produced a post-initial-ETE assemblage with abundant theropod dinosaur tracks including *Eubrontes giganeteus* as well as the crocodylomorph track *Batrachopus* and a new species of *Rhynchosauroides*. Also present is a *Brachyphyllum*-*Clathropterus* macrofossil assemblage with a *Classopollis*-dominated sporomorph assemblage.

Cores 6, 7, 8: Feltville Formation, Washington Valley Member and Orange Mountain Basalt (Figure 10) and Shuttle Meadow Formation (Figure 18).

Van Houten cycles in the lower Feltville Formation (Washington Valley Member of Olsen 1980a, 1980c) are unusual in the Newark basin in having deeper water intervals that are very carbonate rich, often limestones. These cores display the lower Feltville Formation in the Martinsville no. 1 core of the NBCP (Core 6) and PT-14 of the ACE cores (Core 7) , and the correlative part of the Silver Ridge B-1 core of the Shuttle Meadow Formation of the Hartford basin (Core 8). These cores were essential to understanding the cyclostratigraphy of the Feltville and Shuttle Meadow Formation and hence to the chronology of the continental Triassic-Jurassic boundary. In fact, it is doubtful that the basic stratigraphy could have been worked out in outcrop without the model and testable hypotheses provided by both the ACE and Silver Ridge cores.

Correlation between cores and outcrops of the lower Feltville Formation is now relatively straightforward. But that was not the case in 1980 when there were only outcrops to work with. Having not seen more than one lake level cycle in superposition I had a stratigraphic model in mind in which there was only one lake level cycle in the Feltville Formation. I interpreted various outcrops along strike accordingly and assumed the dramatic differences between outcrops was due to lateral change (Olsen et al. 1980). I had the same model in mind for the lower Shuttle Meadow Formation. However, when the ACE cores of the Feltville were examined in 1985 it was obvious that there were several Van Houten cycles in the lower Feltville (Olsen et al., 1988; Fedosh & Smoot, 1988). Instead of one carbonate-rich lake level cycle there were two. I revisited the type section of the Feltville (our Stop 3) and was able to test the outcrop generality of this pattern by digging that there were indeed 2 carbonate bearing cycles and then in late August, 2010, I did the same at our Stop 2. This has allowed unambiguous correlation of the cycles in the lower Feltville laterally. One pattern emerging from this correlation (Figure 18, 19) is a progressive thinning and onlap of the lower Feltville onto the Orange Mountain Basalt along strike from the northeast to the southwest, prima facie evidence of syndepositional folding in the Watchung syncline.

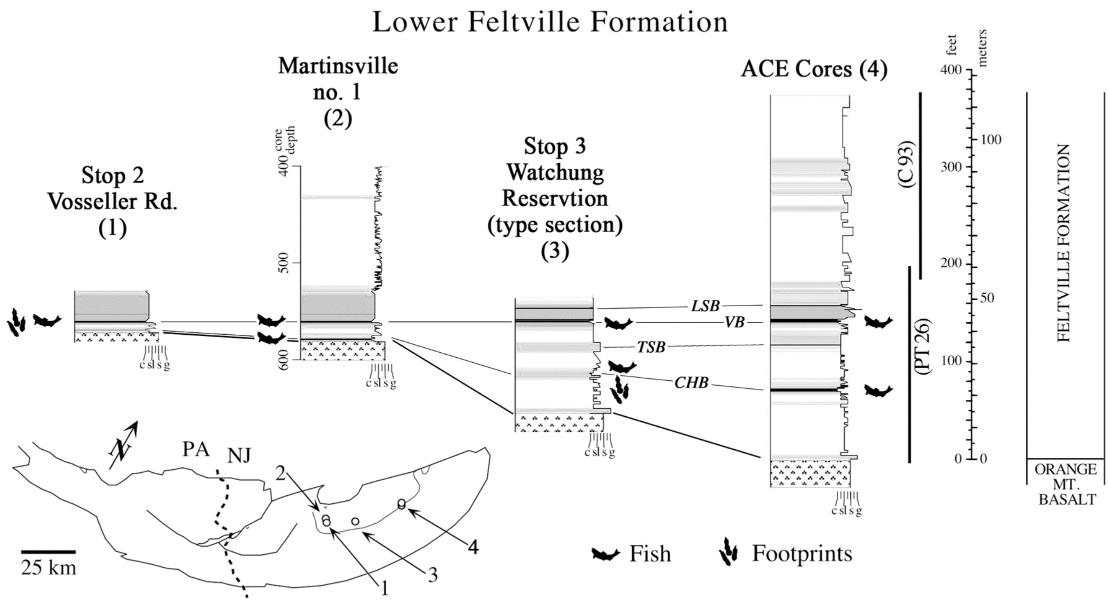

Figure 19: Lateral correlation of basal Feltville showing progressive onlap and thinning. Abbreviations are: *LSB*, Lake Surprise Bed; *VB*, Vosseller Bed; *TSB*, Trailside Bed; *CHB*, Cataract Hollow Bed.

Comparison of the stratigraphy of the lower Feltville and Shuttle Meadow Formation reveals another pattern, and that is that the cyclostratigraphy is more subtle and complicated than expected. Correlation between the Shuttle Meadow Formation and the Feltville is surprisingly obvious (Figure 18), but it makes clear that rather subtle lithological features seen in outcrop and core of the Feltville have important cyclostratigraphic implications. In order to facilitate discussions about Feltville cyclostratigraphy I have given names to the individual deeper water units within the Washington Valley Member (Figures 10, 18). Whiteside et al. (2010) shows that the cyclostratigraphy of the Shuttle Meadow Formation reflects not only the ~20 ky climatic precession cycle but also a ~10 ky hemiprecessional cycle. In specific the Cataract Hollow, Trailside, and Vosseller beds reflect mostly the ~20 ky cycle, while the Lake Surprise Bed reflects largely the ~10 ky cycle. The presence of this hemiprecessional cycle outside the equatorial tropics is not expected of local insolation forcing[15] and

[15] A 10 ky hemiprecessional cycle caused by direct forcing should be limited to ±5° of the equator, while the Shuttle Meadow Formation was deposited at about 21° N.

suggests an expansion or export of equatorial climate variability and an enhancement of the hydrological cycle perhaps due to the super-greenhouse conditions of the syn-CAMP world.

Cores 9 & 10: Preakness Basalt: Flow number 2.

We will see the stratigraphy of the lower Preakness Basalt at Stop 4, but what we will not see there are the gabbroid layers that characterize parts of flow 2. These are important because in some areas they contain zircons that crystallized from the melt and provide high-resolution dates for the flow (e.g. Blackburn et al., 2009). Several segments of ACE cores PT-20 (Core 9) and C-97 (Core 10) will be on display showing these gabbroid layers. These particular layers did produce zircons but these were all apparently assimilated from crustal rocks through which the feeders flowed (S. Bowing, pers. com.) as they have ages 100s of millions of years older than the crystallization age.

Cores 11 & 12: Towaco Formation and Pompton Tuff.

The Towaco Formation is especially well represented in the ACE cores with many parts of the formation being represented by multiple cores. At Stop 5 we will see the 3 prominent Van Houten cycles formed during times of maximum precessional variability in the middle Towaco Formation. The bed names for each of the prominent deep-water units for these cycles are given in Figure 9. Portions of cores C-128 and PT-14 (cores 10 & 11, respectively) that contain the Colfax Road Bed are on display. This bed contains the Pompton Tuff, an airfall ash we will see in outcrop at Stop 5, but here we see it in two cores. The Pompton Tuff also extends to the Hartford basin where it is known from 4 localities a slab from one of these will be on dispaly as well.

Philpotts (in Olsen et al. 2005 and paraphrased here) described this unit in the Hartford basins as thin bed that tends to erode proud of the surrounding deep-water lacustrine unit. The only detectable particles in the layer are all euhedral plagioclase laths with a high aspect ratio that show no signs of rounding. They must have originally been enclosed in glass that is now converted to a clay or to what appears to be chalcedony. There are small circular areas that contain a mottled gray birefringent material that looks very much like chert. These observations are entirely consistent with the small laths of plagioclase being part of a basaltic crystal tuff. There are also small rounded particles that seem to be bits of basalt. The tuff would seem to represent an explosive eruption of the CAMP that is separate from those that produced the stereotypical sequence of known CAMP flows.

What you can see in these cores, the thin sections on display, and slabbed portion of the Westfield bed of the East Berlin Formation of the Hartford basin is a very close match not only between the tuff, but also between the surrounding laminae. As discussed above the similarity is so strong that it seems plausible the lakes in the two basins were connected.

Core 13: Hook Mountain Basalt.

On display is an ACE core C-104 in the Hook Mountain Basalt which exhibits thin gabbroid layers. These too produced only zircons inherited from crustal sources.

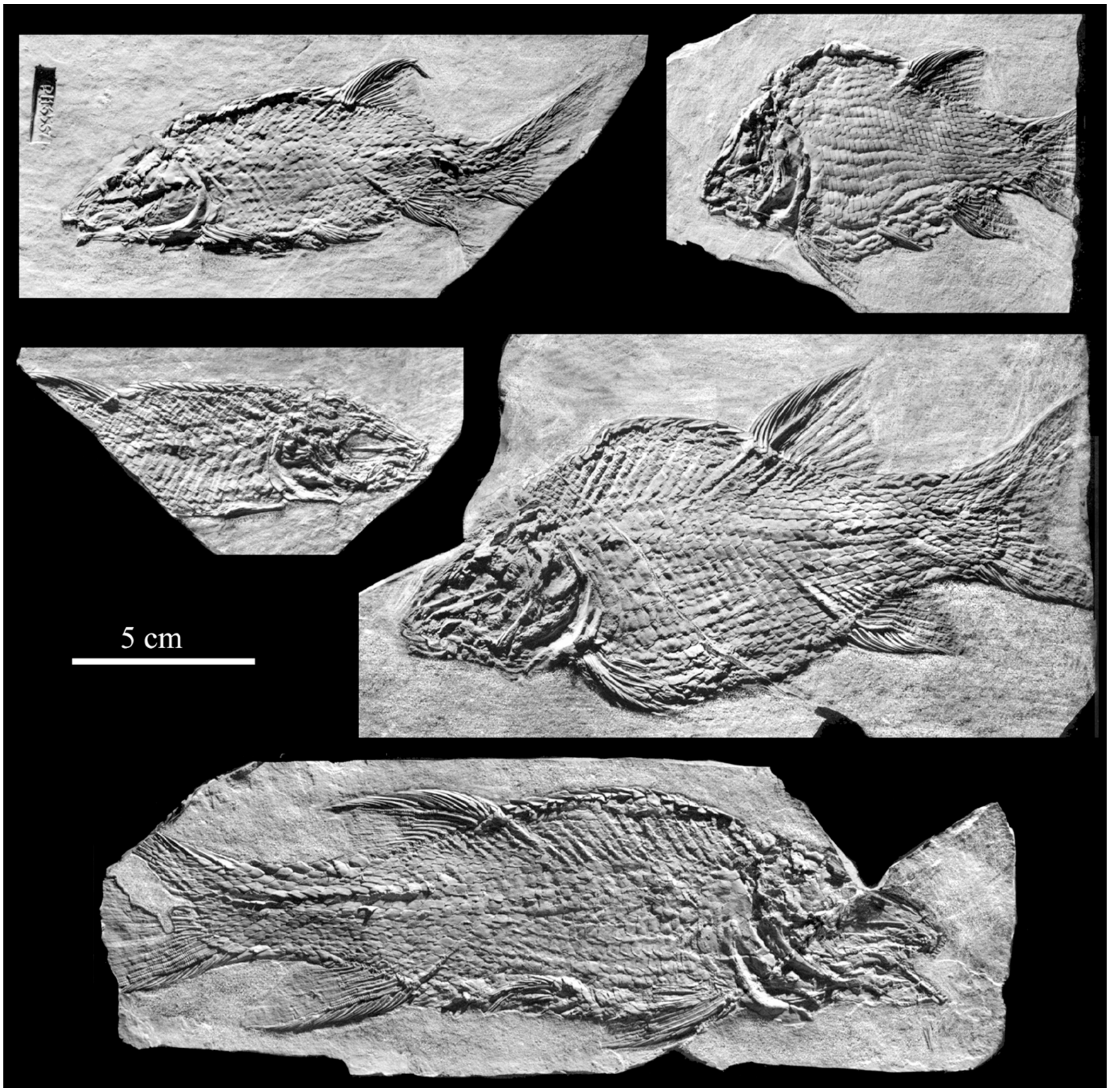

Figure 20: Representatives of the Boonton species flock of *Semionotus*.

Core 14: Boonton Formation, Rockaway River Bed (Boonton fish bed)

The Boonton Formation as sampled in the ACE cores closely resembles the Towaco Formation in its pattern of cyclicity and facies. ACE core PT-6 on display includes the famous bed that produced abundant fish described from the Rockaway River in Boonton, New Jersey, named here the Rockaway River Bed (Figure 10). The semionotids from this bed comprise a distinct species (20).

Lamont Carbonation Experiment Borehole

A borehole drilled at LDEO has been used for experiments in carbon sequestration carbonation using *in situ* carbonation in mafic igneous rocks, in this case diabase (Matter et al. 2007). Palisade diabase is visible behind (west) of the borehole site. The small small-scale CO_2 injection experiment at this site showed large decreases in Mg^{++} and Ca^{++} concentrations over short periods of time (200 hours) suggested both mixing between the injected solution and aquifer water and the release of cations from water–rock dissolution, that neutralized the introduced carbonic acid (Goldberg et al. (2010). This the same kind of reaction that would result in the weathering of CAMP basalt flows and the accumulation of carbonates in the early CAMP lakes of the lower Feltville Formation, which now has the potential to sequester the carbon from the burning of fossil fuels. As Goldberg et al. (2010) point out, if large CAMP reservoirs are found in proximity to major industrial carbon sources huge volumes of CO_2 could be sequestered permanently by the *in situ* carbonation reaction.

Return to vehicles and leave LDEO via the exit to 9W.

0.0 mi. Leave entrance to LDEO, turn left and proceed south on Rt. 9W along the strike of the Palisade sill. There fill be small exposures of the sill along Rt. 9W towards Fort Lee.

10.4 mi. Take the ramp on right onto US-9W S.

10.9 mi. Continue onto Fletcher Ave. Exposures on north side of ramp for NJ-4 of coarse gabbroid of Palisade sill. This gabbroid produces zircons yeilging a very precise age (Blackburn et al., 2009).

11.0 mi. Turn right to merge onto US-1 S/US-46 W/US-9 S toward I-95 S/I-80 W Continue to follow US-46 W. Driving down the dip slope of Palisade sill. Exposures to north along US-80/I-95 reveal contact of the sill and overlying Lockatong Formation.

13.9 mi. Take the I-95 S/I-80 W/N J Turnpike ramp.

14.0 mi. Continue toward I-95 S.

15.0 mi. Take exit 16W toward Rutherford.

15.6 mi. Merge onto I-95 S.

25.3 mi. Take exit 14 to merge onto I-78 W toward US-1/US-9/US-22/Newark Airport.

35.3 mi. Exposures of Orange Mountain Basalt.

37.0 mi Approximate contact between Feltville Formation and overlying Preakness Basalt. Notable is the high degree of distinctive splintery fracturing that characterizes much of the second Preakness flow, which we will see up close at Stop 4.

44.4 mi. Take exit 40 toward The Plainfields/Gillette/Watchung.

44.6 mi. Turn left at Hillcrest Rd.

45.2 mi. Passing into Feltville Formation.

45.7 mi. Turn right at County Route 527 S/Valley Rd. Continue to follow County Route 527 S.

45.8 mi. Slight left at County Route 527 S/Stirling Rd.

45.8 mi. Turn left to stay on County Route 527 S/Stirling Rd.

45.8 mi. Slight right at County Route 527 S/Mountain Blvd. Continue to follow County Route 527 S. Route follows strike of Feltville Formation, mostly on or near the contact with the Orange Mountain Basalt.

49.6 mi. Continue onto Washington Valley Rd. Route now follows along middle Feltville Formation.

51.6 mi. Turn left at Vosseller Ave. Headind down section in Feltville.

52.0 mi. On right is unpaved access to Eastfields Park of Somerset County Park Commission. Entrance is on upper Orange Mountain Basalt.

Park.

Stop 2: Lower Feltville Formation (Washington Valley Member).

Latitude and Longitude: 40.5932°N, 74.5452° W.
Stratigraphic Units: Lower Feltville Formation, Washington Valley Member, Orange Mountain Basalt
Age: Latest Triassic (Latest Rhaetian) to ?earliest Jurassic (201.3 Ma)
Main Points:

1. Condensed section of Washington Valley Member
2. Best exposure of Vosseller Bed
3. Very condensed section of Cataract Bed

4. Onlap onto Orange Mountain Basalt
5. First 100 ky after initiation of ETE
6. Limestones as products of super-greenhouse
7. Intense disruption of Vosseller Bed and “dead horses”
8. Species flocks of semionotid fishes
9. Dinosaur tracks

Walk about 250 ft north to outcrops along Blue Brook and walk downstream past the outcrops along the north bank of the stream about 800 ft west to the contact with the Orange Mountain Basalt.

About 12 m of Feltville section outcrops here. The complete Van Houten cycle containing the Cataract Hollow Bed outcrops along with nearly the complete cycle containing the Vosseller Bed, this being the best outcrop of the latter. The section is closely comparable to what is seen in the Martinsville no. 1 core, except it is even more condensed. The section begins with the vesicular upper surface of the Orange Mountain Basalt overlain by a decimeter of basalt gravel and cobbles in a bluish-gray mudstone. This is followed by a decimeter of gray/white, seemingly? oscillatory rippled sandy calcarenites passing up into two meters of red mudstone. This bed is clearly the Cataract Hollow Bed. In my section of this interval in 1980 I completely missed this bed., because I did not have an expectation of its existence. In August 2010, I looked for the Cataract Hollow Bed in the completely submerged interval near the Orange Mountain Basalt. This illustrates the need to have some kind of rigorous testable hypothesis of what should be present. Without that, why would I explore every hidden nook and underwater cranny.

The red beds above the Cataract Hollow bed is overlain by 20 cm of blue gray mudstone with fabrics like the underlying red mudstone. This is abruptly overlain by ripple crosslaminated gray fine sandstone with sole casts of dinosaur tracks. These tracks are so far all brontozooids (grallatorids).

Above the ripple crosslaminated sandstone is a grey mudstone with floating blocks to flakes of black laminated to microlaminated limestone and claystone. Presumably, these blocks once comprised continuous layers. They are in fact comparable to what I have termed “dead horses” (Olsen et al., 1989), although these are much larger clasts (“horses”). These blocks generally exhibit signs of bedding plane-parallel shear, including folds and imbricate thrust faults. The name derives from the term "horse" for a fault-bounded sliver of rock and their "dead" or dismembered condition and the name was coined to call attention to the phenomenon. The gray mudstone shows few or no depositional sedimentary structures. There is no well-defined upper surface of the bed. In this and other deposits such as the East Berlin Formation (Olsen et al., 2005), or Towaco Formation (Stop 1), the associated massive mudstone suggests that partial liquefaction accompanied the deformation that produced the dead horses, and I interpret them as a form of low-pressure structural mélange. These features are common in fine-grained organic rich rocks but they have usually been interpreted as depositional units, such as muddy turbidites (e.g., Dyni & Hawkins, 1981). Dead horses are still a feature of the Vosseller Bed at Stop 3, but seem absent in the ACE cores.

The limestone blocks themselves can be more than a meter in diameter and 30 cm thick, especially those with folds. Fractures in the limestone contain a tarry bitumen that is liquid, if highly viscous, at body temperature but generally solid in the water. However,

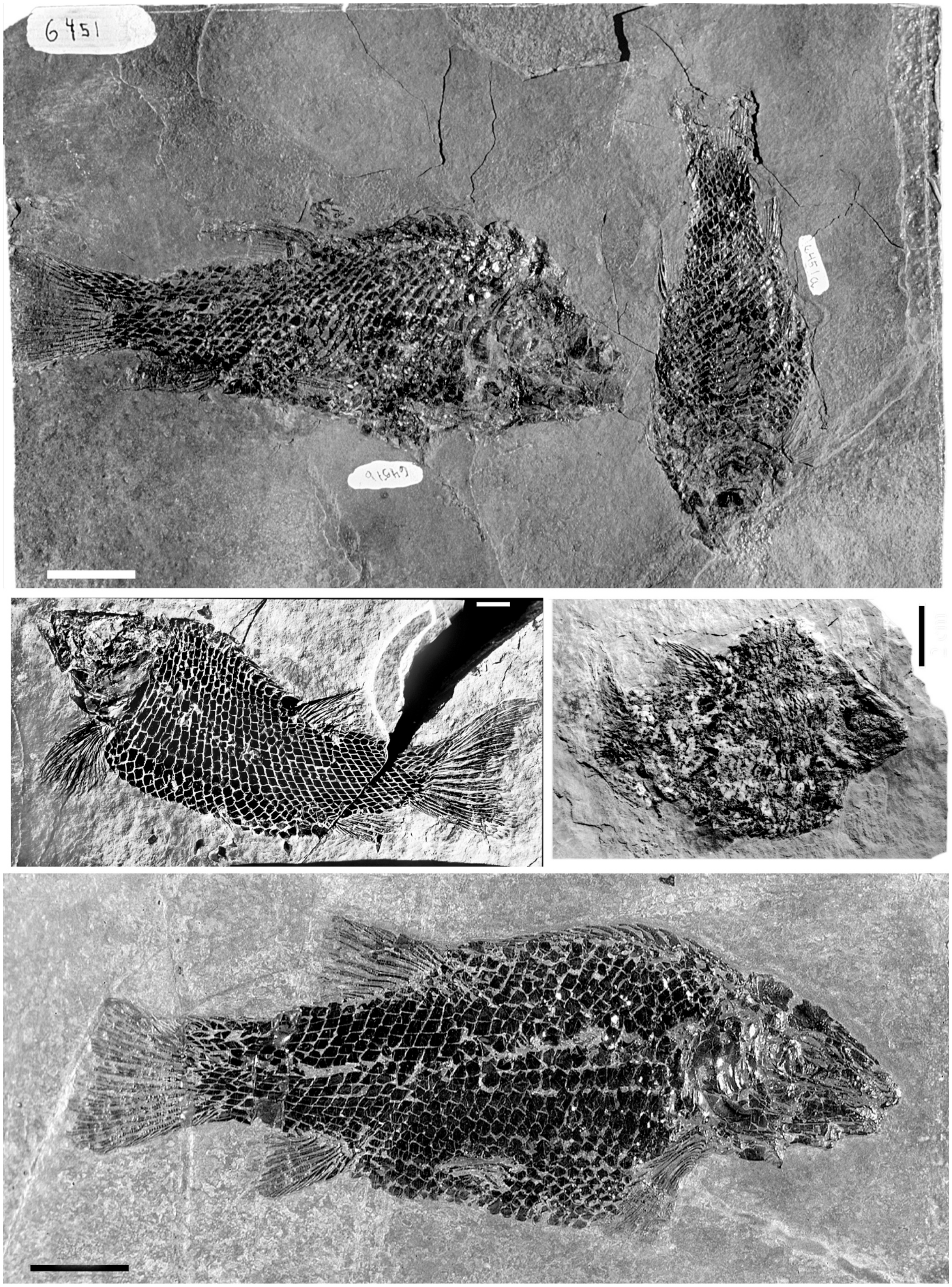

Figure 21: Examples of the Feltville, Vosseller Bed, semionotids from Stop 2.

the oil slicks that appear on the streams surface when blocks are pried loose from the matrix attest to the presence of more fluid hydrocarbons. The thermal maturity of these beds are within the oil window based on a vitrinite reflectance value of 1.18 R_0% from the Vosseller bed of the nearby Martinsville no. 1 core (Malinconico, 2009), and the hydrocarbons present were surely generated from the limestones.

These blocks of limestone themselves contain mostly articulated, and some disarticulated fish, including *Semionotus*, *Redfieldius*, and *Ptycholepis* (Figure 21) The presence of coprolites suggests the presence of coelacanths yet to be found. The semionotids comprise a species flock with most variation being in overall body shape and the morphology of the series of spined scales, called dorsal ridge scales along the dorsal midline (Figure 14). In fact the name *Semionotus* is derived from "signal back". Preservation is often exquisite and with chemical preparation[16] extremely fine details can be seen. McCune recorded six species in the Feltville (McCune et al., 1984; Olsen et al., 1989).

The origin of the carbonate so prevalent in the Vosseller, Cataract Hollow, and Feltville in general is worth commenting on because such carbonate rich beds are virtually absent from the rest of the Newark basin. Carbonate and limestones are not at all uncommon in lakes, and such famous units as the Green River oil "shales" are in fact carbonates. The carbonate was mostly derived from biologically mediated chemical precipitation in the upper part of the water column, rather than from the tests of organisms. As discussed above, the unusual amount of carbonate in the Feltville and in fact in all of the post-initial CAMP basalt flows in central Pangea may relate to weathering of the CAMP basalts themselves under super-greenhouse conditions in the first 100 ky after initiation of ETE.

Proceed back to vehicles.

52.0 mi. Turn left (northeast) onto Vosseller Avenue toward Perrine Rd.

52.5 mi. Turn right at Washington Valley Rd.

55.2 mi. Continue onto County Route 527 N/Mountain Blvd Ext. Continue to follow County Route 527 N.

58.3 mi. Turn left at County Route 527 N/Stirling Rd. Continue to follow County Route 527 N.

58.4 mi. Turn left at Hillcrest Rd.

59.2 mi. Merge onto I-78 E via the ramp towards Newark.

62.9 mi. Take exit 44 toward New Providence/Berkeley Heights.

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[16] In chemical preparation, the side of the fish exposed upon splitting the rock is embedded in a resin such as epoxy or polyester and the maturix is subsequently removed by dissolution in acetic acid, leaving the bone adhering to the plastic.

63.1 mi. Turn left at County Route 527 N/Glenside Ave.

63.5 mi. Take the 1st right onto Cataract Hollow Rd.

63.6 mi. Turn right into parking area for Watchung Reservation, Union County Parks an Community Renewal.

Park

Stop 3: Lower Feltville Formation (Washington Valley Member) Type section.

Latitude and Longitude: 40° 41.009'N, 74° 23.292'W.
Stratigraphic Units: Lower Feltville Formation, Washington Valley Member, Orange Mountain Basalt
Age: Latest Triassic (Latest Rhaetian) to ?earliest Jurassic (201.3 Ma)
Main Points:

1. More expanded section of Washington Valley Member
2. Best exposure of Cataract Bed and superposition of Vosseller bed.
3. Fissures in Orange Mountain Basalt
4. Copper at contact with Orange Mountain Basalt
5. First 100 ky after initiation of ETE
6. Limestones as products of super-greenhouse
7. In situ *Clathropteris*
9. Dinosaur tracks

Walk southeast and then southwest to the beginning of the loop near the end of Cataract Hollow Road about 2090 ft from the parking area. At the trail head, follow trail to southeast 330 ft and then about 600 ft to the northeast along the raised path (old canal levy) to Station 1 from where we will proceed to 3 additional stations to cover the stratigraphy at the type section of the Feltville Formation (Figure 19). Feltville is the name of the abandoned village of Feltville along Cataract Hollow Road, now part of the Watchung Reservation of the Union County Department of Parks and Community Renewal.

<u>Station 1</u> at about 40˚ 40.747'N, 74˚ 22.718'W is an outcrop of pink, gray, and white limestone interbedded with red siltstones and fine sandstones. This is the best vertical outcrop of the Cataract Hollow Bed and its type section. Fragmentary to articulated fish and occasional coprolites occur in the lower laminated part of the limestone. The fish bones and scales are pink and dark brown. The upper carbonate beds at this station are calcarenites that grade into the overlying red beds. The laminated and laterally continuous nature of the base the Cataract Hollow Bed as well at the articulated fish shows that it was deposited in quiet water, below wave base, generally lacking oxygen in contrast to the bed at Stop 1.

To the southeast, on the opposite side of Blue Brook is a ravine that cuts down down the dip-slope of the Orange Mountain Basalt. Thin to thick-bedded red siltstones of the basal Feltville Formation fill what is clearly a fissure in the basalt. Such fissures are common in the Orange Mountain Basalt and can cut down through much of the Formation, as seen at various quarries. In Nov Scotia, such fissures contain tetrapod bones, and it quite likely they do here as well, although this site has not been sufficiently prospected.

Proceed 500 ft northeast to bluff on west side of southeast flowing tributary (out of Cataract Hollow) to east-facing bluff that is Station 2.

<u>Station 2</u> at approximately 40˚ 40.822'N, 74˚ 22.718'W is an outcrop of poorly exposed black laminated carbonate and mudstone of the Vosseller Bed and surrounding gray clastics. Careful examination of the geometric relationship between this outcrop and Station 1 shows that this bed in superposition to the Cataract Hollow Bed, but this can be seen unambiguously at Station 4. Proceed southeast along the trail 270 ft crossing the small south-flowing tributary at Cataract Hollow and proceed about 620 ft to trail heading east up the slope (noting limestone of the Cataract Hollow Bed in the path). Continue 250 ft east to Station 3.

<u>Station 3</u> is at about 40˚ 40.860'N, 74˚ 22.718'W and consists of an outcrop extending from this ridge to the stream below and to the south. The contact between the Orange Mountain Basalt and the overlying Feltville is nearly at stream level and the overlying beds consist a nodular limestone succeeded by the transgressive portion of a Van Houten cycle bearing the Cataract Hollow Bed, the Cataract Hollow Bed itself and the overlying regressive red beds of the cycle, which we saw better exposed at Station 1. The larger blocks of the nodular limestone are very fine-grained and contain darwinulid ostracodes. The gray sandstones and siltstones of the transgressive portions of the cycle conatin plants in groth position including the horsetail *Equisetites* and the fern *Clathropteris* as well as dinosaur footprints. The Cataract Hollow Bed is deeply weathered here and invaded by modern tree roots, and the overlying beds are highly weathered. Just upstream from this station is the old copper mine described by Manspeizer (1980) at this site. I do not think there is evidence of intrusion here and the thermal maturity of the organic material here is not elevated. Proceed north and down the slope about 150 ft to path and walk about 960 ft northeast to Station 4 on the bluff on the north side of the west flowing tributary.

<u>Station 4</u> is at about 40˚ 41.015'N, 74˚ 22.718'W and has good outcrops of the Lake Surprise Bed, Vosseller Bed, through the Trailside Bed and then poor outcrops down through the Cataract Hollow Bed. It was at this site that I was able to confirm the superposition of the Vosseller and Cataract Hollow beds by digging for the latter after seeing the ACE cores. The Trailside Bed consists of purple and gray siltstomnes containg abundant *Brachyphyllum*. The cyclostratigraphic equivalent Stagecoach Road bed of the Shuttle Meadow Formation is a black calcareous mudstone with whole fossil fish and conhcostracans. The Lake Surprise Bed is a finely laminated black shale bed here with partly articulated *Semionotus*. The Vosseler bed is unfortunately very weathered and crumbly, although I have found articulated fish in it here.

Return to vehicles.

63.6 mi. Turn left from parking are onto Cataract Hollow Rd.

63.7 mi. Turn left at County Route 527 S/Glenside Ave.

64.0 mi. Turn right to merge onto I-78 E.

72.1 mi. Take exit 52 to merge onto Garden State Parkway N.

91.1 mi. Take exit 160 toward Fair Lawn/Hackensack.

91.3 mi. Slight right at Garden State Plaza Parkway.

91.3 mi. Take the 1st left onto Paramus Rd/W Passaic St. Continue to follow Paramus Rd.

91.8 mi. Merge onto NJ-4 W via the ramp to NJ-208 N/Fair Lawn/Hawthorne.

92.1 mi. Continue onto NJ-208 N.

97.5 mi. Take the Grandview Ave exit toward Wyckoff.

97.6 mi. Turn right at Grandview Ave.

97.9 mi. Turn right at Goffle Hill Rd.

98.5 mi. Continue onto Sicomac Ave.

98.8 mi. Turn left at Mountain Ave.

99.7 mi. Continue onto Sicomac Rd.

100.2 mi. Turn left at High Mountain Rd.

100.3 mi. Slight right at Belmont Ave/Passaic County 675.

101.7 mi. Turn right at W Overlook Ave.

101.8 mi. Turn left into parking area for strip mall.

Park

Stop 4: Upper Feltville Formation and Preakness Basalt, William Paterson University.

Latitude and Longitude: 40° 57.083'N, 74° 11.524'W.
Stratigraphic Units: Upper Feltville Formation, Preakness Basalt (flows 1 and 2)
Age: Earliest Jurassic (201.1 Ma)
Main Points:

1. Upper sandy Feltville Formation
2. Contact between Feltville and Preakness Basalt
3. Basal pillowed flow of Preakness Basalt (P-1)
4. Flow P-2 of Preakness basalt with low paleomagnetic inclinations
5. Largest single flow in world?
6. CO_2 sequestering possibilities in basalt

Walk west from parking area about 290 ft to entrance to abandoned quarry on south side of West Overlook Ave head towards stream walking about 300 ft to open area of the old quarry.

Manspeizer (1980) described this locality in that year's NYGSA guidebook noting the three basic units present. The upper Feltville Formation, a lower pillowed flow of the Preakness Basalt and a second massive and highly fractured flow of the Preakness Basalt. The Feltville Formation exposed in the old quarry and adjacent stream consists of interbedded tan and red sandstones and red and gray and purple siltstones. The latter contain sporomorphs and conifer fragments. In terms of the cyclostratigraphy of the Feltville this sequence is the interval of maximum precessional variability in the 100 ky cycle exhibiting the lowest precessional variability of its 405 ky cycle. The contact with the Preakness has a cyclostratigraphic age of about 201.1 Ma.

The basal flow of the Preakness as exposed here consists of a complex of pillowed basalts and the more massive flow lobes that fed the pillows all of which show considerable vesicularity. This is Preakness flow P-1 of Tollo & Gottfried (1992) and is a distinct and mappable unit extending at least from the ACE core transect to the north to near the border fault. It seems to absent from at least West Orange (I-280) to near Somerville, NJ, but reappears along I-275 at Pluckemin, NJ. I (Olsen, 1980a, 1980d) described outcrops of the lower flow but did not different it from the overlying flow P-2 of Tollo & Gottfried (1992). At this locality there is little or no visible metamorphic effect on the underlying Feltville Formation, which to be expected because P1 was extruded into water, presumably as deep as the thickness of the pillowed flow itself.

Proceed from the quarry up the hill on the north to West Overlook Avenue and then to the west to exposures of basalt. The contact between P-1 and Tollo & Gottfried's (1992) flow P-2 can be seen here. P-2 is the very thick flow that is present all over the entire extent of the Preakness Basalt and is characterized by having an intense splintery or prismatic fracture (Faust, 1975) and gabbroid layers (Puffer & Volkert, 2001). Prévot & McWilliams (1989) noted that this flow has unusually low magnetic inclinations compared to the other Newark basin flows. They also found the same low inclinations in the second flow of the Hartford basin Holyoke Basalt and the Deerfield basin Deerfield Basalt. Hozik (1992) showed that the Sander Basalt shared the low inclinations unlike all the other basalts of the Culpeper basin. All of these low-inclination flows have the same chemistries (Puffer, 1992, 2003) and all have gabbroid segregations and all but the Deerfield Basalt have the characteristic splintery fracture. In the Hartford and Deerfield basins, a flow of similar chemistry but having a tendency to be pillowed is present below the flow with the low inclinations. As pointed out by Prévot & McWilliams (1989) these directions would seem to indicate correlation, and correlation within the time frame of secular variation, suggesting that these flows represent the same eruptive event of no more than 10s or 100s of years. The Sander flow with the low inclinations is over 200 m thick, and P-2 of the Preakness Basalt is more than 90 m thick, and if the flow extended from Massachusetts to Virginia, it would be on of the largest lava flows known on Earth. In terms of environmental effects, it is surely the rate as well as the magnitude that matters, and the eruption of this flow, if it was indeed one eruption, would have had significant environmental effects. The largest single flow of the Columbia River Basalt is on the order of 5000 km^3 (Tolan et al., 1989), but if the Preakness and equivalents averaged 100 m in thickness, spanned 800 km along strike, and were 100 km wide prior to erosion, it would have a volume of 8000 km^3. This does not include the area spanned by dikes of the same composition that extend well into Canada (McHone, 1996).

Flow P-1 is highly vesicular at this locality, and P-2 has very significant fracture porosity and permeability. As previously discussed, Goldberg et al. (2009) argue that carbonation reaction in porous zones of basalt could provide a significant locus for carbon sequestration. The large amount of porosity and permeability is obvious here, but whether that holds at depth has yet to be demonstrated.

Return to vehicles.

101.8 mi. Turn left (west) from parking are onto W Overlook Ave toward Lenox Ave.

102.1 mi. Turn right at Mills Dr.

102.4 mi. Continue onto College Rd.

103.2 mi. Turn right at County Rd 504 W/Hamburg Turnpike. Continue to follow Hamburg Turnpike.

108.1 mi. Turn right at Terhune Dr. (US-202).

108.5 mi. Turn right at Brook Terrace.

108.8mi. Take the 1st right onto Green Knolls Dr/Pines Lake Dr W.

108.8 mi. Green Knolls Dr/Pines Lake Dr W.

Park

Stop 5: Middle Towaco Formation, The Glen, Pompton, NJ.

Latitude and Longitude: 40° 59.467'N, 74° 16.083'W.
Stratigraphic Units: Middle Towaco Formation, Colfax, The Glen, and Pines Lake Beds
Age: Early Jurassic (200.8 Ma)
Main Points:

1. Classic 19th century site never exploited
2. Well developed cyclcity
3. Basin margin facies
4. Lower trophic levels well represented
5. Species flocks of semionotid
6. Tracks
7. Pompton Tuff
8. Size of lakes

The Glen is a community reserve managed by the Pines Lakes Association. Visits to this site should always be coordinated through them.

About 65 m of the middle Towaco Formation extensively outcrops along the creek and its tributaries joining Pines Lakes and Pompton Lake (Figure 22), including the top of one Van Houten cycle, two complete cycles, and the lower half of another. Three deep-water units of the upper three successive Van Houten cycles outcrop in their entirety, with the cycles averaging about 25 m thick. These three beds are termed in succession, the Pines Lake Bed, the Glen Bed, and the Colfax Bed, corresponding to the designations P-5, P-4, and P-3 in Olsen et al. (1989) and papers by McCune, respectively.

The Glen was made known as a fossil locality by W. C. Redfield (1842) who reported on the lower two shales in succession producing fossil fish and intermediate strata producing reptile tracks. It is worth mentioning that William C. Redfield, was the first president of the American Association for the Advancement of Science in 1848 and great grandfather of Alfred C. Redfield of Redfield Ratio fame.

From the parking area, walk 0.2 mi southeast across bridge and turn west into woods to the west at the small tributary to the main stream. Proceed down hill about 135 ft along tributary to Station 1.

Station 1 is at approximately 40° 59.467'N, 74° 16.083'W. Exposed in the small run is the lowest deeper water unit, the Pines Lake Bed (P-5). This unit lies on gray sandstones and conglomerates with organically preserved roots, which in turn lie on red clastic rocks. The Pines Lakes Bed is a laminate, but is not microlaminated and has rare whole fish and fish fragments. The most distinctive aspect of the bed is the presence of stromatolites around trees (Figure 23). These have been termed arboreal stromatolites by Whiteside (2004), which were a part of the assemblage of lacustrine primary producers. The Pines Lake Bed is presumably the lower of the two beds described by Redfield (1842), although he gives the distance to the next fish-bearing unit is 60 m (200 ft) while it is certainly closer to 25 m. Proceed downstream about 150 ft south to near where the stream turns west and then northwest to Station 2.

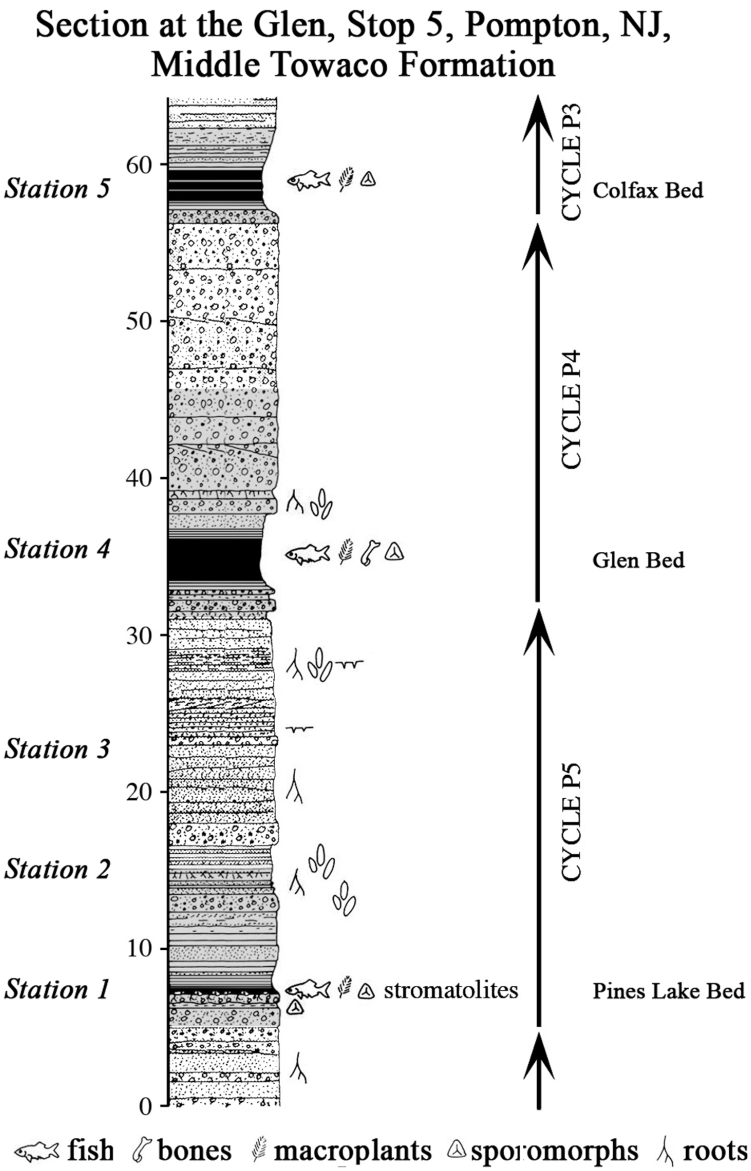

Figure 22: Section at Stop 5. White denotes red beds. Modified from (Olsen et al., 1989).

<u>Station 2</u> is at about 40° 59.442'N, 74° 16.080'W. We have passed up section through gray mudstones, sandstones and minor conglomerate into reddish thin-bedded siltstones and very fine sandstones with abundant dinosaur footprints. Proceed downstream about 200 ft.

<u>Station 3</u> is at 40° 59.454'N, 74° 16.116'W. Upsection from Station 3 we pass into presumably fluvial red sandstones and minor mudstones comprising the shallow water portion of the lower Van Houten cycle. The upper beds of this red sequence becomes finer grained and more highly organized. This red sequence is the interval mentioned by Redfield (1842) as being quarried by Peter M. Ryerson, Esq., of Pompton, NJ. This interval produced several dinosaur tracks, all brontozooids.

Continue downstream about 500 ft.

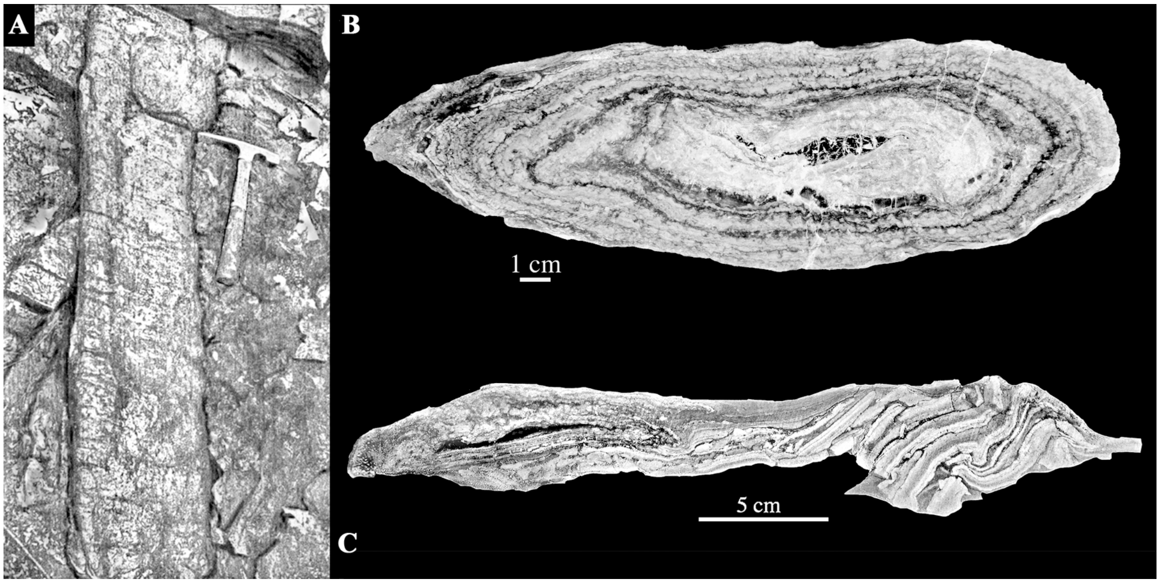

Figure 23: Stromatolites at Station 1: A, *in situ* +1.5 m stromatolite around tree in 1971 at Station 1, Stop 5; B, crossection of same stromatolite with organically preserved tree in center; C, imbricate slumped microbialite mat and associated stromatolite around small tree.

Station 4 is at approximately 40° 59.532'N, 74° 16.164'W at the point that the bluff comes down to near the level of the creek. Here, The Glen Bed is exposed at an accessible level. This is the upper fish-bearing unit mentioned by Redfield (1841). It is also the site of a quarry opened in 1980 by McCune and PEO, and excavated by Amy R. McCune (McCune, 1982, 1986, 1987a, 1986b, 1990, 1996, 1997, 2004).

The lower transgressive portion of this Van Houten cycle (paraphrased from Olsen et al., 1989) is characterized by an overall fining-upward trend from wave-reworked conglomerates into oscillatory rippled sandstones and siltstones. The high-stand sequence begins with oscillatory-rippled siltstones, passes rapidly into laminated siltstones with pinch-and-swell laminae, and then up into a thick microlaminated interval showing silt-carbonate couplets and many sub-millimeter to decimeter thick graded distal lacustrine turbidites. One turbidite has rare 1-2 cm diameter pebbles at its base. The total organic carbon content of these beds is relatively low compared to the finer facies of the Towaco, averaging less than 2% by weight. Complete, well-preserved fossil fishes are abundant in these beds (see below). The transition into overlying non-microlaminatcd silts is abrupt, as is the correlative disappearance of fish.

A peculiar feature of several decimeters of this sequence is the presence of white to tan irregular flattened objects in some of the finest microlaminated beds. Many shapes have cuspate edges or apparent holes, while others are more rod, vermiform, or even hair-like (Figure 24). The simplest explanation of this material is that it represents devitrified and flattened tephra, largely pumice particles that fell into the lake or were washed in from accumulations along shore. The latter interpretation makes more sense because these particles are so abundant on so many laminae. However, the presence of such foamy pyroclastics implies silicic eruptions which have not yet been documented in the CAMP

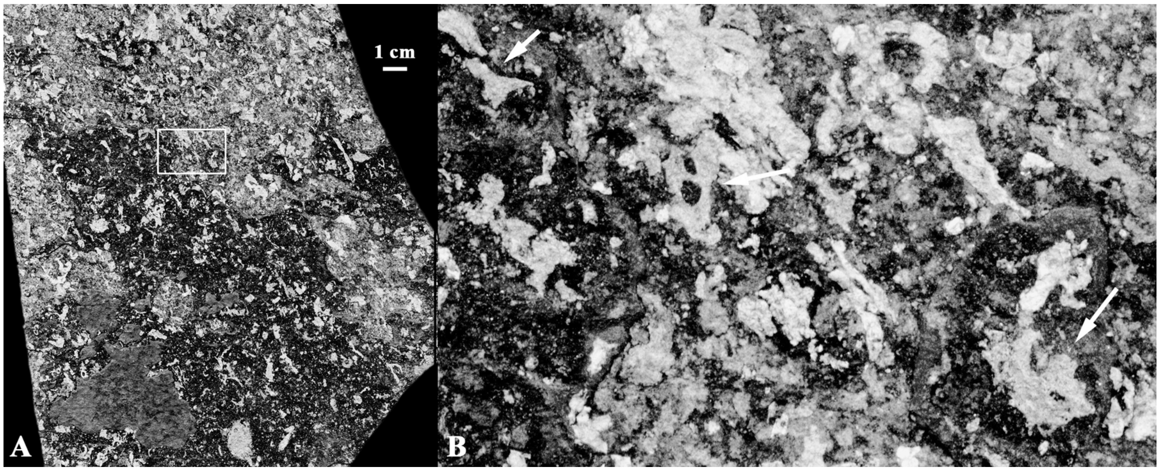

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Figure 24: Devitrified and flattened tephra fragments in microlaminated strata at Station 4. A, slab showing blobs, shards, and filaments; B, inset of A showing details with arrows pointing to vesicular blobs.

Sandstone and conglomerate at the transition into the strongly regressive and low-stand portions of the cycle exhibit features suggesting wave-reworking (Smoot, 2010), including virtually all the features described by LeTourneau (1985a,1985b), such as oscillatory ripples, "fitted fabric" of pebbles and cobbles, oriented plant debris. and well-sorted layers and patches of sand, granules, and pebbles. Also present are small lenses of oscillatory-rippled sandstone nestled in well-sorted conglomerate beds. Organically-preserved roots, and dinosaur footprints are present. Higher in the low-stand interval, the conglomerates appear fluvial in origin, although they remain to be examined in detail. These are comprise the thick sequence of conglomerate (Figure 22) referred to by Redfield (1842) as the, “variegated calcareous conglomerate" of Rogers.

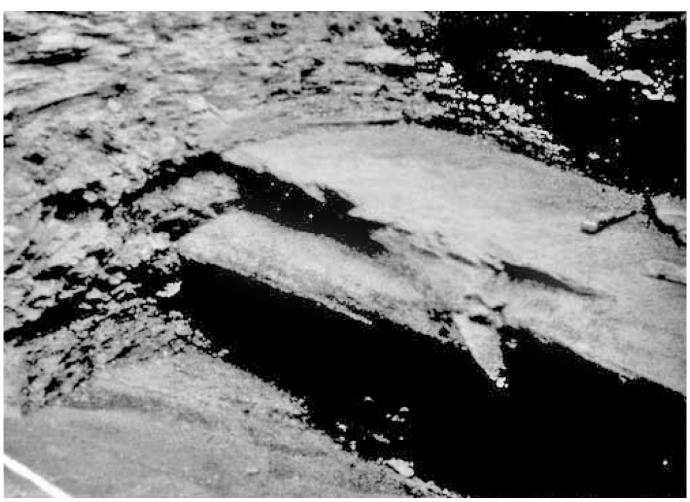

Figure 25: Nappe-like folds associated with bedding plane-subparallel thrust faults in Towaco Formation, The Glen, Stop 5, Station 4. Photo by A.R. McCune from Olsen et al. (1989).

Decimeter-scale folds are prominent in the microlaminated portions of this cycle. They could be subaqueous slump folds, but additional work done by McCune and PEO in the early 1980s, shows they are nappe-like drag folds of thrust faults propagating upward through major portions of the microlaminated unit. These folds show large amounts of thickening in their hinges, which are often cut by small faults. The fault-adjacent limbs are often sheared out, although completely oveturned bedding is locally present. Fish are deformed within these folds, shortened where they are perpendicular to the fold axis and elongated when they are parallel to it (Figure 25). The thrusts themselves are slickensided and sometimes polished, and they sole into bedding plane shear zones. The ductile behavior of the beds and the extremely low thermal maturity of the contained hydrocarbons (Pratt el al., 1988) demands that this thrust faulting and folding be early in the burial history of the units, prior to complete lithification, but after significant burial. An implication of these folds is that the deep-water unit is considerably structurally thickened.

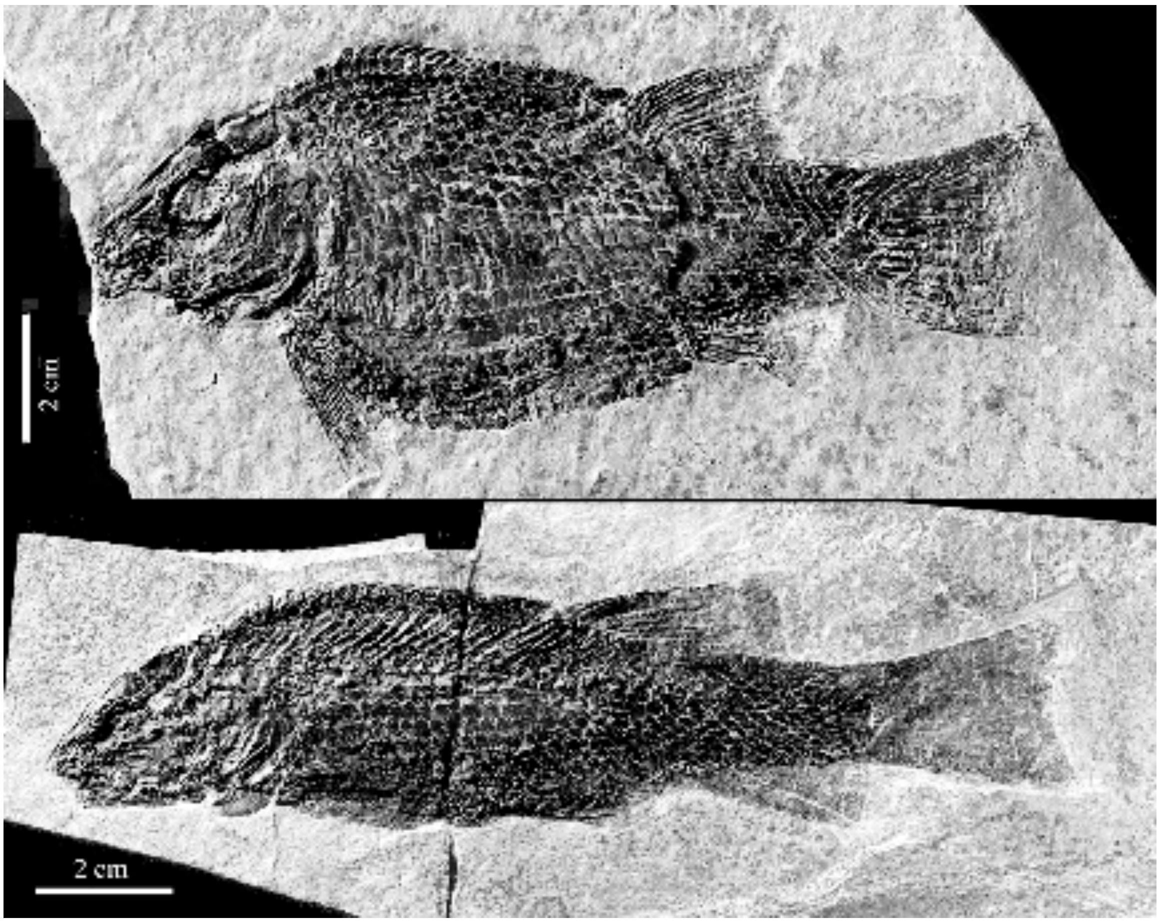

Figure 26: Two examples of the *Semionotus* species flock from the Glen Bed collected by McCune, Stop 5, Station 4.

On the basis of the collection of over 1700 fish made here (Figure 26), McCune described 21 species of *Semionotus* from the microlaminated deep-water part of this cycle, and suggested that a large proportion of them might have evolved *in situ*, endemic to this single lake, analogous to the endemic cichlid fishes of the African great lakes. McCune surveyed more than 2000 museum specimens from 45 localities in eastern North America and showed that six of the species in the Glen Bed were not found in deposits equal in age to or older and thus evolved within the lake level rise and high stand of the Glen Bed, with eight species occurring in older deposits and plausibly supplying the colonizing species. The evidence for seven species was equivocal. McCune (1990) showed that about 5.5% of the specimens have anomalous dorsal ridge scale patterns mixed in with otherwise stereotypic dorsal-ridge-scale patterns, and that dorsal-ridge-scale anomalies are significantly more frequent in older than in younger sediments of the Glen Bed, which she interpreted as being the result of relaxed selection during the early colonization of the lake. Results of McCune's analysis from the Glen Bed are shown in Figure 27.

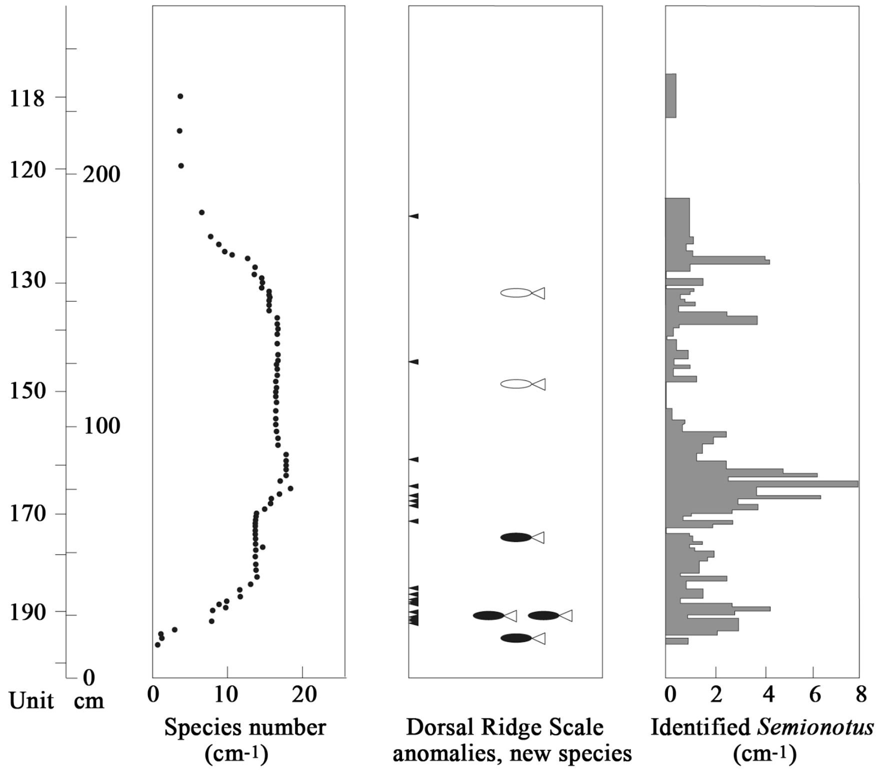

Figure 27: Microstratigraphic distribution of species number and dorsal ridge scales anomalies through 3 m of sediment from the Glen Bed. "Units" are of varying thickness (generally 1–2 cm) with each centimeter of sediment corresponds to about nine years of time (McCune 1990). The right panel plots identified *Semionotus* showing the number of individuals per centimeter of sediment, for which the dorsal ridge scales are visible. The left panel shows the number of species ranges that pass through a given microstratigraphic unit. Arrows in the middle panel indicate layers producing fish with dorsal ridge scales anomalies. First appearances of species are indicated in the same panel by filled fish symbols. Open fish mark singletons; filled outlines are species known from multiple individuals. Adapted from McCune (1990).

Proceed over the ridge in which the quarry is located and walk about 240 ft to the west to the large north-northwest-flowing tributary to the main creek and then proceed about 360 ft south-southwest along the tributary to the large outcrops of the Colfax Bed laminite.

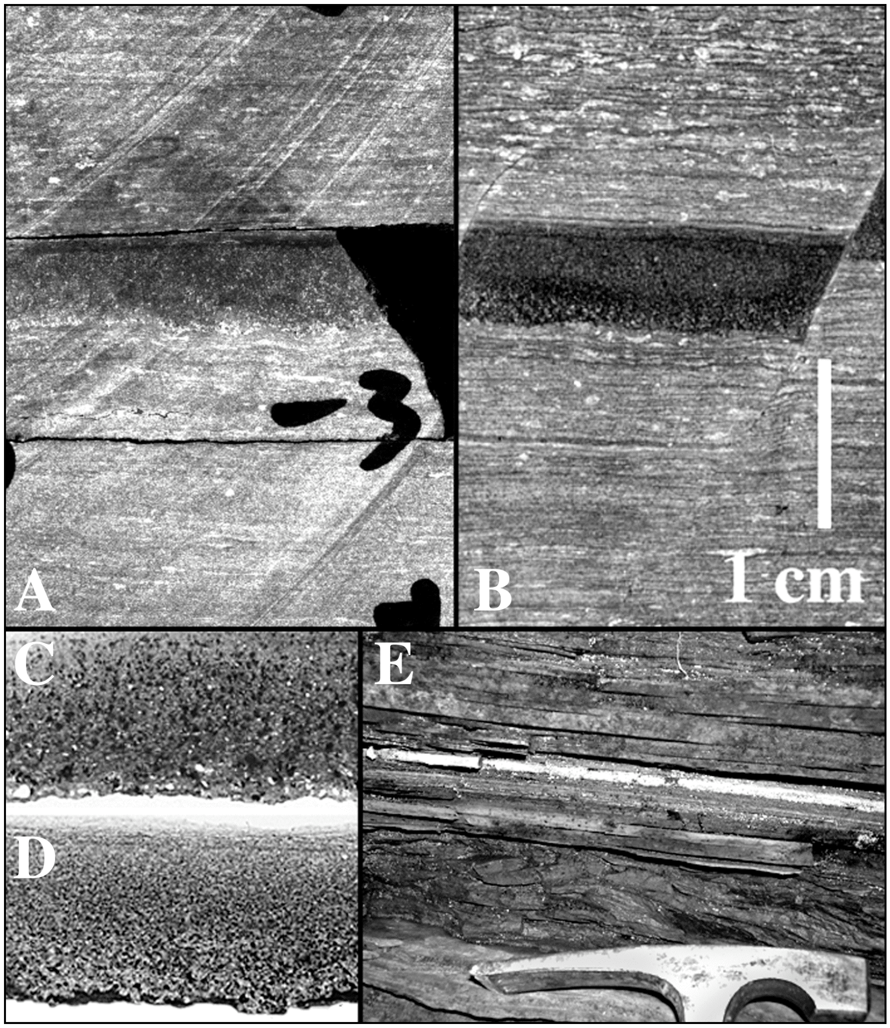

Figure 28: Pompton Tuff: A & B, Pompton Tuff in Towaco ACE core C-128 and slabbed section from the East Berlin Formation (Hartford Basin) "Stevens locality", Parmele Brook, Durham, CT, respectively; C & D, thin sections from the Colfax Bed, Stop 5 (type section) and ACE core C-128, respectively; E, outcrop of the Pompton Tuff (tuff is orange weathering, looking white here), the Colfax Bed, Stop 5, Station 5 (type section).

Station 5 is at about 40° 59.454'N, 74° 16.164'W. Here are outcrops of the Van Houten cycle bearing the Colfax Bed. The Colfax bed is similar to the Glen Bed. It is, however, much less studied. The lower few meters of the laminite have more silt and are thrown into even more folds than are present in the Glen Bed. Fish are not abundant in the Colfax Bed, but this could be because so much of the outcrop is deformed. So far only semionotids have been found.

This is the type section of the Pompton Tuff. After discovery of the small airfall ash in the East Berlin Formation (Olsen et al., 2005), I hypothesized it should be present in the Towaco Formation in the Colfax Bed, based on correlations described by Olsen et al., (1989) and Olsen et al., (1996b). Despite the fact that I had examined the Colfax Bed many times in the last 30 years, I had not noticed any ash layer. However, once knowing where to look, Gustaf I. Olsen and I found it on October 9, 2008 at this site (Figure 28). The ash is thus named after this locality, the Pompton Tuff. Subsequently, I examined the same cycle in the ACE cores and identified the tuff in ACE cores C-128 and PT-14. In the Hartford basin there are abundant conchostracans belonging to the genus *Bublimnadia*. Conchostracans had never been found in the Towaco Formation before but careful examination of the microlaminated interval in the homotaxial position above the Pompton Tuff revealed the same forms (Figure 29). These are representative of the lower trophic levels of consumers.

The extreme similarity in details of the Colfax Bed in terms of lake level phase, conchostracans, and the position of the Pompton Tuff in the correlative Westfield Bed in the Hartford basin would seem unusual in two isolated lakes with separate watersheds. Such similarity would not be surprising if the Colfax and Westfield beds were deposited by the same lake. If they were the deposited by the same lake and this lake extended to the Culpeper basin, it would be larger than the American or African great lakes, or Lake Baikal, as previously discussed.

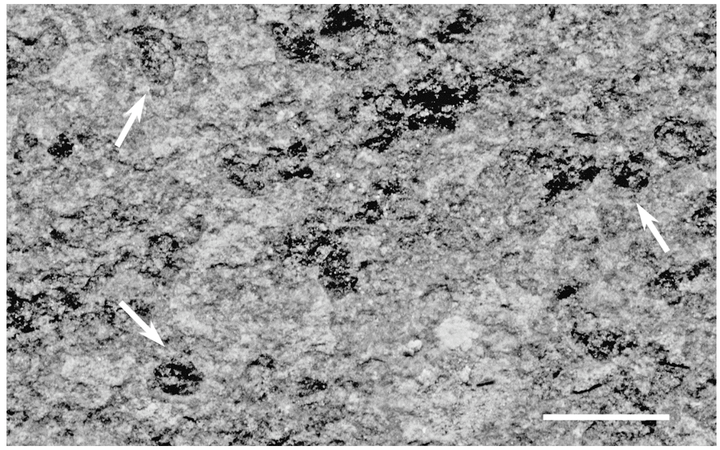

Figure 29: Conchostracans of the genus *Bulblimnadia* from the Colfax Bed, Stop 5, Station 5.

108.8 mi. Head north on Green Knolls Dr/Pines Lake Dr W toward Brook Terrace (80 ft).

108.8 mi. Turn left at Brook Terrace.

109.1 mi. Turn right at Terhune Dr. (US-202).

110.5 mi. Turn left to stay on Ramapo Valley Rd (US-202).

112.4 mi. Take the ramp onto I-287 N.

121.2 mi. Take the I-87 S/I-287/New York Thruway exit toward Tappan Zee Br/New York City.

121.8 mi. Merge onto I-287 E/I-87 S.

125.2 mi. Extensive exposures of Passaic Formation conglomerates.

133.4 mi. Exposures of Palisade diabase.

133.6 mi. Take exit 11 toward US-9W/Nyack/S Nyack.

133.8 mi. Turn left at NY-59 E.

134.2 mi. Turn right at S Highland Ave.

134.4 mi. Continue onto US-9W S/Hillside Ave. Continue to follow US-9W S following strike of the Palisade sill.

141.1 mi. Turn left into entrance of LDEO. Proceed to parking area.

END OF FIELD TRIP

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