

TRIP A6 AND B5: NEW INSIGHTS ON RIFT BASIN DEVELOPMENT AND THE GEOLOGICAL CARBON CYCLE, MASS EXTINCTION, AND CARBON SEQUESTRATION FROM OUTCROPS, AND NEW CORE, DRILL HOLES AND SEISMIC LINES FROM THE NORTHERN NEWARK BASIN (NEW YORK AND NEW JERSEY)

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INTRODUCTION

The Carbon Cycle, Global Environments, and the Sedimentary Record

It is a viable theory that in the big picture, much of Earth's geological record of is, in one way or another, a reflection of changes in the carbon cycle, itself uniquely (as far as we know in THIS solar system) and profoundly a reflection of life. Plausibly, because of the interplay between photosynthesis, weathering, CO₂, radiative balance, and the thermal gradient of the crust and mantle, the Earth is the only body around our Sun with granite and plate tectonics (Sleep, 2012) and the only body with an atmosphere with significant O₂ (Sleep, 2012). This field trip will view the Newark Basin (Fig. 1) and its sedimentary and biotic record through this lens of the carbon cycle. The fact that there is a Newark rift valley at all and that it is filled with largely red beds reflects this profound difference of our planet compared to our companions. Even the packaging of the Newark Basin sedimentary sequence, which is so profoundly cyclical and clearly paced by variations in the Earth's orbit, seemingly entirely extrinsic to the intrinsic Earth system, must reflect a very strong, nonlinear amplification by feedbacks between the carbon cycle and climate. Finally, the end-Triassic mass extinction (ETE) appears to be an example of a carbon cycle perturbation caused by tectonics – the eruption of the giant Central Atlantic Magmatic Province (CAMP) pumping CO₂ and SO₂ into the atmosphere. These same rocks may be part of our effort to sequester anthropogenic CO₂, itself a product of combustion of a half a billion years of photosynthetic natural carbon sequestration. The fact that the eruptions could occur at all may be a function of the same thermal gradient reflecting the radiative balance that makes our planet hospitable. Much of the theory of the Earth as it relates to life and the carbon cycle is on the fringe of knowledge, but it is useful to ponder how these things might work in order to frame testable hypotheses that may not only help us understand the Earth system from the seemingly arcane vantage of the Newark basin, but also frame hypotheses that have practical consequence for our future.

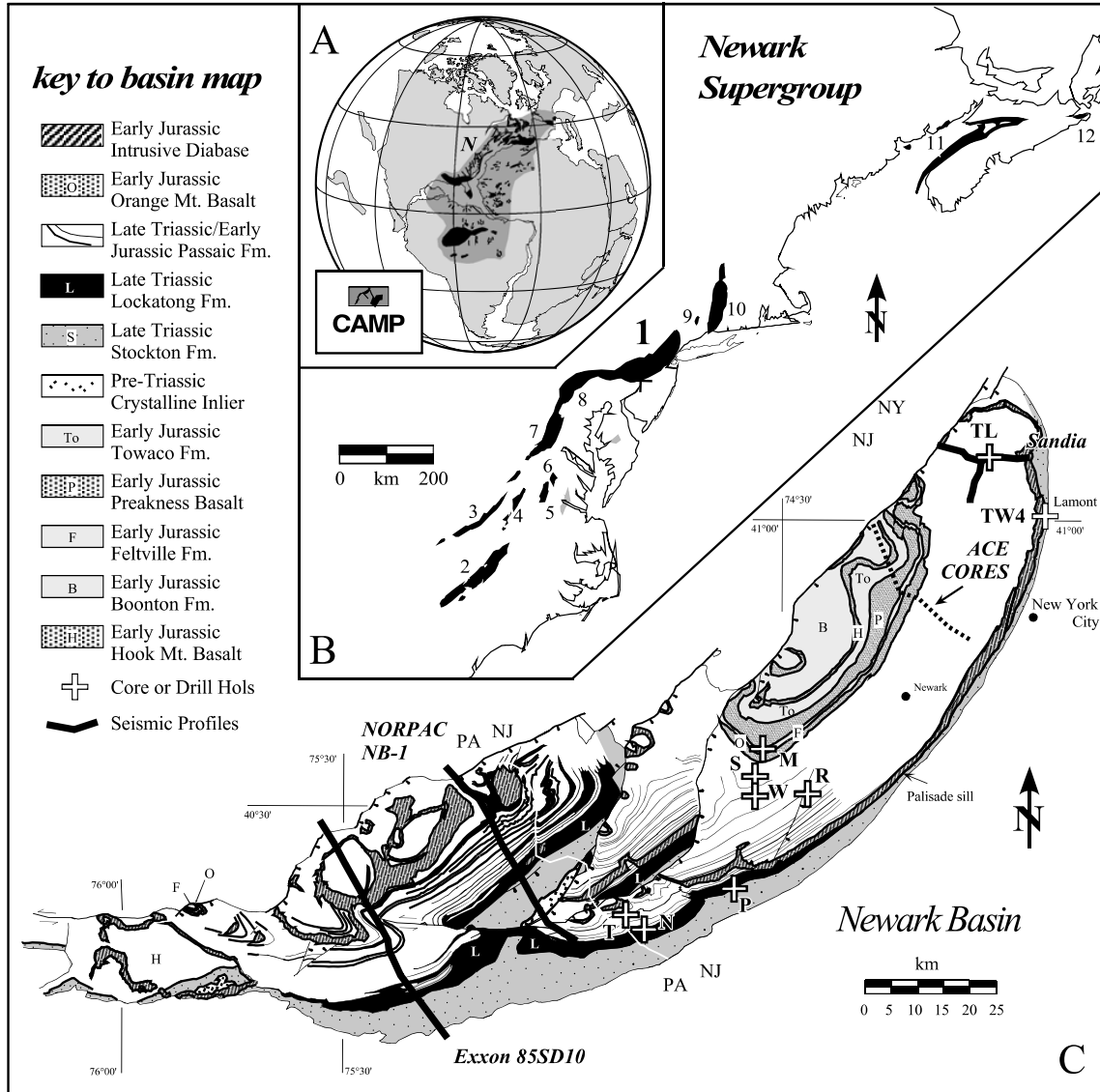


Figure 1: Location Map. **A**, Earth at 201 Ma showing the distribution of CAMP igneous rocks: A, Newark Basin. **B**, Newark Supergroup basins of eastern North America: 1, Newark; 2, Deep River, 3, Dan River; 4) Farmville; 5, Richmnd; 6, Taylorsville; 7, Culpepr; 8, Gettysburg; 9, Pomperaug; 10, Hartford/Deerfield; 11, Fundy; 12 Chebucto. **C**, Newark Basin with positions of core and drill holes and seismic profiles: M, Martinsville; N, Nursery; P, Princeton; R, Rutgers; S, Somerset; T, Titusville; TL, Tandem Lot; TW4, Test Well 4 (LDEO); W, Weston Canal.

The Newark Basin is a continental, compound, half graben rift basin (Fig. 2). While large compared to modern continental half grabens, it is a small part of very broad, still incompletely delineated rift zone that minimally stretched from the parts of central Pangea that would become northern Europe through western Africa, and eastern North America (Fig. 1). It is plausible that it extended all the way through to Panthalassa on the southwest, to the Tethyan rifts on the northeast, and between Greenland and Norway to the north. It is the largest continental rift complex we know to have existed, dwarfing those existing today, and formed as the African and north American plates began to pull apart from one another, plausibly as early as 260 Ma.

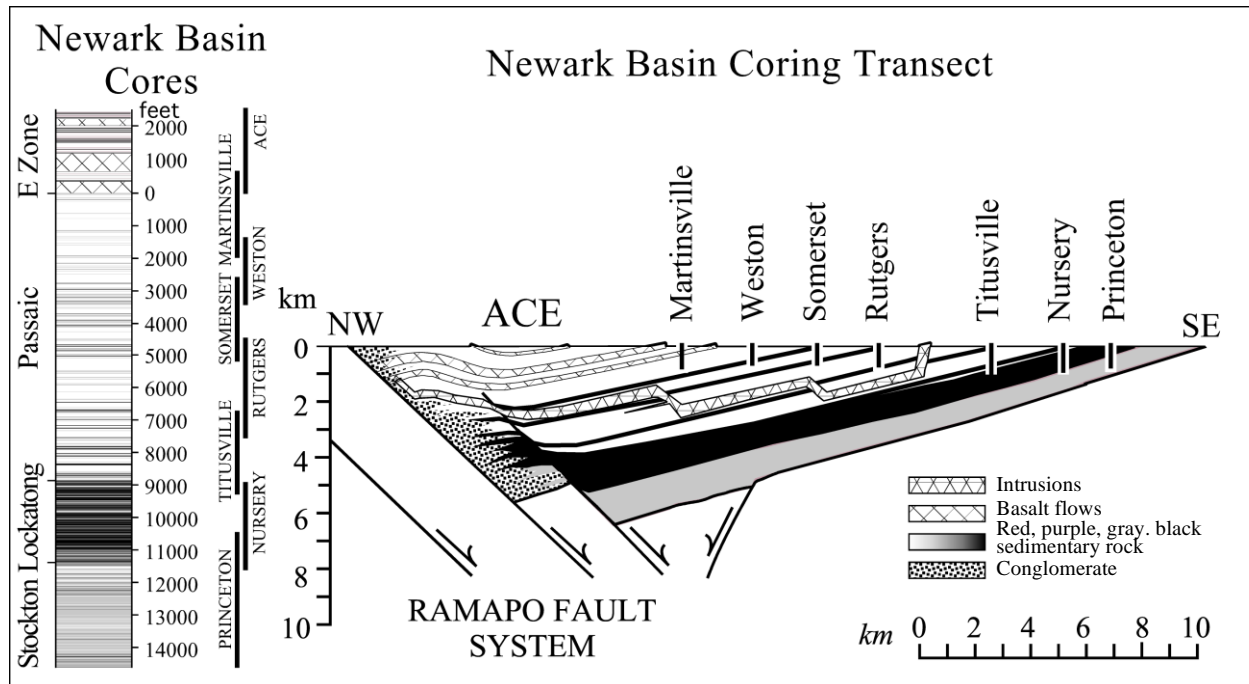


Figure 2: Diagrammatic cross section of the Newark Basin half graben showing the offset coring method. All the cores are projected onto one cross-section with no vertical exaggeration. Feet are shown because that was the original driller's units.

Continental rifting and basin filling seems to have begun in the latest Permian, ~40 Myr after the termination (~300 Ma) of the Permo-Carboniferous docking of North America and Africa, the last eastern North American Orogeny that completed the supercontinent of Pangea. The later orogeny temporally overlaps the Permo-Carboniferous Ice Ages. Multiple theories have been proposed for cause of the glaciations, but proxies (e.g., leaf stomata) indicating low CO₂ are roughly temporally correlated with the physical evidence of glaciation (Fig. 3). Current thinking is that in the long-term, or geological carbon cycle, variations in CO₂ are a function of the balance between the magnitude of carbon input, volcanic and metamorphic outgassing, and the removal of this carbon by carbon burial of organic photosynthetic products and chemical weathering of mineral silicates followed by precipitation and burial of carbonates. Because the proxy data for low CO₂ is correlated with high ¹³C values in the Late Paleozoic, enhanced organic carbon sequestration resulting from the spread of vascular plants in wetlands and over the continents has been a leading hypothesis for the glaciations (Berner, 1997) – and thus plant evolution drove the Late Paleozoic Ice Ages. However, both orogeny-driven increased chemical weathering sequestration of CO₂, (Raymo, 1991) and a reduction of arc volcanism lowering CO₂ input (McKenzie et al., 2016) have also been offered as explanations, as well as a plethora of other explanations including continental position, ocean gateway changes, and elevation (Montañez & Poulsen, 2013).

The oldest putative rift strata found so far is present in Morocco and Atlantic Canada, the former based on diagnostic skeletal, tetrapod remains, the latter on as yet unpublished paleomagnetic data (Olsen & Et-Touhami, 2008; Sues & Olsen, 2015). Because the oldest strata in most of the rift basins in the US are undated, a geographic progression of rifting cannot be assumed on the basis of this information. In addition the plate configuration at this time is still poorly constrained (Domeier et al., 2012). Angular

unconformities separate these Permian strata from definitive Triassic strata, but it unknown if the Permo-Triassic boundary (252 Ma) is recorded in any of these strata. This boundary (more correctly the end-Permian) marks what is supposed to be the largest mass extinction of the Phanerozoic, including on the continents, but low latitude, continental records of this extinction are extremely scarce, and therefore it is hard to assess its global magnitude. That said, at least the marine extinction was profound and seems to overlap the emplacement of the Siberian Traps and related pyroclastics and intrusions. Both a super-greenhouse effect from volcanogenic and metamorphic CO₂ and CH₄ (methane) and cooling from similarly derived sulphur aerosols have been proposed as drivers of this extinction (Svensen et al., 2009; Jones et al., 2016). These concepts have very strong parallels for the end-Triassic

extinction, for which an excellent record exists in the Newark and other Central Atlantic

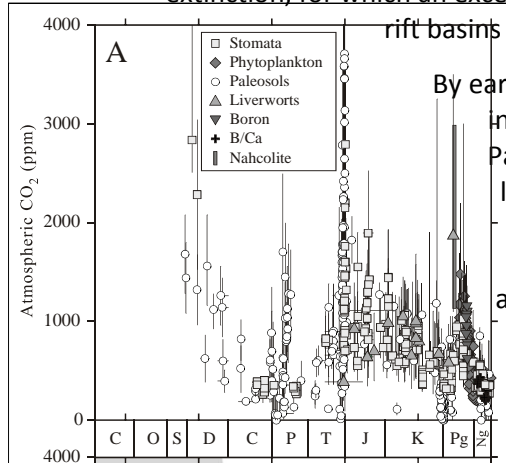


Figure 3: History of atmospheric CO₂ from proxies for the Phanerozoic (Royer, 2014). A, Individual CO₂ estimates. B, CO₂ estimates averaged into 10 Myr bins (white circles). Bins represented by single estimates are excluded; darker gray band captures ±1σ of the binned data set. Lighter gray band is GEOCARB model.

rift basins (see Stops 1.1, 1.3, 1.5 and below). By early Late Triassic time (Carnian, ~230 Ma) definitive deposition in many rift basins, including the Newark Basin, in Central Pangea had begun. Although a case can be made that at least locally strata of minimally Middle Triassic age are present at the base of the sequences (Olsen & Et-Touhami, 2008; Sues & Olsen, 2015). Although it is possible that Middle Triassic age strata occur in the basal part of the Newark basin (Olsen & Rainforth, 2003), definitive geochronological or paleontological data are lacking (Kent et al., 2016) (Fig. 4).

From Carnian (~232 Ma) through early Early Jurassic (~199 Ma) time, the record of the Newark Basin is dominated by lacustrine deposition with very good geochronological control from Milankovitch lake-level cycles (astrochronology) based on continuous coring and U-Pb dates (zircon geochronology) from lava flows and associated intrusions (Fig. 4) and is the foundation for the Newark-Hartford Astrochronostratigraphic Polarity Time Scale (APTS) (Kent et al., 1995, 2016; Kent & Olsen, 1999; Olsen and Kent, 1999; Olsen et al., 2011). This timescale uses the large-scale lithological sequences, termed McLaughlin cycles, paced by the 405 kyr long-eccentricity-cycle of the Earth's orbit as its metronome, but a full hierarchy of precession-related Milankovitch cycles are represented from 20 kyr to 1.8 Myr (Olsen, 1986; Olsen and Kent, 1996; 1999; Olsen, 2010) (see Stops 1.1, 1.3, 1.5).

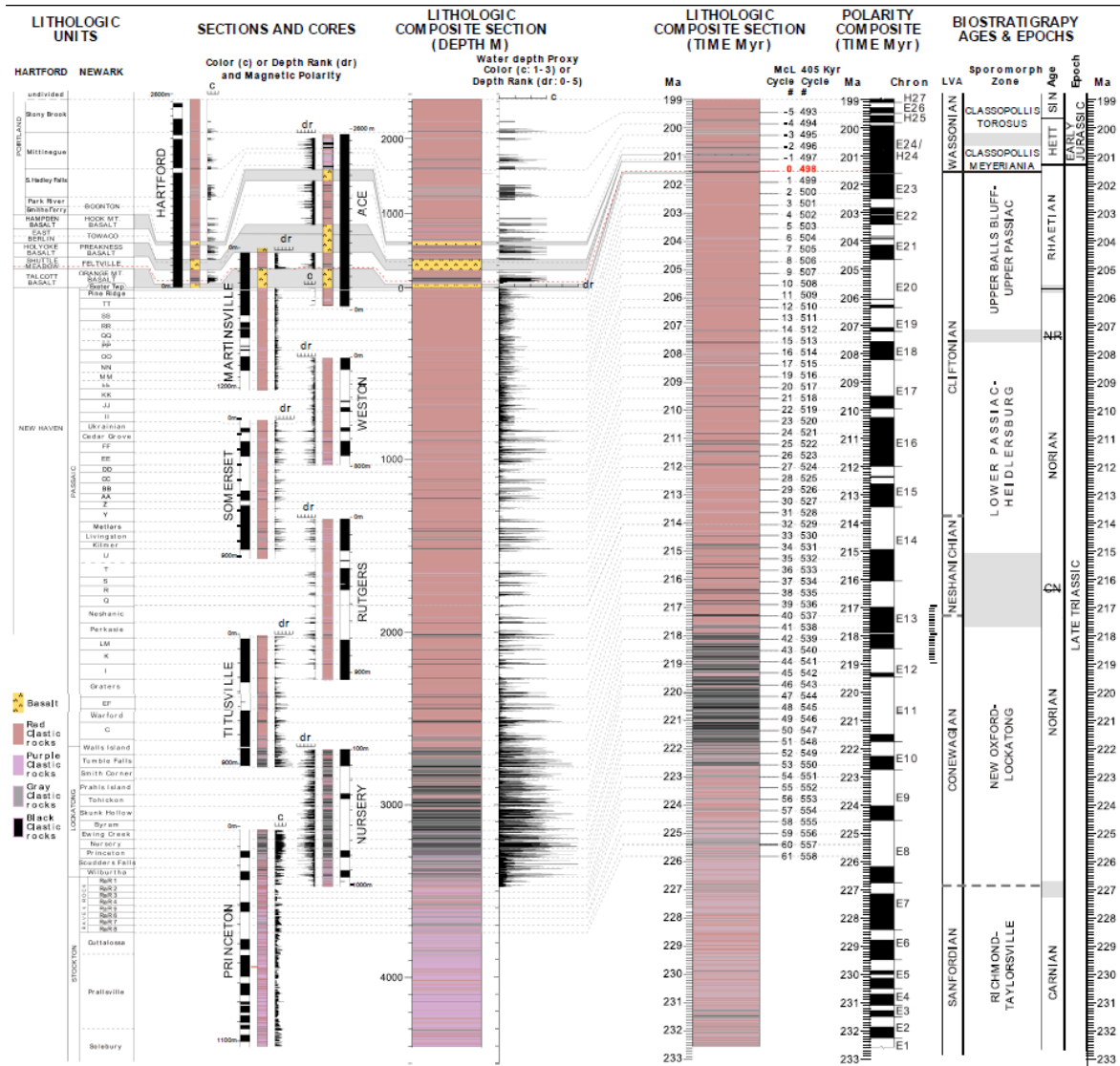


Figure 4: Newark Hartford APTS (from Kent et al., 2016). Stratigraphic framework for the Newark-Hartford APTS. Geologic columns show subdivisions of Hartford and Newark basin sections into formations and members (Olsen et al., 1996a). Magnetostratigraphy and lithostratigraphy are based on outcrop studies and short geotechnical cores from the Hartford Basin (Kent & Olsen, 2008) and in the Newark Basin, outcrop studies and numerous short Army Corps of Engineers (ACE) cores (Olsen et al., 1996b; Witte & Kent, 1990) plus seven long scientific drilling cores from the Newark Basin Coring Project (NBCP) (Kent et al., 1995; Olsen et al., 1996). Magnetic polarity is indicated by black and open bars for normal and reverse polarity. As seen in the cores, dominant Stockton lithologies are sandstones and minor mudstones and strata above that are lacustrine mudstones and fine sandstones with colors indicative of their actual hues. (Olsen et al., 1996a; b). The composite section is scaled from depth to time by assigning each of the 66 McLaughlin cycles to the 405 kyr eccentricity climate cycle anchored to 201.5 Ma at the base of the Washington Valley Member, corresponding to 201.6 Ma for the onset Chron E23r (= base of the Exeter Twp. Mb). Column labelled Ecc405 is the number of 405 kyr cycles counting back from the present to match peak eccentricity at the base of the Washington Valley Mb. at 201.5 Ma.

In the central parts of the Newark basin, based on outcrops near the Delaware River and cores, the basal part of the section (lower Stockton Formation) is fluvial and the upper Stockton Formation (Raven Rock Member) is marginal fluviolacustrine. Based on seismic profiles (Reynolds, 1994), the entire Stockton Formation may become lacustrine in the central deep, buried parts of the basin. The overlying Lockatong and succeeding sedimentary Formations of the basin are lacustrine, except along the northeastern and southeastern termini of the basin where they are replaced by largely fluvial strata (see Stops for Day 2). Sedimentary formations above the Passaic Formation are interbedded with three flood basalt formations of the CAMP that provide U-Pb zircon CA ID-TIMS dates. The continental record of the ETE begins in the uppermost tens of meters of the Passaic Formation of latest Triassic (Rhaetian) age (at 201.564 ± 0.015 Ma: Blackburn et al., 2013) and the Triassic-Jurassic boundary projects from its marine, ammonite-based Global Boundary Stratotype Section and Point (GSSP) in Austria to the Middle Feltville Formation at 201.3 to 201.4 Ma (Wotzlaw et al., 2014; Sha et al., 2015). The youngest strata in the Newark Basin (Boonton Formation) for which there is time control is about 200.6 Ma, but the lacustrine record in the nearby Hartford Basin of Connecticut and Massachusetts extend to about 199 Ma and is mostly if not entirely fluvial above that. The Hettangian-Sinemurian boundary is recorded in those strata at 199.7 Ma in the Newark-Hartford APTS (Fig. 4).

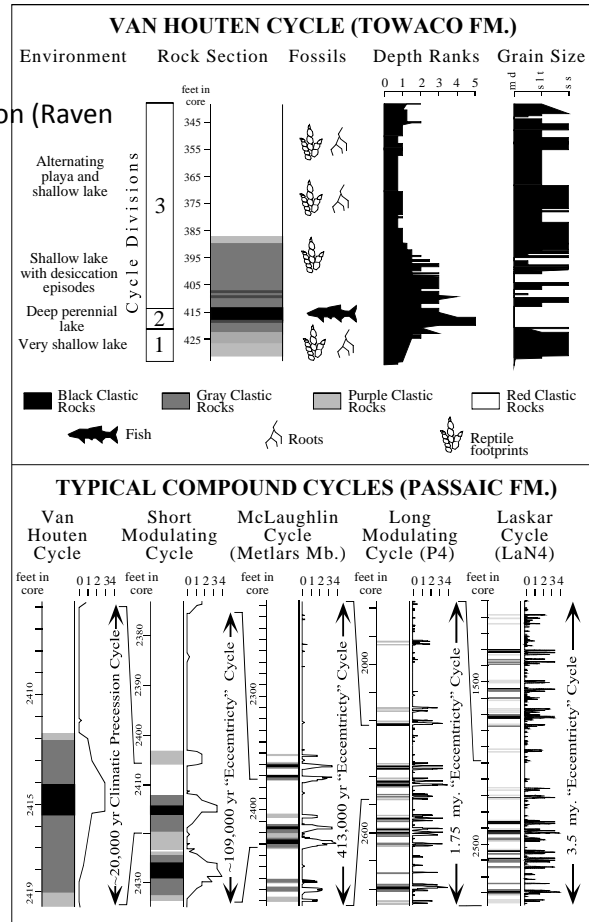


Figure 5: Van Houten and compound cycles (modified from Olsen and Kent, 1999).

Van Houten (1962; 1964; 1969) was the first to recognize the hierarchical cyclicity of the Newark lacustrine strata. Concentrating on the Lockatong and to a lesser extent Passaic formations, including the exposures at Stop 1.5, he correctly attributed the cyclicity to a celestial mechanical origin. Olsen (1986, 2010), Olsen & Kent (1996; 1999), Olsen et al. (1996a, 1996b), Kent and Olsen (2008), and elsewhere, showed that this astronomically paced cyclicity is characteristic of the entire lacustrine sequence (Fig. 5). Kent et al. (1995), Olsen et al. (1996), Kent and Olsen (1997), Kent and Olsen (2000; 2008), Deenen et al. (2011), Husing et al. (2014), Sha et al. (2015) and Kent et al. (2016) have shown that this cyclicity and associated paleomagnetic polarity zones are repeatable between different cores in the Newark basin and between basins on three continents. This cyclicity is remarkable, not so much because it influenced lake depth and sedimentary environments – it would be surprising if it did not, but rather because it is so extreme. The only Neogene cyclicity that is as obvious is in the Mediterranean sapropel record and related deposits (e.g., Rossignol-Strick, 1985; Grant et al., 2016) and those cover a much shorter period. Originally, it seemed that the extreme expression of cyclicity in the Newark (and other) Triassic-Jurassic basins might be due to the rifts forming in a particularly climatically sensitive tropical zone (Olsen, 1991); however, the intensity of that cyclicity changes dramatically through the section and

seems to correlate at the grossest level with $p\text{CO}_2$ (Fig. 6). Based on the soil carbonate proxy of atmospheric $p\text{CO}_2$ (Schaller, 2011; 2012; 2015) concentrations fluctuated about a mean of roughly 2500 ppmv during most of the Late Triassic (Fig. 6) (see Stop 2.2). However, from the late Norian (at about 210Ma) through nearly all of the Rhaetian, concentrations drop to about 1000 ppmv. During this time, the lacustrine cyclicality, so obvious previously, dropped in its apparent amplitude to barely discernable, but still there. The cyclicality then abruptly increased in amplitude during the emplacement of the CAMP and ETE when CO_2 reached nearly 6000 ppm and dropped down again afterward. This is especially remarkable in that not only is it seen in the Newark Basin, it is seen in every rift basin sequence in North America, Morocco, and even Europe – including in marine sequences (Olsen et al., 2015). While of much lower temporal resolution and with lower absolute values, the leaf stomatal record of $p\text{CO}_2$ agrees in outline (e.g., McElwain et al., 1999; Steinthorsdottir et al., 2011).

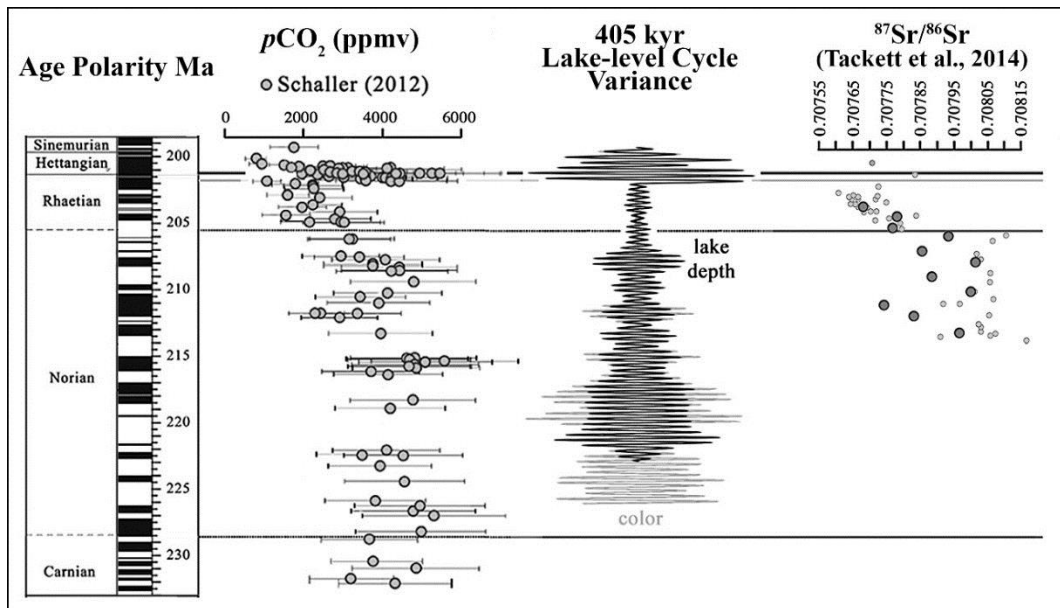


Figure 6: $p\text{CO}_2$ estimates based on the soil carbonate proxy from (Schaller et al., 2012; 2015) compared to the envelope of variance at the 405 kyr level (Olsen et al., 1999) and the strontium isotope data of Tackett et al. (2014). Note the correlated drop in $p\text{CO}_2$, lake level variance, and $^{87}\text{Sr}/^{86}\text{Sr}$ during the Rhaetian and the rise during the CAMP episode at around 201.6 to 200 Ma.) compared to the envelop of variance at the 405 kyr level (Olsen et al., 1999) and the strontium isotope data of Tackett et al. (2014). Note the correlated drop in $p\text{CO}_2$, lake level variance, and $^{87}\text{Sr}/^{86}\text{Sr}$ during the Rhaetian and the rise during the CAMP episode at around 201.6 to 200 Ma.

Models of increasing atmospheric anthropogenic $p\text{CO}_2$ predict an intensification of the hydrological cycle, coupled with warming and an implied amplification of the effects of orbitally-forced precipitation fluctuations, with wet areas will growing wetter and dry areas drier (Stocker et al., 2013). There is also some evidence that higher CO_2 amplifies the effects of Milankovitch orbital forcing, including the effects on the monsoon (Araya-Melo et al., 2013). In addition, strontium isotopes ($^{87}\text{Sr}/^{86}\text{Sr}$) in marine strata apparently track the CO_2 proxy with a dramatic decrease from about 0.70795 to 0.70765 (Tackett et al., 2014) suggesting both a mechanistic link though weathering and that the relative changes in the soil carbonate and stomatal proxy data are meaningful, even if the absolute values of the two differ.

Viewed this way, one can argue that many of the primary features of the variability of these Triassic sequences are controlled by climate variations amplified by effect of background levels of atmospheric CO₂. High climate variability would be expected at precessional frequencies during very high CO₂ and low climate variability at times of low CO₂. This parallels the concept that the overall context of very low CO₂ during the Neogene greatly amplified climate variability in the obliquity (41 kyr) and later eccentricity range (~100 kyr) by allowing the development of glaciers at high latitudes. The Milankovitch cycles in insolation did not CAUSE the extreme variability, rather they PACED the variability of an unstable, non-linear climate system that tunes to different frequencies depending on pCO₂, ice or lack thereof, and plausibly other contexts (e.g., Huybers, 2004; Daruka & Ditlevsen, 2016), analogous to how a medical pacemaker stabilizes the rhythm of a chaotically beating heart. Such hypotheses require testing by more direct climate indicators over a broader geographic spread as well as more sophisticated modelling that can take advantage of the climate proxies. If indeed this view of the interplay between the quasiperiodic orbital cycles and a highly sensitive, non-linear system is correct, it does not bode well for forecasting the effects of pCO₂ outside present boundary conditions.

Generalizing a bit, the high CO₂ times of the Triassic (and Permian) are often referred to as arid times, supposedly indicated by the abundance of red beds. Although globally that certainly was not the case. The Late Triassic and earlier Early Jurassic (Hettangian-Sinemurian) are a time for which there is no evidence for ice at the poles – no dropstones, glendonites, or tills (Frakes et al., 1992). In fact, the North Pole was located in northeast Siberia within a few degrees of the present Lapdev Sea where there are Triassic-Jurassic gray, coaly beds, and abundant plant debris and microfossil plants indicating a warm to cool temperate flora (Dobruskina, 1988; Ziegler et al., 1993; Ilyina & Egorov, 2008). Similarly, at 70° S latitude in what is now New Zealand similar gray deposits yield plant fragments of mesophytic (temperate) floral assemblages and relatively diverse sporomorph assemblages (Retallack, 1985; Zhang & Grant-Mackie, 2001). Evidently, both polar regions were temperate and relatively humid. In contrast, today's polar areas have extreme aridity with the two polar areas comprising the largest deserts on Earth (Loewe, 1974; Smiley & Zumberge, 1974; Doran et al., 2010; Callaghan et al., 2006), comprising about 28.6 million km² or 19 % of the area of land on Earth¹. With the polar continental areas being relative humid during the Triassic-Early Jurassic, why the bias towards thinking the Triassic was an arid time? One important reason, is that many, maybe most, locations of post-graduate education were located in the tropics to subtropics of early Mesozoic Pangea, at one time or another. The second major reason is that central Pangea drifted northward during the Triassic-Jurassic, producing a highly diachronous swath of red beds and aeolianites (Kent & Tauxe, 2005), with the diachroneity not visible until non-biostratigraphic means of correlation became available (e.g., Kent & Tauxe, 2005; Irmis & Whiteside, 2010; Irmis et al., 2011; Olsen et al., 2011; Kent et al., 2014). In combination, it means that nearly every major location of higher learning, from Cape Town to New York, Berlin, to Stockholm has a nearby deposit from one part of the Triassic-Jurassic or the other, with red beds. A corollary to this is that the supposed aridification trend (Parrish, 1993) through the Triassic actually reflects the northward migration of central Pangea as well as southern high latitude sites from zonal more humid to arid climate belts. Thus, northern Europe and Greenland actually show the reverse trend, going from arid to humid though time, exactly what one would expect from a northward translation of those areas from the subtropics into the temperate latitudes. The null climatic hypothesis is that nothing changes, and it cannot be falsified until time and plate position are accounted for. When time and plate motion ARE accounted for there is no evidence of global aridification and no evidence that the arid belts were any wider in the Triassic-Early Jurassic, than at the present (Kent & Tauxe, 2005).

¹ Derived from <http://geology.com/records/largest-desert.shtml>

Thus, the famous red beds of the Newark Basin are NOT typical of the Triassic, anymore that the climate of Timbuktu is typical of the Earth today. We can go further and state quite the opposite of the usual bias for Triassic climate. It is actually, on the whole, wetter than today due to the lack of polar deserts. Once we account for plate position and time, there is a residuum of change that then may require special explanation, such as at the ETE (see below).

The Broad Terrane

Historically, the northern Newark basin has been a conceptual battleground area. It has been a locus of the “broad terrane” debate began by I. C. Russell in 1880 (although there were predecessor concepts, as he noted), which may be the most persistent controversy involving these basins and it melds into very substantive debates on not just the interconnectedness of the presently individual basins, but also on the area and volume of the CAMP, thereby bearing on the magnitude of CO₂ and SO₂ release and the cause of the ETE and other mass extinctions. Because the northeastern terminus of the Newark Basin is the closest part of the basin to the Hartford Basin, new data on its three dimensional geometry and the sedimentary and volcanic facies bear directly on the problem and we will examine that data on stops on Day 2.

Russell argued that the Newark and Hartford basin were once part of a much larger more symmetrical basin, now arched in the middle, and very deeply eroded. The basic evidence for this he listed in 1879, 1880, and 1892 (with later additions by others) paraphrased as follows:

- 1) The strata in the Newark Basin largely dip to the west and those of the Hartford Basin deep to east.
- 2) The distribution of marginal conglomeritic facies in the basins is highly asymmetrical and generally limited to one side.
- 3) There is a small basin in between – the Pomperaug Basin,
- 4) There are boundary faults² that determine the margins of opposing sides of the basins, and that on the west of the Newark Basin seems traceable up the Hudson River (for which there is no modern evidence).
- 5) The strata of the basins are very deeply eroded.

After Russell, many authors have discussed and debated the broad terrane hypothesis. Wheeler (1938) summarized Russell’s arguments and cited evidence from a well boring that suggested that there is a continuation of the Hartford Basin in the subsurface to the south, below Long Island, and also provided a clear conceptual diagram to illustrate the overall concept (Fig. 7). The more recent proponents have been Sanders and colleagues (Sanders, 1960; 1963; 1974; Sanders et al., 1981; Friedman et al., 1982), and Hutchinson and Klitgord (1988). The basic 20th century concept is that the Newark and Hartford basins formed as a single large full graben within an extensional regime that was then arched and eroded, producing 2 asymmetric half-grabens (Fig. 7). The opposing concept has been termed the “isolated basin” model, which asserts that the strata of the various Newark Supergroup basins never extended much further than they are preserved today. Proponents of this model have included W. B. Rogers (1842; 1860, and elsewhere), Glaeser (1966), Savage (1968), Klein (1969), Abdel-Monem and Kulp (1968), and Faill (1973; 2003).

Because the strata and hanging wall onlap surfaces of the Newark and Hartford basins project into

² This was not clear in I. C. Russell’s time but was introduced by Barrell (1915) and followed up on by W. L. Russell (1922) and Longwell (1922, 1937).

space, and are unquestionably deeply eroded, it is clear that the basins must have been to some extent larger than now, the questions are: 1) “how much bigger?”, and 2) “did they connect by sedimentary or volcanic plains, or water?” However, it seems clear that the specific geometry expressed in the diagrams of Wheeler (1938) and Sanders (1963) (Fig. 7) cannot be correct, because the Pomperaug Basin section is very thin compared to counterparts in the Newark and Hartford basins, exactly where it should be thickest or at least intermediate in thickness between correlative strata in the Newark and Hartford basins. From the basal (East Hill Basalt), overlying sedimentary Cass Formation, to the top of the second basalt formation (Oreanaug Basalt) only about 110 m of strata is present, and that is correlative to about 560 m in the Newark basin and 400 m in the Hartford Basin. Furthermore, the basal formation of the basin, the South Britain Formation underlying the East Hill Basalt comprises ~270 m of red beds while strata below the Orange Mountain Basalt (correlative to the East Hill) is ~3000 m in the northern Newark Basin (Olsen and Rainforth, 2003) and 2000 m below the Talcott Formation (correlative with the East Hill) in the Hartford Basin (Kent & Olsen, 2008) (thickness of Pomperaug strata from Huber and LeTourneau, 2006). Therefore, the geometry of the so-called Danbury anticline of Sanders (1960; 1963; 1974) must have been in existence during sedimentation or developed during sedimentation if it is an anticline at all.

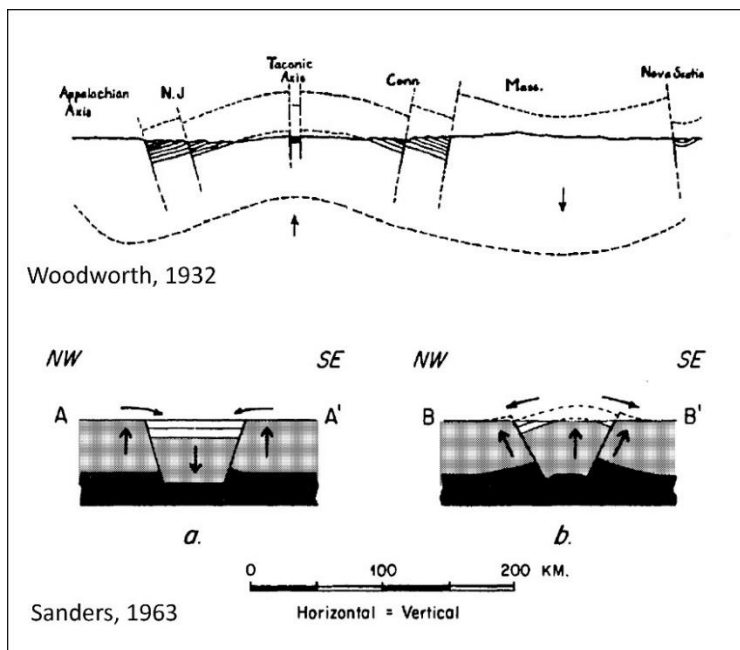


Figure 7: Broad Terrane hypothesis as expressed by Woodworth (1932) and Sanders (1963).

Olsen and Schlische (1987) proposed that the Pomperaug Basin was a crustal collapse graben (cf., McClay & Ellis, 1987), associated with rotation and extension on the conjoined hanging wall blocks of the Newark and Hartford basins. While the basic geometry is similar to what might be expected of a crustal collapse graben, we would argue that the length scale is inappropriately long compared to the thickness of the brittle crust. In this case, the Pomperaug Basin is more simply explained as just another of the many half grabens developed in the wide rifting zone of central Pangea, and has no special generative relationship with the Hartford and Newark basins.

That said, even if the strict geometrical construction of the Broad Terrane model is falsified by what is

known now of the thicknesses of the strata in the Pomperaug Basin, that does not mean there was not a sedimentary connection between the Newark and Hartford basins, or between them and any of the other basins. Likewise, a local source for sediment or clasts from presumed highlands between the basins also does not obviate the possibility of sedimentary connections between those same basins.

McHone (1996) argued for a modified broad terrane model, in which the preserved basins formed more or less as half grabens and there was at least local relief between basins; however, the basins could have been interconnected by sediments. Additionally, the CAMP lavas covered vastly larger areas than today, with their former extent indicated by the numerous dikes of the CAMP intruded into basement rocks

now currently lying outside the areas of the preserved sedimentary basins.

Recently, Withjack et al. (2013) using a combination of seismic, field, core, borehole, and vitrinite-reflectance data have shown that the Newark basin was initially narrow and markedly asymmetric with significant thickening toward the basin-bounding faults. As the basin became wider through time the fanning towards the border fault became spread out over a longer distance and hence subtler.

Projection of the cross-sectional geometry from the northern Newark Basin to the southeast, approximately perpendicular to the border fault system suggests that the Newark basin may have merged with the New York Bight Basin, which has a northwest-dipping border fault, producing a full-graben as much as 100 km wide and up to 10 km deep (along the border faults) by the cessation of rift sediment accumulation and before continued tilting and post-rift erosion. The New York Bight Basin (Hutchinson et al., 1986), lies along strike of the Hartford basin and may still be connected to it in the subsurface.

That the Newark basin was much wider than it is presently is consistent with studies of other Triassic-Jurassic basins in Eastern North America and Morocco (Withjack et al., 1995; 2009; 2010; 2012; 2013; Withjack & Schlische, 2005; Letourneau, 2003; Malinconico, 2003; 2010).

New data from the northern Newark basin

Four previous guidebooks have dealt with the Northern Newark basin in Rockland County, New York (Savage, 1968; Sanders, 1974; Olsen & Rainforth, 2003). All are based on scant surface information, and all of them discuss the Broad Terrane model, at least to some extent. Acquisition of two seismic profiles, and drilling of a deep (6800 ft) test boring, and recovery of 352 m of core from a 550 m core in this area provides insights into the Broad Terrane issue.

Core TW4 and Surface Data

It has been apparent for over 30 years that the lower part of the Newark Basin section thins towards the east and the northeast (Manspeizer and Olsen, 1981; Olsen et al., 1996; Olsen & Rainforth, 2003; Withjack et al., 2013). This is most obvious in the thickness of the Stockton and Lockatong formations that thin by a factor of 6 from the central to the northeastern part (Bergen County) of the Newark Basin as well as in the thickness of lacustrine cycles in the Lockatong Formation that thin by of four towards the northeast (compare core at Stop 1.1 to exposures at Stop 1.5). It is also clear that the proportion of tan and gray arkose in the Lockatong increases in the same direction with the most northeastern outcrops of the Lockatong (northeastern Bergen County) being almost entirely composed of tan sandstone with only minor gray mudstone (Parker et al., 1988; Olsen and Rainforth, 2003; Monteverde, 2011). Rockland County outcrops have proved difficult to interpret in terms of the standard Stockton, Lockatong and Passaic formations and the same rock units have been variously labeled each of these names by various authors, with single authors such as PEO changing formational designations through time.

Core TW4 (Fig. 8) (41.002928, -73.910618) was acquired just north of the New Jersey-New York border on the campus of Lamont Doherty Earth Observatory (Zakharova, et al., 2016), funded by the TriCarb Consortium for Carbon Sequestration and EPA STAR grant 834503. It was spudded in the Palisade Sill and ended in gneiss basement. The sedimentary succession in the core consists mostly of white, tan and pinkish arkosic sandstone and red to purplish mudstone. The only candidate for Lockatong Formation is a few meters of gray massive mudstone and white arkose in contact with and as a xenolith within the sill. The most distinctive facies in the rest of the section (except at the bottom) is abundant purplish-red

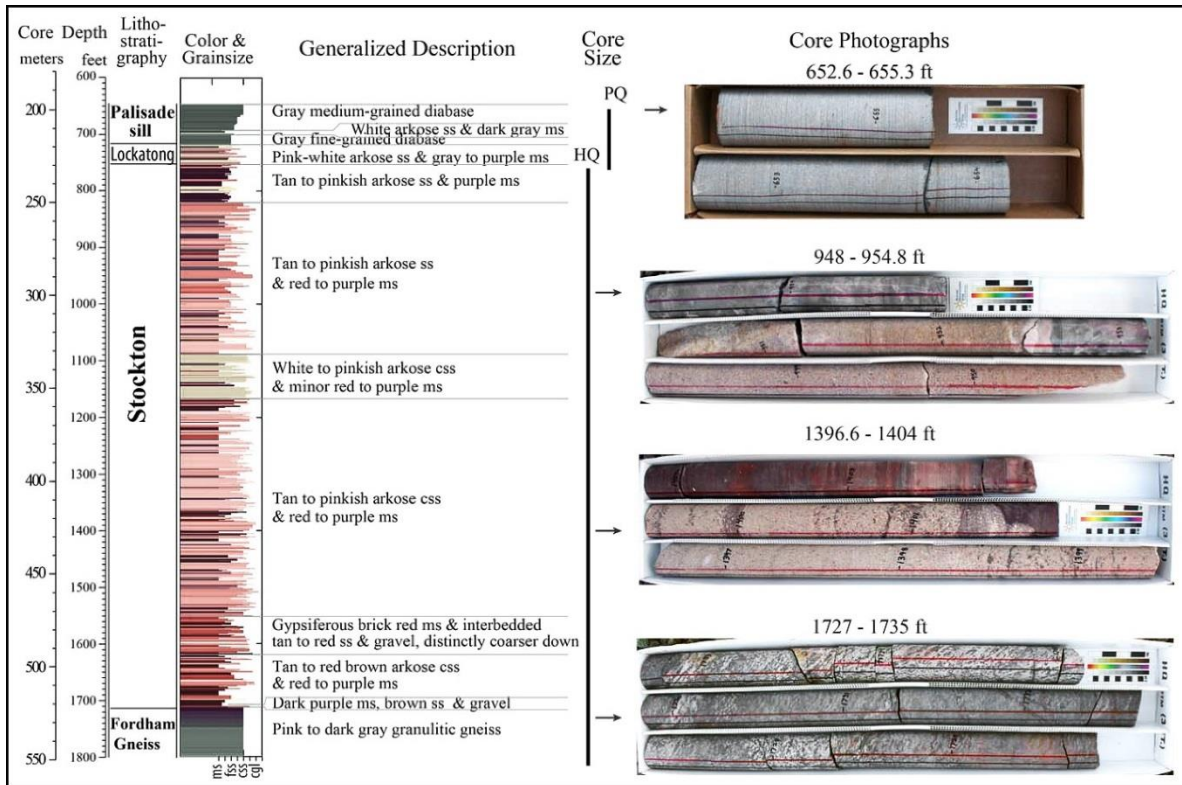


Figure 8: TW4 core from Zakharova et al. (2016).

massive mudstone with very large ptygmoidal dessication cracks filled with grayish coarse arkosic sandstone seen at Fort Lee and just to the north to Snedens Landing (Savage, 1968; Van Houten, 1969) and in the core, but never in the Passaic Formation. Associated sedimentary rocks include white, tan and pinkish arkosic sandstone and red to purplish mudstone. In total, this facies does not resemble coarser grained Passaic Formation but finds its closest match in what is identified as Stockton Formation below undoubted Lockatong Formation in Fort Lee, NJ (Stop 1.2) and Alpine, NJ on the upper surface of the sill (Fig. 9) (Monteverde, 2011). Prominent benches of tan arkosic sandstone are present and exposed near the level of the Hudson River, sporadically outcropping going north from Snedens Landing to Tallman Pool (41.032113°, -73.912819°) and in Piermont (41.041373°, -73.917571° and 41.041846°, -73.917116°), where the exposed section appears to be climbing stratigraphy as does the sill. Given that the sill jumps up section at Alpine, NJ (Monteverde, 2011), cutting through the tan, arkosic remnants of the Lockatong Formation, these tan sandstone at Tallman and Piermont may also pertain to that formation below the sill and its appearance before passing below the river. There are good exposures of red sandstones and mudstones closely resembling normal sandstone and mudstone facies of the Passaic Formation in northern Piermont in small stream gorges (41.051588°, -73.921605°, 41.056002°, -73.921888°), at Grand View-On-Hudson along a path (41.060802°, -73.921601°), and in several abandoned quarries (e.g., 41.071227°, -73.921185°). There are no significant tan sandstones and no hint of the purplish-red massive mudstone with the very large ptygmoidal dessication cracks, suggesting the sections are all above what is present in the TW4 core (as suggested by Van Houten, 1969). From South Nyack north to Upper Nyack, the sill dramatically rises and falls in stratigraphy making an arcuate outcrop pattern (Fig. 9), as stated by Kümmel (1900). Near the waters edge in Nyack (41.087859°, -73.917730°) the outcrops resemble normal Passaic Formation as do the low cuts for the New York State

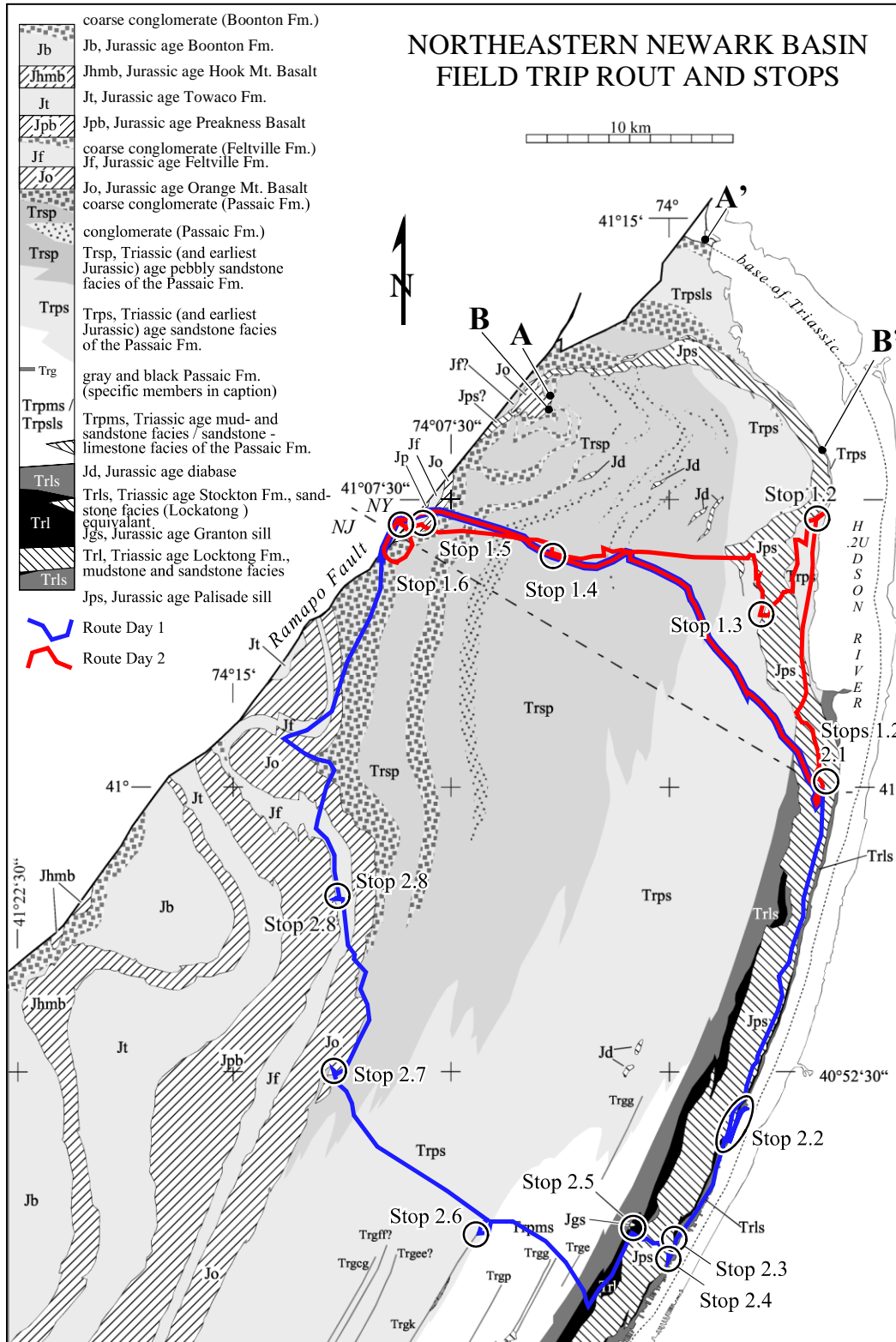


Figure 9: Geologic map of the northern Newark Basin with field stops. Modified from Olsen et al (2003).

Thruway. At Nyack Beach State Park (Stop 2.2) the long red bed section on the site of the former Manhattan Trap Rock Quarry looks like Passaic Formation and largely lacks tan arkosic sandstones. There is no hint of the purplish-red massive mudstone with the very large ptygmoidal desiccation cracks typical of the section in most of the TW4 core. Near the very top of the section at Nyack Beach State Park there are purplish mudstones and tan sandstones, but these are difficult to separate from contact metamorphism due to the adjacent sill. More convincingly, at an exposure just to the north of 9W along the northwest terminus of Upper Nyack (41.112127°, -73.927668°), there are tan and gray strata consistent with coarser facies of gray portions of the Passaic Formation. These exposures have produced several small brontozoid (c.f. *Grallator*) (theropod dinosaur) footprints, possible *Atreipus* (non-dinosaurian, dinosauromorph) footprints, bone fragments, and fish scales. Savage (1968) documented that the arkose, interbedded with the unusual purplish-red massive mudstone seen at Snedens Landing, contains abundant zircon, rutile, and anatase, which consistent with TW4. Whereas, the Passaic Formation in Rockland has abundant tourmaline, garnet, and metamorphic rock fragments.

Going farther north along the Hudson, Kümmel (1900) documents the sill cutting up section progressively along Verdrietege Hook (Fig. 10) to the Rockland Lake trap rock quarry to north of Trough Hollow to Waldberg Landing in Haverstraw, the site of a small sandstone quarry that has produced many fossils over the last several decades (Olsen and Rainforth, 2003). According to Kümmel, who had good access to the exposures due to the quarrying activities and denudation at the time, the increase in height of the stratigraphic section exposed near waterline is about 56 to 170 m / km which means that from just north of Nyack Beach to Haverstraw, the section traversed is very roughly about 650 m. Some of this traverse has been described by the field guides of Savage (1968) and Sanders (1974). Along this traverse, now a lovely bike and pedestrian path in Hook Mountain State Park, there are several gray to buff sandstone and siltstone intervals separated by red beds that are most simply interpreted as coarser facies of gray sequences occurring at ~405 ky intervals in the Passaic Formation. Exposures and outcrops in Haverstraw near the level of the river (41.202875°, -73.973494°) and at higher elevation (41.200319°, -73.984453°) are of typical Passaic Formation facies.

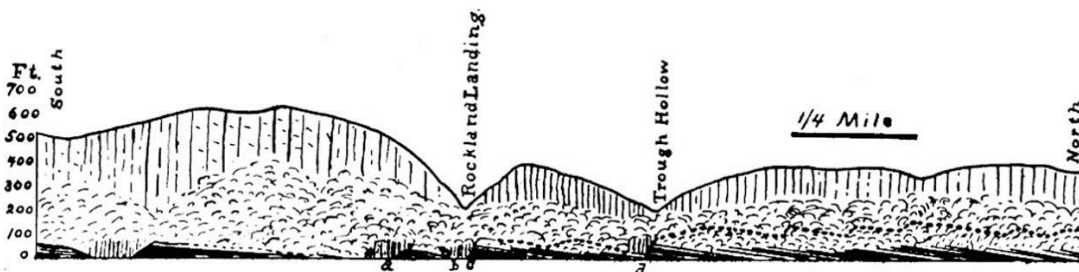


Figure 10: Section showing the step-wise climbing of stratigraphic level northward of the contact between Palisade Sill trap and Passaic Formation along the Hudson River near Rockland landing (from Kümmel, 1900).

Sanders (1974) provided a discussion of what he regarded as diametrically opposed models of how the stratigraphy of the northern terminus of the Newark basin relates to the former continuation of strata beyond the present limits of the basin and the Broad Terrane model (Fig. 11). He termed these the “shelving-basin school” (c.f., Savage, 1968) and the “transverse-anticline school” (Sanders, 1974). We are now in a position to evaluate these models with data, a goal Sanders explicitly desired.

According to Sanders' view of the "shelving-basin school," the strike of the sedimentary strata continues constant at about N 20° to 25°E striking directly into basement as the basement contact swings to the northwest at the basin's north end. In current parlance, we would call that onlap onto the hanging wall (Fig. 11A). According to Sanders, a corollary is that the Palisades sheet is a sill in southern Rockland County but becomes a dike cutting cut across Newark Basin strata as its outcrop expression swings to strike to the west (e.g., Lowe, 1959). Sanders viewed this concept as antithetical to the Broad Terrane model.

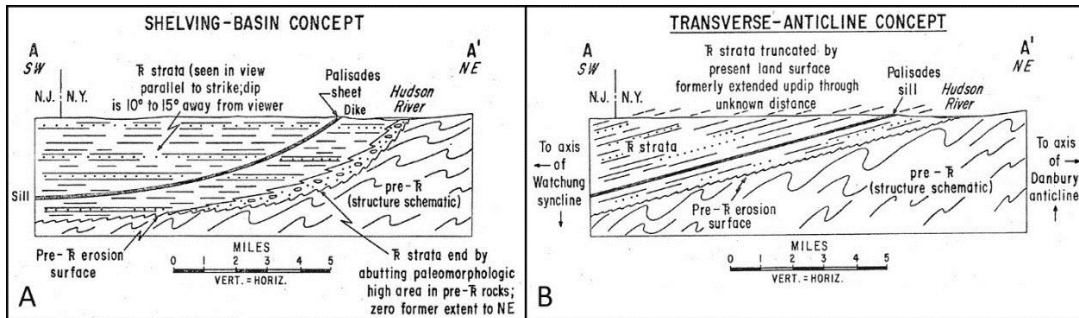


Figure 11: Sanders' (1974) two models for the northern terminus of the Newark Basin.

According to Sanders' view of the "transverse-anticline school", for which he claimed to be the only proponent at the time, the strike of the sedimentary strata follows the course of the Hudson River and the outcrop attitude of the Palisade sheet reflects that it remains a sill as it curves towards the west and is folded. Thus the sedimentary strata, "...strike directly and perpendicularly into the Ramapo Fault and dip away from [the] major anticlinal axis lying to the northeast (named the Danbury anticline by Sanders, 1960)" (Sanders, 1974), consistent with the strict version of the Broad Terrane model.

Data on strike and dips and the mapped distribution of lithofacies in Rockland County, available before 1974 (Kümmel, 1899; Perlmutter, 1959; Savage, 1968) and accumulated since (Ratcliffe, 1980; 1988; Yager and Ratcliffe, 2010; Heisig, 2011), clearly show strike curving from the south to the east very roughly paralleling the river until Stony Point where the strike is east-southeast, perpendicular to the Ramapo fault system (Fig. 9). This is consistent with Sanders' "transverse-anticline school". However, it is also clear from the map relations of lithofacies (Ratcliffe, 1988; Yager and Ratcliffe, 2010) and compilations of strikes and dips (Heisig, 2011) that the Palisades Sill cuts dramatically up section though much of the Passaic Formation. In addition (Kodama, 1983) has shown that the Palisade sheet continues to the west of its last contiguous outcrop at Mount Ivy and connects to the Ladentown basalts, which is also consistent with the chemistry of the two units and Ratcliffe (1988) (a detailed discussion of the relationship is given in Blackburn et al., 2013, supplemental materials; DOI: 10.1126/science.1234204). These observations are consistent with the "shelving-basin school".

Outcrops at the northern terminus of the Newark Basin are mostly consistent with Passaic Formation. There is no sign at all of Lockatong Formation or Stockton facies (even as seen in core TW4), including near Stony Point, where section below the Hudson is present on land. In fact, along Cedar Pond Brook at Stony Point, there are outcrops of limestone and carbonate-rich sequences unlike any elsewhere in the Newark Basin, except in the Jacksonwald Syncline. The facies typical of the Newark Basin in Bergen County or the bulk of the basin do not project into the northern terminus of the basin, which is inconsistent with a simple version of the "transverse-anticline school".

From these observations, it is clear that if hanging wall onlap is present, it must be more subtle than Sanders' "shelving-basin school", but the geometry of the strata relative to the Palisades Sill is

incompatible with simple post-sill folding and the “transverse-anticline school”. However, onlap does predict that there should be less thickness of strata present from the Ladentown basalts to the basement contact at Stony Point than between these basalts and the basement contact to the southeast in the Hudson River. Stratigraphic thicknesses estimates, based on published dip and strike data from two orthogonal transects of the northern Newark Basin (Fig. 9), show that the thickness along the transect from the Ladentown Basalt (= Orange Mountain Basalt) to basement parallel to the Ramapo Fault system (A-A’) is projected to be significantly less than that perpendicular to the fault system (B-B’), consistent with a modified version of the “shelving-basin school” (Table 1). A model of dip change with distance along the transects is required to calculate stratigraphic depth and three simple models were used here (Table 1) to honor the field data or their dip-domain equivalents. A simple average of the field data is inappropriate because of the uneven spacing of the field data. The average of the three stratigraphic thickness models along traverse B-B’ is 2986 m and for section A-A’ it is 2266 m, which is an average difference of 302 m (Table 1). We regard the linear regression model (Table 1) as both the simplest and most reasonable and that predicts a difference of 269 m between the transects, but all three models produce reasonably similar results. However, these models are some kind of prediction of what the depth to basement near the middle of the transects might be. Simply projecting the 10.5° average dip (dip-domain data of Helsig, 2012) from the Ladentown Basalt to the basement contacts of each traverse gives us a difference of 798 m, still fully consistent with a modified version of the “shelving-basin school”. Projecting from the Ladentown Basalt area is very sensitive to the specific dip value used and we really do not know what the appropriate value is within 10°. With all these caveats, however, all existing data are consistent with hanging wall onlap in exactly the directions there should be none according to the strict Broad Terrane model or Sanders’ “transverse-anticline school”.

Table 1
Stratigraphic Thickness Models³

Section	Linear interpolation Projection	Linear regression	4 degree polynomial	Average	Simple
Ladentown Basalt to Ossining Basement (Section B-B’) ¹					
Average interpolated dip at 1 m	10.5°	10.6°	10.6°	10.5°	10.5°
Stratigraphic thickness	2953 m	3010 m	3005 m	2986 m	2983 m
Ladentown Basalt to Stony Point Basement (Section A-A’) ²					
Average interpolated dip at 1 m	12.6°	13.2°	13.2°	13.0°	10.5°
Stratigraphic thickness	2604 m	2740 m	2718 m	2688 m	2185 m
Difference between sections	349 m	269 m	287 m	302 m	798 m

¹ Traverse B-B’ is 16369 m long with 10 dip values based on the dip-domain map of Heisig (2012) except for the two most two dips of 13.5° extrapolated from the most eastern dip-domain. Width of Palisade sheet has been removed from the traverse between the two dip points where it lies.

² Traverse is 11987 m long with 7 dip readings from Yager & Ratcliffe (2010) except for the two most southern dips of 10.5° which are from the dip-domain map of Heisig (2012). Width of Palisade sheet has been removed from the traverse between the two measured points where it lies.

³ Stratigraphic thickness models were derived from dip models produced in Excel™ that are regressions of the original dip data against their projected linear distances along the traverses. The regression equations were then applied to a 1 m delta distance series of the transects from which a running sum of stratigraphic thickness was derived (in Excel™). The linear interpolation models were produced by interpolation of dip values between measured points along the traverses using Analyseries™ that were then applied to a 1 m delta distance series of the transects from which a running sum of stratigraphic thickness was derived (in Excel™).

15, 2011 (Fig. 14). About 150 ft of core was recovered along with cuttings, 50 sidewall cores, and an extensive suite of wireline geophysical logs. This hole, along with the surface data, ground truths the seismic line. The description of the seismic lines and the Tandem Lot drill hole will be superficial here; there is far more data available than can be covered, but the highlights are summarized. The most obvious features on the profiles are the pair of strong reflectors crossing the basin, making a trough- or scoop-shape (Fig. 14). Prior to drilling these were interpreted as demarcating the Palisade sheet, which proved to be correct. The hole was spudded in middle Passaic Formation. Visible metamorphism and metamorphic minerals (e.g., epidote) were encountered in reddish Passaic Formation gradationally at about 4500 ft, which is more intense downwards until the strata are drab reddish greenish gray. The Palisades Sill was encountered at (4992.25 ft) and the underlying metamorphosed Lockatong Formation was entered at (6567 ft). The drill hole reached total depth (T.D. = 6881 ft), still in the Lockatong. The border fault is not visible in the seismic profiles, but it can be projected from the surface on the seismic traverse to depth to the west of the faint bedding reflectors to the west of the Palisade sheet. At depth, strong discordant reflectors demarcate basement structures, plausibly Paleozoic thrust sheets incorporating Paleozoic carbonates, as are visible in other seismic lines across the basin (Fig. 15).

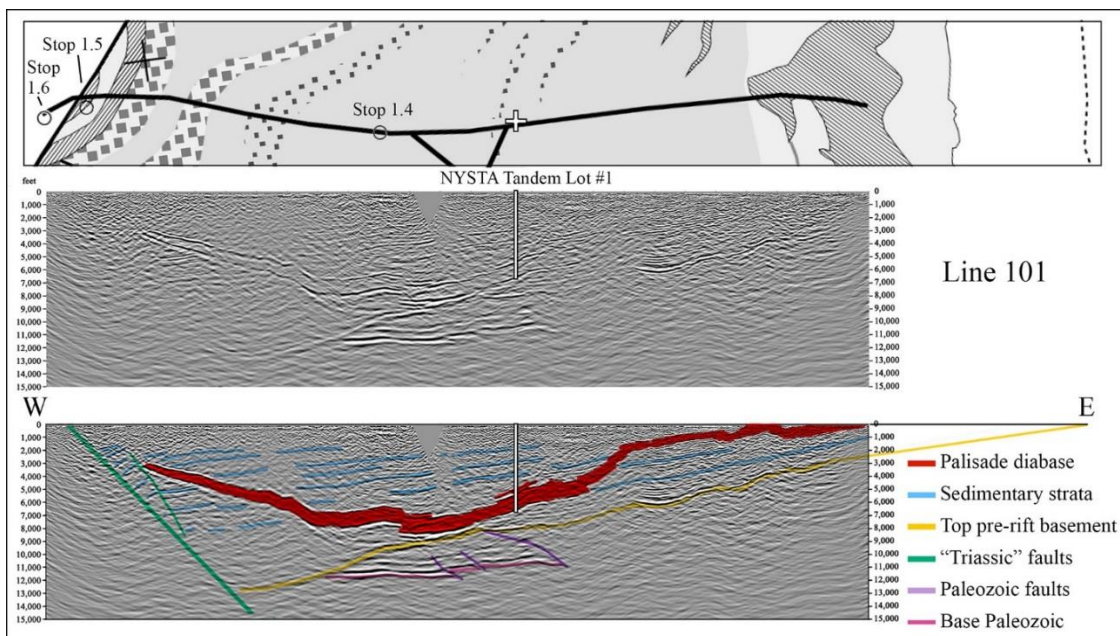


Figure 13: Sandia line 101, data and interpretation. Interpretation is consistent with drill hole and outcrop data.

Drilling cuttings and Schumberger’s FMI (Fullbore Formation Microimager) and other geophysical wireline logs plus 150 ft of core near the base of the Passaic Formation allow the section encountered in the Tandem Lot core to be understood in some detail. The upper part of the section of Passaic Formation, made up of largely red interbedded conglomerate, sandstone, and mudstone is similar to what is exposed along the NYST between exits 14A and 15, although those exposures are higher stratigraphically. Many of the red muddy units have carbonate nodules of pedogenic origin. There is a largely white to tan sandstone sequence between 2230 and 2470 ft, a largely red sandstone sequences at 2790-3228 ft, and a cyclical appearing sandstone and mudstone unit at 3540-4250 ft. Peaks in natural

gamma radiation between 3228 ft and the Palisade Sill suggest the presence of cyclical red and some gray mudstone intervals in the lower Passaic Formation.

We can make a direct comparison of the gamma ray data between the lower Passaic of the Tandem Lot hole and the cores of the Newark basin coring project (NBCP) (Fig. 16). When this is done quantitatively, it appears as though the Lower Passaic section at the Tandem Lot is about 68% of the thickness of roughly correlative strata in the NBCP cores. The 405 kyr (58 m) and the double ~100ky eccentricity peaks (around 17 m) can be seen in the gamma ray data of the Tandem Lot hole. The ~20 kyr cycle is also present, but it is very smeared out in the gamma ray data of the Tandem Lot hole. However, that is true for gamma ray data of the NBCP cores examined here as well, even though it is evident in the depth rank data that when scaled to the Tandem Lot, it is about 3.4 m (11 ft) with broad dispersion.

The Palisades Sill exhibits much lower gamma values than the surrounding sediments. While the overlying Lower Passaic Formation averages 114 api-gamma-units, the sill averages 44 api-gamma-units. This is typical of basaltic rocks in general and the sharp upper and lower boundaries are typical of intrusions. The relatively higher values between 5100 and 5625 ft and especially between 5100 and 5375 ft, with peaks reaching nearly 100 api-gamma-units probably reflects enrichment in Uranium, plausibly in zircons, within a gabbroid (coarse grained) zone called the “sandwich horizon” (Shirley, 1987; Block et al., 2015) seen in outcrops. Zircons approximately from this level in outcrop have produced a very high precision zircon U-Pb CA-ID-TIMS date of 201.520 ± 0.034 Ma (Blackburn et al., 2013). There is no indication in the geophysical logs of the famous olivine zone (Lewis, 1908; Walker, 1940), but they may not be diagnostic. The cuttings, still unstudied, should provide definitive data. In fact the olivine zone has been seen at Tallman Mountain State Park (Savage, 1968; Van Houten, 1969), so it should be present in TW4. We will see the lower part of the Palisades Sill at Stops 1.1-1.4, and an offshoot of it at Stop 1.5.

Theories on the origin of the various layers in the Palisade Sill are both contentious and important. Early ideas on the olivine zone regarded it as an early fractionation product following the Bowen Reaction Series (Lewis, 1908; Walker,

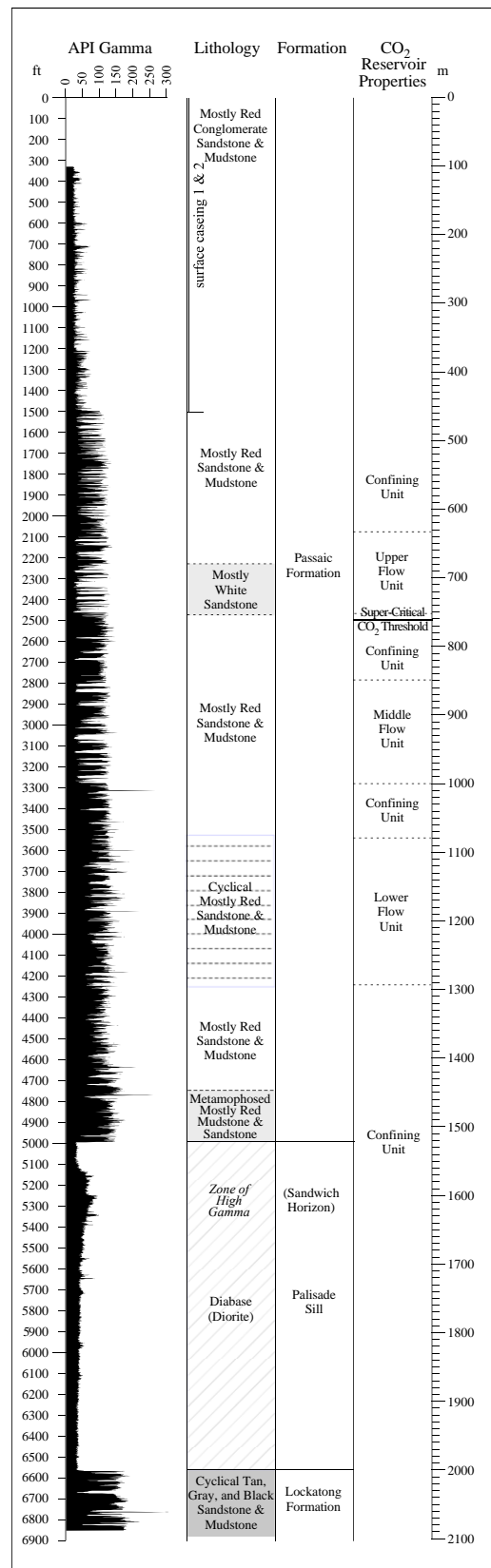


Figure 14: NYSTA Tandem Lot no. 1.

1940), but more recently it has been argued that it represents a separate injection (Husch, 1990) or the product of flow differentiation (Steiner et al., 1992; Goring and Naslund, 1995). As mentioned, the Palisade sheet appears to have been a feeder and directly connected to the Orange Mountain Basalt (Ladentown Basalt). Most recently, Puffer et al. (2009) and Block et al. (2015) have made a case for multiple injection events within the sill spanning the ~600kyr duration of the CAMP in the Newark Basin and may have been a feeder to all of the flows. To have received multiple injections within the sill over ~600 kyr requires the interior of the sill to have been incompletely solidified over the duration, only possible if there was a continuous flow of magma within the sill to some extent over that period. There is some evidence for this in the form of scoria at multiple levels in the Towaco Formation.

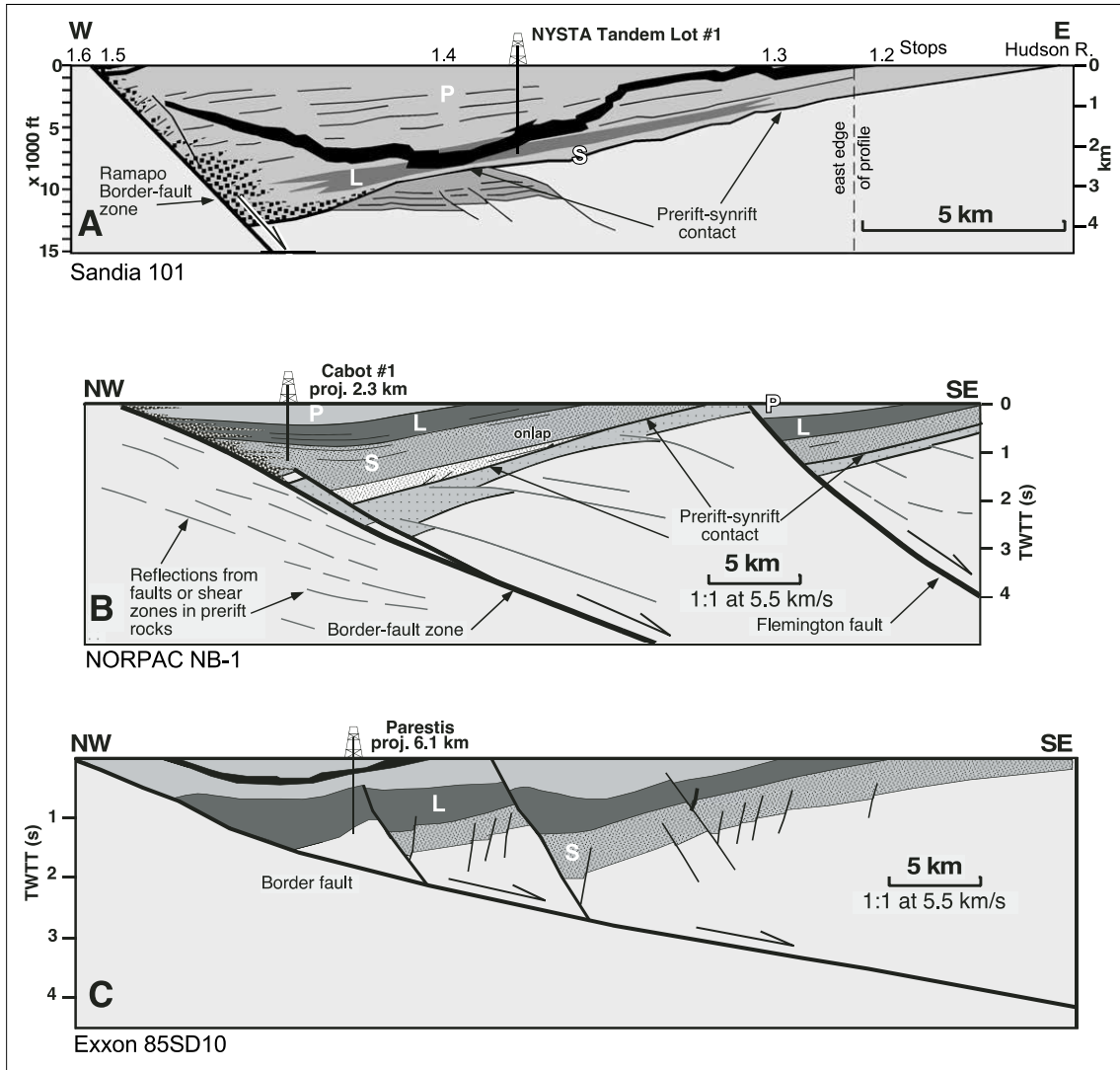


Figure 15: Interpretations of seismic profiles across the Newark Basin. A) Sandia Profile 101 with firdtrip stops shown; B) and C) are from Schlische & Withjack (2005). See Figs. 1 and 12B for locations.

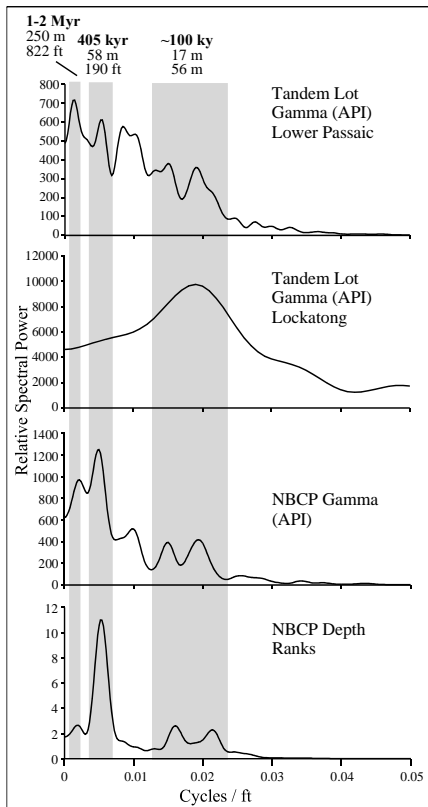


Figure 16: Blackman-Tukey spectral analysis of the NBCP cores (lower Passaic: Mb. C-Perkasie Mb.) and NYSTA Tandem Lot no. 1 natural gamma ray data. NBCP data has been scaled to the Tandem Lot section. Tandem Lot and NBCP data were filtered at the precessional cycle thickness based on previous analysis (Olsen & Kent, 1999) and then AM filtered (Hilbert Transform) and then subjected to Blackman-Tukey analysis.³

Strata below the Palisades Sill are drab in color to T.D. (based on cuttings and sidewall cores) and are metamorphosed. Based on the FMI images, they are cyclical alternations of laminated mudstone, massive mudstone, and arkosic sandstone and belong to the Lockatong Formation, similar to what we will see at Stops 1.2 and 1.4. Based on time series analysis of gamma logs from this interval, the periodicity of the cyclicity seen in this interval is similar to what is seen above the sill (Fig. 16) and the NBCP data scaled to the Tandem Lot. The short length of this data results in better resolution of the ~20kyr climatic precession cycle with two peaks around 3.1m (3.5 and 2.85), and there is a peak at 17m, as in the overlying Passaic Formation, consistent with the ~100kyr eccentricity cycles. The section is too short to resolve the 405 kyr cycle.³

With this “ground truth” in hand, it is relatively easy to interpret the Sandia seismic lines. While subtle, divergence of bedding is apparent in the seismic lines and agrees with the down-hole dipmeter logs and published surface measurements. The Palisade sheet clearly has a scoop shape in the seismic line, which

³ Blackman-Tukey spectral analysis of the NBCP cores (lower Passaic: Mb. C-Perkasie Mb.) depth rank data shows a prominent spectral peak at 84.7m (277.8ft) that corresponds to the thickness of the 405 kyr orbital cycle. To compare with the much noisier gamma ray data from the same sequence the latter has to have the high values truncated (clipped) at 175api-gamma-units, because of the huge dynamic range of the high gamma peaks and their erratic positions. Blackman-Tukey spectral analysis of gamma over this interval also has a prominent peak at 84.7m (277.8ft) reflecting the lithological variations that covary with depth ranks and color corresponding to the same orbital cycle as well as an additional peak near infinity reflecting unremoved trend in the length of the data. The gamma ray data from the Tandem Lot, clipped at 115 api-gamma-units, has a prominent peak at 57.5m (188.7 ft) as well as another much longer period also reflecting the length of the data. Assuming the 84.7m and the 57.5m peaks represent the same lithological cyclicity, we can scale the NBCP data to the Tandem Lot data (NBCP thickness times 0.679245283) to make direct comparisons between spectra. An additional manipulation is needed however, because the gamma ray data in both the NBCP cores and the Tandem Lot are so noisy compared to the depth rank data in the NBCP cores, consisting of filtering the gamma ray data at 3.4m (11.1ft), and then using a Hilbert transform to recover the envelope of the data (AM filter in Analyseries™). When this is done to both the Tandem Lot and NBCP cores, the resulting spectra are remarkably similar showing the expected peaks around 58 m as well as two peaks centered around 17m. If the 58m peak is 405kyr, then the two peaks should correspond roughly to 136kyr and 104ky. In the depth rank an color data of the NBCP core, these twin cycles are identified as the double peak of the ~100kyr orbital cycles.

possible environmental effects of the event. In order to address the hypothesis of Broad Terrane lavas, and perhaps get a handle on the distribution of lava shortly after the CAMP eruptions, an experiment was pursued in the late 1990s (Fairfield, 1998; Fairfield et al., 1999; Olsen et al., 1999) that looked for ghosts of the former extent of the lava. The experiment was predicated on the idea that if the lavas covered extensive areas outside the presently preserved basins, especially in the watersheds in higher elevations between basins, the basalts themselves should have contributed sediment to the basins beginning shortly after eruptions. A geochemical tracer of basaltic eruptions is needed because basalt debris would be very quickly weathered to clay in the tropical depositional climate of these basins. Because Nd, Sm/Nd, and Nd isotope ratios, particularly expressed as ϵ_{Nd} (epsilon Nd), differ so much between basement ($\sim -15 \pm 5 \epsilon_{Nd}$) and CAMP rocks ($\sim 0 \pm 6 \epsilon_{Nd}$) (Pegram, 1990; Dorais et al., 2005) they would make good tracers. Therefore, sediments derived from pre-rift basement should differ from those derived from the degradation of the CAMP flows. However, a preliminary assessment of a suite of Hartford basin mudstones (Fig. 18), from all of the exposed formations, shows no overall trend in the epsilon Nd or Sm/Nd isotopic ratios from pre-flow through syn-flow and post-flow units indicative of basalt input ($y = -0.8161x - 8.9683$; $r^2 = 0.24$). Older pre-CAMP sediments have values of $-10.0 \pm 1.1 \epsilon_{Nd}$ and younger syn- and post-CAMP sediments having values of $-11 \pm 1.9 \epsilon_{Nd}$ (2σ error), indicating very little if any input from the basalts. We had included one sample of highly altered (paleo-weathered) Hampden Basalt to assure that a material we *knew* had a very high CAMP content would show it (which it did: Fig. 18). In retrospect, the Hartford Basin may have been a poor choice for the experiment, because it received most of its syn- and early post-CAMP input from the western, hanging wall, side of the basin, where there are no CAMP dikes. However, the data would seem to indicate that any flows emplaced between the Pomperaug Basin and the Hartford Basin could not have contributed significantly to the Hartford Basin strata, either because they were not there at all, or because they were quickly buried by a veneer of sediment and therefore not available for erosion into the basin.

Motivated by the negative results of the first experiment and by the question of whether completely weathered, essentially invisible, CAMP ashes could be identified at the Newarkian ETE (see below), we looked at a single small (~ 5 mm), definitive airfall ash in the middle Towaco and middle East Berlin Formation called the Pompton Ash (Olsen et al., 2012; 2016). The andesitic to basaltic ash occurs in microlaminated, fish-bearing units that are in the same 20kyr cycle in the two basins. The ash and about 1.5 cm of surrounding microlaminated strata were analyzed for ϵ_{Nd} to see if the ash showed a distinct igneous signature (Fig. 18). From the Hartford basin, the strata surrounding the ash are within the

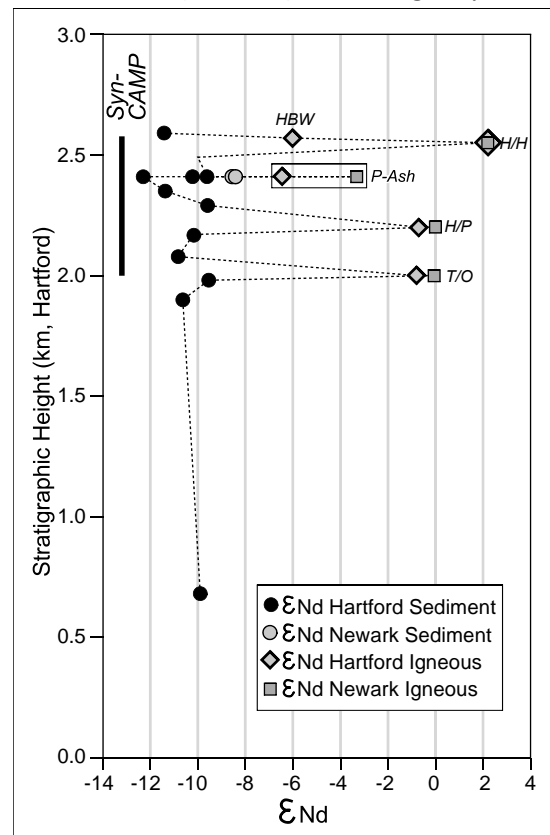


Figure 18: Epsilon Nd results for the Hartford and Newark Basin. Data for Hartford sediments are from Fairfield (1998). Data for Basalts from Pegram (1990). Data for ashes courtesy of S. Jaret (pers comm., 2016).

range of the rest of the East Berlin Formation, and the ash is nearly 4 ϵ_{Nd} units closer to the CAMP basalts, although not as igneous-looking as anticipated. Clearly the ash as preserved is an admixture of igneous and non-igneous material and is very close to the altered Hampden Basalt previously mentioned. In contrast, the sediments around the Newark Basin sample of the ash is about 1 ϵ_{Nd} unit “more igneous looking” than exactly correlative Hartford sedimentary strata and is shifted nearly 4 ϵ_{Nd} units towards CAMP basalts compared to its Hartford counterpart (Fig. 18). This may indicate a larger CAMP contribution to Newark Basin syn-CAMP sedimentary strata, consistent with it receiving much of sediment from the hanging wall side of the basin, which is in the direction of known CAMP on the conjugate margins of eastern North America and Morocco. Further work on Newark basin sediments should reveal how much contribution of CAMP there is from outside the basin to the preserved sediments.

A related issue is the size of the lakes that occupied the basins. It has been known for more than 30 years (e.g., Olsen et al, 1996b) that there is a very close correspondence between the cyclostratigraphy of the very latest Triassic age and Early Jurassic age lacustrine sequences in the Newark and Hartford basins and it is reasonable to ask if the lakes that made those sequence were connected by open water at least during highstands. The cyclicity was caused by variations in the hydrologic balance paced by variations in the Earth’s orbit and axial orientation (Olsen, 1986; Olsen & Kent, 1996; 1999; Blackburn et al., 2013), and we know that the lacustrine cyclicity was synchronous in the two basins at the finest observable levels. The Pompton Ash is in the same position with cycles correlated long before the ash was known (e.g., Olsen, 1984 vs. Olsen, 2010), and additionally, there is a very close match of laminae around the ash between the two basins. Because both the cyclicity and the laminae were paced by climate change (seasonal to millennial), it really is not surprising that the cycles and laminae match, but the match at multiple scales is close, much closer than anything documented in Pleistocene Great Basin lakes, for example. This raises the question of whether the lakes are one during high stands. Olsen (et al, 1982) originally thought that the lakes were not connected (or even correlated), because the fish assemblages seemed so different. Further collection proved that the differences were apparent, not real, and due to sampling biases (Olsen, 1983). The cyclostratigraphy of what is known of syn-CAMP strata of the Culpeper Basin of Virginia and Maryland and the Deerfield Basin of Massachusetts, suggests a match with the Newark and Hartford basins and opens the possibility that during high stands, a single lake may have been present from northern Massachusetts to south central Virginia, a distance of nearly 700 km and significantly larger than Lake Tanganyika, and longer than Lake Superior, making it one of the largest lakes known. But testing this hypothesis is not easy and remains for future work. Perhaps comparison of the strontium isotopic composition of the carbonates in exactly contemporaneous laminae around the Pompton Ash could provide a test. They should differ if the lakes had different watersheds with different Sr isotopic compositions, and they should have the same Sr isotopic values if the lakes were connected by open water.

The Broad Terrane model thus survives in a modified form. The simplest interpretation of the existing data is that the presently preserved remnants of basins formed mostly as half graben growth structures that linked up through time with thinner and plausibly patchy cover of sediments and lavas covering intervening basement rocks. Water may have connected the basins by rivers, or open water with huge lakes, during lake highstands. However, there is no evidence supporting the strict Broad Terrane model. We can imagine a landscape comprised of a vast rifting zone with many depositional basins, many connected by flat plains, lakes or rivers, looking much like the Basin and Range does today. However, despite considerable progress, we are as yet unable to constrain the area occupied by CAMP lavas, within two orders of magnitude (0.3 – 10 Mkm²: McHone, 2003) and thus constrain the volume and

direct volcanic output of volatiles and their environmental effects.

Record and Causes of the End-Triassic Mass Extinction

Formerly usually referred to as the Triassic-Jurassic boundary event, the end-Triassic mass extinction (ETE) is one of the “big-five” extinction events of the Phanerozoic, as first pointed out by literature reviews of the ranges of marine and continental animals (Raup and Sepkoski, 1982). The detailed review and update by Benton (1995) (Fig. 19) shows that, depending on the metric used, the ETE can be as large or larger in magnitude as the K-Pg (Cretaceous-Paleogene, or K-T) or even Permian-Triassic for continental organisms. The ETE was formerly often considered synonymous with the Triassic-Jurassic boundary. However, the recent establishment and ratification by the IUGS in 2010 (Morton, 2010) 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* (Morton, 2012; Hillebrandt et al., 2013) defines the ETE as a Late Triassic event. The Triassic-Jurassic boundary is thus part of the tenuous recovery identified in the marine realm occurring about 100-200kyr after the ETE (Schoene et al., 2010; Sha et al., 2015). Furthermore, because ammonites were cephalopods completely restricted to marine environments, recognizing the Triassic-Jurassic boundary in continental deposits is highly inferential, based on one or two provincial sporomorphs or extrapolation of astrochronologies from zircon U-Pb dates.

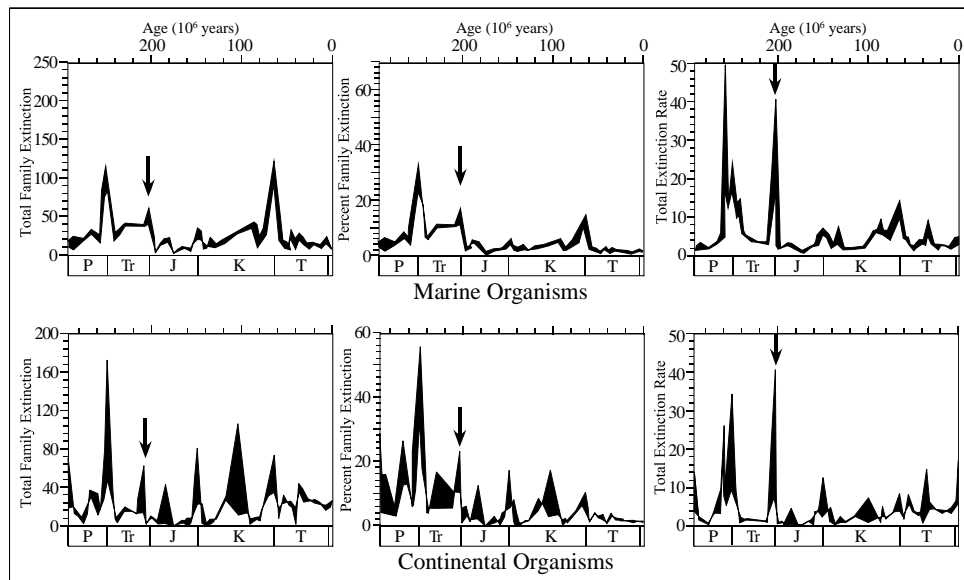


Figure 19: Various metrics of extinction for marine and just continental organisms (from Benton, 1995).

For those not intimately familiar with the literature, it is worth noting that this final (for now) definition of the Triassic-Jurassic boundary has produced considerable confusion in the literature because the meaning of the term Triassic-Jurassic boundary or Rhaetian-Hettangian boundary has changed since 2010 when it was equated with the ETE. Several relatively recent papers conflate the pre-2010 meaning of the Triassic-Jurassic boundary with its present GSSP meaning, arguing that pre-2010 correlations of the Newarkian continental to marine environments were incorrect (e.g., Cirilli et al., 2009; Lucas and Tanner, 2015), when in fact the correlations of the strata have not changed, only the definitions of the ages have.

In eastern North America, the most dramatic biotic change in the entire sedimentary record of the rifts

occurs very close to, but not at, the base of the oldest CAMP basalts (Fig. 20). This change involves the abrupt last appearances of all the Triassic-type footprint taxa (Fig. 21A) and nearly all Triassic-aspect pollen and spores, as documented by Olsen et al. (2002). Included are the previously abundant footprints *Brachychirotherium* that is representative of the non-crocodile, “crocodile-line” top terrestrial predator and possibly herbivore pseudosuchian, and *Apatopus*, a representative of the phytosaurs, the semi-aquatic, non-archosaur, crocodile-mimic pseudosuchians (see Stops 1.6 and 2.2). Amongst sporomorphs, the most prominent extinction includes the vesiculate pollen forms, including previously very abundant *Patinasporites*, produced by an extinct, largely tropical conifer group. Many other pollen and spore taxa disappear as well. Above the level of the extinctions, the only representative of the pseudosuchians is *Batrachopus* (Fig. 21), a track that could have been made by small protosuchians or sphenosuchians, the only two crocodile-line lineages that survive the ETE, the former implicated in the origin of the modern crocodilians (see Stop 1.7). Below the extinction level, the only dinosaur footprints present are small brontozoids (*Grallator* and *Anchisauripus*), representing small carnivorous theropod dinosaurs, and *Evazoum* (e.g., Stop 1.6), made by a basal dinosaur (which is rare and disappears well below the ETE) (see Stop 1.6). Above the extinction level, the much larger theropod dinosaur brontozoid footprint *Eubrontes giganteus* abruptly appears and is abundant along with the smaller brontozoids. The dinosaurian herbivore tracks *Anomoepus* (made by small ornithischians) and *Otozoum* (made by basal sauropodomorphs = “prosauropods”) appear at slightly higher levels (Olsen et al., 2002) (Figs. 21).

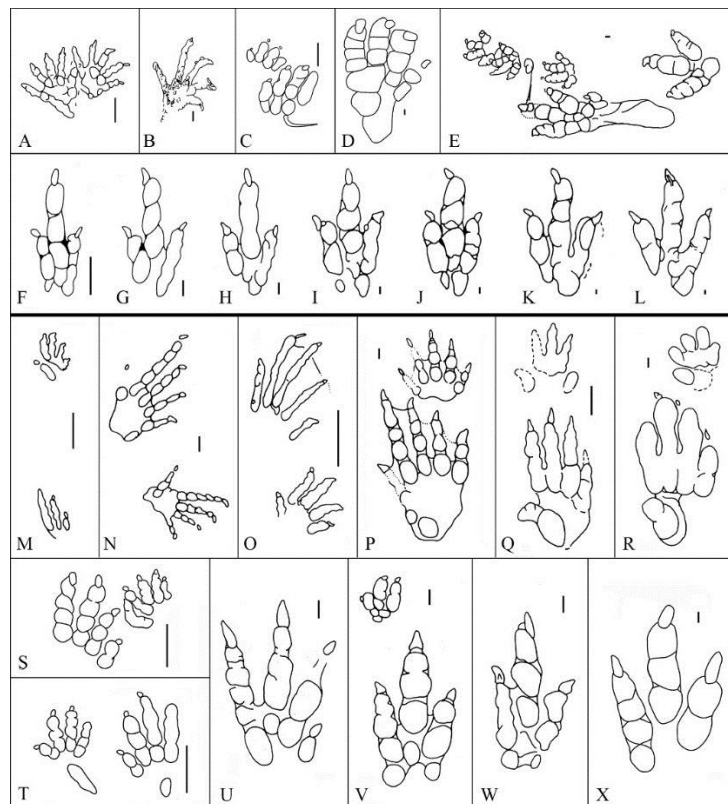


Figure 21: Post ETE taxa are A-L and pre-ETE fprms are M-X. Taxa are A, *Ameghinichnus* n.sp.; B, *Rhynchosauroides* n.sp.; C, *Batrachopus deweyii*; D, *Otozoum moodii*; E, *Anomoepus scambus*; F-G, different sizes of brontozoid with *Grallator* spp.. on the left, *Anchisauripus* in the middle and *Eubrontes giganteus* on the right; M, *Rhynchosauroides brunswickii*; N, *Rhynchosauroides hyperbates*; O, *Gwyneddichnium* sp.; P, *Apatopus lineatus*; , Q, *Chirotherium lulli*; R, *Brachichirotherium parvum*; S, “new taxon B”; T, “*Batrachopus*” *gracilis*; U, *Evazoum* n.sp.; V, *Atreipus milforensis*; W, small brontozoid, *Grallator* sp.; X medium sized brontozoid *Anchisauripus* c.f. *tuberatus*. From Olsen and Rainforth (2003) and Rainforth (2003).

Lucas and Tanner (2015, and papers by the same authors cited within) argue that the ETE was not a major extinction event at all, pointing out the small number of tetrapod forms that go extinct at the Newarkian ETE. It is true that the skeletal material is relatively rare in Eastern North America, and that while tracks are very common, only 4 footprint taxa have their last appearances at the ETE in eastern North America. However, the footprint taxa may be representatives of several or even many biological species and as a proportion of what is present, the extinction level sees a 44% (4/9) diversity drop, much higher than documented anywhere else in the +30 Myr record of early Mesozoic faunal change in eastern North America. The change in the footprint diversity is also completely consistent with the global record of taxa represented by bones. One can argue that the footprint record in the Triassic-Jurassic is analogous to what one might see along the edges of a Serengeti (Tanzania and Kenya) lakeshore. Cohen et al. (1993) document that the abundant Serengeti animals leave the abundant footprints but not in proper proportion to their abundances. While it is possible to recognize the species that left these tracks, it is because we know what species are present and most have a more or less consistent adult body size – constraints not present for our Triassic-Jurassic assemblages. One footprint genus, for example, might be analogous to footprints of ALL feline and canine species or ALL artiodactyls (even-toed ungulates - antelope, wildebeest, giraffe, pigs, etc.) present in the Serengeti (Fig. 22). Thus, the footprint record sees biological diversity as, “through a glass darkly”, surely massively underrepresenting the change in biological species diversity.

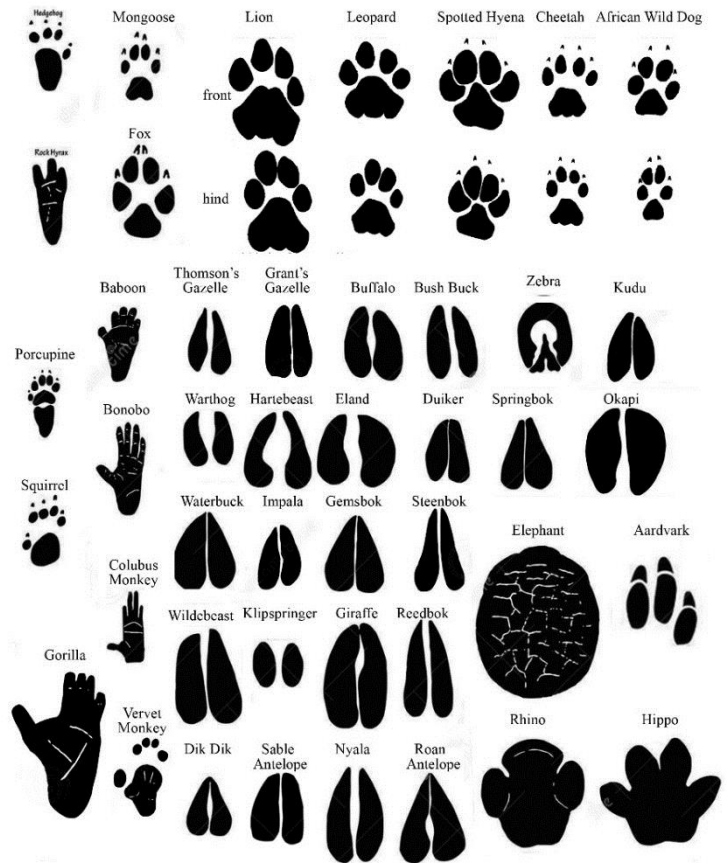


Figure 22: East African animals likely to leave footprints (<https://www.dreamstime.com/royalty-free-stock-photography-african-animal-tracks-image7728337> and https://www.google.com/search?q=leopard+footprint&hl=en&biw=1436&bih=776&source=lnms&tbm=isch&sa=X&ved=0ahUKewj7jvup0IPPAhVKdz4KHSRzBo0Q_AUICSgC#imgrc=5bfA6tSd)

With plants, as seen via pollen and spores, the extinction level in eastern North America is dramatic and overlaps precisely with that of tetrapods, although with tighter stratigraphic and temporal precision.

A high diversity assemblage dominated by the vesticulate forms, especially *Patinasporites*, as well as bisaccate pollen, is replaced by an assemblage extremely strongly dominated (+90%) by the conifer pollen form *Classopollis meyeriana* with a net loss of 60% of the diversity (Fowell et al., 1994; Olsen et al., 2002). There is a laterally continuous “fern spike” exactly at the extinction level in the southwestern Newark Basin similar to that at the K-Pg boundary (Tschudy et al., 1984) in which fern spores constitute

50 to nearly 100% of the assemblage (Fowell, 1994; Olsen et al., 2002). While ferns might be expected to be dominant at any time locally, no other such level is known through the Newark Basin succession.

There is at least one major extinction event seen in the track record prior to ETE. The track taxon *Atreipus* is the most common three-toed form prior in much of the Triassic record (see Stops 1.6 and 2.2). *Atreipus* was originally interpreted as either a very “precocious” hadrosaur-like ornithischian dinosaur or a primitive dinosaurian belonging neither to ornithischians nor saurischians (Olsen & Baird, 1986). In modern context, the latter would be a non-dinosaurian dinosauroform. The simplest interpretation of all the available evidence is that it was made by an herbivorous silesaurid, dinosauriform, dinosauroform (c.f., *Silesaurus*; Dizk, 2003). The last known occurrence of this form is in member JJ in the Newark Basin (Olsen et al., 2002), which based on the newest version of the Newark APTS (Kent et al., 2016), is about 206 Ma. At the level of sampling of the Newark Basin section, this is indistinguishable from the 205.5-205.7 Ma age of the Norian-Rhaetian boundary (Wotzlow et al, 2015; Maron et al., 2015; Kent et al., 2016). Thus far, this is the only evidence in eastern North America of a tetrapod event correlating with that boundary.

The palynological transition seen between the Lower-Passaic-Heidlersburg and the Upper-Balls-Bluff-Upper-Passaic assemblages of Cornet (1977) was correlated to the marine Norian-Rhaetian boundary (Cornet, 1977; Fowell et al., 1994) and accepted as such by Kent and Olsen (1996) and Olsen et al. (1996) in the Newark APTS. This boundary was picked by Cornet (in Olsen et al., 1996a) at the first occurrence of *Classopollis torosus* and the last appearance of a few sporomorph taxa. This pick was criticized by Lucas & Tanner (2007) because *C. torosus* occurs in the Norian of Europe. However, there is a larger palynological extinction event identified by Fowell (1994), corresponding to about 206 Ma on the most recent Newark APTS. Based on the sampling density of palynological levels, it is indistinguishable from the the Norian-Rhaetian boundary age (Wotzlaw et al, 2015; Maron et al., 2015). Because *Atreipus* was such an important component of Norian assemblages, and there is a significant correlative palynological event at current sampling density, the idea that Rhaetian-Norian boundary represents an important extinction or turnover event (Wotzlaw et al, 2015; Maron et al., 2015) is supported, although it is much smaller in magnitude than the ETE.

Identifying the cause of the ETE has been contentious. An iridium (Ir) anomaly occurs at the same level and sections as the fern spike (Olsen et al., 2002a; b), and the similarity to the K-Pg (K-T) extinctions, Ir anomaly, and fern spike was consistent with the idea that the ETE was the result of a major asteroid or comet impact (Olsen, 1999). A similar, perhaps correlative section with a fern spike in the Hartford basin (Olsen et al., 2003) has recently been shown also to have a significant Ir anomaly (Tanner & Kyte, 2016). Uncertainties in the age of the giant Manicouagan impact had earlier allowed it to be a candidate for the event (Olsen et al., 1987). However, the hypothesis that Manicouagan could be the cause was effectively tested and falsified by Hodych and Dunning (1995), who showed that the age of Manicouagan was far too old (214 +1.25/-1.03 Ma; corroborated by the ~215.5Ma date of Ramezani, 2005) to be implicated in the cause of the ETE, or the Norian-Rhaetian boundary, for that matter, although Manicouagan may have been involved in a lesser turnover event earlier in the middle Norian (Whiteside & Ward, 2011; Parker & Martz, 2011; Olsen et al., 2011; Onoue et al., 2016).

While shocked quartz has been reported from two marine sections at or near the ETE (Austria- Badjukov et al., 1987; Italy-Bice et al. 1992), these reports are uncorroborated by additional, more diagnostic, analyses (Olsen et al., 2002b). Furthermore, multiple Ir anomalies have been found at ETE sections since 2002, both above and below the extinction level (Tanner & Kyte, 2005; 2016; Tanner et al., 2008; Olsen, 2010), and a minor Ir anomaly has been identified in the GSSP section of the base Hettangian at

Kuhjoch, Austria above the base of the ETE (Tanner et al., 2016). Proposing multiple impacts based on Ir-alone seems unparsimonious.

Although there is thus no compelling evidence for the role of an impact at the ETE, there is nonetheless a small impact structure Rochecouart (pre-erosion diameter ~40-50km: Sapers et al., 2014), where if its date is accurate (201 ± 2 : Schmieder et al., 2010), it should have left some kind of record near the ETE level even if it was too small to have caused extinctions. Because impact rocks at Rochecouart are of reverse magnetic polarity (Eitel et al., 2014), it would have most likely hit during E23r, apparently before the extinction. Schmieder et al. (2009) has proposed that the Rochecouart impact produced a mega-seismic event that produced a well-known deformed unit present over a large part of the UK described as a mega-seismite by Simms (2007), present just below the ETE. The lack of impact debris (thus far) at these sections and the large uncertainties in the impact date make the association between Rochecouart and levels near the ETE only possible, not definitive. There is also no evidence of reverse polarity in strata in contact with upper surface of the seismite (Hounslow et al., 2004).

In contrast, Marzoli et al. (1999; 2004) argued that the close association of the ETE with the oldest lava flows of the CAMP (Marzoli et al. named the CAMP) indicated that the environmental effects of the emplacement of that large igneous province was the cause (Olsen et al., 1999). The two principle proposed triggers for the extinction are: 1) CO₂-induced global warming and ocean acidification, and/or methane release; and 2) H₂SO₄ (sulphuric acid) aerosols causing global dimming and cooling. These are not mutually exclusive, although their effects should operate on vastly different time scales and differ depending on the temporal concentration of the magmatism (mass CO₂ or SO₂ per unit time). Similar effects have been proposed as causes or contributors to the K-Pg (McLean, 1985; Caldeira & Rampino, 1990; Self et al., 2006) and the end-Permian (see reviews of Saunders & Reichow, 2009; van de Schootbrugge, 2016) mass extinctions, among others.

The CAMP could theoretically inject CO₂ through four main, not mutually exclusive, mechanisms: 1) direct outgassing (McHone, 2003); 2) thermogenic CO₂ and thermogenic methane (oxidizing to CO₂) (cf., Svensen et al., 2004; 2007); 3) triggering outgassing from methane clathrates in the oceans (Hesselbo et al., 2002); 4) destruction of the biological pump and turnover of the global oceans as a consequence of the mass-extinction (cf., McLean, 1985; Knoll et al., 1996). Each of these mechanisms is complex with complex predicted sequelae and a detailed exegesis of them is beyond this guidebook. However, there IS direct evidence of pulsed doublings to triplings of CO₂ in the Newark and Hartford basin section as seen in the soil carbonate proxy directly associated with CAMP eruptive events (Schaller et al., 2011, 2012; 2015). Carbonate proxy records of massive increases in CO₂ are consistent with plant leaf stomatal data from Greenland and Sweden (McEwain et al., 1999; 2009) and Germany (Bonis et al., 2010).

The killing mechanism for CO₂ would be expected to be different on land vs. the ocean. The ocean would see three possible effects: 1) ocean acidification; 2) warming with reduction of dissolved O₂; and 3) nutrient increase by increase continental weathering. Ocean acidification would be expected to occur only if the CO₂ doubling occurred in less than 10kyr or so, in which case surface seawater aragonite saturation would drop and pH would drop as the oceans absorbed atmospheric CO₂, resulting in a bio-calcification crisis (Hönisch et al., 2012). While the entire CAMP episode took nearly 1 Ma, it was pulsed, and there is cyclostratigraphic and paleomagnetic evidence that at least some of the CO₂ doublings occurred within thousands of years with individual giant eruptions occurring in less than 100 years, so this is possible (Kent et al., 2009). This hypothesis is consistent with the preferential extinction of aragonitic and/or high-Mg calcitic marine organisms, such as virtually all ammonites, scleractinian corals, and sphinctozoid sponges (Hautmann et al., 2008; Martindale et al., 2012). If the CO₂ doublings take

longer, >50 ka, seawater aragonite saturation actually goes up as pH goes down and then both eventually go up and carbonate precipitation will increase. The decrease in greenhouse effects are not as sensitive to injection rate. CO₂ concentration and the temperature increase exponentially decrease for about 1Ma until background levels are reached with normal continental weathering. However, they would drop below background levels if weathering was enhanced by large areas of highly reactive exposed basalt flows in the tropics. Most of the CO₂ however would be consumed in about 300 ka or less depending on the acceleration of weathering.

The IPCC (AR5: Collins et al., 2013) consensus predicts about a 3°C (1.5°-4.5°) increase in global average temperature with each doubling of CO₂. But some models predict more than 6°C. Over geological time and with high initial *p*CO₂ (+1000 ppm), the temperature sensitivity is poorly constrained. The increase in temperature would be reached quickly and decrease slowly over many 10s to several 100s of ka. Multiple injections over tens of thousands of years would be additive in their effects on concentration and temperature effect, but not on acidification. Under higher ocean temperatures, O₂ concentration would decrease on the long term, which would also negatively impact pre-CAMP communities and add to any ocean acidification effects. On land, one would expect an amplification of the hydrological cycle, with drier areas becoming drier and wet areas becoming wetter (as discussed above), but some continental areas would become lethally hot with those zones greatly expanded compared to today. Increased heat (and associated fire) has been discussed both as a damper to dinosaur diversity in the tropics (Whiteside et al., 2015) and as major driver of plant extinctions during the ETE (e.g., McElwain et al., 1999). However, Late Triassic tropical areas were dominated by non-dinosaurian archosaurs and other reptiles, forms that would seem to be particularly resistant to additional heat, yet they were preferentially effected, while the pre-ETE dinosaurs at high latitudes, in presumably more temperate Triassic climes, were hardly effected.

On scales of millions of years, flood basalt eruptions such as the CAMP would be expected to produce huge continental areas that were more reactive to chemical weathering than normal continental crust, leading to long term cooling. The CAMP erupted primarily in the tropics where weathering (carbonation of basalt) would have been particularly intense, leading to a drawdown of CO₂ below pre-CAMP levels. Evidence for this is also seen in the Newark Supergroup carbonate proxy of CO₂ (Schaller et al., 2015), and a similar drawdown for the Cenozoic has been argued for the Deccan (Kent & Muttoni, 2008; 2013).

Sulfuric acid (H₂SO₄) aerosols, usually termed sulfate aerosols, the other main proposed CAMP effects on climate. In contrast to CO₂ (or methane), which are greenhouse gasses that cause warming, sulfate aerosols cause cooling by increasing planetary albedo. Most of the negative deviations from the CO₂ greenhouse warming over the past 2500 years have been due to volcanic sulfate aerosols (Sigl et al., 2015). Some spectacular examples include the historical eruptions of Laki in Iceland in 1783-1783, implicated in the death a huge number of people, 10,000 to millions globally by crop failures and social unrest (Thordaldson & Self, 2003); Tambora, Indonesia, in 1815 resulting in what was called in “The Year Without a Summer” and “Eighteen Hundred and Froze to Death” (Stothers, 1984); the eruption and explosion in 1883 of Krakatoa resulting in significant cooling and floods (Bradley, 1988) as well as the Edvard Munch painting “The Scream” ten years later (Olson et al., 2007); and the 1991 eruption of Pinatubo in the Philippines that caused a 3-year, 0.5°C drop in global temperatures (Self et al., 1996), as well as major pandemics, famines, and socioeconomic disruptions in Eurasia and Mesoamerica in the 6th century. All of these eruptions are, however, at least two-orders-of-magnitude smaller than CAMP eruptions, and it seems parsimonious to hypothesize that the temperature decreases were at least an order of magnitude greater (McHone, 2003). Unlike the three major pulses of volcanism seen in outcrop and core and the consequent three or four CO₂ pulses, there were likely hundreds of brief (3 to 10 year)

volcanic winters, perhaps with freezing in the tropics, for which the tropical pre-ETE continental biota would have had no evolutionary adaptive experience. While the marine extinctions are consistent with the effects of increased CO₂, the continental biota pattern at the ETE seems more consistent with intense volcanic winters that the insulated (protofeathered) avemetatarsalians (Pterosauria + Dinosauromorpha) could survive (Olsen et al., 2013; Olsen, 2015).

Sedimentary and Igneous Facies and Carbon Sequestration

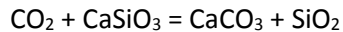
The Newark Rift Basin is the nearest best hope for carbon storage for the major CO₂ emission sources in New York, New Jersey and Pennsylvania. The TriCarb Consortium for Carbon Sequestration (Collins et al., 2014) was an organization of geologists and engineers from Sandia Technologies (Houston, TX) and LLC and Conrad Geoscience, Corp. (Poughkeepsie, NY) and academic advisors (LDEO, represented by DG, DVK, PEO) and the New York State Museum (BS) that undertook a basin characterization study that combined geophysical seismic profiling with data derived from a stratigraphic drilling, coring and logging program. American Recovery & Reinvestment Act (ARRA) funding was provided by a 2009 DOE Project Award co-funded with New York State Energy Research & Development Authority (NYSERDA) and DOE/NETL, with leveraged technical services partner, Schlumberger Carbon Services. This project was linked with a longer term set of projects at LDEO exploring CO₂ sequestering, including drilling, logging, and analysis of the TW4 core and hole, already described, funded by TriCarb and EPA STAR grant 834503 (Zakharova et al., 2016). Stops 1.1 and 2.2 will examine data derived from these projects and earlier drilling and logging projects on the LDEO campus (Yang et al., 2014).

The TriCarb project, core, and hole TW4 focused on characterizing conventional sandstone reservoirs for sequestering anthropogenic CO₂ in the northern Newark Basin. The basic concept is that CO₂ recovered from point sources such as power plants is injected into a porous sandstone reservoir sandwiched between much less permeable confining layers, such as mudstone or igneous rock. The CO₂ is generally injected in liquid form with low temperature and modest pressure, but at depth and increasing pressure, it transforms into a supercritical fluid and warms to the ambient temperature. A supercritical fluid has properties of both a liquid and a gas; it can diffuse through solids like a gas, but dissolves things or into things like a liquid. In non-saline waters at depth, supercritical CO₂ is denser than the water and thus tends not to ascend once it is pumped down. In the Newark Basin CO₂ would be supercritical deeper than 800m (2625ft) and would require thick sandstones with sufficient porosity for large volumes and permeable enough to permit injection at high flow rates without requiring overly high pressure. 800 m is also deep enough to protect drinking water resources.

Based on log analysis, but particularly side wall cores, several sandstone intervals were identified in the Passaic Formation in the Tandem Lot well with porosities exceeding 10% and permeabilities exceeding 100 millidarcys (Collins et al., 2014). Abundant low-permeability mudstones and the Palisades Sill are potential confining units. Specifically, Slater et al. (2013) reports over 300 m total of a potential sequestration zone, identified based on log data that was confirmed with core and thin section analyses. Additional laboratory measurements indicate porosity ranging from 3.6 to 15% (averaging 11.4%) and highly variable permeability values averaging 213 millidarcys but with a good correlation between porosity and permeability. Two relatively high porosity and permeable zones totaling 365 m were identified below the supercritical depth (Collins et. al., 2011). Below 1520 m (4990 ft), metamorphism and the Palisades Sill limited porosity and permeability. Measured porosities and permeabilities were not as high in sandstones in the nominal Stockton Formation in the TW4 core, all of which are relatively close to the Palisades Sill, and the correlation between measured porosity and permeability, although positive, is verging on non-existent ($r^2=0.08$).

Results from the Tandem Lot well suggest there are potential conventional sandstone sequestration zones and confining units below the supercritical level. However at the Tandem Lot site, these sequestration zones, if they are laterally continuous, tilt towards the surface, lack apparent structural closure, and are thus most simply interpreted as unsuitable for sequestration targets. However, they could be targets if more is known about their lateral distribution and if they intersect the discordant Palisades Sill in the up-dip direction.

Less conventional sequestration targets in the Newark basin are the mafic rocks of the CAMP, particularly lava flow units. The carbonation reaction of CO₂ (in water) is as follows in abstracted and abbreviated stoichiometric form:



The CO₂ as carbonic acid reacts with mafic mineral silicates (here as Wollastonite) to produce bicarbonate ions that react with Ca ions to precipitate limestone. Mg and Fe carbonates can also be produced from reaction with minerals (Takahashi et al., 2000). Once in the form of limestone, the CO₂ is effectively locked up on geological timescales of millions of years. This is part of the reaction series for chemical weathering that on geological time scales removes CO₂ from the atmosphere. But it also offers a fast way of sequestering CO₂ without the risk of long-term leakage Matter & Kelemen (2009).

Newark basin CAMP flows are obvious targets for sequestration (Goldberg et al., 2010). They can have porosity and permeability and are amongst the most reactive basaltic rocks. They were specifically examined for their potential for carbon sequestration Goldberg et al. (2010) using the NBCP cores and logs, specifically Martinsville no. 1 (Fig. 23). They showed that the density and porosity logs vary from 10% to 20% porosity over the 15-m (50-ft) thick flow-top boundary zone between the first and second flows of the Orange Mountain Basalt, which amounts to $\sim 2.25 \times 10^6 \text{ m}^3$ open pore volume per km². Results of direct carbonation experiments on various continental basalts by Schaef et al. (2009; 2010; 2010; 2014) show that the carbonation rates were highest for a sample of the upper (2nd) flow of the Hartford Basin Holyoke basalt. This is particularly interesting because that particular flow has a characteristic and highly unusual joint pattern consisting of extremely abundant and dense vertical fractures that Faust (1978) termed platy-prismatic jointing that might represent a permeating anisotropy increasing mineral effective surface area. The platy-prismatic jointing is characteristic of the entablature of the second flow of the Holyoke basalt, nearly everywhere, from the southern to the northern Hartford Basin and it is also characteristic of that flow's apparent exact correlatives in other Newark Supergroup basins in the Deerfield (2nd flow of the Deerfield Basalt), Newark (2nd flow of the Preakness Basalt), and Culpeper Basin (1st flow of the Sander Basalt). That this is the same eruptive unit is supported by both its chemistry (Puffer & Philpotts, 1989) and peculiar paleomagnetic directions (Prevot & McWilliams, 1989). If borne out by additional experiments this could represent a huge sequestration resource because the principle problem with basalts (or diabase for that matter) is the lack of dense permeable networks with reactive surface, which this basalt seems to have (Fig. 23). We will see this basalt and its platy-prismatic fracture at Stops 1.8 and 2.5.

While it has been recognized that in situ basalt carbonization could be a way of sequestering carbon safely for the long term, the speed of the reaction below the supercritical zone was very poorly constrained and there was concern it was slow (hundreds to thousands of years). However, the recent spectacular test described by Matter et al. (2016) at the CarbFix site in Iceland showed that the in situ carbonation reaction in basalt could be surprisingly fast. They found in their injection experiment that over 95% of the injected CO₂ was mineralized to carbonate minerals in less than 2 years. The experiment was terminated early because the sub surface pump became clogged with carbonate minerals. We do

not know how reactive basalts like the Preakness or Orange Mountain will be because they are much less fresh than those in Iceland, but the results thus far are promising. There is some irony in the promise of geological carbon sequestration in the Newark Basin in that the very source of CO₂ implicated in the ETE could be a viable sink for anthropogenic CO₂.

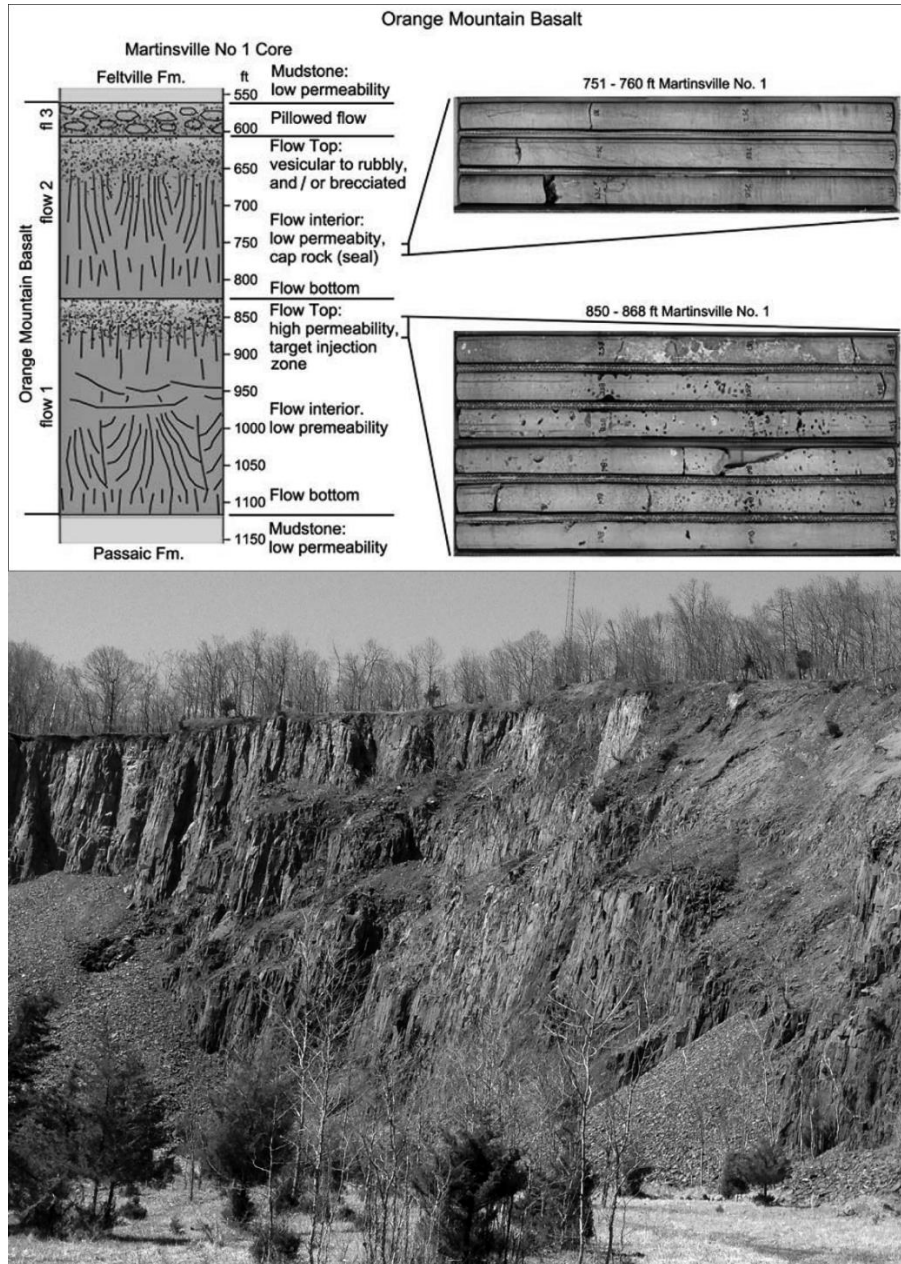


Figure 23: Above, schematic profile of multiple flow units and core photographs from the Orange Mountain basalt, modified from ref. 28. Flow-top boundary zones show considerable vesicular and rubby pore space as compared to the dense, low-positivity flow interior. Scales are in feet). From Goldberg et al. (2009). Below, Preakness Basalt with platy-prismatic jointing, at Dock Watch Hollow, Martinsville, NJ.

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FIELD GUIDE AND ROAD LOG

DAY 1

Meeting Point: Parking Lot of DoubleTree by Hilton Hotel, 425 State Route 59, Nanuet, New York, 10954.
Access is from the east-bound lanes of State Route 59.

Meeting Point Coordinates: 41.090694°N, 73.995438°W

Meeting Time: 8:00 AM (Both Days)

Distance in miles (km)		
Cumu- Lative	Point to Point	Route Description
0.0 (0.0)	0.0 (0.0)	Assemble in the parking lot of the DoubleTree, Nanuet. Leave Parking lot, turning right at entrance on to eastbound State Route 59, keep in right lane.
0.1 (0.2)	0.1 (0.2)	Get on entrance ramp to Palisades Interstate Parkway South from NY-59 E.
0.3 (0.5)	0.2 (0.3)	Merge onto Palisades Interstate Parkway South.
8.4 (13.5)	8.1 (13.0)	Follow Palisades Interstate Parkway South to US-9W N/N Rte 9W N in Alpine. Take exit 4 from Palisades Interstate Parkway South.
8.5 (13.7)	0.1 (0.2)	Turn right onto US-9W northbound.
9.4 (15.1)	0.9 (1.5)	Pass traffic light and turn right into campus of Lamont-Doherty Earth Observatory of Columbia University just before New York-New Jersey State line (Ludlow Lane).
9.7 (15.6)	0.3 (0.5)	Proceed to the front of the Geoscience building. Park and you will be guided on foot to the core repository.

STOP 1.1: Cores, Seismic Lines and Posters at Lamont-Doherty Earth Observatory, Palisades, NY

NYSGA: Geologic Diversity in NYC

Location Coordinates: 41.004471°N, 73.908982°W: Core repository Laboratory, Geoscience Building, Lamont-Doherty Earth Observatory.

Duration: 1:00 hr.

Here we will examine cores, down-hole logs, seismic profiles, and some fossils from the New Jersey portion of the Newark Basin and in a more central position. These will relate directly to the outcrops will see after this stop.

Newark Basin Coring Project (NBCP) and Army Corps of Engineers (ACE) Cores

Stockton Formation - Princeton NBCP core and TW4: The facies of the Stockton Formation as seen in the Princeton core is closely comparable to the outcrops of the type area of the Stockton Formation near Stockton, New Jersey. Thick intervals of purplish, tan, and gray sandstone and minor conglomerate are interspersed with bioturbated red mudstone and muddy sandstone. Pedogenic features including cracks, slickensides, carbonate nodules, rhizoliths (root traces), and burrows are abundant.

Compare the red mudstone facies of the Stockton Formation in the Princeton Core with typical red mudstone from the nominal Stockton Formation of the TW4 core. Note that the huge sand-filled cracks seen in the TW4 mudstones are not seen in Princeton core mudstones. Also note the difference in color with the higher color saturation and slight purple tint of the TW4 mudstone. The TW4 facies resembles the nominal Stockton we will see in outcrop at Stop 1.2 (Fig. 24).

Lockatong Formation - Nursery and Princeton NBCP cores: Cores and logs: intervals of cores of exactly correlative lacustrine (Van Houten) cycles will be on display. Observe the decrease in thickness and change in facies from the Nursery core to Princeton, corresponding to the up-dip relationship between Nursery and Princeton and how that projects towards the hanging-wall side of the basin – the Granton Quarry area (Stops 1.3 and 1.5).

Also on display will be examples from these same cores of the lower Lockatong Formation, which changes dramatically in facies between Nursery and Princeton with correlation based on the magnetostratigraphy.

The trend from Nursery to Princeton is from thicker to thinner cycles, from finer to coarser grain, and a tendency for the lower Lockatong cycles to be replaced by red beds and tan and gray sandstone. This trend continues and is exaggerated from the Princeton area towards the northeast towards our field stops.

Passaic Formation – Somerset, Rutgers, and Titusville NBCP cores: Examples of stratigraphic overlap between the Rutgers and Titusville cores illustrates an extreme examples of lateral continuity. In particular are two siltstone layers only a few cm thick correlative between the two coring sites separated by 42 km. There is also a polarity transition just below these siltstones that serves as a corroboration of the hypothesis of correlation.

Examples of the single black-mudstone-bearing Van Houten cycle near the base of the Kilmer Member from the Somerset and Rutgers cores (11.5 km apart) will be on display. Note slight change in facies. As in the case above there is a polarity transition below and close to the base of this cycle and the base of the Member.

Passaic Formation and Orange Mountain Basalt – Martinsville core and ACE core PT-38: The upper few tens of meters of the Passaic Formation and its contact with the overlying Orange Mountain Basalt will

be shown. The interval on display in the two cores contains the base of the Exeter Member of the Passaic Formation, the remarkably continuous and thin polarity zone E23r, and the projected level of the ETE. At first glance there is not much to suggest anything particularly momentous is going on here. However, correlation between the Jacksonwald Syncline section (Exeter) in the southwestern Newark Basin, from where most of the pre-ETE and ETE biological data come from, and the Martinsville, ACE cores, and exposures (Stop 1.7) is highly corroborated by cyclostratigraphy (the deep water phase of lacustrine cycle at base of the Exeter Member, polarity zone E23r, and the base of the Orange Mountain Basalt).

There is scant biological data from the cores themselves, no sporomorphs, for example; however, the spinocaudatan (clam shrimp) *Shipingia olseni* is present within the fine mudstones in E23r in the Martinsville core, consistent with its presence in the Jacksonwald Syncline section (Kozur and Weems, 2005). Supposedly, the presence of this taxon is indicative of a Norian age and is the principle basis for the argument that a major hiatus, spanning millions of years, is present between the base Exeter Member and what we identify as the ETE (Kozur and Weems, 2005; 2010; Tanner and Lucas, 2015).

The key observation supporting the hiatus as stated by Kozur and Weems (2011) is that, "...the upper Norian faunas were dominated by very large conchostracans, while the Rhaetian (and Hettangian) conchostracan faunas are everywhere composed of very small forms." The lack of the early and middle Rhaetian clam shrimp zones and the presence of the large *Shipingia* just below the ETE in Newark Supergroup strata led to the hypothesis of this major hiatus. However, their hypothesis is falsified by the discovery of abundant large cf. *S. olseni* in late Rhaetian, largely marine strata in the North Germanic Basin underlain by strata contains the middle to late Rhaetian bivalve *Rhaetavicula contorta* and overlain by basal Liassic strata with psilocerid ammonites (Olsen and Kent, 2016).

Kent et al. (2016) note that, "...the recorded presence of the very short (~10 ka) Chron E23r immediately below CAMP basalts in three entirely separate basins, namely Newark (Kent and Olsen, 1999; Kent et al., 1995; Olsen et al., 1996a), Fundy (Deenen et al., 2011) and Argana (Deenen et al., 2010), makes such a major regional hiatus untenable. This is because the hiatus would require essentially identical intervals of nondeposition and/or erosion in all of these basins, so that somehow deposition halted just after or else each section was eroded to variable depths exactly down to just above Chron E23r, the shortest identified chron in the Newark-Hartford APTS. But not to be deterred, (Tanner and Lucas, 2015) do not regard such an argument as precluding a regional unconformity and speculate that a late Rhaetian episode of uplift along the entire rift axis might somehow have caused such precisely timed nondeposition or erosion of the exposed strata in all of the rift valleys prior to the flood basalt eruptions."

Examine the cores, and see if you can spot if there is any physical evidence for a hiatus.

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 Jacksonwald syncline to the southwest, to obscure cyclicity 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 crocodile-line tracks such as *Brachychirotherium* and *Apatopus*, while outcrops adjacent to the location of PT-38 have produced a post-initial-ETE assemblage of sparse sporomorphs and microfossil plants, and abundant footprints (Stop 1.7).

Preakness Basalt – ACE core PTI-3 and Lucent Technologies core MW-21D: ACE core PTI-3 was described by Tollo et al. (1990) and Tollo & Gottfried (1992). It contains both the pillowed lowest flow of the Preakness Basalt and the lower part of the second flow with the characteristic platy-prismatic jointing

that we will see in exposures at Stops 1.8 and 2.5. Lucent Technologies core MW-21D was cored in New Providence, NJ adjacent to the Lucent Technologies (formerly Bell Labs) complex at Berkeley Heights, NJ at 40.688342°, -74.396989°. This core site is 1.7 km northeast from the site (40.676546°, -74.409439°) from which Blackburn et al. (2013) obtained a zircon U-Pb CA ID-TIMS, Thorium-corrected ²³⁸U-²⁰⁶Pb age of 201.274±0.032 Ma from a gabbroid float in a stream cut of Preakness Basalt. Several well-developed gabbroid layers are present in Core MW-21D and samples from this core produced an indistinguishable, although as yet unpublished, age (T. Blackburn and J. Ramezani, pers. com., 2014). There is a small gabbroid roadside exposure directly in-line and between these two sites at (40.677774°, -74.408121°) from which we have a sample that will be on display.

This Preakness date along with others from intrusions associated with the other basalt flow formations of the Newark Basin, tested and precisely corroborated the hypothesis that the cyclicity seen in the syn-CAMP lacustrine strata was astronomically paced. The durations of the sequences between basalt flow formation based on U-Pb dates provided by Blackburn et al. (2013) are indistinguishable from the astrochronological estimates derived from Olsen et al. (1996b) and Whiteside et al. (2007).

Distance in miles (km)		
Cumu- Lative	Point to Point	Route Description
10.0 (15.6)	0.3 (0.5)	Return to vehicles and proceed to exit, back to 9W southbound. Passing TW4 core site at left at (41.002829°, -73.910658°). Turn left onto Route 9W southbound.
10.7 (17.2)	0.7 (1.1)	Proceed south and turn left onto ramp for Palisades interstate Parkway (PIP) south.
10.8 (17.4)	0.1 (0.2)	Take ramp and enter PIP southbound. PIP follows ridge crest of the Palisades cuesta at about 1/3 stratigraphic thickness of the sill.
18.7 (30.1)	7.9 (12.7)	Take PIP southbound and take exit 1 on right for E Palisade Ave in Englewood Cliffs.
18.9 (30.4)	0.2 (0.3)	Take ramp for exit 1, keeping left.
19.0 (30.6)	0.1 (0.2)	Turn left onto E. Palisade Avenue
21.2 (34.1)	2.2 (3.5)	Turn right onto Hudson Terrace.
21.4 (34.4)	0.2 (0.3)	Continue straight onto Main St. This road follows a small fault with normal separation that offsets the Palisades ridgeline, dropping down to the left.
22.3 (35.9)	0.9 (1.4)	Turn left onto Henry Hudson Drive and into Palisades Interstate Park.
22.6 (36.4)	0.3 (0.5)	At circle take first right exit. Proceed downhill.
22.7 (36.5)	0.1 (0.2)	Turn right at Ross Dock Picnic area and park.

STOP 1.2: Ross Dock Picnic Area. Fort Lee, NJ

Location Coordinates: 40.860247°, -73.956157°: Parking for Ross Dock Picnic Area. Nominal Stockton Formation, olivine zone in Palisade Sill.

Duration: 1:30 hr.

Here we will examine strata below locally exposed Lockatong Formation that is intruded by the Palisades Sill. Then we will look at the olivine zone on the way back to the bus.

Walk south along the path next to the river to its termination at the south tip of the park. At several places we will look at mudstones and sandstones of the Stockton. Walk 960 yards south along the river path to beneath the George Washington Bridge. Continue 887 yards farther south looking for outcrops at head height on the right (west) of the path. Depending on the plant cover you will see variegated (tan purple and red) cross-bedded arkosic sandstones and minor conglomerate interbedded with massive, intensely red-purple mudstone. The mudstone is cut by numerous large and deep ptygmatic sandstone-filled cracks that are often vertically boudinaged. This is the same facies we see in TW4 below the possible Locketong (Stop 1.1).

Figure 24: Ptygmatic arkose-filled crack in massive, purplish-red mudstone of nominal Stockton Formation at Stop 2.2. Cracks are about 54 cm wide but look wider because of “bleaching” of mudstone around cracks.



Phytosaur locality (40.846613°, -73.963452°): At the end of the path along the Hudson is part of a set of outcrops of some notoriety, having been the subject of page one of the 1910 Christmas Day pictorial section of the New York Times, entitled, “When the Giant Dinosaur Walked Down Broadway”. Never mind that it was not a giant, or a dinosaur, or on Broadway! It is in fact a partial disarticulated postcranial skeleton of a large phytosaur (semi-aquatic, non-archosaur, crocodile-mimic pseudosuchian) that was recovered at the water's edge (Fig. 25) on private land then owned by the “three Goetschius brothers” just south of the park boundary. It appears to have come from the transition between an overlying arkose and red mudstone. The specimen was discovered in March or April of 1910 by Jesse E. Hyde,

Daniel D. Condit, and Albert C. Boyle Jr., who were at the time graduate students of Prof. James F. Kemp of Columbia University (Hyde, 1911). They contacted Barnum Brown and W. D. Matthew at the American Museum of Natural History (AMNH) in New York City. The specimen was collected by Brown for the AMNH over a two-week period in late December 1910, after a few months of negotiations with the land owners. The phytosaur locality has been misidentified in most published reports, and was cited by the New York Times (December 21, 1910) as being “a half-mile south of the George Washington Bridge, opposite 155th St.”, even though 155th St is closer to one mile south of the bridge! However Matthew (1911) and Hyde (1911) state that the specimen is from opposite 160th St., which is approximately 1/2 mile south of the bridge and there are outcrops still present on private land at (40.844472°, -73.965003°); The specimen is also commonly referred to as the “Fort Lee phytosaur”, although it is actually from Edgewater. The specimen (AMNH 4991; Fig. 27) consists of several posterior dorsal, sacral, and anterior caudal vertebrae, both femora, tibiae, and fibulae, a few dorsal ribs, many gastralia, and numerous osteoderms. Huene (1913) described the specimen and named it *Rutiodon manhattanensis*. Based on the structure of the ilium, the generic assignment is correct (Huber and

Lucas, 1993; Huber et al., 1993) but the specimen is indeterminate at the species level. This is the only vertebrate reported from this facies, but it suggests that further exploration might prove fruitful. AMNH 4991 is currently on exhibit in the Hall of Vertebrate Origins at the American Museum of Natural History in New York.

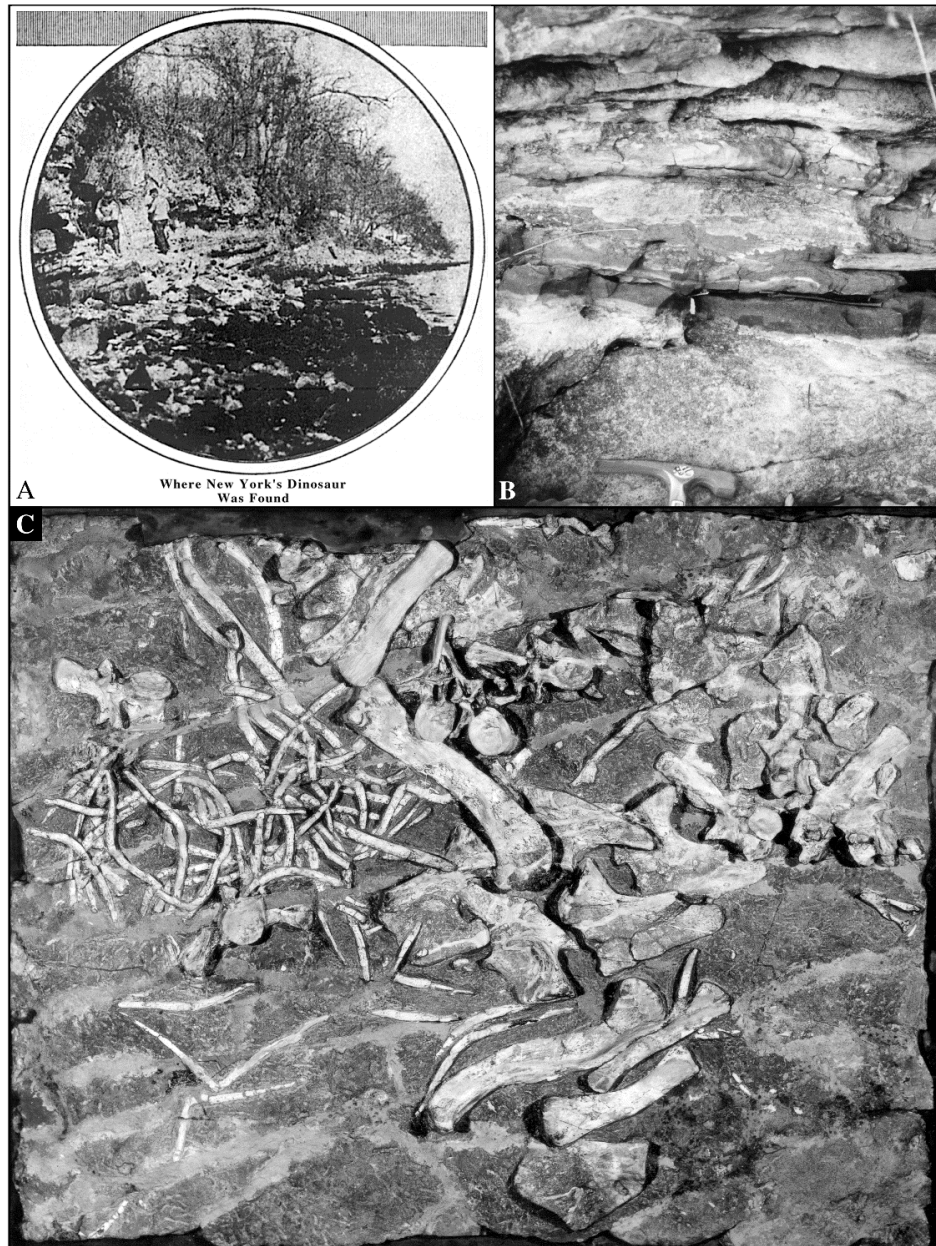


Figure 25: The phytosaur *Rutiodon manhattanensis* and outcrops at Stop 1.2, Stockton Formation: A, photograph from the front page of the magazine section of the New York Times for December 25, 1910, showing the location of the phytosaur skeleton just south of the boundary with the Palisades Interstate Park (Stop 3g) (with permission of the New York Times); B, photograph of typical lithologies (purple and red mudstones and tan arkose) at the north end of the outcrops shown in A; C, disarticulated partial skeleton of the large phytosaur *Rutiodon manhattanensis* (AMNH 4991) (courtesy of the American Museum of Natural History)

Take the switchback path that works its way up the hill to Henry Hudson Drive and proceed south toward the intersection with Main Street (Fort Lee) and the entrance to the park.

Weathered olivine zone of the Palisade Sill (40.846664°, -73.965107°): Exposure on the north side of road has weathering profile of the olivine zone that can be clearly seen here. The olivine zone is weathered to a very crumbly diabase that, according to Naslund (1998), still has many fresh looking olivine crystals. The olivine zone is about 10 to 15 m above the base of the Palisades Sill and its weathering produces an obvious bench along the escarpment, essentially paralleling the lower contact of the sill, as observed by Walker (1969).

Proceed north along the base of the sill on Henry Hudson Drive, keep track of the bench and note where sedimentary strata at the base of the sill are exposed. Also note the faint and fine layering visible on the diabase surfaces near the base of the sill. Beware of vehicles!

Slightly discordant base of sill and metamorphosed Lockatong Formation (40.851995°, -73.960580°): The thin-bedded to massive metamorphosed mudstones and siltstones, interbedded with tan arkosic sandstone, is typical of the lower Lockatong in this part of the basin, and it shows unequivocally that the red and tan units along the river are stratigraphically lower than Lockatong Formation. Of interest here is that as the sill cuts lower in the section, at either side of this outcrop, the strata become disrupted and the finer-grained appear to behave more competently while the coarser beds appear to have liquefied and partially and chaotically mixed with the diabase (Fig. 26). Note that the topographic bench tracking the olivine zone appears to bump upward here.

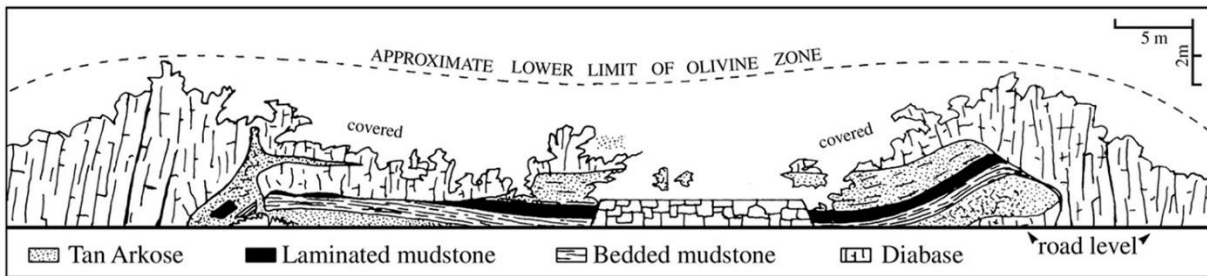


Figure 26: Sketch of discordant contact of Palisade sill and Lockatong Formation along Henry Hudson Drive just south of the George Washington Bridge at 40.851995°, -73.960580° (from Olsen, 1980).

Proceed north under the George Washington Bridge. The road basically follows the base of the sill.

Razor-edged contact between sill and Lockatong Formation (40.854159°, -73.959776°): Probably the best and most accessible contact between the Palisades Sill and underlying meta-sediments is here, where the contact tends to be planar, and very sharp, and although predominantly concordant, locally hops up or down in stratigraphy tens of centimeters. Note that over most of the exposure there is very little evidence of melting or assimilation and that fine-scale sedimentary structures are well-preserved despite proximity to the sill.

While these rocks are gray to nearly black, they contain no organic carbon. North along this same road, past the circle (at 40.862042°, -73.955783°), are exposures of similarly metamorphosed Lockatong Formation in a part of the stratigraphy (cycles W-5 and W-6 of the Nursery Member) that has been

traced along the base of the sill to Hoboken and correlated with the NBCP cores (Olsen, 1980c; Olsen et al., 1989; 1996a). Away from the sill, the fossiliferous (clam shrimp, ostracodes, fish, and rare reptiles), microlaminated portions of the cycles have total organic carbon contents of 3 to 8 % (Manspeizer & Olsen, 1981), but here have essentially 0% organic carbon. Prior to larger scale metamorphism by the intrusions, the organic carbon content was probably in excess of 10%. This carbon, originally in the form of kerogen, was thermally cracked eventually to gas, finally methane and CO₂, presumably with ¹³C-depleted isotopic ratios (δ¹³C of ~-28‰). These greenhouse gasses made their way into the atmosphere, but whether they did so catastrophically, contributing to the ETE greenhouse crisis and “initial isotopic excursion”, or whether they leaked out slowly and assimilated without incident is unknown, although signs of disruption by escaping gas are certainly not obvious.

Proceed north to traffic circle.

Reclaimed quarry in lower Palisades Sill at traffic circle (40.857748°, -73.959041°): In this old quarry, the olivine zone is marked by a zone of deflected columns in the cliff face in fresh rock. Naslund (1998) notes flow banding is present within the olivine zone. You can project that zone of deflected column into the benchmarking the weathered olivine zone. What might the evidence be that the olivine zone represents a separate intrusive event or separate major pulse as suggested by Husch (1990), Puffer et al. (2009), and Block et al. (2015), and can you see any of the evidence here?

Proceed down steps at northwest side of circle and back to parking area.

Distance in miles (km)		
Cumu- Lative	Point to Point	Route Description
23.1 (15.6)	0.4 (0.6)	Return to vehicles and proceed to exit, back to Main street, Fort Lee. Turn right and Main Street, Fort Lee, becomes River Road, Edgewater, NJ.
26.3 (42.3)	3.2 (5.1)	Proceed south on river road. There are several exposures of Lockatong and Stockton formations along the base of the Palisades Sill along this route, described in Olsen (1980c). Keep left to stay on River Road. On Gorge Road to the right are excellent exposures of Lockatong Formation described by Olsen (1980c, 2003) and Colbert & Olsen (2001). These allowed the section at Granton Quarry above the Palisade Sill (Stop 1.5) to be concatenated with that of the sections below the sill.
26.7 (43.0)	0.4 (0.6)	Turn Right onto Church Road and Park along road.

STOP 3: Churchill Road Exposure of Dramatically discordant sill and Lockatong, North Bergen, NJ

Location Coordinates: 40.803242°, -73.994221°:

Duration: 0:30 hr.

This is a very large, virtually unstudied exposure of a dramatically discordant contact between the Palisade Sill and the Lockatong Formation (Fig. 27). “Measured” using GoogleEarth™ (which has large uncertainties in altitude) the shear face of the wall is about 39 m. Using this as a guide, and accounting for parallax (non-quantitatively), about 23.5 m of Lockatong is exposed and the sill cuts down through it all towards the southeast. About three large cycles, each consisting of a couplet of a complex of prominent tan sandstone beds and complex of mudstone beds, are present averaging about 7.8 m thick.

Given that the average thickness of a 20 kyr cycle in this area is about 1.5 m (Olsen, 2001), consistent with smaller scale alternations of laminated and non-laminated mudstones at this outcrop, it would suggest the larger cycle is an expression of the short eccentricity (~100 ka cycles). We will see more of this pattern at Granton Quarry at Stop 1.5.

Strata near the contact are deformed in a variety of ways. There are a number of blobby diabase intrusions that warp the surrounding strata, as well as several sheet-like ones that extend far from the main diabase body, and a series of small thrust faults (Fig. 27). This outcrop is very similar to that at Kings Bluff at Weehawken (Darton, 1890; Olsen, 1980c), except even more dramatic (Fig. 27). From here to the south near Weehawken, the base of the sill sits below the lowest well-developed mudstones on nominal Stockton Formation (Stop 1.4, below).



Figure 27: Dramatically discordant contact between Lockatong Formation and Palisade Sill. Sill cuts down section here to base of Lockatong at Stop 1.3.

Distance in miles (km)		
Cumu- Lative	Point to Point	Route Description
27.3 (43.9)	0.6 (1.0)	Return to vehicles and proceed back to River Road, turning right and proceed south to Walgreens parking lot. Turn right into parking lot of the Walgreens Pharmacy and park.

STOP 1.4: Palisades Sill contact with nominal Stockton Formation, North Bergen, NJ

Location Coordinates: 40.803242°, -73.994221°:

Duration: 0:30 hr.

This is another very large virtually unstudied exposure this time of largely concordant contact between the sill and underlying Stockton Formation strata. “Measured” by GoogleEarth™, the cliff face is about 51 m high (consistent with topographic map) (Fig. 28). Based on photographs and accounting for parallax (non-quantitatively) about 14 m of mostly tan and white arkosic sandstone and minor pebbly sandstone with a minor bed of olive-gray sandy mudstone are exposed here below the sill. Several tan arkosic xenoliths are present, one of which remains attached at its south end, making a flame-like structure (Fig. 28). This outcrop has been described by Zakharova et al. (2016) and this specific xenolith compared with the arkosic xenolith in core TW4 (Stop 2.1).

This facies of tan arkosic sandstone is common below and in part above, typical Lockatong Formation. To the north it seems to interfinger with purple and red mudstones. While we are calling this Stockton

Formation, it is unclear if it is a lateral facies equivalent of the lowest Lockatong or time equivalent to part of the Stockton as it is represented in the central Newark basin.

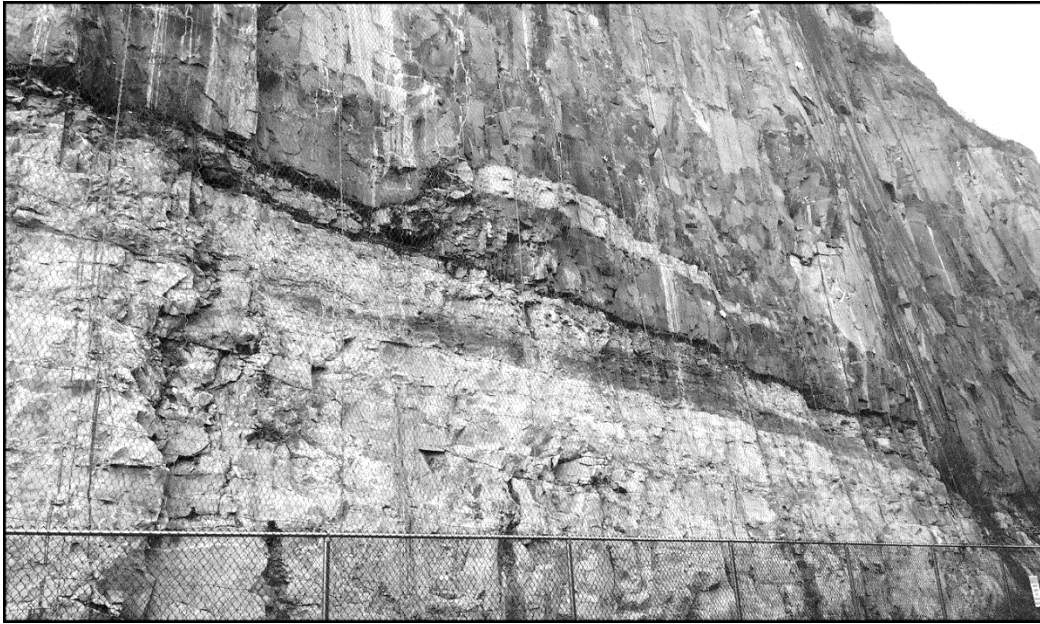


Figure 28: Xenolith still attached to underlying strata at Stop 1.4.

Distance in miles (km)

Cumu- Lative	Point to Point	Route Description
28.1 (45.2)	0.8 (1.3)	Return to vehicles and proceed to exit and turn left onto River Road, North Bergen. Turn sharp left onto River View Drive south.
28.7 (46.2)	0.6 (1.0)	Follow River View Drive south to 17 th Street. Continue straight on 79 th Street, North Bergen, NJ.
29.0 (46.7)	0.3 (0.5)	Continue on 79 th Street. Make left and right “dog leg” on John F. Kennedy Boulevard, continuing on 79 th Street.
29.4 (47.3)	0.4 (0.6)	Continue on 79 th Street to Tonnelle Avenue (US Route 1 & 9) driving down dip-slope of Palisade cuesta, and cross into parking lot for Lowes Home Improvement Center, North Bergen, NJ and park.

STOP 1.5: Granton Quarry exposures of Lockatong Formation and Granton Sill, North Bergen, NJ

Location Coordinates: 40.807058°, -74.017908° and 40.807380°, -74.017774°.

Duration: 1:00 hr.

Leave vehicles and proceed to northeast corner of store. The following is adopted from Olsen (2003). In the northeastern Newark basin, the outcrop belt of the Lockatong on both sides of the Palisades sill along the Hudson River is remarkably rich in vertebrate fossils, despite varying degrees of contact metamorphism. In this region, virtually all fine-grained facies and all cycles have some vertebrate body fossils. Here, only the Princeton, Nursery, and Ewing Creek members of the Lockatong Formation have

been positively identified. The Granton Quarry is the most famous of all the fossil localities in this belt and, in fact, the entire Newark Basin.

Remnants of the old Granton Quarry are preserved between the new Lowes Home Building Center on the south and Tonnelle Plaza (Hartz Mountain Industries) on the north. Granton Quarry was actively quarried for road metal, fill and rip rap during the 1950s and 1960s and was abandoned by 1970, whereafter it was slowly consumed by commercial developments and warehouses. Nonetheless, excellent exposures remain. The site has produced, and continues to produce, extraordinarily abundant fossils, especially vertebrates, and it is certainly one of the richest sites in North America for the Triassic (Fig. 29). This is also the best locality on this trip to see the details of Lockatong-type Van Houten cycles. Eleven such cycles with a thin-bedded to laminated division 2 are exposed on the sill-capped hill: seven are exposed on the south-facing exposure (Fig. 30) where we will see them first (40.807058°, -74.017908°), three additional cycles are exposed on the east-facing exposure; and all 11 cycles are exposed on the north-facing exposure, which is where we will examine them in more detail (40.807380°, -74.017774°). The base of the section appears to be 38-46 m above the contact with the Palisades Sill (Van Houten, 1969). This contact may be close to what was, prior to intrusion, the local Stockton-Lockatong formational contact. This section has been described in several papers including Van Houten (1969), Olsen (1980c; 2003), Olsen et al. (1989), and Colbert and Olsen (2001).

According to Van Houten (1969), these Lockatong hornfels include calc-silicate varieties in the middle carbonate-rich part, and extensively feldspathized and recrystallized diopside-rich arkose in the upper part. Some beds of arkose show well-developed cross-bedding. Because of the buff arkose at the top of nearly every cycle, these are the most visually-graphic of the detrital cycles seen on this field trip. Here, the many correlated changes occurring though individual cycles can be easily seen (Fig. 30).

Cycles G3 and G7 (Fig. 30) have produced representatives of all the known skeletal remains of Lockatong vertebrates, except definitive examples of the holostean *Semionotus*. The basal portions of division 2 of both of these cycles have extremely high densities of fossil fish,

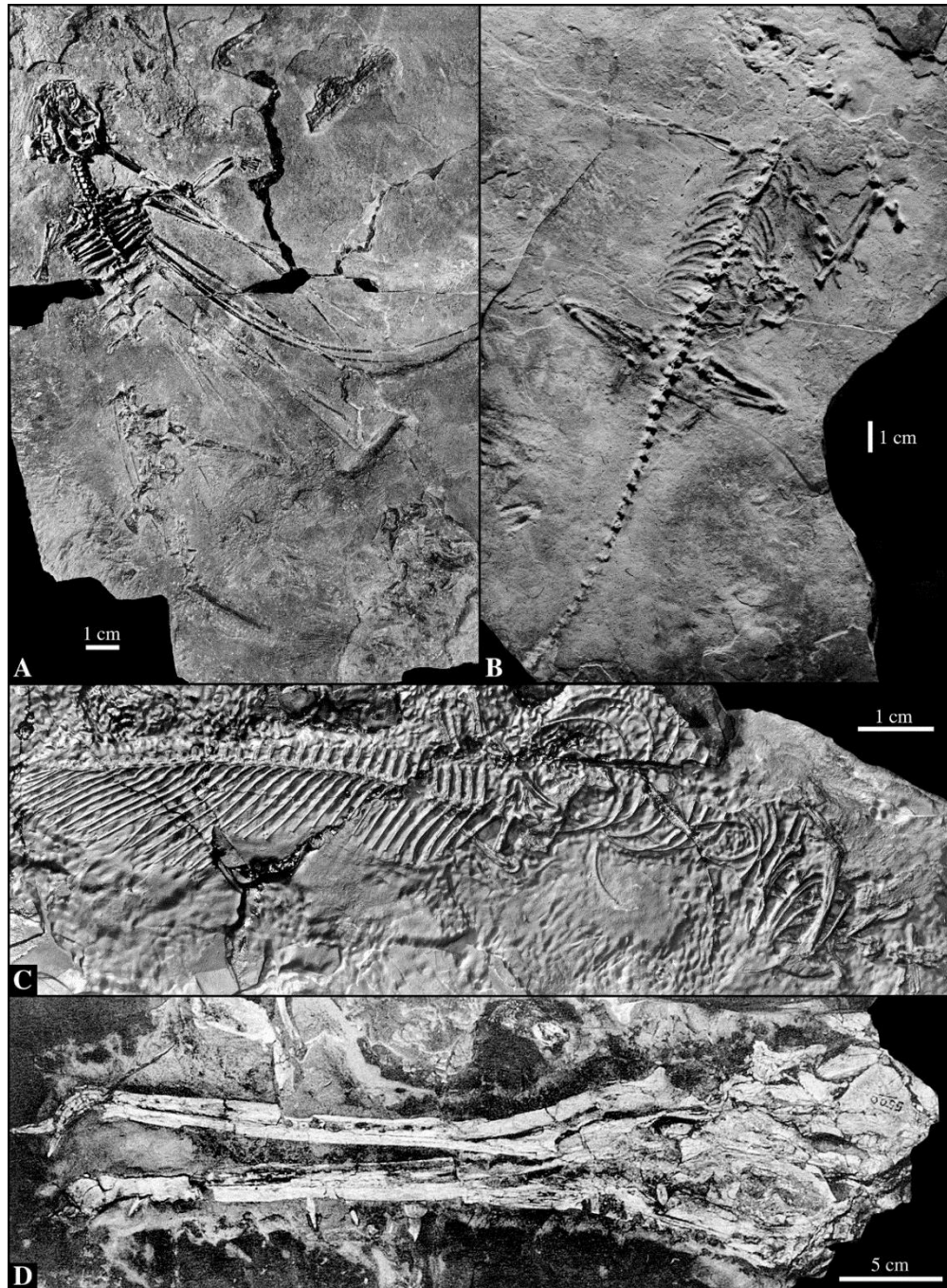
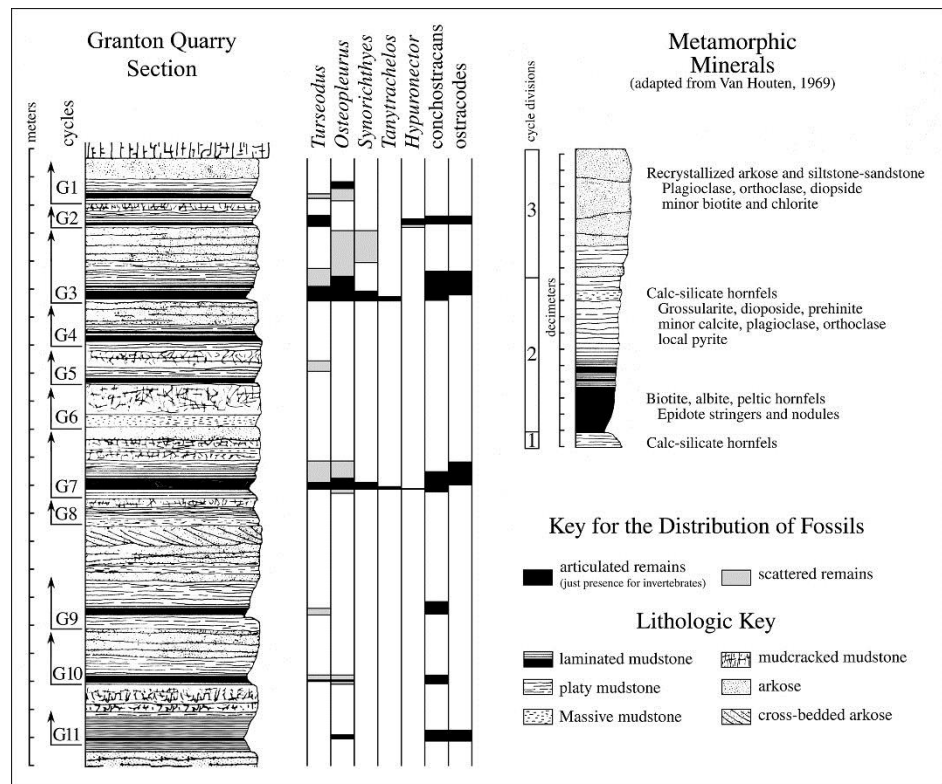


Figure 29: Reptiles from the Lockatong Formation at Granton Quarry (Stop 1.5): A, holotype specimen of *Icarosaurus seifkeri* from cycle G-3 (uncataloged AMNH specimen) (from Colbert, 1966; with permission of the American Museum of Natural History); B, female *Tanytrachelos ahynis* found by Steven Stelz, Trinny Stelz and James Leonard (New Jersey State Museum GP 22356); C, holotype specimen of *Hypuronector limnaios* from cycle G7 (AMNH 7759) (from Colbert and Olsen, 2001; with permission of the American Museum of Natural History); D, skull of cf. *Rutiodon carolinensis* found in float (AMNH 5500) (from Colbert, 1965; with permission of the American Museum of Natural History).

especially the coelacanth *Osteopleurus newarki* (Schaeffer, 1952). Small reptiles are also surprisingly abundant, and many important fish and unique reptile skeletons have been discovered here by dedicated amateurs and donated to various museums through the years (Colbert, 1965; 1966; Colbert and Olsen, 2001; Olsen et al., 1989; Schaeffer, 1952; Schaeffer and Mangus, 1971). Without a doubt, the three most spectacular skeletons of small reptiles found in the Lockatong Formation come from this site. These include the type specimen of the bizarre "deep-tailed swimmer", *Hypuronector limnaios* (Colbert and Olsen, 2001), the peculiarly-abundant and sexually dimorphic, tanystropheid *Tanytrachelos ahynis* (Olsen, 1979), and the gliding lepidosauromorph *Icarosaurus seifkeri* (Colbert, 1966) (Fig. 29). Larger remains occur as well, of which the most spectacular is the skull of a juvenile rutiodontine phytosaur (Fig. 29), but isolated phytosaur bones and teeth are fairly common and isolated vertebrae of a metoposaur amphibian also been found. This biotic assemblage is very different than that seen in post-ETE strata, but some of the elements are present just before the ETE, including phytosaurs and tanystropheids, both based on footprints in the Newark – *Apatopus* and *Gwynnedichnium*, respectively. Cycles G8 through G11 overlap with the section on the east side of the Palisades Sill as exposed at Stop 4, as has been previously noted. A prediction of this correlation is that cycle W0 should be equivalent to G11. Examination of the easternmost outcrops at Granton Quarry of cycle G11 show that this is indeed the case. In fact, this cycle is distinctive in having a very pyrite-rich division 2 that has strikingly bright yellow and orange clay seams on weathering, a feature not seen in other Granton Quarry cycles. With the sections from both sides of the Palisades Sill combined, it is now possible to look at trends in

Figure 30: Measured section at Granton Quarry, Stop 1.5. Section is based on measurements from both sides of the promatory. Based on Olsen (1980b) and Olsen (2003). Metamorphic Minerals from Van Houten (1969). All of these cycles are still exposed.



lithology and biota at the scale of from a few thousand years (within one Van Houten cycle) to over 1 million years (i.e. three McLaughlin cycles) (Fig. 31).

Van Houten cycles thin to an average of about 1.5 m in this region, and at least some cycles from the drier phases of the 404 ky McLaughlin cycles appear to be entirely missing or replaced by tan arkose, accounting for the disproportionately low (18m) thickness of McLaughlin cycles in this area. If they scaled to the Van Houten cycles that are easily recognized, the McLaughlin Cycle should be $\sim 20 \times 1.5 \text{ m} = 30 \text{ m}$ thick. The couplets (i.e. varves) of microlaminated mudstones are thinner than their counterparts towards the center of the basin, but not in proportion to the thickness of the cycles, again suggesting a preferential omission of drier facies in each cycle.

This is an extension of the same thickness and coarsening trend in the Lockatong Formation that we have seen going from the Titusville to Nursery to Princeton NBCP cores (Stop 1.1). It is consistent with thinning onto the hinge margin of a half graben growth structure and incompatible with the strict version of the Broad Terrane hypothesis. The “missing” cycles and corresponding thinning of the McLaughlin cycles is consistent with offlap and bypassing on the hinge margin during lake lowstands and is also inconsistent with the strict version of the Broad Terrane. In fact, projecting that trend to the east and north, one would expect the Lockatong Formation to thin further and disappear entirely into sandstones, becoming unrecognizable, which is exactly what it does.

The stratigraphic sequence in the Hackensack Meadowlands, underlying the Granton sill to the west of the sill’s dip slope, consists of arkosic tan sandstones, overlain by black shales (that surely represent much of the remainder of the Lockatong Formation), which are in turn overlain by red mudstones of the Passaic Formation (Parker, 1993). If something like the average accumulation rate, based on the

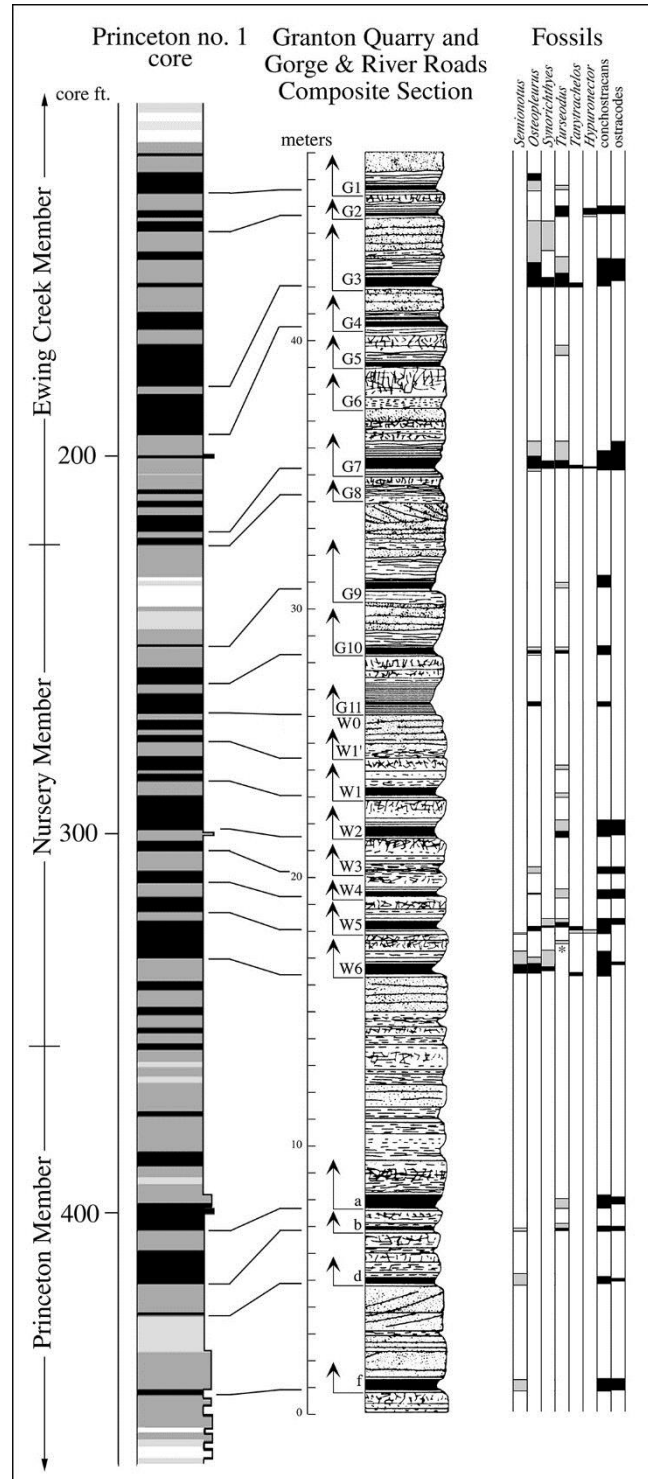


Figure 31: Composite section in vicinity of stop 1.2-1.5 with the distribution of fossils and correlation to the NBCP Princeton no. 1 core. Modified from Olsen (2003).

thickness of the Princeton, Nursery, and Ewing Creek members in the vicinity of North Bergen and Edgewater (i.e. 18 m/McLaughlin cycle), was maintained upward to the position of the Graters Member of the Passaic Formation (encountered in a boring; Lovegreen, 1974 cited in Parker, 1993), there is sufficient stratigraphic thickness in this area for the rest of the Lockatong and basal Passaic Formation.

At the south-facing exposures, cycles G1 and G2 are injected by diabase of the 20 m thick Granton sill (Van Houten, 1969), another component of the CAMP, which has protected the Lockatong Formation from erosion in this area. Because this sill is thin, and the Palisade Sill fairly remote stratigraphically, much of the sedimentary rock is not as metamorphosed as under the sill; some cycles still have considerable organic matter, especially noticeable in cycle G3 which on the south side of the Quarry is black with significant organic matter (Maliconico, 2010), but has little or none at the north end, where we are looking. The change in bone color tracks the thermal maturity as well: black where relatively low and white where high. Perhaps, the sandstones interbedded with the formerly organic-rich mudstones could act as conduits in the up-dip direction, venting methane and CO₂ into the atmosphere.

Distance in miles (km)			Route Description
Cumu- lative	Point to Point		
32.0 (51.5)	2.6 (4.2)		Return to vehicles and proceed to exit and turn right onto Tonnelle Avenue (US Route 1 & 9) south. Proceed southbound to ramp for NJ 3 on right.
32.6 (52.5)	0.6 (1.0)		Take ramp for US 3 and merge with US 3.
36.5 (58.7)	3.9 (6.3)		Continue on US 3 to exit for Polito Avenue on right.
37.4 (60.2)	0.9 (1.4)		Follow signs for Polito Avenue. Turn left onto Polito Avenue.
37.8 (60.8)	0.4 (0.6)		Turn right into parking area for 165 Polito Avenue, Lyndhurst, NJ 07071

STOP 1.6: Passaic Formation and Copper Prospect at Lyndhurst, New Jersey

Location Coordinates: 40.807357°, -74.108448°.

Duration: 0:45 hr.

Leave vehicles and proceed to view the adjacent exposures. The following is modified from Olsen et al. (2004).

We have moved considerably up section from earlier stops as well as deeper towards the depositional center of the basin, although we are still in a marginal facies. These exposures reveal most of the Kilmer Member of the Passaic Formation (~210 Ma, NBTS) (Fig. 4) exposed on the east side of a prominent ridge marking the western border of the New Jersey Meadowlands. This ridge is characterized by a heterogeneous assemblage of red mudstones and sandstones. However, there are a few purple and gray units present, and the eastern and stratigraphically lowest of these is parsimoniously interpreted as marking the base of the Kilmer Member (Fig. 4).

The Passaic Formation, characterized by extremely widespread units with a particularly well-developed “layer cake”-style stratigraphy, marks a stage in the tectonic evolution of the Newark rifts. Based on the very slow rates of thickening towards the faulted margins of the basin (Stop 1.1; Withjack et al., 2013), extension rates were slowing, the basin was filling towards its outlet, and the basin floors were extraordinarily wide and flat. The red Passaic and other central Pangean equivalents are the strata that most people think of when they think

Triassic. At yet a broader scale, the Passaic marks an interval where basins deeper within the arid belt and closer to the rifting axis began to receive considerable amounts of brine of marine origin with the consequent development of thick evaporite sequences (e.g. Osprey Salt of the Canadian Maritime margin).

Based on correlation to the central Newark Basin, albeit tentative, the exposures at this stop reveal the uppermost few meters of member T-U and the lower half of the Kilmer Member (Fig. 4). In the central Newark basin, the basal Kilmer Member includes a prominent Van Houten cycle with a well-developed black division 2. In the region around New Brunswick (NJ), this black shale and the underlying division 1 of this cycle are often rich in copper minerals, particularly chalcopyrite. At these outcrops in the northeastern Newark basin, the same Van Houten cycle apparently lacks black shale, instead having a purple shale with associated tan or white sandstones. The unit is still copper-mineralized, at least locally, and where intruded by thin diabase sills (2.2 mi) to the south-southeast in North Arlington (NJ), it was commercially exploited in what is supposed to be the oldest copper mine in North America – the Schuyler mine (Lewis, 1907). The exposures at this stop are almost certainly the prospect mentioned by Woodward (1944) on the Kingsland estate (located at - 40.8064893°, -74.1109759° http://geonames.usgs.gov/apex/f?p=gnispq:3:0::NO::P3_FID:877576), inspired by the Schuyler mine but never worked extensively. An exploratory shaft at least was opened, and the now-cemented entrance is still visible. Tan and white sandstones associated with purple and gray mudstone are exposed and mineralized with the same minerals as at the Schuyler mine, including chalcocite (black copper sulfide), chrysocolla (bluish-green copper silicate), malachite (green copper carbonate), and azurite (blue copper carbonate). It is noteworthy that there is no evidence of diabase at this location because the presence of an intrusive diabase sill has been used as an explanation for the copper mineralization at the Schuyler mine (e.g., Woodward, 1944).

The overall section at this stop consists of lower red massive mudstones of member T-U, followed by the tan and white sandstones surrounding a purple well-bedded mudstone of the basal part of the Kilmer Member. This is succeeded by massive red mudstones, a well-bedded interval, and then red mudstones and fine sandstones with gypsum nodules. The overall stratigraphy is very similar to the expression of member TU and the Kilmer member in the NBCP cores we have seen at Stop 1.1.

A large collection of very well-preserved reptile footprints was made near here by Lawrence Blackbeer in the late 1960s (pers. comm., 1985; Olsen and Baird, 1986) (Fig. 32). Although the exact location was not recorded in detail, the lithology of the footprint slabs is consistent with the local expression of the Kilmer Member. This was confirmed by the discovery in 2002 of two well-preserved tracks during a field trip for IGCP 458 (Olsen, 2002) (by PEO and A.V. Hillebrandt,

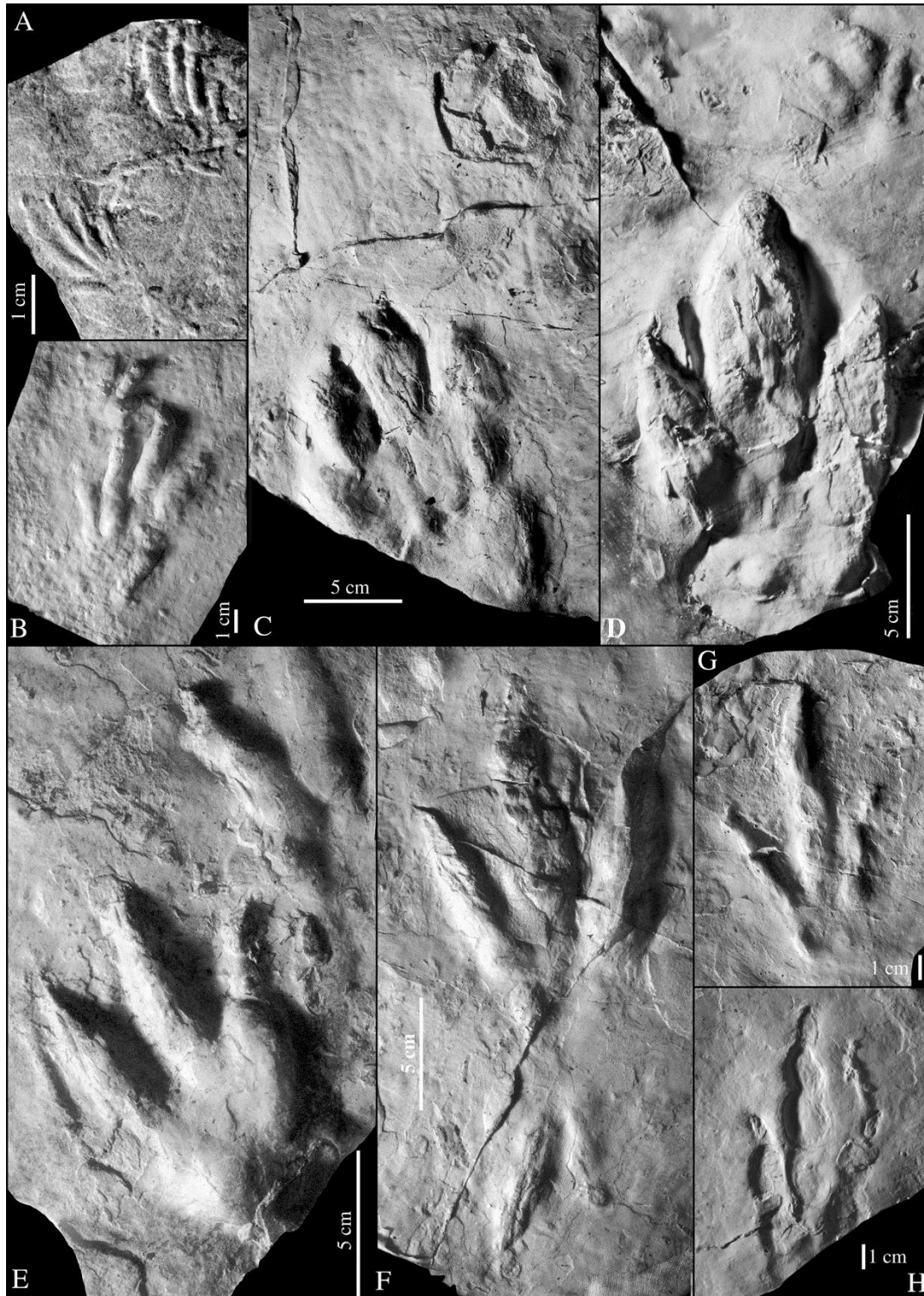


Figure 32: Footprints from Lyndhurst (Stop 1.6) or vicinity, Lawrence Blackbeer collection (all polysulfide casts): **A**, lepidosauromorph track *Rhynchosauroides* sp; **B**, probable suchian track *Brachychirotherium* sp.; **C**, probable suchian track *Brachychirotherium parvum*; **D**, silesaurian dinosauriform *Atreipus milfordensis*; **E**, ?saurischian dinosaurian track *Evozoum* sp.; **F**, theropod dinosaur tracks (brontozoid)

Anchisauripus sp. (above) and *Grallator* sp. (below); **G**, theropod dinosaur track (brontozoid) *Grallator* sp.; **H**, theropod dinosaur track (brontozoid) *Gral c.f lator* cf. *G. parallelus*.

June 8, 2002) in appropriate lithology from between the two prominent copper-bearing sandstones at this spot (Fig. 32).

This Norian age assemblage is distinguished in the Newark Basin by the presence of relatively large grallatorid (theropod dinosaur) footprints, up to the size of *Anchisauripus tuberosus*; this is the oldest level in the basin with such tracks. Relatively large examples of the plausibly non-dinosaurian dinosauriform (silesaurid?) track, *Atreipus milfordensis*, are present, along with a the possibly dinosauriform or poposaurid track *Evozoum* sp.⁴ (“*Coelurosaurichnus*” sp. of Olsen and Flynn, 1989), as well as crocodile-line pseudosuchian track *Brachychirotherium parvum*, and the small lepidosaurian track *Rhynchosauroides* spp. This track assemblage records the beginning of the rise to ecological dominance of the dinosaurs, which in lower horizons are conspicuous by their relative rarity and small size. This assemblage is similar to that recovered from the Passaic Formation of Rockland County (see Stop 2.2).

Return to vehicles.

Distance in miles (km)		Route Description
Cumu- lative	Point to Point	
32.0 (51.5)	2.6 (4.2)	Return to vehicles and proceed to exit and turn left onto Polito Avenue. Proceed northbound to ramp for NJ 3 on right.
39.1 (62.9)	0.9 (1.4)	Take NJ 3 to merge with US Route 46, keeping right, take the exit immediately right for Valley Road.
46.2 (74.3)	0.2 (0.3)	Take ramp for Valley road keeping right and turn right onto Valley Road.
46.7 (75.2)	0.5 (0.9)	Proceed north along Valley Road (County Road 621) and turn left onto Four Seasons Boulevard.
47.0 (75.6)	0.3 (0.5)	Follow Four Seasons Boulevard and park in front of the “Club House”.

STOP 1.7: ETE and E23r in uppermost Passaic Formation and Overlying Initial CAMP Basalt. Former Tilcon Quarry, Woodland Park, New Jersey⁵

Location Coordinates: 40.875964°, -74.188455°.

Duration: 0:45 hr.

Leave vehicles and proceed to view the adjacent exposures. The following is modified from Olsen et al. (2004).

The “Four Seasons at Great Notch Spa and Club” (K. Hovnanian) “adult community” occupies the site of a quarry that produced a very large amount of important stratigraphic, sedimentological, and

⁴ Lockley & Lucas (2013) figured two polyurethane molds of PEO as *Evazoum* sp. (their Fig. 1A1, 1B1) but list the locality incorrectly as “from the Upper Triassic Wolfville Formation, Nova Scotia”, when in fact they are from the Passaic Formation of New Jersey.

⁵ It is possible that there may be no available exposures at this site at the time of our field trip. If that is the case we will visit an alternative exposure nearby possibly near Garrett Rock or Lambert Castle (40.899848°, -74.172896°).

paleobiological information, especially during the last few years as construction proceeded for the development. The quarry was owned and run by a series of operators and is therefore known by various names, including: Little Ferry Asphalt, Union Building and Construction (UBC) Corporation, R.A. Hamilton, Dell Materials, and most recently Tilcon (the last operator). It has also been referred to as the “Clifton Quarry” and West Patterson Quarry (Gallagher & Hanczaryk, 2006).

The end Triassic extinction has left a rich, albeit asymmetrical, record in the northern Newark Basin. While most of the biological record below the ETE is from the southwestern Newark Basin, that above the initial ETE is from the northern part. Fortunately, the very short (~10m, <20ka) interval of reverse magnetic polarity lies in close proximity under the ETE allowing correlation that is independent of biostratigraphy (although in agreement with it).

Two main localities have provided nearly all of the biotic data in the Northern Newark Basin, both of which are now-abandoned quarries incorporated into Montclair State University and the now-abandoned quarry (this stop). The Montclair State locality (40.868368°, -74.194389° and vicinity) produced a good collection of footprints in the 1970s into the early 1980s with some excellent examples being represented by actual specimens or latex molds ending up in the Princeton collection at Yale, Peabody Museum (New Haven, CT) or the Donald Baird mold collection, now with PEO (the latter badly in need of archiving). Overwhelmingly dominant are brontozoid tracks, including a full size range from *Grallator* to *Eubrontes giganteus*. *Batrachopus cf. deweyii* is the only other form recognized. The Woodland Park quarry site, described by Olsen et al. (2004) and Gallagher & Hanczaryk (2006), produced large numbers of a full size range brontozoid dinosaur tracks including *Eubrontes giganteus*, several with unique preservation styles, abundant *Batrachopus deweyii* tracks, and the lizard-like track, *Rhynchosauroides* sp. Possibly thousands of tracks from this site have been examined by amateur collectors, with many finding their way into museum collections, notably the Princeton collection at Yale, the New Jersey State Museum (Trenton, NJ), and the Morris Museum (Morristown, NJ). Despite the very large numbers of tracks this site has produced there have been no examples of any other footprint taxa, nor any typical Triassic forms, some of which have been found nearby in lower strata (e.g., Baird, 1986) in Essex County.

As it was in 2011, when Tilcon was winding down its activity, the quarry exposes about 50 m of uppermost Passaic Formation and most of the 55 m thick lowest of three major flows of the 150 m thick Orange Mountain Basalt (Figs. 33 and 34). In this area, the uppermost Passaic Formation consists of two units of strongly contrasting facies: a lower interval of fluvial facies, and an upper unit of marginal lacustrine facies. The bulk of the upper Passaic in this region consists of upward fining cycles of relatively poorly sorted pale red conglomerate and pebbly sandstone with poorly defined trough cross bedding grading upward into massive red mudstones and sandstones (described by Parker et al., 1988). The units are intensely bioturbated, which is what obscures the bedding. This facies is overlain by red and gray mudstones and sandstones with more distinct bedding and excellent preservation of small-scale sedimentary structures, which are more heterogeneous than underlying units. There are cross-laminated sandstones with channel morphologies, tilted thin beds that toe laterally into mudstones suggestive of small deltas or crevasse splays, tabular beds of climbing ripple cross-lamination, and thin bedded mudstone beds suggestive of suspension deposits in standing water. Many sandstone beds are bound by clay drapes, and many surfaces are covered by ripples, desiccation cracks, and trace fossils, notably reptile footprints. The uppermost surface of this interval is covered by the Orange Mountain basalt and locally seems to have preserved some depositional relief, including small channels (Fig. 33).

Vertebrate footprint assemblages (Fig. 35) and floral remains from the upper facies completely lack all Triassic forms, indicating that the interval postdates the ETE.



Figure 33: Stop 1.7 as it appeared in 2004. North wall of quarry showing the lower flow of the Orange Mountain Basalt resting on Passaic Formation containing the ETE level and polarity zone E23r. Double arrow incated footprint-bearing levels in sandstone-and mudstone facies, a few meters below which is E23r in mixed pebbly sandstone-minor mudstone facies.

The thin reverse polarity interval E23r was identified in the north Quarry wall at about 16 m below the Orange Mountain Basalt in the coarser, fluvial, conglomeritic facies (Fig. 34) that is similar to normal Passaic formation in this area (Parker et al., 1988).

The coring transect of the Army Corps of Engineers (ACE) Passaic River Diversionary Tunnel Project passed through this area, and a series of short cores (<200 m) were recovered through the uppermost Passaic and lower Orange Mountain Basalt (Fedosh and Smoot, 1988; Tollo & Gottfried, 1992; Olsen et al, 1996a; Olsen et al., 2002). All of them show the same facies transition as seen at this stop. Reverse polarity zone E23r has been identified in these cores at the appropriate levels as well (e.g., ACE core PT-38: Stop 1.1).

The floral assemblage from discontinuous pods of gray mudstone and sandstone at this stop include abundant remains of *Brachyphyllum*, a cheirolepidiaceous conifer and less common fragments of the dipteridaceous fern *Clathropteris meniscoides*, a large variety of casts of stems and rhizoliths and poorly preserved pollen mostly of the genus *Classopollis*. This is a typical post-ETE assemblage very similar to that in the overlying Feltville Formation. The fern *Clathropteris meniscoides* is the source of the spores at the spore spike in the southwestern part of the Newark basin.

A quite remarkable feature of the assemblage from this site is that it records the appearance of the first truly large theropod dinosaurs, and increase in size (length) of over 20% that occurs at or just after the extinction level. A 20% increase in track length should scale to a doubling of mass, and thus this represents a very significant change in the top predators. As spelled out by Olsen et al. (2002), there are two obvious possible scenarios to explain the abrupt increase in size in theropod dinosaurs across the Triassic-Jurassic boundary in eastern North America (and globally). First, that the appearance of the much larger theropods represents a dispersal event from some unknown location, or second, that it

represents an evolutionary event. Thus far, we favor the second, evolutionary hypothesis for the appearance of large theropods and suggest that the abrupt increase in size is most easily explained by a sudden evolutionary response of the theropod survivors (which may have been quite small) to ecological release, operating at time scales of thousands of years. We hypothesize a response similar to that inferred for reptiles on modern islands lacking competitors. This evolutionary response hypothesis could be falsified by the discovery of equivalently large theropod bones or diagnostic *Eubrontes giganteus* (emphasis on the ichnospecies is important) tracks in unquestionably pre-ETE Triassic strata. Thus far these are wanting, contrary to assertions of, for example, Thulborn (2003), Lucas et al. (2006), and Lucas & Tanner (2008). In any case, this track assemblage suggests that the dramatic drop in non-dinosaurian diversity was caused by an extrinsic environmental catastrophe such as the CAMP, and the resulting drop in competitive pressure was the trigger for the global *spread* of large theropods. Sauropodomorph dinosaurs also survived the ETE, joining theropod and ornithischian dinosaurs to establish the familiar dinosaurian-dominated ecological pattern that dominated the terrestrial world for the next 135 million years.

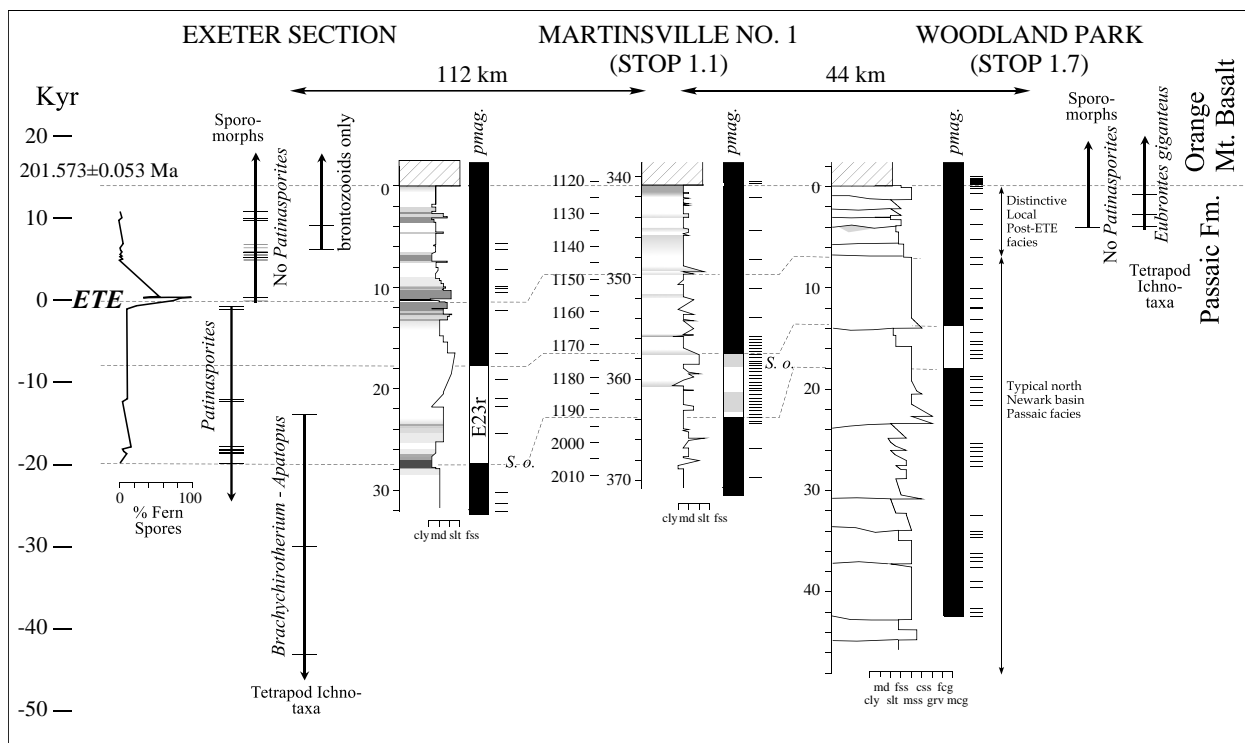


Figure 34: Physical, astro-, bio-, and magnetic polarity stratigraphy of three key sections around the ETE in the Newark Basin. Note the close correspondence with different lines of evidence. Black is normal polarity and white is reverse. *S. o.*, marks occurrences of the spinocaudatan *Shipingia olsenii*

The Orange Mountain basalt overlies the Passaic Formation and is a high-Ti quartz normative tholeiite (Puffer and Lechler, 1980; Tollo & Gottfried, 1992), a basalt type extremely widespread

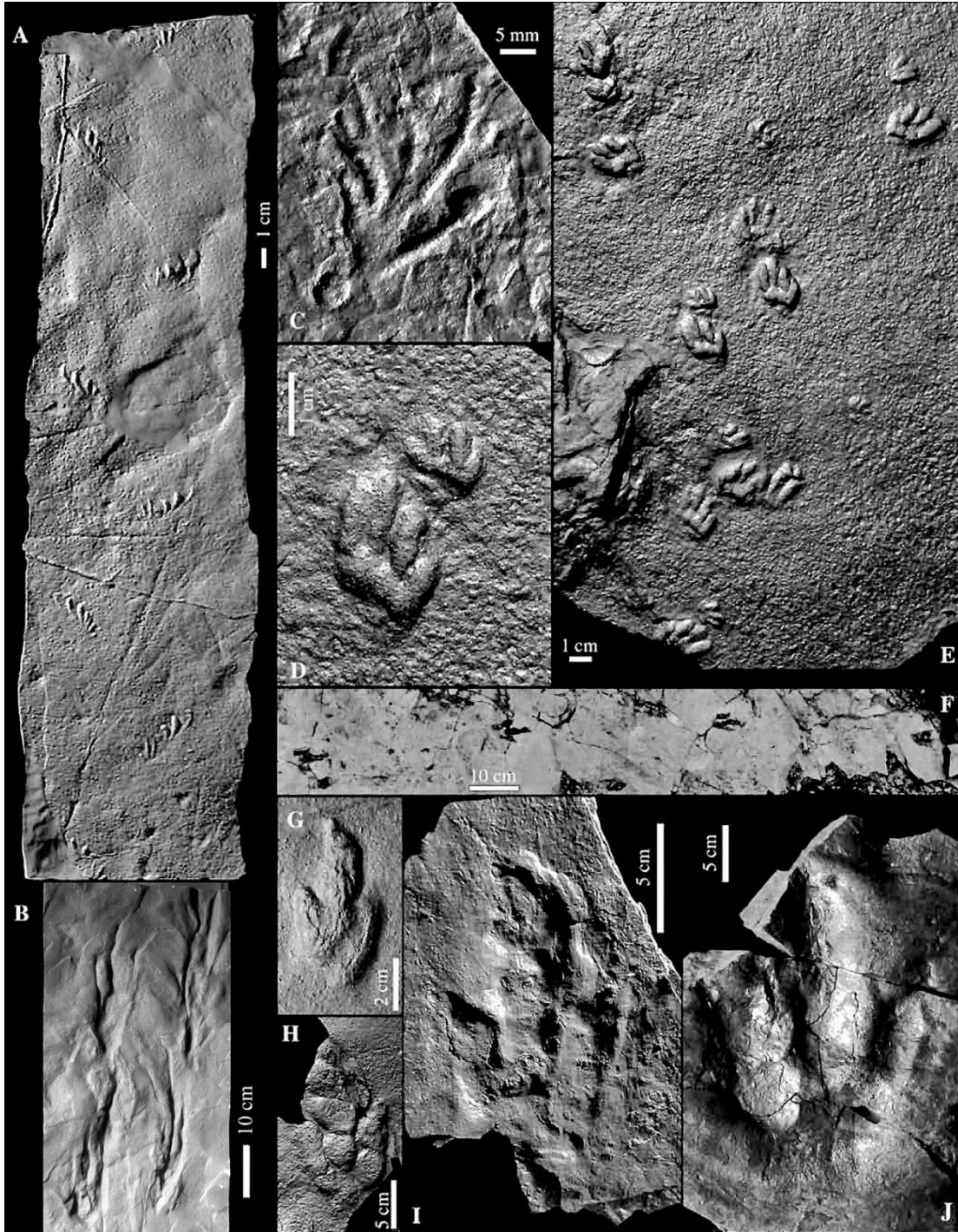


Figure 35: Post-EETE (latest Rhaetian) reptile footprints from the uppermost Passaic Formation at Stop 1.7: A, C, *Rhynchosauroides* n. sp.; B, swimming brontozoid tracks; D-E, *Batrachopus deweyii*; F, trackway of medium-sized brontozoid *Anchisauripus sillimani*; G, smaller brontozoid *Gallator* sp.; H, medium-sized brontozoid *Anchisauripus* sp.; I, medium brontozoid, *Gallator tuberosus*; J, largest and characteristic post-EETE brontozoid, *Eubrantes giganteus*.

in the CAMP as the initial flow type. While it is the oldest flow sequence in the Newark Basin, slightly older flows (by thousands to tens of thousands of years), occur in Nova Scotia and Morocco (Blackburn et al., 2013). Based on the connection of the zircon-bearing Palisade intrusion to the Orange Mountain Basalt flows in Rockland County, the ^{238}U - ^{206}Pb age of the basalt is 201.520 ± 0.034 Ma and that of the ETE is 201.564 ± 0.015 Ma based on the zircon ages and astrochronology (Blackburn et al., 2013).

Distance in miles (km)		
Cumu- lative	Point to Point	Route Description
47.3 (76.1)	0.3 (0.5)	Return to vehicles and proceed to exit and turn left onto Valley Road (Passaic County Road 621).
49.3 (79.3)	2.0 (3.2)	Proceed north on Valley Road (Passaic County Road 621) to left turn onto New Jersey Route 19.
50.5 (81.3)	1.2 (2.0)	Proceed on New Jersey Route 19 to end and turn left onto Main Street, Patterson, NJ.
50.8 (81.8)	0.3 (0.5)	Follow Main Street north to left-right dogleg via Broadway to West Broadway (Passaic County Road 673).
51.3 (82.6)	0.5 (0.8)	Proceed on West Broadway to right turn onto Belmont Avenue (Passaic County Road 675).
53.4 (86.0)	2.1 (3.4)	Follow Belmont Avenue north to West Overlook Avenue, North Haledon, NJ and park in strip mall on left.

STOP 1.8: Upper Feltville Formation and lower Preakness Basalt, William Paterson University.

Location Coordinates: 40.951754° , -74.189353° (parking); 40.951150° , -74.192169° (quarry)

Duration: 1:00 h.

Leave vehicles at strip mall and walk about ~145 yards (~133 m) and enter woods down to creek). Proceed upstream ~130 yards (~ 119 m) observing outcrops of upper Feltville Formation to abandoned quarry in upper Feltville and basal Preakness Basalt. The following is modified from Olsen et al. (2003).

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 (Fig. 36). 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, conifer fragments, and wood. In terms of the cyclostratigraphy of the Feltville, this sequence is the interval of maximum precessional variability in the 100 ka cycle exhibiting the lowest precessional variability within its 405 ka cycle. The second flow of the Preakness has a ^{238}U - ^{206}Pb age of 201.274 ± 0.032 Ma, which is in complete agreement with its cyclostratigraphic age (Blackburn et al., 2013).

The basal flow of the Preakness as exposed here consists of a pillowed basalt complex and 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 near the border fault. It seems to be absent from at least West Orange (I-280) to near Somerville, NJ, but reappears along I-287 at Pluckemin, NJ. Olsen (1980a; 1980b) described outcrops of the lower flow but did not differentiate it from the overlying flow P-2 of Tollo & Gottfried (1992). At this locality there is little 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.



Figure 36: Preakness Basalt at Stop 1.8. Left, contact between upper Feltville Formation and pillowed and curvilinear flow 1 of the Preakness Basalt. Right, contact between flow 1 and 2 of the Preakness Basalt – John Puffer for scale.

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), 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 one of the largest lava flows known on Earth.

In terms of environmental effects, rate as well as magnitude matters, and the eruption of this flow, if it was indeed one eruption, would have been among the largest flows known on Earth in terms of volume, and would have had significant global environmental effects. The largest single flow of the Columbia River Basalt is on the order of 5000 km³ (Tolan et al., 1989), but if the Preakness and its 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 km³. 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. The well-developed, platy-prismatic fracture in Flow P-2 is very well displayed here and could be very significant for sequestration. As previously discussed, Goldberg et al. (2009) argue that carbonation reaction in permeable zones of

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basalt could provide a significant locus for carbon sequestration. However, while there is a large amount of porosity and permeability obvious here, it has yet to be demonstrated that it exists at depth in these ancient basalts.

Return to vehicles.

Distance in miles (km)		
Cumu- lative	Point to Point	Route Description
54.7 (88.1)	1.3 (2.1)	Proceed to exit onto West Overlook Avenue eastbound and turn left onto Belmont Avenue (Passaic County Road 675), heading north, and at end turn slightly left onto High Mountain Road.
55.6 (89.5)	0.9 (1.4)	Proceed on High Mountain Road and turn right onto Ewing Avenue.
57.6 (92.7)	2.0 (3.2)	Follow Ewing Avenue north and turn left onto ramp for New Jersey Route 208 N.
57.8 (93.1)	0.2 (0.3)	Take ramp and merge onto New Jersey Route 208 N
59.2 (95.3)	1.4 (2.3)	Proceed on New Jersey Route 208 N to merge onto US I-287 N.
59.6 (96.0)	0.4 (0.6)	Follow merge onto US Interstate I-287 N.
67.0 (107.9)	7.4 (11.9)	Proceed on I-287 N and rise the right 2 lanes to merge onto I-287 E/I-87 S toward Tappan Zee Br/New York City.
74.7 (120.3)	7.7 (12.4)	Follow I-287 E to take exit 14 on right for New York Route 59 toward Spring Valley/Nanuet.
74.9 (120.6)	0.2 (0.3)	Follow ramp keeping left, then turn left onto NY-59 East.
76.6 (123.3)	1.7 (2.7)	Follow NY-59 E to right hand turn to entrance for Double Tree, Nanuet.

End of Day 1.

DAY 2

Meeting Point: Parking Lot of DoubleTree by Hilton Hotel, 425 State Route 59, Nanuet, New York, 10954.
Access is from the east-bound lanes of State Route 59.

Meeting Point Coordinates: 41.090694°N, 73.995438°W

Meeting Time: 8:00 AM (Both Days)

Distance in miles (km)		
Cumu- Lative	Point to Point	Route Description
0.0 (0.0)	0.0 (0.0)	Assemble in the parking lot of the DoubleTree, Nanuet. Leave Parking lot, turning right at entrance on to eastbound State Route 59, keep in right lane
0.1 (0.2)	0.1 (0.2)	Get on entrance ramp to Palisades Interstate Parkway South from NY-59 E
0.3 (0.5)	0.2 (0.3)	Merge onto Palisades Interstate Parkway South
8.4 (13.5)	8.1 (13.0)	Follow Palisades Interstate Parkway South to US-9W N/N Rte 9W N in Alpine. Take exit 4 from Palisades Interstate Parkway South
8.5 (13.7)	0.1 (0.2)	Turn right onto US-9W northbound.

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9.4 (15.1)	0.9 (1.5)	Pass traffic light and turn right into campus of Lamont-Doherty Earth Observatory of Columbia University just before New York-New Jersey State line (Ludlow Lane).
9.7 (15.6)	0.3 (0.5)	Proceed to the front of the Geoscience building. Park and you will be guided on foot to the core repository.

STOP 2.1 - Cores, Seismic Lines and Posters at Lamont-Doherty Earth Observatory, Palisades, NY

Location Coordinates: 41.004471°N, 73.908982°W: Core repository Laboratory, Geoscience Building, Lamont-Doherty Earth Observatory.

Duration: 1:00 hr

Here we will examine cores, down-hole logs, seismic profiles, posters, and some fossils mostly from the Rockland portion of the Newark Basin near the northern terminus of the basin. These exhibits are meant to compliment the Stops of Day 1 and are related directly to the new seismic profiles across the basin in Rockland. These will relate directly to the outcrops will see after this stop.

Core TW4 on the Lamont Campus

We will examine the following features of Core TW4: 1) Diabase and xenolith; 2) Possible Lockatong Formation; 3) Nominal Stockton Formation; 4) High gamma sandstones without organics; 5) Gypsum-bearing red beds; Odd purple interval; Contact with Fordham Gneiss.

Core from the Tandem Lot

Metamorphosed lower Passaic Formation: Approximately 150 ft of core was collected from above and close to the Palisade Sill. Core is characterized by the presence of metamorphic minerals, most obviously epidote, and by peculiar colors, such as green and various shades purple and orange (Slater et al., 2012). Anhydrite, quartz, calcite, and feldspar cements are present. Sedimentary fabrics suggest pedogenic nodular carbonates and fluvial or marginal lacustrine environments.

Cuttings and side wall cores from the Tandem Lot

Compare these cuttings and sidewall cores with the equivalent footages on the logs and with the seismic line to get a comprehensive perspective on the 6880 ft Tandem Lot drill hole.

Passaic Formation: Most Passaic formation washed cuttings look like they are nearly 100% sandstone because the mudstones wash out. We will show examples of the comparison between cuttings and sidewall cores.

Palisades Sill: Cuttings displayed will be from: 1) the upper part of the sill near the contact with the Passaic Formation; 2) the high gamma part of the sill plausibly from the zircon-bearing “sandwich horizon”; 3) from the middle part of the sill; 2) and the lower part of the sill, comparable to what is seen in the TW4 core.

Lockatong Formation: The cuttings are easily mistaken for gneiss, but the sidewall cores and FMI logs show that the section below the sill is similar to what we saw at Stops 1.3 and 1.5 consisting of alternating gray and dark mudstones and tan sandstones.

Fossils from the Passaic Formation of Rockland County

Fossils from the old Manhattan Trap Rock Company Quarry at Nyack Beach State Park (Stop 2.2): Large tetrapod, probably reptile burrows; a small theropod dinosaur (brontozoid) track, *Grallator* sp.; *Scoyenia*

(arthropod) burrows; and rhizoliths (root traces). The *Grallator* sp. track may be the only definitive dinosaur footprint from New York State.

Fossils from Snedeker's (or Waldberg) Landing, Haverstraw (41.173200°, -73.934633°): A small sandstone quarry located south of Haverstraw Beach State Park has produced a series of quite good and diverse fossils over the decades. On display will be the tetrapod trace fossils and fish scrap shown in Fig. 37, as well as additional *Cynodontipus* burrows, small very clear *Brachychirotherium* sp. tracks, and complex rhizomorphs (root traces).

Distance in miles (km)		
Cumu- lative	Point to Point	Route Description
10.0 (15.6)	0.3 (0.5)	Return to vehicles and proceed to exit. Passing TW4 core site at left at (41.002829°, -73.910658°). Turn right onto Route 9W northbound.
15.5 (24.9)	5.5 (8.9)	Proceed north to slight right turn onto Broadway, South Nyack.
19.0 (30.6)	3.5 (5.6)	Proceed north on Broadway though Nyack to end of Broadway and entrance to Nyack Beach State Park, keeping right entering park. Keep right, driving down to parking area near beach and Park.

STOP 2.2: Passaic Formation and Palisade Sill, Old Manhattan trap Rock Quarry, Nyack Beach State Park, Upper Nyack, NY.

Location Coordinates: 41.120639°, -73.911824° (parking, Station A); 41.121467°, -73.912095° (sandstone quarry, Station B); 41.121902°, -73.912009° (trap rock quarry, Station C).

Duration: 2:30 hr

Park vehicles at lot near beach on the Hudson and walk toward the brownstone Dutch colonial-looking “bathhouse” and the adjacent outcrops of Passaic Formation.

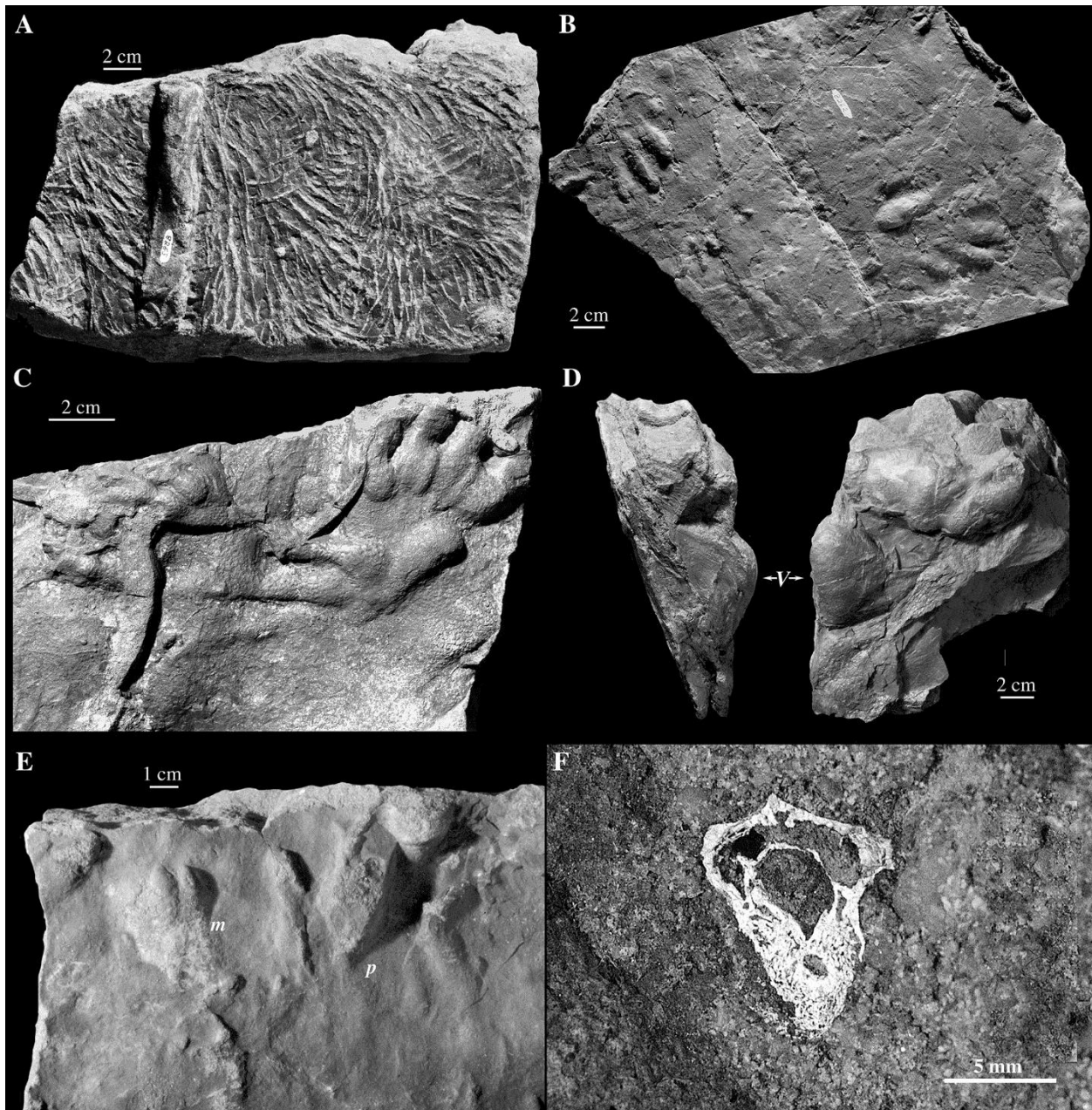


Figure 37: Fossils from near Haverstraw beach, Haverstraw, New York: A, the terminal marks of the tetrapod burrow *Cynodontipus* (YPM 8263); B, natural cast of partial trackway of *Brachychirotherium* cf. *B. eyermani* (YPM 8262); C, natural cast of manus-pes set of *Apatopus lineatus* and partial pes of *Brachychirotherium* cf. *B. eyermani* (YPM 7731); D, very deep natural cast of *Brachychirotherium* cf. *B. parvum* (AMNH uncataloged), V indicates digit V (five); E, natural cast of poor manus (m) and pes (p) set of ?*Atreipus* sp. (YPM 8553); F, partial *Semionotus* braincase (AMNH uncataloged) thought originally to be a tetrapod.

Nyack Beach State Park, is the southeastern component of a series of contiguous state parks including Hook Mountain State Park, Rockland Lake State Park, and Haverstraw Beach State Park all contained within Palisades Interstate Park. It hard to see it now, but this park was a huge active trap rock quarry in the nineteenth and early twentieth century (Fig. 38). Quarrying

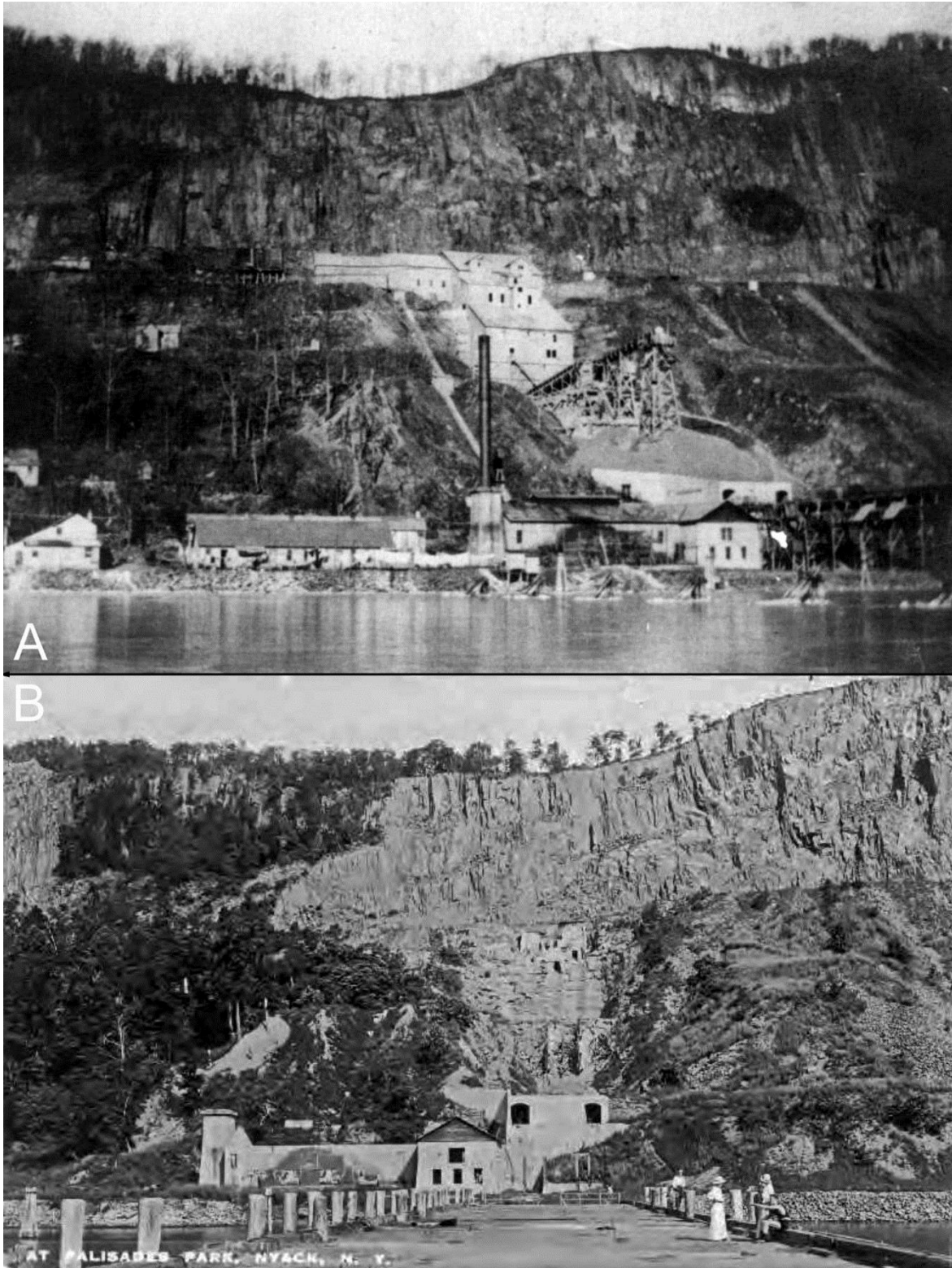


Figure 38: View of Nyack Beach and Hook Mountain state parks. A, is the Manhattan Trap Rock Company quarry and associated structures in full operation ca. 1909. B is the same area after 1912. Both from Nyack Public Library.

may have begun for the red sandstone at this site by the middle 18th century and the quarry at Station B may have been part of the string of at least 30 (!) quarries extending from Upper Nyack to Piermont at the peak of the operations in the middle 19th century. The “Voorhis” Quarry at this site may have begun as one of these, but by the later 19th century, the quarry was focused on the diabase instead; the production of a higher quality “brownstone” had shifted to the Connecticut Valley (mostly of Jurassic age). In 1872, a huge stone crusher complex was built, eventually extending from the trap rock quarry floor to the river, filling the sandstone quarry as part of the Manhattan Trap Rock Company (OPRHP, 2013). The blasted diabase was dumped from the quarry floor (Station C) into the steam-engine-powered-crusher and sorting system and conveyed via belts down to a cement structure (still extant) on the north side of the brownstone powerhouse and then out on a long raised double pier where the crushed stone was loaded onto freighters on the Hudson (Fig. 38A). By the 1890s, the destruction of the Palisades cliffs and the associated blasting and crushing noise at Hook Mountain and elsewhere had become a conservation cause. In 1906, the authority of the Palisades Interstate Park Commission (the impetus for which was led by The New Jersey Federation of Women’s Clubs) was extended by legislative action to Hook Mountain and north, and by contributions from members of the Rockefeller (the estate – Kykuit - of which was opposite the quarry), Harriman, and Perkins families, the Hook Mountain area and its quarries were purchased by the commission – ending quarrying by 1912. Many buildings of the complex were removed as was the rock crushing equipment, making the sandstone quarry again accessible (Fig. 38B). The Work Progress Administration (WPA) repurposed and rehabbed the powerhouse in 1936 as a bathhouse with a large fireplace and ballroom (OPRHP, 2013), which is the structure visible today at Station A.

The most extensive description of the geological aspects of the quarry was made by Kümmell (1900) when the Manhattan Trap Rock Company was operating at full-bore. Kümmell (pp. 27-28) notes: “At the time of my visit, the trap at the south end of the quarry could be seen cutting across the sandstone obliquely downward to the north, a distance of about 35 feet in 60. Thence it followed conformably the bedding planes and gradually regained its former elevation above sea level, though still at a lower geologic horizon. Its general elevation at the quarry is about 260 feet A. T. If the base of the trap kept to the same horizon which it has where it turns so abruptly eastward at Verdrietege Hook, the rise of the underlying sandstones up the dip would carry it from 620 to 6601 feet higher above sea level than it actually is at the quarry. The evidence seems conclusive therefore that the abrupt turn at Hook mountain north of Nyack, is due to the descent of the base of the sheet to a geologic horizon 600 or 700 feet lower than that occupied where it crosses the Rockland Lake-Nyack road.” These observations are very important because they show that the crescentic offset of the Palisade ridge is not due to an anticline as sometimes presumed, but rather by the change in stratigraphic level intruded. Unfortunately, very few other detailed observations have been published on the extensive remaining aspects of the site, except for Savage (1968), and in fact the large sandstone quarry at Station B, seems to have escaped geologists’ attention entirely until brought to the attention by Dr. William Menke (Columbia University) who found it while hiking. There are no published observations on the paleontology for the Passaic Formation at this site.



Figure 39: Fossils from the former Manhattan Trap Rock Quarry, Nyack Beach State Park. From left to right they are: small brontozoid theropod dinosaur track, *Grallator* sp.; large tetrapod burrow; complex bone fragment; bifurcating rhizomorph (root trace).

Station A, 41.120639°, -73.911824°: The handsome brownstone “bathhouse” is the repurposed part of the Manhattan Trap Rock Company complex of buildings. The tall brownstone tower-like structure on the south side of the building was the base of a large smokestack of the steam powerhouse of the plant (anonymous, 2011). We presume the sandstone for this building came from the quarry at Stop 2.2, Station B (below). Exposures behind the “bathhouse” are the lowest in this area and consist of a complex of bioturbated mudstones, micaceous reddish-brown sandstone, and some greenish-gray arkose (Savage, 1964). Some surfaces have desiccation cracks and less disrupted clay bedding surfaces that could have tetrapod tracks. Carbonate nodules are present and carbonate and mud pellet intraformational conglomerates are present at the bases of some of the sandstones. According to Savage (1964) the paleocurrents are towards the northeast here, characteristic of many exposures in the Rockland part of the Newark Basin where they generally have easterly directions. This is distinctly different, as is the mineralogy, from the Stockton Formation as seen at Snedens landing where the paleocurrents are largely towards the west (Savage, 1964; Olsen, 1980b). Completely absent at this and the other exposures at this site are the red-purple massive mudstones with the very large ptygmatic arkose-filled desiccation cracks seen in TW4 core and exposures at Stop 2.2, consistent with this section being upsection from those near the state-line.

From Station A walk 112 m SW along entry road to end of wall on right side of road. Turn northeast and walk up macadam path 39 m to entrance to trail at 41.120394°, -73.912492°. Walk 54 m northeast and turn sharp left up switchback trail. Walk 57 m southwest to right turn. Walk 90 m northeast to path to quarry. Go 41 m northeast into middle of old sandstone quarry.

Station B, 41.121467°, -73.912095°: This sandstone quarry was previously filled by the crusher complex of the Manhattan Trap Rock Company of the late 19th and early 20th century and prior to that it may have been the Voorhis sandstone quarry. More than 50 m of red mudstone and red and some tan sandstone are exposed. Fossils are abundant in the rubble from the quarry along its edges and down the slope to the east. Beware of broken glass, etc. Fossils are common (Fig. 39). *Scoyenia* (arthropod) burrows and clay rhizoliths (root traces) are abundant, sometimes obliterating bedding. At least one

brontozoid dinosaur track, *Grallator* sp., has been found along with one bone fragment. But most striking are the large flattened burrows, 10 to 15 cm wide, that are surprisingly common (four visits with four finds). We identified similar flattened cylindrical features as large roots at Stop 2.3, but they differ from roots in maintaining a relatively constant diameter and not branching. Instead they are consistent with tetrapod burrows. Such burrows indicate that these strata have intervals when they were above the water table for significant amounts of time. It is possible that skeletal remains could be found in such burrows and one should be alert to that possibility when looking about.

The section has several laterally persistent intervals of thin bedded siltstone and sandstone suggestive of shallow water lacustrine intervals seen in the Passaic Formation near Milford, NJ., and it is possible these were deposited during the wet phases of climatic precessionally paced Van Houten cycles.

Looking west and up towards the top of the quarry, you can see the remains of the cement portals for chutes from the trap rock quarry itself. These can be seen in Fig. 38B. The adjacent mudstones have a gray or purplish tint and the sandstones are large tan. We will see this again at Station C and the colors might be most simply explained by metamorphism approaching the sill contact.

Time permitting, we will leave the sandstone quarry and continue up the trail to the road up to the upper parking area.

Return 90 m southwest from middle of quarry and then 138 m southwest to switchback turn. Turn right and go 119 m to chair at last switchback turn before road. Turn left and go 98 m to entrance onto road. Turn right and walk northeast up road to parking area.

Station C, 41.121883°, -73.912208°: The upper parking area at Nyack Beach State Park occupies the former floor of the main part of the Manhattan Trap Rock Company Quarry. Observe jointing in sill: is this columnar jointing? At the far northern end of the quarry a “rotten” zone of diabase typical of the olivine zone of the sheet (Walker, 1969; Steiner, 1989).

Walk 90 m southwest to electrical service boxes. On right (northwest) 26 m is a continuous exposure on a promontory of metamorphosed Passaic Formation to the oblique, down to the west, contact with the Palisade Sill. Exposure is at 41.121486°, -73.913278°.

From the electrical boxes, proceed 44 m southwest to stone bathroom. Most of the blocks of this building are metamorphosed Passaic Formation. Many blocks exhibit nodular zones with epidote, presumably replacing carbonates. Walk back to electrical service boxes and go right down steps to road.

Walk 316 m southwest to guardhouse at entrance to park. Turn left and walk 353 m back to the parking area.

Distance in miles (km)

Cumu- lative	Point to Point	Route Description
19.3 (31.1)	0.3 (0.5)	Leave Stop 2.2 onto N Broadway and turn right onto Larchdale Ave.
19.5 (31.4)	0.2 (0.3)	Continue on Larchdale and turn left onto N Midland Ave.
20.2 (32.5)	0.7 (1.1)	Continue on N Midland and turn right onto Christian Herald Rd.
20.3 (32.7)	0.1 (0.2)	Proceed on Christian Herald Rd and turn left onto N Highland Ave (US Rt. 9 S).
21.7 (34.9)	1.4 (2.3)	Follow N Highland Ave (US-9 S) to slight right onto S Highland Rd.
21.9 (35.2)	0.2 (0.3)	Take S Highland Rd and turn right onto Highland Ave.

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22.1 (35.6)	0.2 (0.3)	Follow Highland Ave and turn right onto Bradley Parkway.
23.4 (37.7)	1.3 (2.1)	Continue on Bradley Parkway and turn right onto N Greenbush Rd. Park on left near fence for Stop 2.3.

STOP 2.3: Passaic Formation above Palisade Sill, large bedding surfaces with footprints and large burrows, Blauvelt, NY.

Location Coordinates: 41.085697°, -73.946738° (parking); 41.084403°, -73.946810° (main exposure of bedding planes).

Duration: 2:00 hr (Lunch)

Park vehicles on unpaved road and proceed 294 m south along N Greenbush Rd to break in vegetation at 41.083084°, -73.946247°. Proceed southwest 34 m at 232° (true N) to 41.084403°, -73.946810°, adjacent to large bedding surfaces. Proceed 54 m due west to a point (41.083117°, -73.946887°) on the bedding plane surface near the “discovery site” for the area.

Reptile footprints were originally found during reconnaissance of excavated rock surface by Paul Olsen and Robert F. Salvia in fall of 1972. Fossils were removed shortly thereafter and deposited in the State Museum of New York, in the case of footprints, and Yale University, in the case of bone and teeth fragments. After initial discovery and salvage, no more material was found until the preparation of this report. Excavation of the site for fill occurred prior to and had ceased by the time of the fossil discovery and since that time the site has lain fallow and is now partly covered by vegetation. However, much bedrock remains exposed.

The site consists of series of large bedding plane (dip slope) exposures comprised of variegated (gray, tan, purplish, greenish, and red) slightly metamorphosed mudstone and siltstone and arkosic sandstone of the Passaic Formation. The stratigraphic level may be correlative with one of the gray sequences further to the southwest within the basin.

The Palisades Sill outcrops to the east, north, and south of the site. From the attitude of the outcropping adjacent diabase on north, south, and east sides, it appears to be strongly discordant, cutting through the Passaic Formation at a high angle. Judging from what is known elsewhere where there are similar contacts and based on the visible metamorphism of the bedrock, the Palisades diabase probably underlies the Passaic Formation at this site at some depth, probably more than 10 feet and probably less than 500 feet.

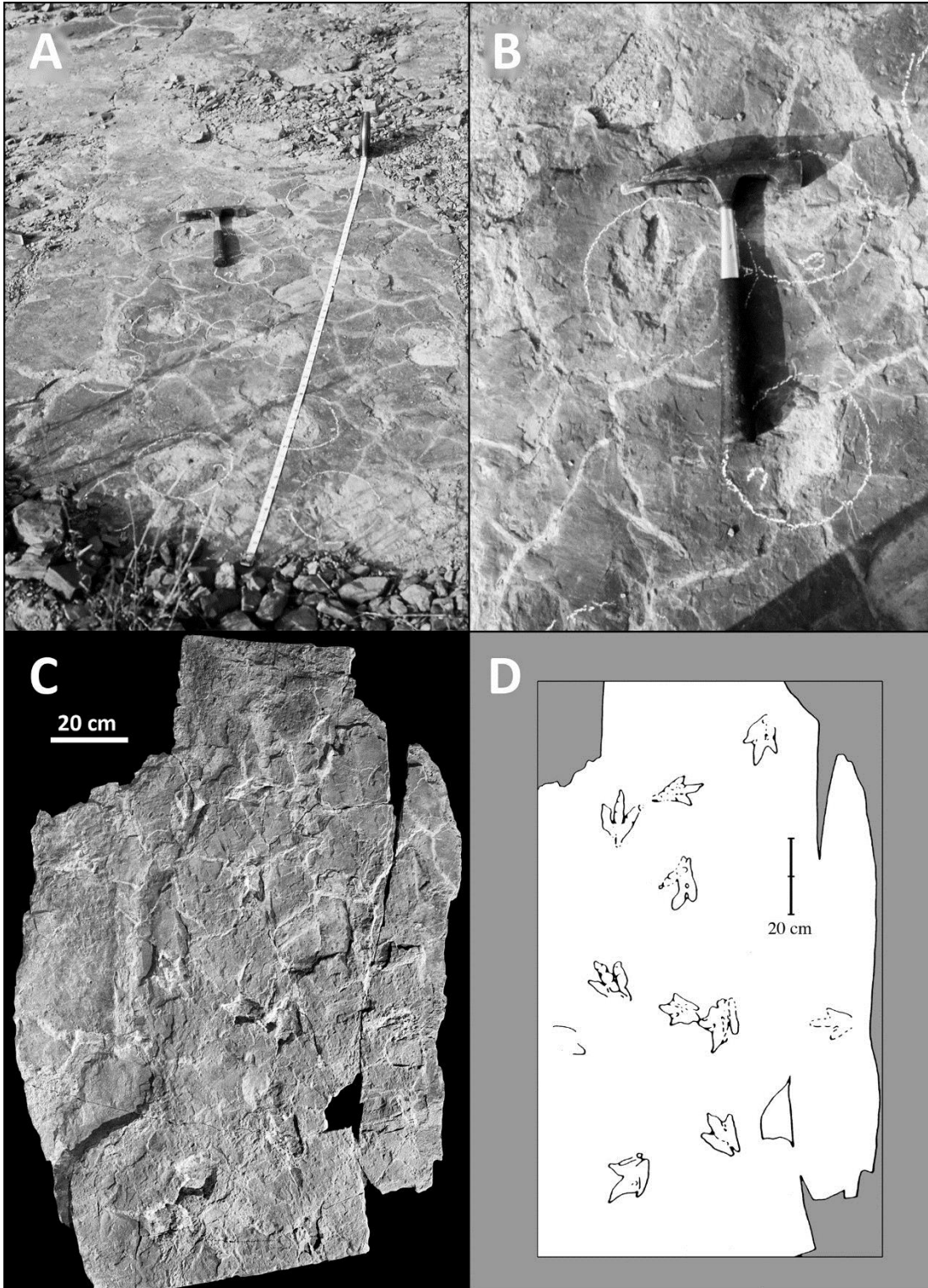


Figure 40: “Discovery slab” of trachs from Stop 2.3, Blauvelt , NY. A, Slab *in situ* prior to excavation. B, single dinosauriform track prior to excavation. C, after excavation preparation and montion at New York State Museum. D, drawing of same.

Four basic classes of fossils were found at the site: 1) reptile footprints; 2) tetrapod burrows; 3) possible teeth and bone scraps; and 4) rhizomorphs (natural casts of plant roots). Two areas you can see most of these features are centered at a north area (A: 41.083347°, -73.947186°) and a south area (B: 41.082287°, -73.946835°) with a 25 m radius.

1) Reptile Footprints: Reptile footprints were found in three places at the site. The first, “the discovery slab” was found in place with all of the tracks filled with white sandstone (Fig. 40) close to 41.083117°, -73.946887°. It was quarried out, taken to the New York State Museum and reassembled by the exhibition staff (briefly described by Olsen & Flynn, 1989). We (Paul Olsen and Robert F. Salvia) subsequently removed the white sandstone fill of the tracks. The slab in its present state reveal at least 12 three-toed footprints, each measuring about 12- 13 cm long (Fig. 40). The tracks have a shape typically associated with small dinosaurs and in fact they have been cited numerous times in the literature (e.g., Fisher, 1982) as the tracks of the small theropod dinosaur *Coelophysis*, which they are certainly not. Small coelophysid theropod dinosaurs such as *Coelophysis*, have a relatively long digit III (middle digit) on the foot producing bronozoid type tracks of the ichnogenus *Grallator*, but digit III on the tracks in the discovery slab is relatively short. On this basis alone, *Coelophysis* can be discounted as the track maker, even though coelophysid dinosaurs are known from strata both older and younger than this site in the Chinle Formation of the Southwestern United States (Irmis et al., 2011; Ramezani et al., 2014), and a small bronozoid was been found at stop 2.2 (Fig. 39).

The tracks on the discovery slab (Fig. 40) could belong to one of non-coelophysid dinosaurs known from the late Triassic, a non-dinosaurian dinosauriform, or even a crocodile line archosaur convergent on dinosaurian form. Non-coelophysid dinosaurs such as the herrersaurid *Chindesaurus* do occur in North America (Irmis et al., 2007) and something like that could have made the track. However, only *Herrersaurus* proper has well preserved feet and these have a large digit 1 on the pes that should have impressed, but we do not see that in these tracks.

Atreipus (Olsen & Baird, 1986) is fair match to the footprints on the “discovery” slab in having a tulip-shaped hind track with a relatively short digit III for its size. Although *Atreipus* is usually found with forefoot impression, it is not rare to see trackways in which no forefoot impressions are present. Unfortunately, the tracks on the discovery slab are not well preserved enough for a positive identification. In any case, *Atreipus* can now, with some confidence, be assigned to a recently discovered group of non-dinosaurian dinosauriforms, the silesaurs (such as *Silesaurus*). It is possible that poposaurs, a group of crocodile-relatives (suchian pseudosuchians) that evolved bipedal very dinosaur-like forms such as *Effigia okeeffeae* (Nestbitt, 2007), could have made these track, but *Effigia* too has a large digit 1 and so cannot be simply hypothesized to have made the discovery slab tracks. In summary, the tracks were not made by *Coelophysis* and may not have been made by a dinosaur.

Two occurrences of probably small quadrupedal forms were found on blocks that had been transported to the west edge of the property abutting Route 303 (Fig. 41). These were collected and given to the New York State Museum, although their present location is unknown. These small tracks probably belong to *Brachychirotherium* (a distant crocodylian relative that was not dinosaur-like), but they are actually too poorly preserved to identify.

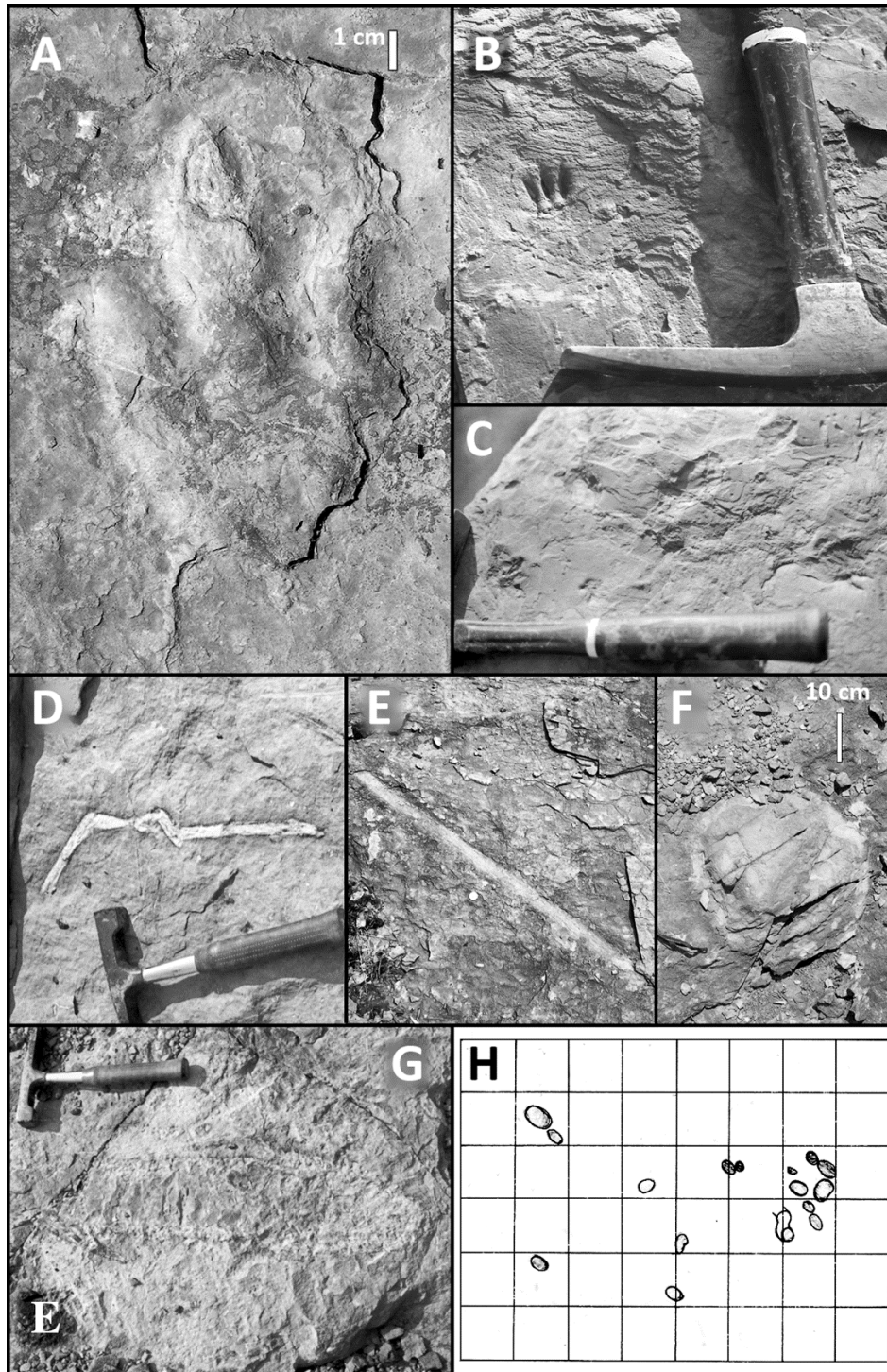


Figure 41: Other fossils from Stop 2.3: A, dinosaur-like possible *Atreipus* footprint (from area B); B, probable quadrupedal track (only one impression present, foot?), hammer for scale; C, very poor quadrupedal trackway, hammer for scale; D, whitish rhizomorph (root natural cast), hammer for scale; E, large root or stem cast, quarter (1.9 cm) for scale; F, large diameter tetrapod vertical burrow, same layer as D and E; G, large diameter horizontal burrow, hammer for scale; H, field sketch of possible tree trunks, same layer as F, grid is 2 ft (61 cm).

2) Tetrapod burrows: Large cylindrical casts are present (Fig. 41) in a greenish gray silty to fine sandy massive mudstone in the southwestern outcrops. Originally, Paul Olsen thought they were tree roots or trunk, but like the similar forms from Stop 2.2, they do not branch and are more simply interpreted as tetrapod burrows.

3) Possible teeth and bone scraps: A few small teeth and bone fragments were found. The shape of the probable teeth is superficially consistent with a phytosaurian reptile, but the metamorphism renders identification extremely difficult. The possible bone fragments are unidentifiable. These specimens were deposited in the vertebrate paleontology collection of the Peabody Museum of Yale University.

4) Rhizomorphs: A variety of rhizomorphs (root traces) are present throughout the exposed bedrock. These consist of whitish claystone natural casts in a greenish gray, silty to fine sandy massive mudstone (Fig. 41). They are present in most layers, but are especially obvious in the layers with the mapped tetrapod.

There have been several attempts at preserving this area via cooperative agreements between the land owner and Orangetown (the municipality), spearheaded by local residents (e.g., Town of Orangetown, 2009). Thus far, a firm arrangement has not been reached because of unfortunately timed economic downturns.

Backtrack to parking area.

Distance in miles (km)		
Cumu- lative	Point to Point	Route Description
23.9 (38.5)	0.5 (0.8)	Return to vehicles and leave Stop 2.3 onto N Greenbush heading north and Turn sharp left onto ramp towards NY-303, and keep right.
24.0 (38.5)	0.1 (0.1)	Go 0.06 m (335 ft) and turn right onto NY-303, and keep left.
24.1 (38.7)	0.1 (0.2)	Go north on NY-303, keeping left and turn onto N Palisades Center Dr, keeping left, and turn left onto ramp and merge onto I-287W/I-87 N.
27.8 (44.7)	3.7 (6.0)	Go west on I-287 W/I-87 N and take exit 14 for NY-59. Tandem lot on left was drilling site. Continue straight onto Forman Dr.
28.0 (45.0)	0.2 (0.3)	Follow Forman Dr and take right onto S Pascack Rd.
28.1 (45.2)	0.4 (0.6)	Take S Pascack Rd north and turn left onto Pipetown Hill Rd.
28.5 (45.8)	0.4 (0.6)	Proceed on Pipetown Hill Rd and turn left onto S Central Ave.
30.2 (48.5)	1.7 (2.7)	Go south on S Central Ave and continue onto Old Nyack Turnpike to left turn onto Saddle River Rd.
30.2 (48.7)	0.1 (0.1)	Take Saddle River Rd. 446 ft south and turn around and park to Stop 2.4.

STOP 2.4: Passaic Formation sandstone and conglomerate on New York State Thruway, Monsey, NY.

Location Coordinates: 41.101932°, -74.069205° (parking); 41.101801°, -74.067179° (main cut).

Duration: 0:45 hr.

NB! This site cannot be visited without a permit from the New York State Turnpike Authority and without proper safety attire.

Park vehicles near overpass for turnpike on east side of Saddle River Road. Leave the vehicles and proceed to north side of overpass. Go over fence and proceed east up to highway. At highway, stay to the outside (north) of the barricade and walk along cut to the east.

This cut exposes roughly 25 m of red pebbly sandstone and conglomerate and very minor mudstone. Conglomerates are dominated by Paleozoic sedimentary and metamorphic clasts. Highly irregular bedding with almost no preserved fine scale sedimentary structures, most simply explained by bioturbation by roots and burrows. This section was certainly deposited by fluvial processes, based on the apparent channel forms. This facies is typical of most of the Passaic Formation in the western half of Rockland County.

Backtrack towards vehicles.

Distance in miles (km)

Cumu- lative	Point to Point	Route Description
30.6 (49.3)	0.4 (0.6)	Return to vehicles and proceed north on Saddle River Rd and turn left onto NY-59 west.
34.7 (55.9)	4.1 (6.6)	Proceed on NY-59 west and turn right onto Tilcon Rd.
34.8 (56.0)	0.1 (0.1)	Go north to gate on Tilcon Rd and park for Stop 2.5.

STOP 2.5: Upper Feltville Formation sandstone and conglomerate and lower flows of the Preakness Basalt, at former Tilcon, Union Hill Quarry, Village of Sufferen, Town of Ramapo.

Location Coordinates: 41.116395°, -74.145690° (parking); 41.117561°, -74.143347° (Station A); 41.117010°, -74.143702° (Station B); 41.116978°, -74.142732° (Station C).

Duration: 1:30 hr.

As mapped by Volkert (2011) Union Hill is underlain by an outlier of Preakness Basalt in a syncline with the Feltville Formation, Orange Mountain Basalt and uppermost Passaic Formation. Only the upper Feltville and lower Preakness Basalt, in which the trap rock quarry was developed, are exposed here, the rest being covered by Pleistocene till and Holocene deposits except for another small outlying area of basalt. We will examine the section from the top down in three stations, A, B, and C.

Station A, Preakness Basalt: The two lower flows of the Preakness Basalt are well exposed here and very similar to what we saw at Stop 1.8. The lowest flow is irregularly pillowed and the succeeding flow has a massive base and a typical, for the Preakness, entablature with well-developed, platy-prismatic jointing. The top of second flow is not exposed. Ratcliffe (1988) suggested that the western part of the quarry occupies a lower structural level than what is exposed on the east side. He suggested two possibilities: 1) a basalt flow complex flowing in a topographic depression west of the conglomerate; or 2) a westward-dipping or vertical fissure feeder surfacing in flow rocks. Ratcliffe notes that the pillowed zone was exposed at the deepest parts of the quarry so that (2) seems contradicted. This will be discussed on the exposures. However, there is little doubt that distinctive features of the Preakness Basalt, particularly its platy-prismatic jointing are very laterally persistent.

Station 2, Contact between Preakness Basalt and Feltville Formation: The contact between the uppermost Feltville Formation was well exposed during quarrying operations and was described by Ratcliffe (1980; 1988). Presently only some ledges of coarse dolostone conglomerate are exposed near where the contact projects. According to Ratcliffe (1980, p. 297), "Superb exposures at the south end of the quarry ... show a vertical wall of coarse conglomerate, enclosing a large boulder of brecciated dolostone in an apparent channel that, in turn, is overlain by an upward-fining cycle ending in fine red shale with greenish slate chips. Vesicular pillow lava directly overlies this shale." Based on stratigraphic

position directly below the lowest flow of the Preakness Basalt, this is time-equivalent to the reddish and gray sandstones and mudstones at Stop 1.8.

Station 3, Wall of conglomerate of the upper Feltville Formation: This wall of brown conglomerate is a noticeably different color than typical red conglomerate of the Passaic Formation. According to Ratcliffe (1980, p. 297), “Clasts in the conglomerate include epidote-rich hornblende granite gneiss, Silurian Green Pond conglomerate, dolostone clasts, and several cobbles of basalt pillows.” Clasts of basalt other than pillows are present as well. The presence of these basalt clasts are evidence that these units are upper Feltville Formation. Volkert (2011) map the presence of Orange Mountain Basalt in this area on the basis of extrapolation from the Cushtunk, NJ area, although there is no direct observational evidence, close to Union Hill. However, Ratcliffe (1988) note that the chemistry of a small basalt outlier 2.4 km northeast of Union Hill (at around 41.133685°, -74.126122°) is indistinguishable from that the upper Ladentown flows to the northeast. The Ladentown flows are in turn part of the Orange Mountain Basalt (Blackburn et al., 2013; cf., Puffer, 1987; Puffer et al., 2009). This small basalt outlier projects to the conjectural Orange Mountain Basalt on the northeast limb of the Union Hill Syncline. That the Orange Mountain Basalt does not outcrop in this area is not indicative however, because it does not outcrop all along the east side of Camgaw Mountain, where it is mapped by records of numerous water wells (Volkert, 2011). The hypothesized presence of the Orange Mountain Basalt in the Union Hill area could easily be tested by drilling.

Walk back towards vehicles.

Distance in miles (km)		
Cumulative	Point to Point	Route Description
34.9 (56.1)	0.1 (0.1)	Return to vehicles, turn around, and head south on Tilcon Road turning right onto Lafayette Ave (NY-59) west.
35.3 (56.7)	0.4 (0.6)	Take Lafayette Ave west (NY-59) and turn left onto Chestnut St.
35.5 (57.1)	0.4 (0.6)	Proceed on Chestnut Street and turn right onto Ramapo Ave.
35.5 (57.2)	0.1 (0.1)	Head toward Brook St and park for Stop 2.6.

STOP 2.6: Footwall of the Ramapo Fault Zone, Suffern, NY.

Location Coordinates: 41.115209°, -74.154462°.

Duration: 30 min.

Proceed to parking lot of train station and to walkway to Station (41.115038°, -74.154749°). Walk east and southeast 42 m and turn left before the chain link fence. Walk 36 m north-northwest to outcrop.

This outcrop is described as follows by Ratcliffe (1980) as cataclasite and fault fabric in cataclastic gneiss at Ramapo fault: “These small exposures of black, chlorite-coated, cataclastic, hornblende-granite gneiss are much more instructive than the exposures north of Suffern commonly visited by field trips. Excellent chlorite-slick surfaces dip in conjugate fashion northwest and southeast and exhibit down-to-the south right-oblique slip on southeast-dipping surfaces and down-to-the-west and left-oblique movement on southwest-dipping surfaces. The near-vertical attitude of the extension fractures here ... suggests that the rocks of the footwall block ... have not undergone rotation after formation of the cataclastic fabric.”

Volkert (2011) mapped this outcrop as Mesoproterozoic hornblende granite of uncertain affinity with abundant xenoliths of well-foliated gneiss. According to Volkert’s map, it is about 330 m northwest from

the mapped position of the Ramapo Fault and 75 m northeast from a NW-SE right lateral strike slip fault. Mapped foliations in the hornblende granite are to the southeast 30°-60° and a southeasterly dipping normal fault is shown on the immediate northwest.

It is worth noting that the mapped position of the fault does not correspond to the apparent escarpment of the Ramapo Mountains, and that nearly all outcrops are hundreds of meters within the footwall from it. Because the border faults of the Newark Basin tend to follow pre-existing structures, especially of Paleozoic age, brittle features that may have formed prior to the formation of the Newark basin proper may be hard to differentiate from Late Paleozoic structures. It is noteworthy that the position of the Ramapo Fault cannot be easily seen seismic profile, although that in part may be do to its relatively high angle here, structure parallel to the boundary fault are not as obvious at this site as one might hope.

Distance in miles (km)		
Cumu- lative	Point to Point	Route Description
35.9 (57.7)	0.2 (0.3)	Return to vehicles, turn around, and head south on Ramapo Ave, turning right to continue of Ramapo Ave.
36.5 (58.7)	0.6 (1.0)	Continue on Ramapo Ave and right on top US-202 south (Ramapo Valley Rd).
37.3 (60.0)	0.8 (1.3)	Go south on US-202 to left turn onto ramp for merge onto NJ-17 north.
37.6 (60.4)	0.3 (0.5)	Head northwest on NJ-17 N and use left lanes to merge onto I-287 E/NJ-17 N towards Tappan Zee Bridge.
38.5 (61.9)	0.9 (1.4)	Take I-287 E/NJ-17 N north and use right two lanes to merge onto I-287 E/I-87 S towards Tappan Zee Bridge.
46.2 (74.3)	7.7 (12.4)	Take I-287 E/I-87 S towards Tappan Zee Bridge and take exit 14 for NY-59.
46.4 (74.6)	0.2 (0.3)	Take ramp and turn left onto NY-59 E from left lanes.
48.1 (77.4)	1.7 (2.7)	Proceed on NY-59 and turn right onto Rose Rd.
48.1 (77.4)	0.0 (0.0)	After 165 ft, turn right into Double Tree Inn.

END OF FIELD TRIP