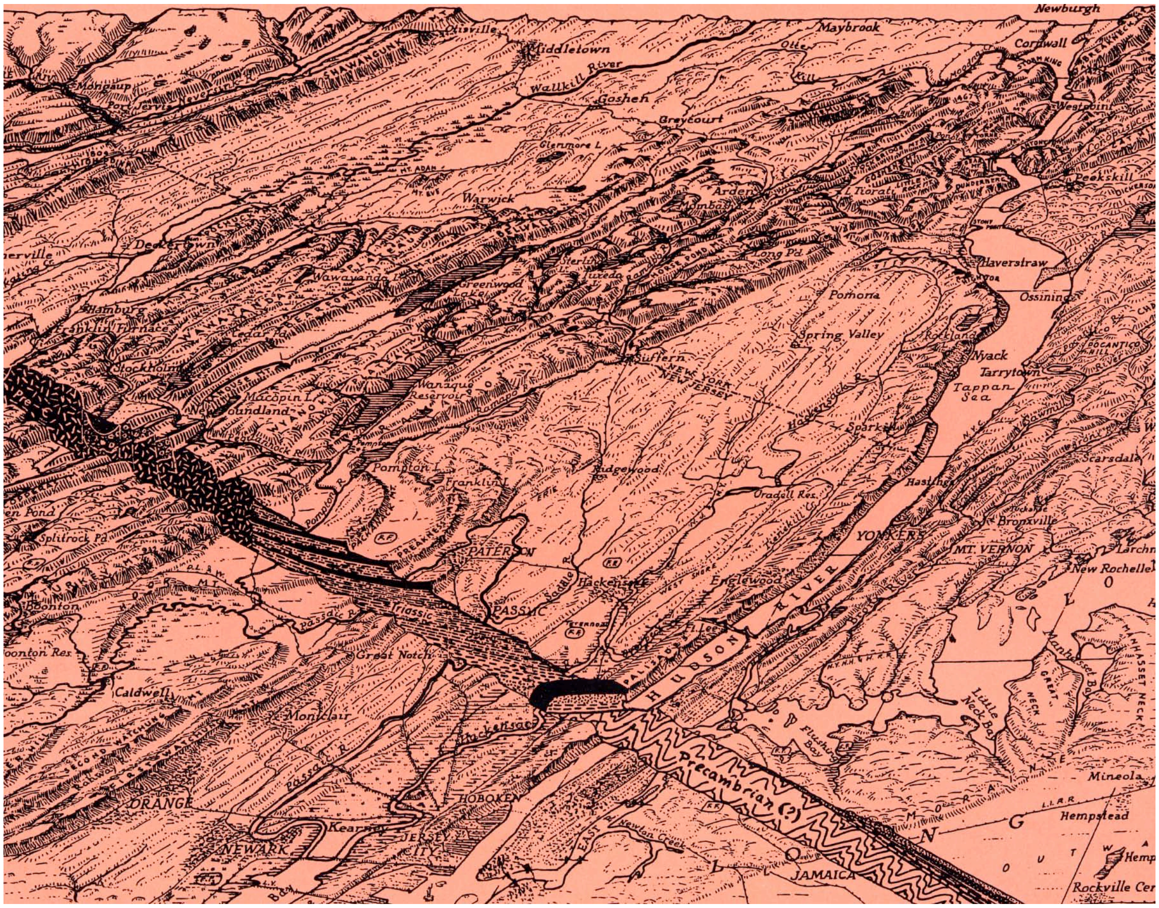
2001 New York State Geological Association Field Trip Guide Book



GEOLOGY OF THE

LOWER HUDSON VALLEY

Organizer: Paul Olsen
Lamont-Doherty Earth Observatory
Editor: Alexander E. Gates
Rutgers University-Newark

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FIELDTRIP GUIDEBOOK:

GEOLOGY OF THE

LOWER HUDSON VALLEY

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Prologue

The New York State Geological Association fieldtrip of 2001 became yet another victim of the September 11 World Trade Center disaster. The original organizer was Dr. Paul Olsen and six trips were to be run from his home base at Lamont-Doherty Earth Observatory in late September. Unfortunately, Paul was stranded overseas for nearly one week by the subsequent suspension of flights after the terrorist attacks. He was unable to make the final preparations for the conference. In addition, the disruption of services and fear that gripped the New York metropolitan area dissuaded many would-be attendees and the registration numbers were way down from previous years. Therefore, the conference was postponed to December 2. Perhaps as a result of continued uneasiness or because of the lateness in the season, registration for the new date was also poor and once again the conference and trips were postponed. A tentative date for the 2001 conference was set for April 5, 2002 but the officers of the New York State Geological Association decided to cancel it instead. Half of the participants opted to withdraw their trips from the guidebook but the other half agreed to include their trips in the guidebook largely to keep the continuity of guidebooks. There was a general consensus that a gap in the series should be avoided. Dr. Olsen did not have time to assemble the guidebook and the editorship was passed on to me. Therefore this 2001 New York State Geological Association fieldtrip guidebook contains three trips on the biostratigraphy and fossils of the Mesozoic Newark Basin, the Mesozoic igneous activity in the Newark Basin and on the regional tectonics of the Grenville event in the Hudson Highlands. Keep in mind that none of these trips were ever run nor even presented at a conference. However, they are presented here so that interested geologists can run their own trips with the latest of ideas.

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The Uniform Composition of CAMP Diabase Types and the Absence of Wall Rock Contamination at Northern New Jersey and Staten Island, New York Locations

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ABSTRACT

This field trip will focus on five contrasting and distinct late Triassic to early Jurassic lithologies. These rocks are trondhjemite, syenite, and their sedimentary protoliths, and diabase sills that are part of the Central Atlantic Magmatic Province (CAMP). All five of these lithologies are exposed at six widely spaced northern New Jersey and Staten Island, NY locations, some of which will be visited on this field trip. Chemical and petrographic analyses of closely spaced samples collected at each of these locations demonstrates a genetic linkage of the syenite and trondhjemite to sedimentary lithologies and a near absence of magma mixing with CAMP diabase. This lack of contamination helps explain the province wide uniformity of CAMP diabase types.

INTRODUCTION

The early Jurassic Palisades intrusion of the Newark Basin crops out from Haverstraw New York to the northwestern part of Staten Island, a distance of 90 km., and underlies a narrow belt along the western part of Staten Island (Figure 1). Detailed studies of the Palisades Sill were made by Lewis, 1907, 1908a, 1908b; F. Walker, 1940; K. Walker, 1969a, 1969b; Pearce, 1970; K. Walker et al. 1973; Puffer, 1984; and Shirley, 1987. None of these earlier studies included the Staten Island portion of the Palisades intrusion probably because the intrusion is poorly exposed on Staten Island. More recent studies of the Palisades and related Mesozoic intrusions and flows within the Newark Basin portion of CAMP were made by Puffer (1992), Steiner et al. 1992, Husch (1990, 1992), Houghton et al. 1992, Tollo, et al. 1992, Puffer and Student (1992), Hozik (1992) and Puffer and Husch (1996). Inasmuch as this field trip deals with syenitic and trondhjemitic fusion at Mesozoic Diabase intrusive contacts with sedimentary rocks of northern New Jersey and Staten Island, New York, the reader is referred to Puffer and Husch (1996) for a recent comprehensive study of the Palisades-Rocky Hill-Lambertville (PRHL) megasheet.

Two contrasting and distinct felsic igneous rocks were generated during the intrusion of early Jurassic diabase into Triassic sedimentary rocks: 1) a trondhjemite characterized by a K_2O and Na_2O contents of about 0.1 and 7 percent respectively and 2) a monzonite to syenitic lithology characterized by K_2O and Na_2O contents of about 4 and 7 percent respectively. One or both of these lithologies are exposed at six widely spaced locations throughout northern New Jersey and Staten Island: 1) the lower contact of the Palisades sill along a series of road-cuts at and north and south of Ross Dock where Lockatong Argillite has fused into syenite and trondhjemite; 2) the interior upper portion of the Palisades sill at a road-cut near the George Washington bridge where trondhjemite dikes are exposed; 3) near the northern and southern contacts but within the Palisades related Snake Hill diabase intrusion where intergrown or commingling trondhjemite and syenite veins are exposed at a rock quarry; 4) and the interior upper portion of the Palisades sill at a rock quarry near Granitville where the margins of a Lockatong argillite xenolith has fused into trondhjemite; and 5) at the lower contact of a Palisades related sill with Lockatong argillite where a syenitic migmatite is exposed along a stream near Brookville New Jersey. Sites 1 and 3 will be observed on this field trip. If time permits site 4 will be included.

The felsic fusion products of sites 1 -5 are absent at most Palisades diabase contact exposures and may be localized due to concentrations of halite or brackish water contained within the intruded sediments that acted as a flux. Alternatively fusion products at most locations may have been swept away by currents of intruding diabase that finally quenched against dehydrated and refractory residual hornfels. Melting experiments are planned where the fluxing effect of salt on Lockatong argillite will be measured.

The chemical compositions and by-modal nature of the felsic fusion products are due to a combination of controlling factors that are currently being evaluated by Benimoff and Puffer (in prep.). These factors include: 1) contrasts in the chemical composition of the host sedimentary rock layers intruded by the diabase, 2) heat driven metasomatism, particularly potassium, among the sediments during conversion to hornfels, 3) the degree of superheating experienced by the metasediments before fusion was initiated, 4) element partitioning between melt phases and refractory residues during fusion, 5) the degree of partial melting of the metasediments, and 6) the degree of mixing of fusion products with diabase magma after fusion.

Alternative interpretations such as the possibility that the fusion products were the result of diabase magma fractionation or liquid immiscibility were also considered but rejected on the basis of new geochemical data presented herein. It will also be shown that the fusion products occur as distinct igneous bodies that did not mix with CAMP diabase melt or materially alter its composition at any of five Newark Basin sites.

FUSION PRODUCT OCCURRENCES

Walker (l969a) reported the occurrence of (1) leucocratic patches in the normal dolerite of the Palisades Sill that he interpreted as the product of assimilation of Newark group sediments, and (2) reomorphic “veins” and “dikes” up to 15 inches thick in the Palisades Sill. Benimoff and Sclar (l984) reported the occurrence of a xenolith of sodium-rich Lockatong argillite in the Palisades Sill the outer shell of which fused and crystallized to a pyroxene trondhjemite. In addition, two albitite dikes (Benimoff et al. 1988; Benimoff and Sclar, l990) and a trondhjemite dike (Benimoff et al. 1989) were found in the Palisades Sill and these dikes were interpreted on mineralogical and geochemical grounds as products of fusion of Lockatong argillite xenoliths.

Barker and Long (l969) reported the occurrence of syenite along the lower contact of the Stockton Diabase, (a Palisades related diabase intrusion) near Brookville, New Jersey and interpreted it as the product of magma mixing of fractionated Palisades diabase and fused Lockatong argillite. Benimoff et al. (1996, 1997,1998), however, determined that the chemical composition of the syenite closely matched the composition of the intruded Lockatong argillite suggesting high degrees of partial fusion took place with little evidence of diabase mixing. The chemical data including REE's supported phase equilibria studies suggesting that wet fusion of syenitic melt could occur at temperatures approximating those at the base of the intrusive.

Puffer et al. (l994) and Puffer and Benimoff (l997) also reported the occurrence of "microsyenite" and "granitoid" veins and trondhjemite veins intruded into the diabase of Snake Hill, New Jersey and interpreted them as the fusion products of metasomatized hornfels.

The most recent focus of our work takes us back to the Palisades Sill where probably the clearest exposures of both syenitic and trondhjemite occur at the lower contact near Ross Dock. The entire progression of contact metamorphic, metasomatic, fusion, and intrusion processes is being systematically sampled and analyzed along traverses perpendicular to the contact. Preliminary results confirm earlier interpretations suggesting that high degrees of partial melting are responsible for the syenitic and trondhjemitic rocks and that minimal mixing of diabase has occurred.

Field occurrences of leucocratic rocks interpreted as products of sedimentary fusion induced by early Jurassic diabase intrusions include:

1. The base of the Palisades Sill from Englewoods Cliffs to Cliffside Park along the road to Ross Dock near the George Washington Bridge (STOP 1)
2. Dikes within the Palisades Sill I-95 road cut, Fort Lee, New Jersey
3. Snake Hill, New Jersey (STOP 2)
4. Graniteville, Staten Island, New York (Optional Stop 3)
5. At the upper contact of an early Jurassic diabase intrusion exposed in a rock quarry at Brookville, New Jersey.

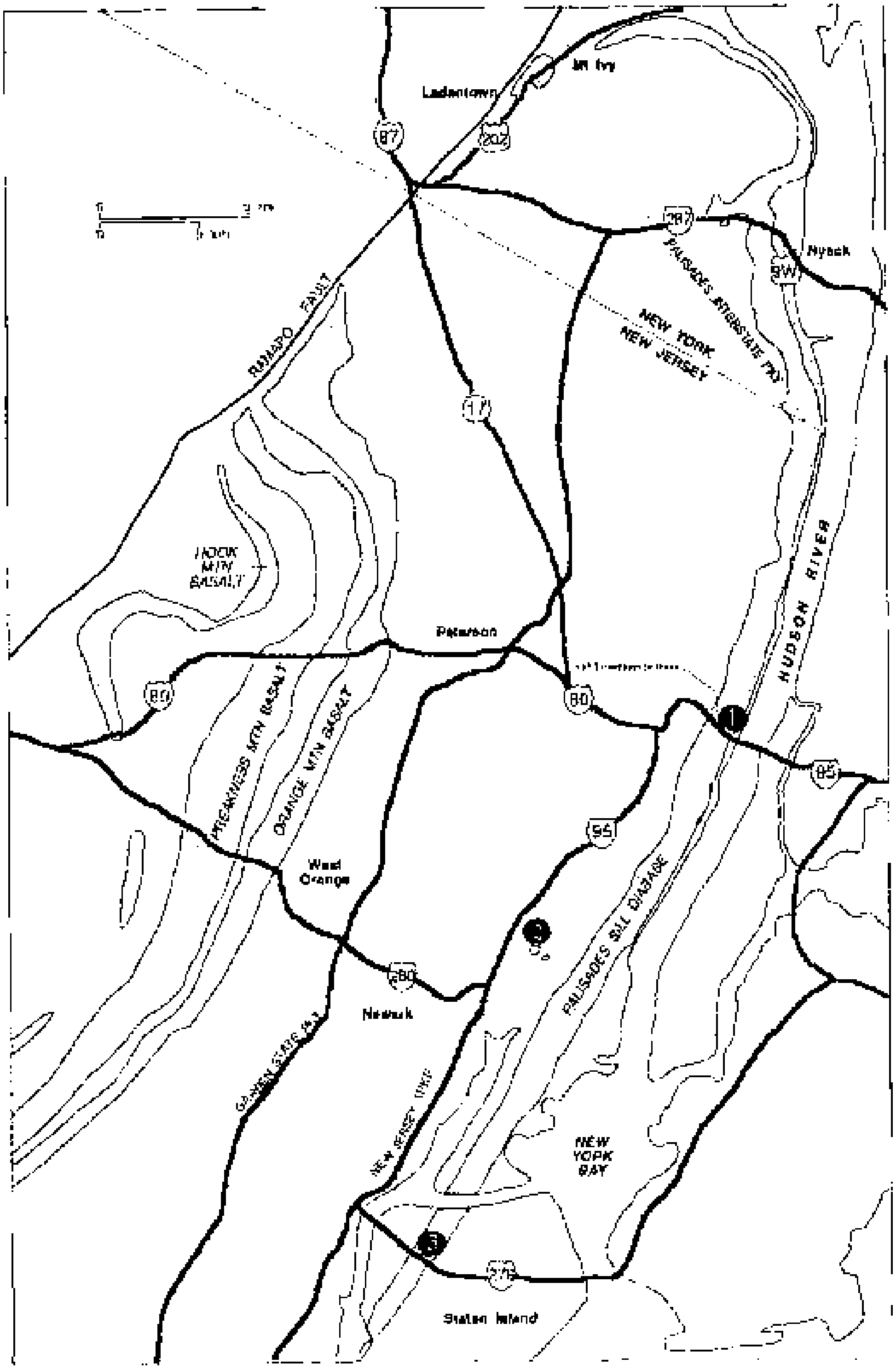


Figure 1: Map showing Jurassic Rocks of the northern Newark basin. The three field trip locations are shown. Modified from Puffer et al. 1992)

1. The Base of the Palisades Sill From Englewood Cliffs to Cliffside Park

A series of road-cuts, active and recent construction sites, and natural outcrops along the west bank of the Hudson River (Figure 1) has resulted in a discontinuous five mile long exposure of the basal contact of the Palisades Sill with the underlying Lockatong Formation and Stockton Formation.

Palisades diabase

The Palisades diabase exposed along the basal contact is fine grained massive chill-zone rock. The diabase is aphyric and subophitic and contains very few xenoliths. A few large xenoliths have been observed but they are typically spaced at least 100 meters apart. The rock is free of any obvious alteration effects except for along widely spaced faults where some chlorite and carbonates are exposed on slickensides. The lower contact is largely parallel to the bedding planes of the underlying metasediments but is locally discordant. Flow of the diabase through Lockatong argillite has excavated a few channel like cuts several meters across that truncate underlying Lockatong bedding plains at about 30 °. In addition, anticlinal dome like structures consisting of Lockatong Formation that rise a few meters into overlying diabase have been observed (Figure 2). It is at these dome-like structures where most fusion seems to have taken place.

Figure 2. Lower contact of the palisades Sill along the road to Ross Dock. Migmatites are present to the right of the meter stick in an anticlinal dome structure.

Lockatong Formation and Fusion Products

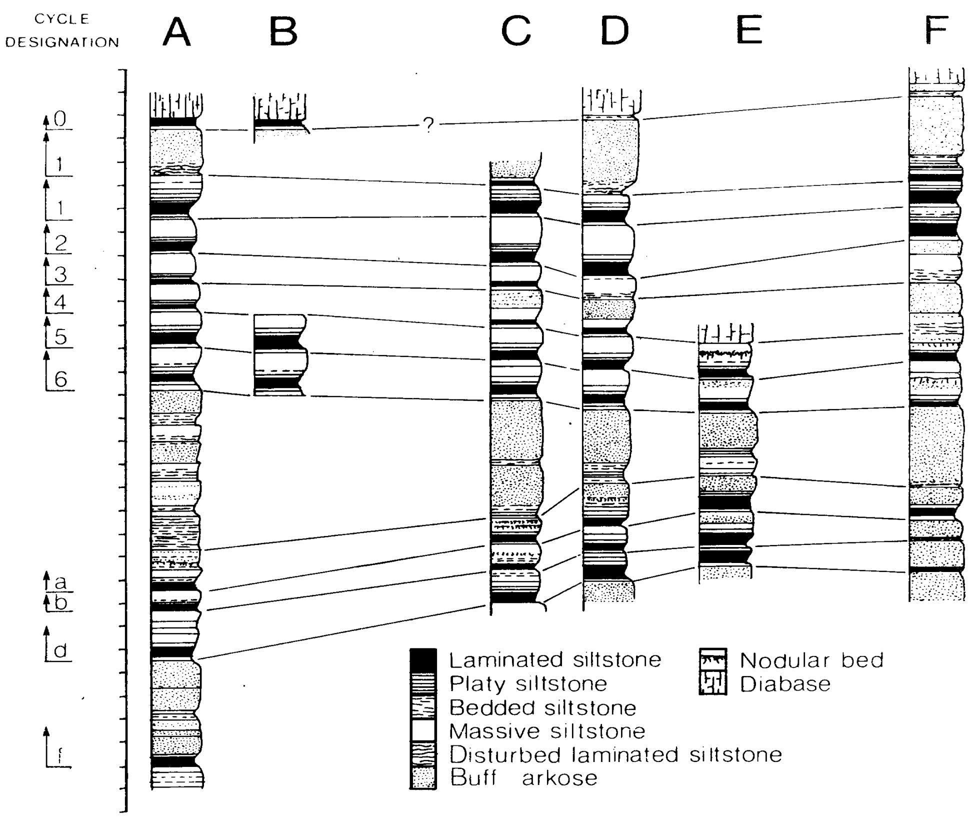
The lithology of the Lockatong Formation has been described in detail by Van Houten (l980) and the exposures along Englewood Cliffs have been described by Olsen (l980). In general the Lockatong formation consists of short detrital and chemical cycles averaging several meters thick. Gray detrital cycles are most common in the northwestern part of the Newark basin including area near Englewood Cliffs unlike the reddish-brown chemical cycles (Van Houten, l980). The detrital cycles are composed of: 1) black pyritic shale, 2) dark gray carbonate-rock mudstone and 3) gray calcareous argillite. These laminated (varved) lake sediments are interbedded with lenses of fine-grained feldspathic sandstone. These detrital sedimentary rocks are composed of sodic plagioclase, illite and muscovite, some K-feldspar, chlorite, calcite, and a little quartz (Van Houten, l980). These sedimentary rocks have been metamorphosed into hornfels near the Palisades Sill and as described by Van Houten (l969) have been recrystallized into: 1) gray layers consisting of biotite and albite with minor analcime and diopside; 2) black layers consisting of biotite and albite with minor orthoclase, calcite and a trace of amphibole; and 3) light greenish gray to pink layers consisting of diopside and grossularite with minor chlorite, calcite, biotite, feldspar, amphibole and prehnite. The gray hornfels is described by Olsen (l980) as "bedded and disturbed laminated siltstone"; the black hornfels as "laminated siltstone" and the light greenish gray to pink hornfels as "platy siltstone". In addition, layers of metamorphosed "buff arkose" are found interbedded with these hornfels types Olsen (l980).

Olsen (l980) has identified several individual meta-detrital cycles consisting of basal laminated siltstone overlain by multiple layers of bedded, platy, and massive siltstones in exposures south of Englewood Cliffs and has correlated nine of these cycles 12 km through six exposures along the base of the Palisades Sill from Kings Bluff, Weehawken to Ross Dock, Fort Lee including the area from Englewood Cliffs to Cliffside Park (Figure 3). In particular, four meters of the section at Ross Dock directly below the Palisades diabase contact where considerable fusion has occurred includes 0.3 m of "buff arkose" in contact with the diabase, underlain by 0.2 m of bedded siltstone, underlain by 2.9 m of "buff arkose" underlain by 1 m of "platy and bedded siltstone" and 0.5 m of "laminated siltstone" (Figure 3).

These lithologies in the sequence described by Olsen (l980) are largely devoid of any clear evidence of fusion. However, at three locations within 2 km of Ross Dock fusion has occurred resulting in syenitic rock intergrown with black laminated siltstone forming a migmatite. This style of fusion is confined to the contact zone within distinctly domed structures (Figure 2) that are characterized by discordant contacts and the uplift of the black "laminated siltstone" to close proximity to the diabase contact. The bedding at these domed structures is disrupted and may have involved movement of volatiles derived from brackish groundwater within the lacustrine Lockatong sediments. The salt content of these volatiles may have helped flux the melting process.

In 1948, the annual GSA meeting was held in NYC, specifically at the old Hotel Penn. One of the field trips associated with this meeting was to the Palisades Sill. The field trip leaders were Walter Bucher, a professor of geology at Columbia University, and S.J. Shand. Charles B. Sclar (former Professor Emiritus, Lehigh University, deceased) was a young geologist and a participant on this trip. At one of the stops at the basal contact of the Palisades Sill 1/4 mile north of the George Washington Bridge was an exposure of Lockatong Argillite impregnated with salmon colored leucocratic granitoid migmatitic material. According to C. B. Sclar (personal communication) Walter Bucher was intrigued and very interested in this occurrence. In his "effervescent" way Bucher called this phenomenon *palingenesis* in a small scale and implied that the homfels was an example of a rock with a low melting component which due to heat of the adjoining Palisades Sill actually melted *in situ.* C. B. Sclar put a labeled and dated specimen that remained largely unstudied until two years ago (1999) The same site that was visited in 1948 was then revisited and found be consistent with the earlier interpretation of Bucher.

Three clearly exposed examples of these domes occur within 2 km of Ross Dock, where fusion has occurred resulting in syenitic rock intergrown with black laminated siltstone forming a migmatite and trondhjemite veins. The major element compositions of these leucocratic rocks and the surrounding dark rocks are given in tables 4 and 5 respectively and their CIPW norms are given in Tables 6 and 7. The bedding at these domed structures is disrupted and may have involved movement of volatiles derived from brackish groundwater within the lacustrine Lockatong sediments. The high salt content of Lockatong Argillite may have helped flux the melting process. Both the syenite and the trondhjemite are sodic, typically containing 4 and 7.5% Na_2O respectively, but the K_2O and Rb is highly partitioned into the syenite typically containing 5 % and 125 ppm vs. only 0.5% and 25 ppm for the trondhjemite. Both the syenite and trondhjemite have similar REE contents that are comparable to the Lockatong Argillite.



Correlation of cycles from Kings Bluff, Weehawken to Ross Dock, Fort Lee; A, Kings Bluff exposure; B, Gratacap's (1886) Weehawken locality and exposure of cycle 0 to the north; C, Gorge and River Roads exposure, Edgewater; D, exposure at east portal northern tunnel for old New York, Susquehanna Southwestern Railroad; E, "old trolley route" below old Palisades Amusement Park, Fort Lee; F, exposures west of Ross Dock, Palisades Interstate Park, Fort Lee. Exposures A and F are 12 km apart; the other sections are positioned to scale.

Figure 3. From Olsen ,1980

Our complete chemical analyses of the leococratic component of the migmatite when plotted(Figures 4, 5, and 6) on primitive mantle normalized spider diagrams (Sun and McDonough, 1989) provide compelling evidence that the igneous components of the migmatite are fused Lockatong argillite. For example, all three lithologies display distinctly negative Nb anomalies. However, the chemistry of these lithologies differs from that of the Palisades Diabase.

Table 1: Major Elements of Leucocratic Rocks

	G2D-L2	G2E-L	G2CL	G127L1	G5L1	G5L2	G2AL1	G3L	G4L	HTQ
SiO_2	65.6	76.2	72.7	71.8	64.6	73.3	75.7	72.8	71.4	52.3
Al_2O_3	12.7	12.7	12.9	10.1	14.5	14.5	11.8	13.6	13.8	14.1
CaO	3.9	0.3	0.5	3.4	2.2	0.6	0.7	1.7	3.1	10.7
MgO	3.4	0.6	1.0	4.5	5.0	1.0	0.9	10.8	2.3	7.8
Na_2O	3.7	4.4	4.2	2.9	7.3	8.5	4.0	5.7	7.8	2.4
K_2O	5.1	4.0	5.4	2.6	1.0	0.4	4.3	2.7	0.1	0.4
Fe_2O_3	1.9	1.2	1.5	1.3	2.8	0.5	0.9	1.3	0.6	11.2
MnO	0.1	0.0	0.0	0.2	0.1	0.0	0.0	0.0	0.0	0.2
TiO_2	0.2	0.2	0.3	0.1	0.4	0.3	0.2	0.3	0.3	1.1
P_2O_5	0.2	0.1	0.2	0.1	0.1	0.1	0.3	0.1	0.1	0.1
Cr_2O_3	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	
LOI	1.8	0.3	0.3	1.8	1.2	0.3	0.5	0.8	0.7	
Sum	98.9	100.1	99.0	98.8	99.3	99.4	99.4	100.3	100.2	100.4

Table 2: Major Elements of dark rock surrounding the leucocratic rock

	G2CD	G2BD	G127D1	G127D3	G5D	G2AD
SiO_2	54.60	55.40	53.00	71.40	54.60	53.20
Al_2O_3	18.60	17.60	17.70	13.40	19.40	17.70
CaO	1.09	0.65	1.62	1.04	1.29	1.08
MgO	3.90	4.59	3.89	0.97	5.60	3.91
Na_2O	3.23	4.68	4.09	3.81	4.59	3.71
K_2O	5.60	5.40	7.09	5.82	6.53	6.62
Fe_2O_3	9.57	8.40	8.89	1.35	4.40	10.10
MnO	0.13	0.08	0.13	0.02	0.04	0.09
TiO_2	0.85	0.91	0.82	0.17	0.86	0.85
P_2O_5	0.27	0.15	0.28	0.17	0.17	0.14
Cr_2O_3	0.02	0.01	0.01	0.01	0.02	0.01
LOI	1.16	1.38	0.60	0.55	1.10	0.85
Sum	99.20	99.50	98.40	98.90	98.80	98.60

Table 3 CIPW Norms of Leucocratic Rocks

	G2D-L2	G2E-L	G2CL	G127L1	G5L1	G5L2	G2AL1	G3L	G4L
Q	15.0	33.89	25.79	33.43	8.13	20.56	33.94	10.96	18.31
or	30.1	23.70	32.09	15.25	5.73	2.36	25.29	16.07	0.53
ab	31.1	37.40	35.29	24.45	61.43	71.50	34.10	47.81	66.26
an	3.1	0.79	0.44	6.97	4.11	0.46	1.31	3.72	2.24
ne	0.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
C	0.0	0.80	0.00	0.00	0.00	0.00	0.06	0.00	0.00
di	11.7	0.00	0.18	7.21	4.70	0.76	0.00	2.32	9.47
hy	3.1	1.52	2.33	7.81	10.27	2.04	2.29	25.82	1.34
wo	0.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ol	0.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ac	0.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
mt	0.2	0.00	0.00	0.25	1.54	0.00	0.00	0.00	0.00
il	0.4	0.02	0.02	0.22	0.84	0.02	0.02	0.09	0.06
hem	1.6	1.16	1.47	1.11	0.88	0.47	0.92	1.34	0.55
ti	0.0	0.00	0.60	0.00	0.00	0.72	0.00	0.72	0.56
ap	0.4	0.23	0.39	0.14	0.25	0.19	0.70	0.25	0.12
cc	0.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
pero	0.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
wus	0.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ru	0.0	0.20	0.00	0.00	0.00	0.00	0.16	0.00	0.00

Table 4 CIPW Norms of dark rock surrounding the leucocratic rock

	G2CD	G2BD	G127D1	G127D3	G5D	G2AD
Q	3.0	0.0	0.0	24.2	0.0	0.0
or	33.1	31.9	41.9	34.4	38.6	39.1
ab	27.3	39.6	23.1	32.2	31.1	29.8
an	3.6	2.2	6.2	2.3	5.3	4.4
lc	0.0	0.0	0.0	0.0	0.0	0.0
ne	0.0	0.0	6.2	0.0	4.2	0.9
kal	0.0	0.0	0.0	0.0	0.0	0.0
C	5.9	3.2	1.0	0.0	2.8	2.8
di	0.0	0.0	0.0	1.0	0.0	0.0
hy	18.5	6.0	0.0	2.0	0.0	0.0
wo	0.0	0.0	0.0	0.0	0.0	0.0
ol	0.0	8.7	12.8	0.0	9.8	14.3
ac	0.0	0.0	0.0	0.0	0.0	0.0
mt	3.4	3.5	3.4	0.0	3.4	3.4
il	1.6	1.7	1.6	0.0	1.6	1.6
hem	0.0	0.0	0.0	1.4	0.0	0.0
ti	0.0	0.0	0.0	0.4	0.0	0.0
ap	0.6	0.3	0.6	0.4	0.4	0.3
cc	0.0	0.0	0.0	0.0	0.0	0.0
pero	0.0	0.0	0.0	0.0	0.0	0.0
wus	0.0	0.0	0.0	0.0	0.0	0.0
ru	0.0	0.0	0.0	0.0	0.0	0.0

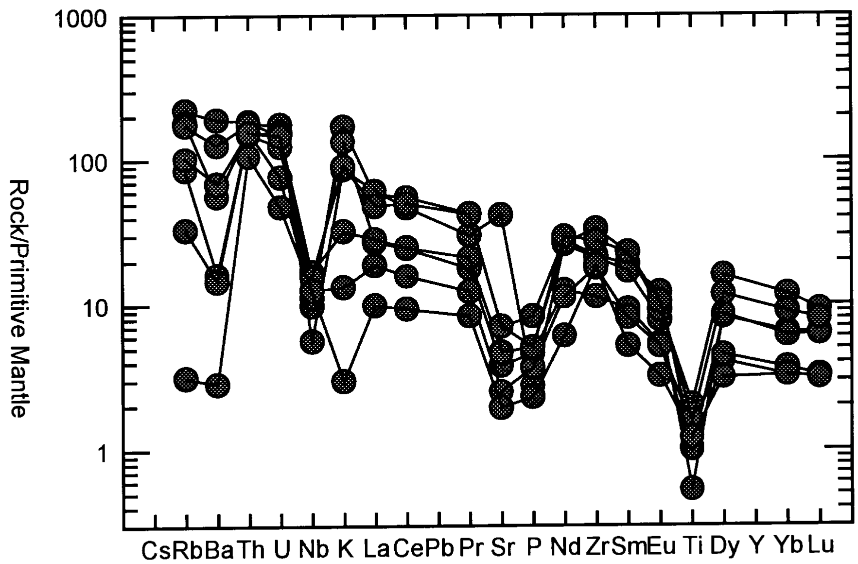


Figure 4. Sun and McDonough(1989) mantle normalized plot of dark rock surrounding the leucocratic rock.

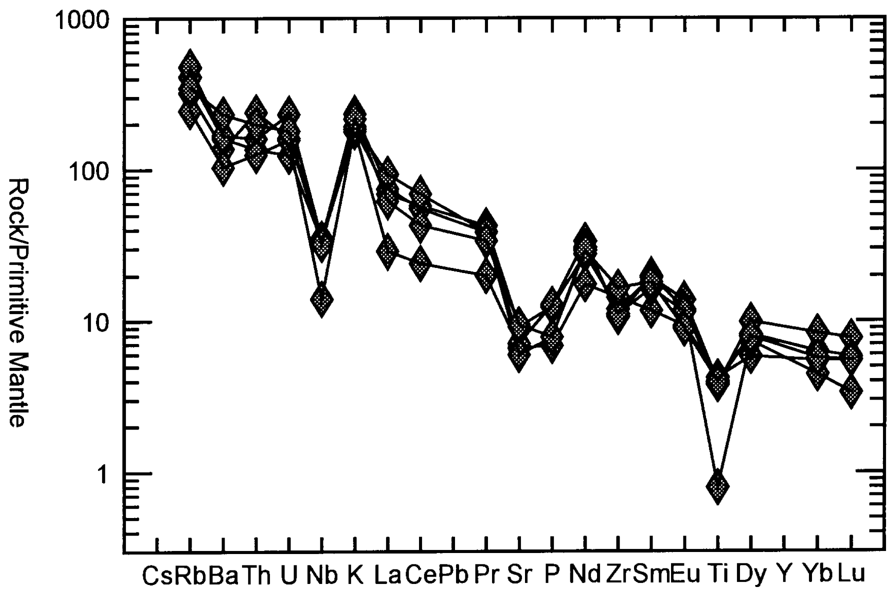


Figure 5. Sun and McDonough(1989) mantle normalized plot of the leucocratic rock.

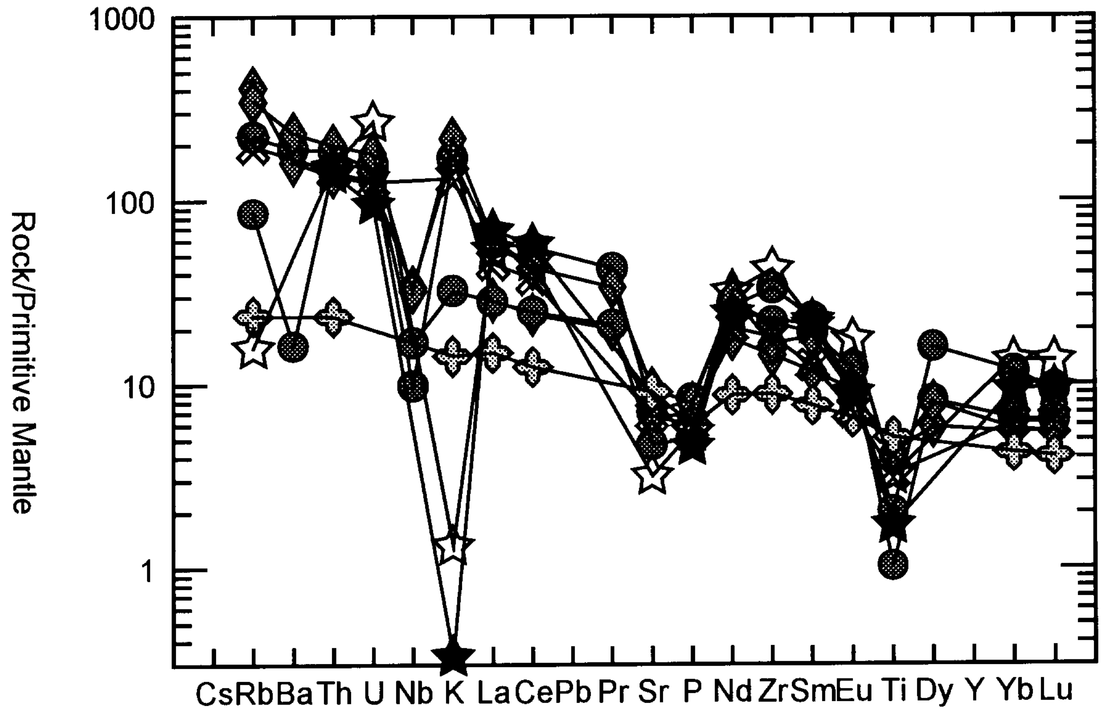


Figure 6. Sun and McDonough(1989) mantle normalized plot. Circles represent leucocratic rocks along road to Ross Dock; Closed diamonds represent the dark rocks that surround the leucocratic rocks along the road to Ross Dock; Solid star represents the I-95 trondhjemite dike.; X represents the North American Shale Composite; Shaded plus sign represents HTQ from Tollo and Gottfried, 1992; Open Star is the Lockatong hornfels xenolith from Graniteville.

Again, our complete chemical analyses of the leucocratic rock and its surrounding dark rock when plotted on primitive mantle normalized spider diagrams(Sun and McDonough, 1989) provide compelling evidence that the leucocratic component of the migmatite and the other leucocratic rocks along this traverse are genetically related to the Lockatong argillite and not genetically related to the Palisades Diabase(see Figures 4, 5, and 6). The leucocratic component is a fusion product of the Lockatong argillite. Also note the similarity of the leucocratic component to the North American Shale Composite(NASC). By comparison, the HTQ magma analysis plot as a line with a different slope, and magnitude. This data rules out magmatic fractionation of a basaltic magma as a process capable of generating the leucocratic rocks of this study.

2. DIKES WITHIN THE PALISADES SILL I-95 ROAD CUT, FORT LEE, NJ

FIELD OCCURRENCE

Exposed in the I-95 road cut within the upper 70 meters of the Palisades Sill (Figure 1) is a group of at least 3 trondhjemite dikes which have been described by Benimoff et al (1989).

Proceeding down section into the I-95 cut from the upper contact of the Palisades Sill a 20 to 30 cm thick leucocratic dike is encountered. A 6 -12 cm thick calcite vein has apparently precipitated into another joint in the dike approximately parallel to its contact. This westernmost dike in a parallel group of leucocratic dikes and calcite veins is designated dike 2. East of dike 2 dike 1 is exposed. Both trondhjemite dikes are near vertical and strike N 27° E which is parallel to a major joint set in the sill. The thickness of dike 1 is about 2 meters. Five samples across the width of the dike were selected for chemical analysis (Table 5) . The network of calcite veins exposed along I-95 may be the result of similar injection into cooling joints but at a lower temperature. Some of the calcite veins contain about 10 percent sulfides and may have precipitated out of hydrothermal vapors injected into cooling shrinkage joints in the Palisades from a Lockatong sedimentary source.

PETROGRAPHY

Dike 1 is a leucocratic holocrystalline microphanerite composed almost exclusively of quartz and albite. The rock is porous and most of the pores are miarolitic cavities. The microtexture observed in thin section is unusual because it consists of a granophyric intergrowth of quartz and albite. Thin sections reveal an intergrowth of clear grains of quartz in a turbid matrix of albite; also present are opaque grains, which, as' determined in reflected light, consist dominantly of discrete grains of ilmenite and subordinately of discrete grains of hematite which are not exsolution intergrowths. Accessory minerals include skeletal sphene, and hematite as minute flakes disseminated in the silicates. Most quartz grains are in approximately parallel optical continuity. This texture is interpreted as an igneous texture resulting from simultaneous crystallization of quartz and albite from a eutectic melt. Chemical analyses of each of the five samples of the dike are presented in Table 5. Total Fe is expressed as Fe_2O_3; FeO was not determined. However the total iron is very low (<1.8% and the silica is very high 77 - 80% (34 - 41.5% normative quartz). Na_2O (6.5-7%; 54-60% normative albite) is exceptionally high and K_2O is extremely low (<0. 8% normative - orthoclase). In addition, CaO and MgO are each less than 1%.

An AFM diagram (Figure 7) compares the dike compositions, samples PD - 71, and PD-72 through PD - 76, with the Palisades differentiation trend of Walker (1969), a leucocratic albitite dike from the Travis quarry on Staten Island (Benimoff et al., 1988), and a "white vein" in the Palisades Sill (Walker 1940). The lower left apex represents the sum of Na_2O plus K_2O. While AFM diagrams are very useful in demonstrating magmatic differentiation, in this case it is misleading because Na_2O and K_2O are combined. In the dike of this study the ratio of Na_2O to K_2O is very high and there is hardly any K,O, whereas in the normal differentiation trend of a tholeiitic magma, the granophyre is composed of K-spar and quartz, in which K_2O is dominant over Na,O. Therefore, it is unlikely that these dikes are late tholeiitic trend or calc-alkaline trend differentiates of the Palisades magma.

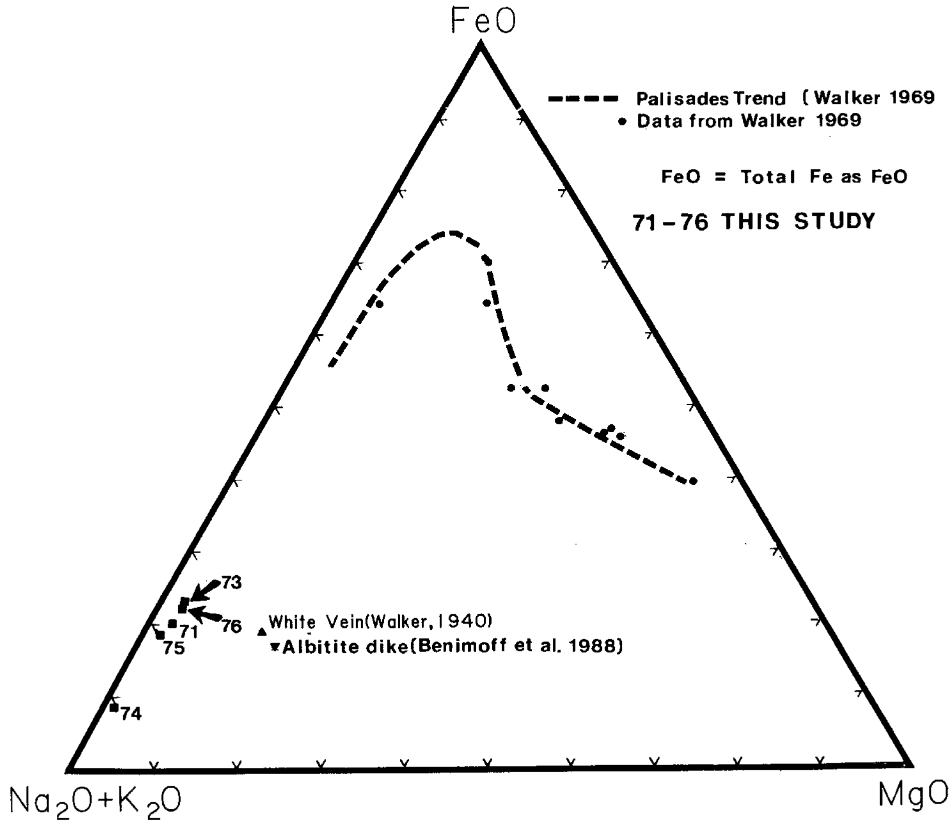


Figure 7. AFM diagram . The dashed line is the Palisades differentiation trend of Walker, 1969a. Closed circles data from Walker, 1969a; 71-76 represent the I-95 Dike 1. Albitite Dike(see table 1) and White vein of Walker, 1940 shown for comparison.

DISCUSSION

The plot of the mafic index verses liquidus temperatures for basaltic magmas of Tilley et al. (1964) is instructive. The chilled margin of the Palisades Sill (W-889LC-60, Walker 1969) has a mafic index of 0.5755 which yields a liquidus temperature of about 1200° C on the basis of the experimental work of Tilley et al. 1964. If this leucocratic dike is not a magmatic differentiate of the Sill then what is its source? At Graniteville in Staten Island, N.Y., Benimoff and Sclar (1984) described a partly fused xenolith of Lockatong Argillite enclosed in the Palisades Sill which crystallized to trondhjemite composed principally of albite-quartz granophyre. The unmelted (hornfesed) xenolith at Graniteville has a normative compositional range between the two vertical lines at the bottom of Figure 13. Clearly, rocks of such composition would melt if immersed in a magma which has a temperature of 1200° C.

Furthermore, the Lockatong xenolith has a very high Na/K ratio, and the same ratio of normative quartz/normative albite that occurs in the trondhjemite dikes. Consequently, the source of the I-95 dike was probably a previously metasomatized xenolith of Lockatong argillite which floated up to about the center of the sill and melted to yield a high Na_2O, low K_2O, silica-rich melt which could have been the parental magma of the dike.

The Lockatong Formation is a complex lacustrine suite(Olsen, 1980) that is dominated by sodic argillite. The Lockatong has been thermally metamorphosed into hornfels up to pyroxene grade facies grade by the intrusion of the Palisades Sill (Van Houten, 1972). The pyroxene facies hornfels exposed above and below the sill along the I-95 section includes a black variety consisting primarily of biotite and plagioclase and a pale green variety consisting of garnet and pyroxene. The black and pale green hornfels are interbedded at irregular 1 to 10 cm intervals. The green hornfels has been interpreted by Van Houten, (1972) as having been derived from carbonate rich beds in the Lockatong argillite. Common discordant pale green bands of calcium rich pyroxene and garnet are interpreted as having been developed from carbonate veining through the argillite. If the liquidus temperature of the sill is 1200° C, then the solidus temperature is about 1050° C (Yoder and Tilley, 1962). The quartz - albite intergrowth texture of the dikes are consistent with the interpretation the magma of the dike was at or near the eutectic composition on the dry albite - SiO_2 phase-equilibrium diagram, and the that eutectic temperature on this diagram is 1062° C. As noted above, the abundant miarolitic cavities in the dike record the presence of a gas phase, probably H_2O and CO_2, which would depress the eutectic temperature.

Therefore, a magma of the trondhjemite composition could have been liquid, while the diabase was solid and jointed. In support of this model, it should be noted that the I-95 dike does not show any chilled margins, which means that it was injected into the diabase parallel to the joints while the solid diabase was still hot. As shown in Table 5, the concentration of Fe, Mg, and Ca is extremely low in the dike. This indicates that virtually no diffusion took place across a liquid - liquid interface such as at Graniteville (Benimoff and Sclar, 1984) where the diabase magma and xenolith derived liquid coexisted and where Fe, Ca, and Mg and other ions diffused across the liquid - liquid interface. Consequently, the diabase magma and the dike magma did not coexist for any great length of time because if they had, there would have been diffusion of ions across the liquid - liquid interface as shown by the Graniteville xenolith. Therefore, the dike was liquid while the diabase was solid.

Chondrite normalized REE plots of dike 1 and dike 2 are shown in Figure 8. The Palisades trondhjemite dikes are very highly enriched in REE (Figure 8) exceeding the chondrite values of) by a factor of 20 to 200. Such high values are commonly interpreted as the result of very high degrees of fractionation (over 90 percent if a Palisades diabase magma source is assumed) but may also be interpreted as the product of very high degrees of partial They have parallel trends and therefore must have a common source. Relative to the diabase host rock the dikes are enriched in light REE's and show a negative Europium anomaly instead of the positive europium anomaly of the diabase. The negative Eu anomaly of the dikes is consistent with the development of a partial melt from the plagioclase rich residue of the Lockatong hornfels.

Table 5: Major Element Analyses and CIPW Norms of Leucocratic Rocks in the Palisades Sill

	I-95 Dike 1					I-95 Dike 2		Travis	Bayonne
	1	2	3	4	5	6	7	8	9
	PD-71	PD-73	PD-74	PD-75	PD-76	PD-84	PD-85	TV01	BBAL
SiO_2	76.89	77.09	78.6	80.17	79.8	57.81	58.03	62.7	60.1
Al_20_3	12.15	11.94	12.18	11.14	11.07	14.71	14.49	17.7	16.2
TiO_2	0.74	0.62	0.64	0.43	0.36	0.5	0.49	0.37	0.85
Mgo	0.19	0.19	0.04	0.12	0.21	2.73	2.61	2.21	0.97
FeO			...	...	...	...	...	1.4	0.7
Fe_2O_3	1.74	1.98	0.65	1.41	1.75	0.61	0.42	0.79	1.36
MnO	0.02	0.02	0.01	0.01	0.01	0.6	0.07	0.07	0.04
CaO	0.94	0.72	0.71	0.51	0.44	6.31	8.14	3.68	4.55
Na_2O	7.03	7	7.25	6.51	6.42	9.01	9.31	8.79	9.39
K_2O	0.13	<.01	<0.01	<.01	<0.01	0.06	0.06	0.48	0.36
P_2O_5	0.18	0.16	0.11	0.12	0.09	0.02	0.08	0.02	0.08
CO_2	...	...	...	...	...	7.32	5.13	...	...
LOI	0.53	0.51	0.86	0.39	0.65	0.97	1.02	1.54	4.7
Total	100.54	100.24	101.06	100.83	100.81	100.65	99.85	99.75	99.3
Q	33.56	34.36	35.59	40.78	40.53	4.32	1.22	0.27	1.59
or	0.77	0.06	0.06	0.12	0.06	0.35	0.31	2.83	2.12
ab	59.38	59.11	61.29	55	54.22	75.31	65.43	74.36	79.46
an	1.21	1.13	0.66	1.12	1.36	...	...	7.42	1
di	1.95	1.21	0.22	0.55	0.2	...	5.04	7.97	...
hy	0.84	1.77	...	1.39	2.41	2.28	2.41	3.1	...
en	...	...	...	...	...	...	...	...	2.42
hm	...	...	0.1	0.42	0.1		...	...	0.42
mt	0.5	0.57	0.04	0.41	0.51	...	...	1.14	2.25
ii	1.41	1.18	1.22	0.82	0.68	0.95	0.77	0.7	...
tn	1.6	0.99	1.51	0.44	0.17	...	...	...	...
ru	0.06	0.2	0.01	0.24	0.28	...	...	...	...
ap	0.4	0.35	0.24	0.26	0.2	0.05	0.16	0.34	0.2
cc	...	...	...	...	...	16.65	10.09	...	6.25
wo	...	...	0.78	...	...	...	...	...	...
ns	...	...	...	...	...	0.12	0.89	...	...
ac	...	...	...	...	...	0.37	11.8	...	...
Total	101.68	100.93	101.72	101.55	100.72	100.4	98.12	98.13	95.71

1 PD-71: I-95 trondhjemite dike (dike 1), New Jersey (Benimoff et al. 1989), west contact.
2 PD-73: I-95 trondhjemite dike (dike 1), New Jersey (Benimoff et al. 1989), 0.3 m from west contact.
3 PD-74: I-95 trondhjemite dike (dike 1), New Jersey (Benimoff et al. 1989), center of dike.
4 PD-75: I-95 trondhjemite dike (dike 1), New Jersey (Benimoff et al. 1989), 0.3 m from east contact.
5 PD-76: I-95 trondhjemite dike (dike 1), New Jersey (Benimoff et al. 1989), east contact.
6 PD-84: I-95 trondhjemite dike (dike 2), New Jersey.(Benimoff et al. 1989)
7 PD-85: I-95 trondhjemite dike (dike 2), New Jersey.(Benimoff et al. 1989)
8 TV01: albitite dike, Travis Quarry, Staten Island, New York (Benimoff et al.1988).
9 BBAL: albitite dike, South End of the Bayonne Bridge, Staten Island, New York, Benimoff and Sclar, 1990

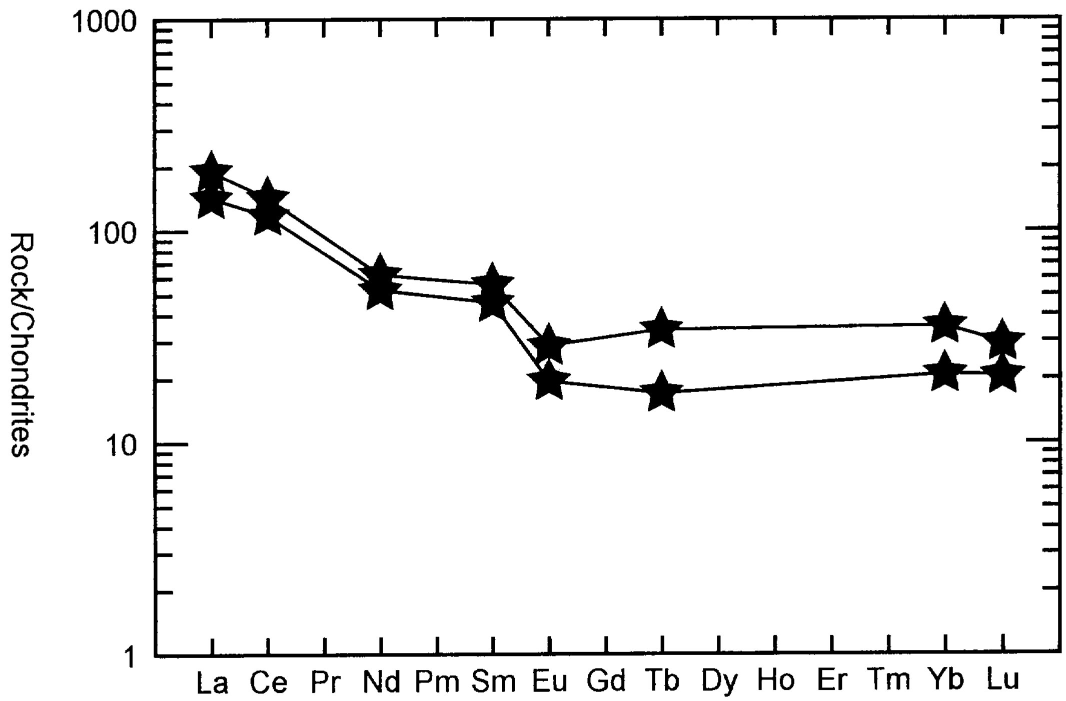


Figure 8. Chondrite normalized REE plot of I-95 Dike 1 and 2 compositions.

Because it is difficult to isolate the mesostasis granophyre which results from the differentiation of a tholeiitic magma for REE analyses, there is a paucity of REE analyses in the literature for any granophyre derived from a tholeiitic magma. The highly fractionated slope of the Palisades trondhjemite REE pattern with distinct enrichment in LREE is consistent with liquid extracted from a garnet rich residue. The REE slope of the Palisades trondhjemite resembles the slope commonly seen in dacites or trondhjemites interpreted as having developed from partial fusion in an island arc or continental arc setting out of an eclogite residue at high pressures. Although the Palisades trondhjemite dikes were presumably generated under low pressures the proposed Lockatong source residue includes a green pyroxene garnet hornfels assemblage that is not unlike the eclogite residue.

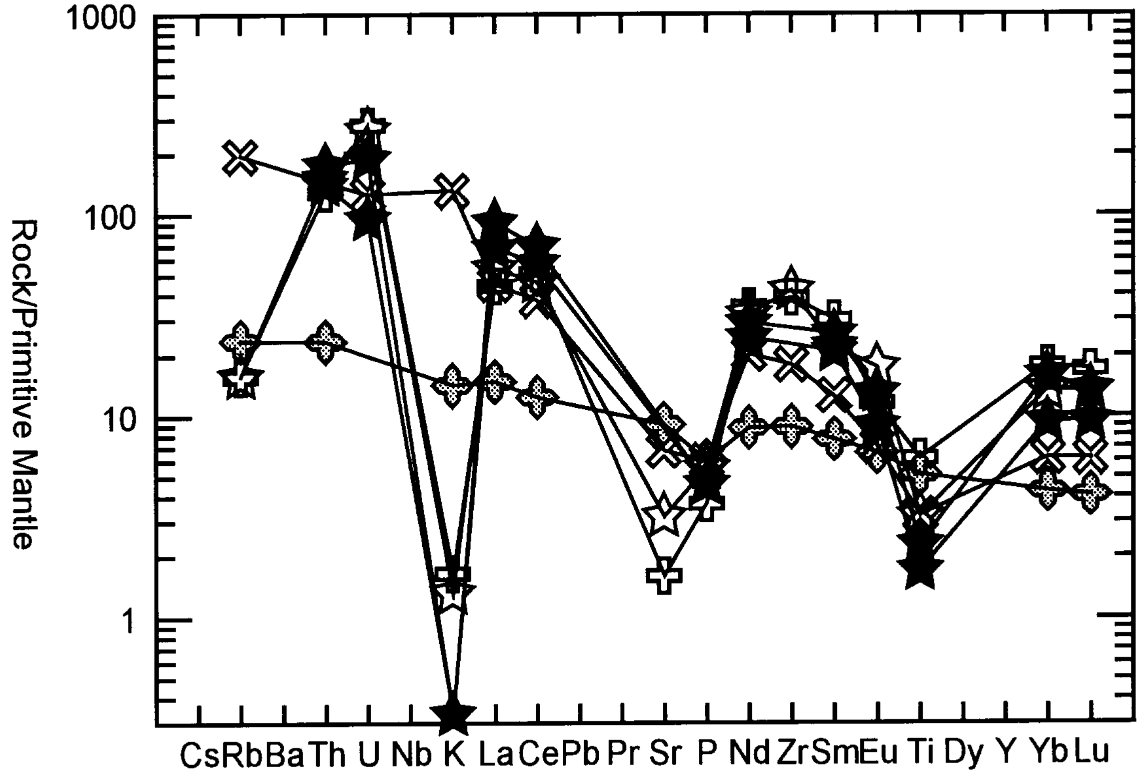


Figure 9. Sun and McDonough(1989) mantle normalized plot. Solid stars represent the I-95 trondhjemite dike.; X represents the North American Shale Composite; Open plus sign is a represents a sample of the Graniteville Quarry Trondhjemite; Shaded plus sign represents HTQ from Tollo and Gottfried, 1992; Open Star is the Lockatong hornfels xenolith from Graniteville. Note the similarity of the chemical signatures of the I-95 trondhjemite and the Graniteville Quarry Trondhjemite.

Our complete chemical analyses of the trondhjemite when plotted (figure 9) on primitive mantle normalized spider diagrams(Sun and McDonough, 1989) provide compelling evidence that the I-95 dikes are intrusions of fused Lockatong argillite. This data is consistent with our previous interpretation(Benimoff and Sclar, 1984);Benimoff et al. 1989). The general slope of the trondhjemite curve is negative. Furthermore, the range of the trondhjemite data plots within the range of the Lockatong argillite data and both display 3 very distinct negative anomalies: potassium, phosphorous and titanium. The trondhjemite data also plot very close to the North American Shale Composite(NASC) except for potassium. By comparison, all Palisades diabase analyses including fractionated granophyres plot as lines with different slopes, magnitudes and do not display the 3 negative anomalies. Our new data rules out magmatic fractionation as a process capable of generating the trondhjemites of this study. The difference in potassium between the trondhjemite and the NASC maybe due to hydrothermal leaching of potassium out of the Lockatong argillite before fusion occurred.

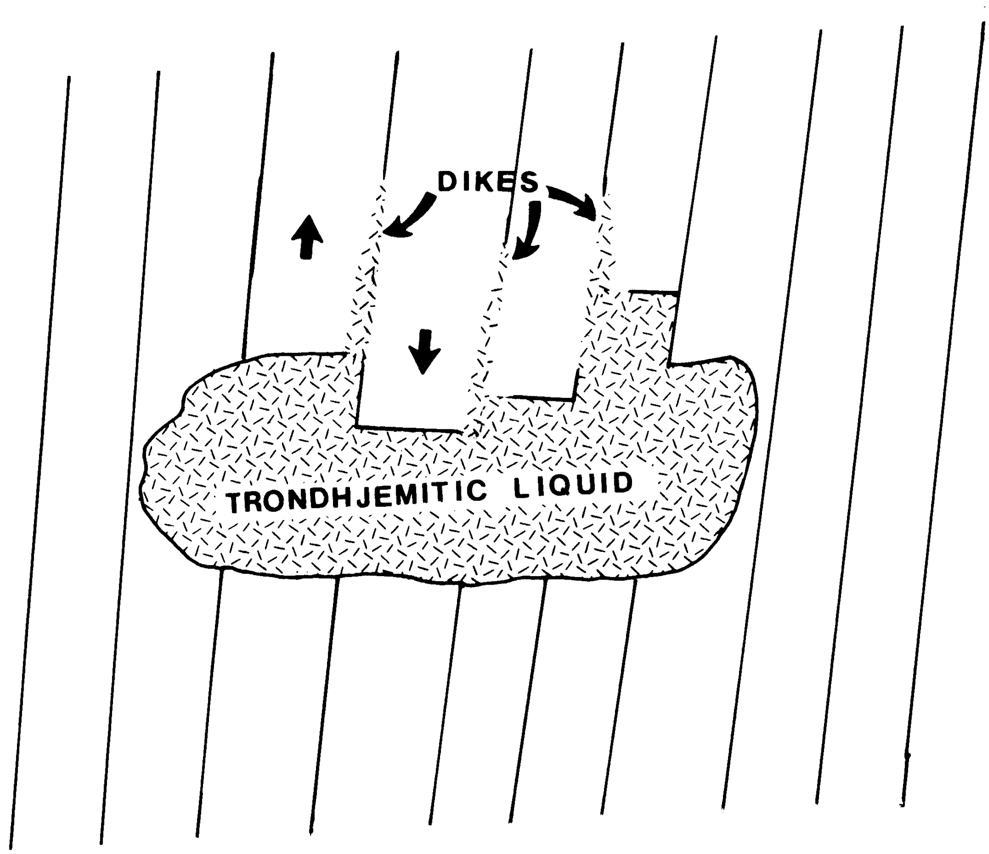


Figure 10. Cartoon showing foundering of joint blocks into pool of trondhjemitic magma resulting in injection of dikes parallel to joint blocks.

The mechanism of injection of the tabular vertical trondhjemite dikes may involve several processes that were occurring simultaneously. Partial fusion of Lockatong argillite would have resulted in a volume increase and a driving force for magma migration into low pressure sites(Figure 10). A low pressure environment was made available as soon as the volume reduction joints associated with diabase crystallization were developed. Since the Palisades intruded under a thick layer of Mesozoic sediments it is at least plausible that a vacuum may have been temporarily maintained long enough to enhance injection.

3. THE LAUREL HILL(SNAKE HILL) DIABASE QUARRY

Laurel Hill (also known as Snake Hill and Fraternity Rock) is an almost completely exposed elliptical body of early Jurassic diabase about 500 m in diameter, located 2.5 km west of the underlying Palisades sill (Figure 1). The diabase body has discordantly intruded the Triassic siltstones of the Passaic Formation and has generated a contact-metamorphosed and partially fused hornfels zone around the margins of the intrusion. On the basis of field observations of exposed discordant contacts and well-log data collected at seven sites near Laurel Hill, the intrusion appears to be a hypabysal volcanic neck. On the basis of chemical analytical data (Puffer and Benimoff , l997) Laurel Hill is co-magmatic with the Palisades Sill.

Field evidence of wall-rock fusion is found near igneous-sedimentary contacts, particularly along the southern contact exposed beneath the New Jersey Turnpike overpass (Figure 1). Passaic Formation red beds exposed <50 m from Laurel Hill diabase have been metamorphosed into a gray, fine-grained hornfels. As the igneous contact is approached the hornfels becomes increasingly mixed with a poorly defined discontinuous zone of leucocratic igneous rock up to 4 m thick that is recognized on the basis of an igneous texture. The leucocratic rock zone is approximately parallel to the contact of the diabase intrusion and is the source of a swarm of veins intruded into the diabase body.

Chemical analyses of Passaic wall rocks collected near Laurel Hill are bimodal with respect to alkali content (Table 3). Beds of potassic hornfels (4-9 %K2O) are interlayered with beds of sodic hornfels (4-9% Na_2O). Fusion of these hornfels produced two contrasting melts: fused potassic and fused sodic sediments.

About half of the leucocratic veins, and most of the fusion zone that crystallized in situ adjacent to the diabase intrusion, are composed of medium-grained potassic granitoids (Benimoff et al. l995). The potassic granitoids, ranging from quartz syenite to quartz monzonite, are equigranular mixtures of K-feldspar and plagioclase in varying proportions, together with about 5-15% quartz and minor amphibole, pyroxene, and ilmenite. The granitoids chemically contrast with the diabase but closely resemble the potassic wall rocks.

The melting of the potassic meta-sediments around the margins of the Laurel Hill diabase was probably fluxed in part the chlorides and sulfates in the sediments of the Passaic Formation and by any brackish groundwater content. In addition, Gardien el al. (l995) have experimentally demonstrated the important fluxing effect of muscovite on the fusion of biotite, quartz and Na feldspar. They have shown that anhydrous fusion of a muscovite + biotite + Na feldspar assemblage can begin at temperatures as low as 750oC at 1 Gpa. The potassic hornfels at Laurel Hill contain each of these minerals, although the pressure during fusion at Laurel Hill was probably closer to 0.05 Gpa corresponding to <2 km of burial. It may also be significant that the

Table 6. Chemical Compositions of Leucocratic Rocks in the Laurel Hill Diabase and Surrounding Wall Rocks. (from Puffer and Benimoff, 1997)

	Passaic Formation Wall-rock			Leucocratic rocks		
sample #		4	1	68	67	P-1
wt.%	ave. 14	potassic	sodic	Granitoid	Trondhj	Pegmatite
SiO_2	61.58	57	67.75	56.01	63.58	53.45
TiO_2	0.88	0.98	0.9	0.95	0.81	2.29
Al_2O_3	15.65	16.31	14.45	16.35	13.66	11.66
FeOt	5.7	7.74	3.75	8.11	5.46	12.27
MgO	3.67	5.71	3.91	5.04	2.34	4.49
MnO	0.09	0.12	0.05	0.05	0.09	0.32
CaO	1.8	2.3	0.51	1.94	1.09	8.78
Na_2O	4.77	3.79	6.54	5.12	7.86	3.83
K_2O	2.55	4.87	0.22	3.51	0.08	0.56
P_2O_5	0.16	0.11	0.17	0.08	0.26	0.4
LOI	2.85	0.92	1.69	1.97	4.58	1.74
Total	99.71	99.85	99.94	99.13	99.81	99.79
ppm						
Cr	87	87	70	111	92	37
Ni	42	52	44	62	54	8
Rb	89	201	38	86	47	nd
Sr	185	213	110	406	188	nd
V	151	166	128	177	149	511
Zn	65	79	60	82	58	nd
Zr	337	177	332	184	415	128

experimental fusion of this mineral assemblage generated a peraluminous melt consistent with the peraluminous norms of Laurel Hill potassic wall rocks and granitoids. Potassic wall-rock sample 4 (table 3) contains 0.9% normative corundum, and potassic granitoid sample 68 (table 6) contains 0.8% normative corundum.

Evidence of a high degree of partial melting includes the close chemical correspondence of the potassic granitoid and the potassic meta-sediments around Laurel Hill (table 6) and the presence of only minor quantities of mafic rock near the diabase that could be interpreted as a mafic refractory residue.

About half of the leucocratic veins, but a minor fraction of the fusion zone that crystallized in place, are medium-grained trondhjemites. Trondhjemite veins are composed of albite with about 10 to 35% quartz and minor quantities of actinolite, pyroxene, ilmenite, and sphene. Some of the quartz is granophyrically intergrown with albite. The chemical composition of the trondhjemites closely resembles the sodic Passaic hornfels and siltstone layers that were the probable sources of the trondhjemite magma. Schairer and Bowen (l956) show that eutectic melting of an albite and quartz assemblage, such as the sodic Passaic beds, occurs at 1062°C to yield a trondhjemite liquid. The resulting liquid contains about 35% normative quartz, consistent with typical normative values found in the Laurel Hill trondhjemites. The experimental work of Schairer and Bowen (l956) was anhydrous at 1 atm pressure, but in the presence of brackish groundwater and pressures corresponding to the approximately 2 km depth of burial, fusion temperatures <1062°C might be expected.

4. THE GRANITEVILLE QUARRY, STATEN ISLAND, NEW YORK

It is a real treat to view the parent of an igneous rock adjacent to that igneous rock. Such is the extraordinary case at the Graniteville Quarry(Figure 11) where a xenolith of sodium-rich Lockatong argillite enclosed in the basaltic magma of the Palisades sill resulted in coexisting silicic and mafic melts. This phenomena has been exhaustively studied by Benimoff and Sclar, 1978, 1980, 1984, 1988, 1992, and Sclar and Benimoff(1993), and their work is presented below.

The occurrence of a coarse-grained igneous rock in direct contact with its protolith is rare. Much of the uncertainty regarding the origin of coexisting silicic and basic igneous rocks arises either because of the absence of contiguous parental material or because we cannot identify unequivocally the parental material through either geochemical and/or petrographic study. Here at the Graniteville Quarry all three contiguous rocks are preserved in the arrested state.

A xenolith of Lockatong argillite is exposed in the Palisades diabase in a quarry at Graniteville, Staten Island. The xenolith has been recrystallized to a hornfels. It is a vertically dipping slab, 0.3 to 0.5 m wide, and some 30 m long. The xenolith strikes N 30° W. The bottom of the xenolith is not exposed. Based on the measured density of 2.60 g/cc and 2.95 g/cc for the xenolith and the enclosing diabase, respectively, we conclude that the xenolith was derived from the Lockatong formation below the sill. Between the diabase and the hornfelsed xenolith is a sharply bounded interface zone of coarse-grained igneous rock. The interface zone ranges from 5 to 12 cm in thickness and completely surrounds the xenolith. We have categorized the coarse-grained rock of the interface zone as a melanocratic pyroxene trondhjemite[1]

The diabase-trondhjemite interface and the trondhjemite-hornfels interface are sharp and irregular. Pyroxene and plagioclase in the diabase within 5 mm of the diabase-trondhjemite interface show the effects of hydrothermal alteration. In the pyroxene, this is manifested by the development of hornblende and actinolite; in the plagioclase, sericite formed.

Mineral compositions were obtained with an ARL microprobe using wave-length dispersive and energy-dispersive analysis. Raw data were corrected by means of the Bence-Albee interactive routine. Natural and synthetic standards were utilized. The bulk chemical compositions of whole-rock samples were obtained by X-ray fluorescence analyses; FeO was determined by wet chemical analysis.

[1]We note that, in the classification of Streckeisen et al. (1973), this rock would be classified as albite granite. However, we consider the term albite granite a contradiction in terms inasmuch as a granite by definition is a K-feldspar-bearing phanerite and K-feldspar is absent in this phanerite. We prefer the classification of O'Connor (1965) in which a quartz-bearing phanerite containing plagioclase of composition Ab_{99} as the dominant feldspar is classified as a trondhjemite.

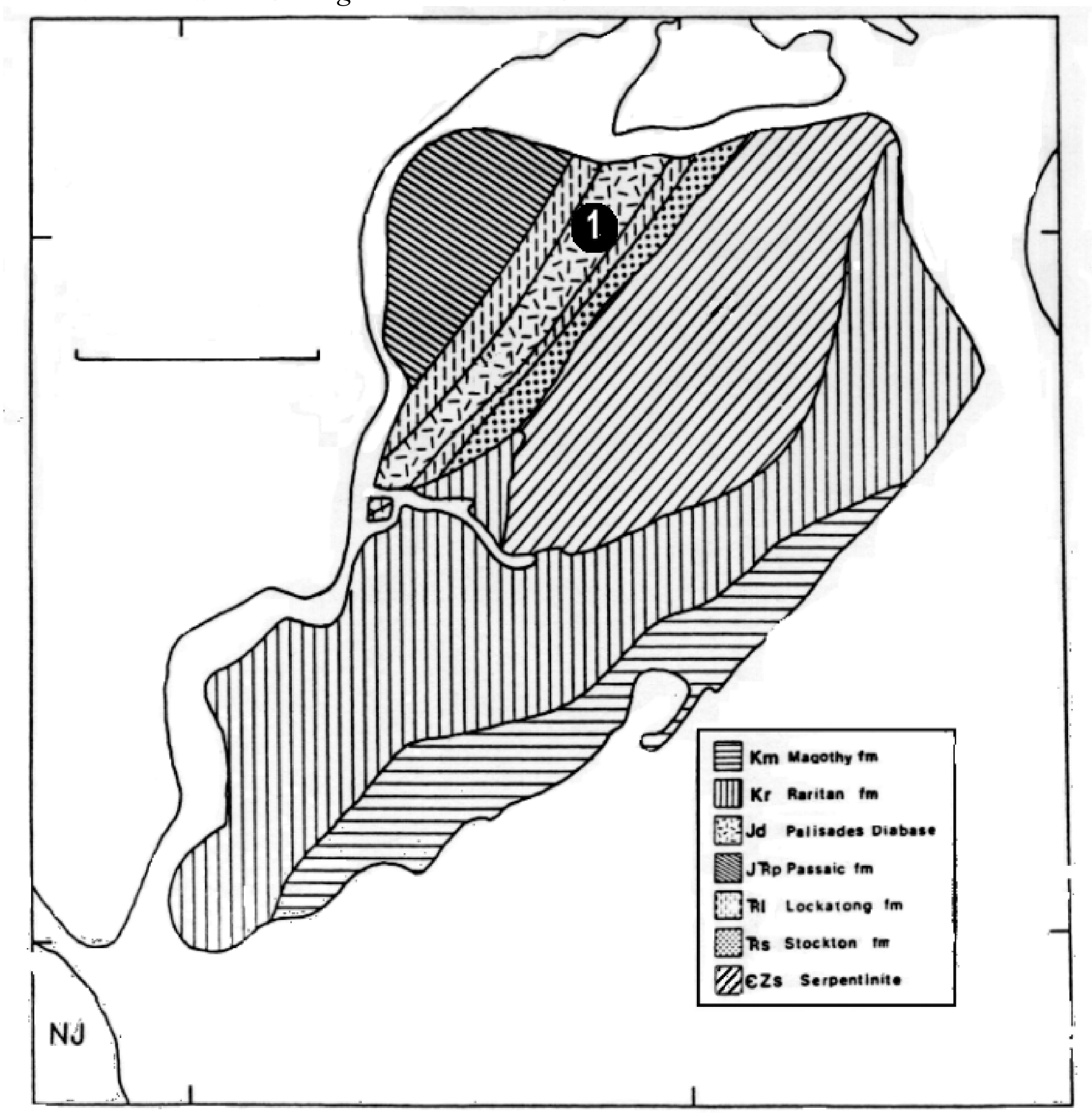


Figure 11. Geologic Map of Staten Island, NY showing the location of the Graniteville Quarry(1).

THE DIABASE

The diabase is composed dominantly of plagioclase ($An_{61}Ab_{38.8}Or_{0.2}$) and augite ($En_{34-44}Fs_{17-31}Wo_{35-42}$). The augite contains exsolution lamellae of pigeonite on (001), and typically exhibits simple contact twinning on (100). A granophyric intergrowth of quartz and K-feldspar is present in minor amounts. Grains of titanomagnetite with oxidation lamellae of ilmenite and discrete grains of ilmenite are common. No attempt was made to use the Buddington and Lindsey (1964) relationship to obtain an fo_2 and temperature of crystallization. However, an independent approach to determining the temperature of the diabase magma will be discussed below.

THE TRONDHJEMITE

The trondhjemite is composed dominantly of quartz-albite granophyre containing large discrete crystals of albite and Ca-rich pyroxene. Minor constituents include interstitial calcite, titanite, ilmenite, optically homogeneous titanomagnetite, nickelian and cobaltian pyrrhotites, apatite, and sphalerite. The modal mineral percentages are clinopyroxene 38, albite 38, quartz 18, titanite 2.7, calcite 1.3, and opaques 2.0.

There is a complete gradation from Fe-rich compositions close to the diabase-trondhjemite interface to Mg-rich compositions close to the trondhjemite-hornfels interface. Pyroxenes adjacent to the trondhjemite-hornfels interface are euhedral crystals 5 to 30 mm in length. Cores of some of these clinopyroxenes are enriched in Mg; Mg/Fe (atomic) ranges from 2.7 in the cores to 2.2 in the rims. Within a distance of 16 mm from the diabase-trondhjemite interface, a few augite crystals occur which are similar to the augite of the diabase inasmuch as they contain lamellae of pigeonite parallel to (001). These crystals are enclosed in the trondhjemite and are probably xenocrysts derived from the diabase. Near the diabase-trondhjemite contact, high-Ca clinopyroxene related to the trondhjemite occurs as overgrowths in optical continuity with cores of pigeonite-augite intergrowths derived from the diabase. The pigeonite-augite intergrowths apparently served as nucleation sites.

Some of the high-Ca clinopyroxene crystals in the trondhjemite were altered hydrothermally by post-magmatic fluids to the assemblage: actinolite + sphene + calcite. There is a complete chemical gradation in pyroxenes from Fe-rich compositions close to the diabase-trondhjemite interface to Mg-rich compositions close to the trondhjemite-hornfels interface. Pyroxenes adjacent to the trondhjemite-hornfels interface are euhedral crystals 5 to 30 mm in length; typical cross-sections are shown in Figure 3. Cores of some of these clinopyroxenes are enriched in Mg; Mg/Fe (atomic) ranges from 2.7 in the cores to 2.2 in the rims. Within a distance of 16 mm from the diabase-trondhjemite interface, a few augite crystals occur which are similar to the augite of the diabase inasmuch as they contain lamellae of pigeonite parallel to (001). These crystals are enclosed in the trondhjemite and are probably xenocrysts derived from the diabase. Near the diabase-trondhjemite contact, high-Ca clinopyroxene related to the trondhjemite occurs as overgrowths in optical continuity with cores of pigeonite-augite intergrowths derived from the diabase. The pigeonite-augite intergrowths apparently served as nucleation sites.

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Albite and quartz

Albite ($Ab_{99}An_{0.52}Or_{0.44}$) occurs as early-formed discrete euhedral crystals and also as a major component of the granophyric intergrowth with quartz. The early albite crystals are localized at the diabase-trondhjemite interface. Some of these large crystals of albite exhibit Carlsbad twinning. D-1 Diabase: Adjacent to Trondhjemite (TA); D-2 Diabase taken 47 meters S30°W of D-1; TA Trondhjemite: North end of xenolith; TB Trondhjemite: south end of xenolith; XA Xenolith north end of outcrop; XB xenolith south end of outcrop

Sphalerite

Sphalerite is present as euhedral crystals embedded selectively in the albite of the granophyre. As seen in thin section dark well-defined cores contain 13-16 mole% FeS, whereas pale yellow-brown rims contain between 0.2 and 2.0 mole percent FeS. Electron microprobe data show that there is a sharp compositional discontinuity between the core and the rim. In reflected light the core-rim relationship is resolvable because of the higher reflectivity of the core, and in transmitted polarized light between crossed nicols, the rim appears illuminated because of internal reflections.

Enclosed selectively in the albite of the albite-quartz granophyre, we found, in thin sections, about a dozen isolated euhedral to subhedral crystals of sphalerite 30-70 µm in diameter. These crystals are transparent, with a high refringence. They are isotropic and optically zoned, with a dark-brown core and pale brown to colorless rim. Also enclosed in the trondhjemite are minute interstitial anhedral grains of Ni- and Co-bearing pyrrhotite, although nowhere is the pyrrhotite in contact with the sphalerite crystals.

The euhedral grains of zinc sulfide typically have the shape of an equilateral triangle, the corners of which are truncated . This results in a six-sided form with alternating long and short sides that resembles most closely the crystallographic form of an isometric positive tetrahedron modified by a negative tetrahedron. This morphology is characteristic of sphalerite. Nevertheless, such a morphology is not incompatible with the hexagonal symmetry of wurtzite if the grains are considered to be basal (0001) sections or near-basal sections. However, the absence of anisotropic longitudinal *c*-axis sections suggests that the crystals that nucleated in the trondhjemite magma were sphalerite and not wurtzite. Some of the euhedral crystals show deep embayments that may be indicative of magmatic resorption.

Compositional data were obtained with an ARL electron microprobe using wavelength-dispersion analysis. Standards were synthetic ZnO for zinc and pyrite for iron and sulfur. Operating conditions were 15 kV and 40 nA. The raw data were corrected for atomic number, absorption, and fluorescence effects (ZAF). Results of the electron-microprobe analysis shows that these isotropic grains are zoned zinc sulfide, with a relatively high-iron core (10-16 mole % FeS) and low-iron rim (1-5 mole % FeS). The core-rim boundaries are abrupt and show concentration gradients in excess of 3 mole % FeS per µm. Some of the grains have an asymmetrically zoned rim composed of alternating relatively high-iron (5 mole % FeS) and relatively low-iron (1 mole % FeS) zones.

Table 7.Chemical analyses and CIPW norms of the xenolith and the associated trondhjemite and diabase from the Graniteville Quarry, Staten Island, New York (from Benimoff and Sclar, 1984)

Chemical Analyses, weight % oxides

	XA	**XB**	**TA**	**TB**	**D-1**	**D-2**
SiO_2	74.8	63.5	58.4	58.2	51.8	52
Al_2O_3	11.6	16.1	6.75	6.91	16.8	16.7
TiO_2	0.61	0.75	1.63	1.48	1.17	1.13
MgO	0.23	0.3	4.89	5.14	4.91	4.74
FeO	0.1	0.1	4.35	5.6	6.7	7.85
Fe_2O_3	0.31	0.47	1.51	1.93	2.75	1.74
MnO	0.02	0.04	0.14	0.16	0.16	0.19
CaO	2.07	5.6	13.1	11.5	8.79	10.2
Na_2O	6.67	9.48	3.76	3.81	3.22	2.58
K_2O	0.07	0.12	0.12	0.14	1.4	0.76
P_2O_5	0.09	0.13	0.3	0.09	0.14	0.15
LOI	1.48	3.85	4.71	2.88	1.15	0.23
Total	98.05	100.34	99.66	97.84	98.99	98.27

CIPW Norms

	XA	**XB**	**TA**	**TB**	**D-1**	**D-2**
Q	35.41	7	19.32	16.27	1.42	4.63
Or	0.39	0.72	0.84	8.29	4.51	3.23
Ab	56.4	80.24	35.41	32.25	27.27	21.82
An			0.39	1.34	27.24	31.74
Wo			13.49	15.46	6.48	7.49
En	0.57	0.74	9.19	9.62	3.67	3.72
Fs			3.25	4.92	2.53	3.62
En			2.99	3.18	8.53	8.08
Fs			1.06	1.62	5.87	7.85
mt			2.2	2.8	3.98	2.52
hm	0.34	0.46				
il	0.26	0.09	3.1	2.81	2.22	2.14
tn		1.2				
ru		0.22				
ap	0.19	0.28	0.65	0.19	0.31	0.34
C	0.2					
cc	3.36	8.76	10.71	6.65		
H_20					1.15	0.23
Total	97.59	100.49	99.68	97.85	98.96	98.69

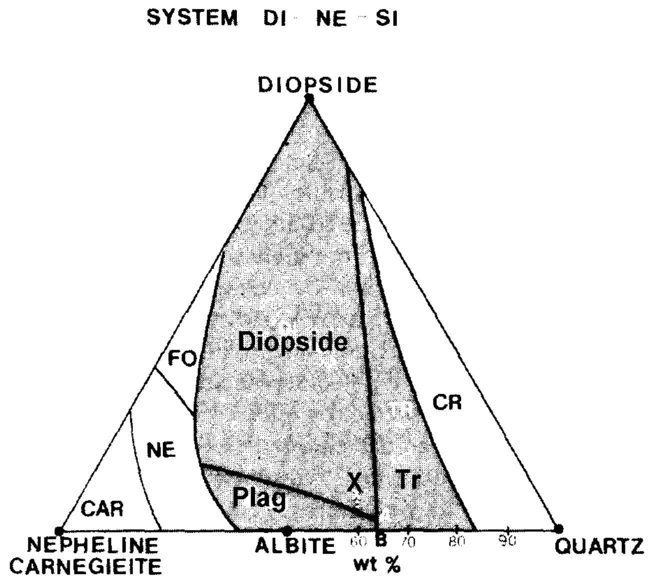

Figure 12. The System diopside-nepheline-silica(Shairer and Bowen, 1960). The deduced composition of Trondhjemite magma is shown by X which lies at a temperature of 1150°C.

Minor minerals

The carbonate phase in the trondhjemite is calcite which occurs as single-crystal interstitial fillings. There is no evidence, such as colliform or crustiform structures, which would indicate that this calcite is a product of cavity filling. The calcite, therefore, appears to be a late igneous mineral.

Aggregates of ilmenite are locally surrounded by euhedral to subhedral crystals of titanite. Titanite is also present as discrete euhedral crystals which appear to be part of the early magmatic suite. Optically homogeneous euhedral grains of titianomagnetite occur as inclusions in the high-Ca clinopyroxene. Nickel- and cobalt-bearing pyrrhotite is present as microscopic grains. Euhedral apatite crystals occur in the granophyre and as inclusions in the high-Ca clinopyroxene.

CRYSTALLIZATION SEQUENCE

The sequence of magmatic crystallization in the trondhjemite, as shown by petrographic relationships is apatite, titanomagnetite, ilmenite, high-Ca clinopyroxene, discrete crystals of albite, sphene, sphalerite, albite-quarts granophyre, and interstitial calcite. The crystallization sequence of the major phases (albite, quartz, and pyroxene) is in accord with the most pertinent ternary phase diagram of this system (diopside-nepheline-quartz) as determined by Schairer and Yoder (1960) (see dark region of Figure 12). If crystallization commenced at x, which is close to 1160°C, the temperature of crystallization of the trondhjemite deduced in subsequent discussion in this paper, diopsidic clinopyroxene would be the first major phase to crystallize. When the diopside-plagioclase boundary was reached on cooling, albite and diopside crystallized. When the temperature reached 1073°C, an albite-quartz granophyre crystallized until the albite-quartz eutectic was reached at 1062°C, which is the dry minimum melting temperature in the quartz-albite system(see figure 4) at one atmosphere (Schairer and Brown, 1956). The composition of this eutectic is 31.5 quartz. 68.5 albite which is close to the normative quartz and albite content of part of the parental hornfels of the trondhjemite.

The xenolith

Petrographic examination shows that the xenolith is now a hornfels and exhibits a hornfelsic texture. The hornfels is composed dominantly of albite and quartz and subordinately of calcite, titanite, apatite, ilmenite, and actinolite. The modal mineral percentages are albite 66, quartz 30, titanite 2.3, calcite 0.9, apatite 0.5, and actinolite 0.3. The bulk composition of the xenolith is variable, as shown in Table 1, which is not unexpected for a meta-sedimentary rock. Normative albite ranges from 56.4 to 80.2 wt.%, whereas normative quartz ranges from 7.0 to 35.4 wt.%.

The hornfels was derived from the Newark Supergroup (Olsen, 1980)of sedimentary rocks which is associated with the Palisades Sill. This group of rocks consists of the Stockton, Lockatong, and Brunswick formations (Van Houten, 1964, 1965, 1969, 1971), The protolith for the xenolith was probably a silty lacustrine sediment rich in sodium and carbonate, but low in potassium, which is characteristic of some parts of the Lockatong formation.

Discussion

It is apparent from spatial relationships and petrochemical data that the margins of the xenolith of Lockatong argillite fused as a result of being immersed in the diabase magma. Based on the dry albite-quartz equilibrium diagram(Figure 13) of Schairer and Bowen (1956), the temperature of the diabase magma surrounding the hornfelsed xenolith must have been about 1160°C in order to have effected melting of a xenolith of the bulk composition shown in Table 7. Hess (1956) noted that dolerites crystallize at a temperature of about 1100°C. Later, Tilley et al. (1964) correlated experimental determinations of liquidus temperatures with an iron-enrichment index [$(FeO + Fe_2O_3)/(MgO + FeO + Fe_2O_3)$] in natural rocks. The iron-enrichment index of the chilled-zone of the Palisades diabase (Walker, 1969a) is 0.58 which correlated with a liquidus temperature of 1220°C. The iron-enrichment index of specimen D-2 is 0.67 which correlated with a liquidus temperature of approximately 1160°C. In the case presented in this study, the presence of pyroxene suggests that the xenolith of Lockatong argillite was dry at the time of fusion. In addition, the position of the xenolith in the middle of the sill suggests that any original water in the sedimentary rock would have been expelled long before it reached this position.

Several lines of evidence suggest that the location of this xenolith is approximately 525±50 feet above the base of the sill. Normalizing the diabase surrounding the xenolith to the Englewood Cliff section of Walker (1969a) shows that:

(1) modal analysis of D-1 lies between W-N-60 and W-R-60 of Walker, thus placing upper and lower constraints of 560 feet and 365 feet above the base of the sill for the xenolith;

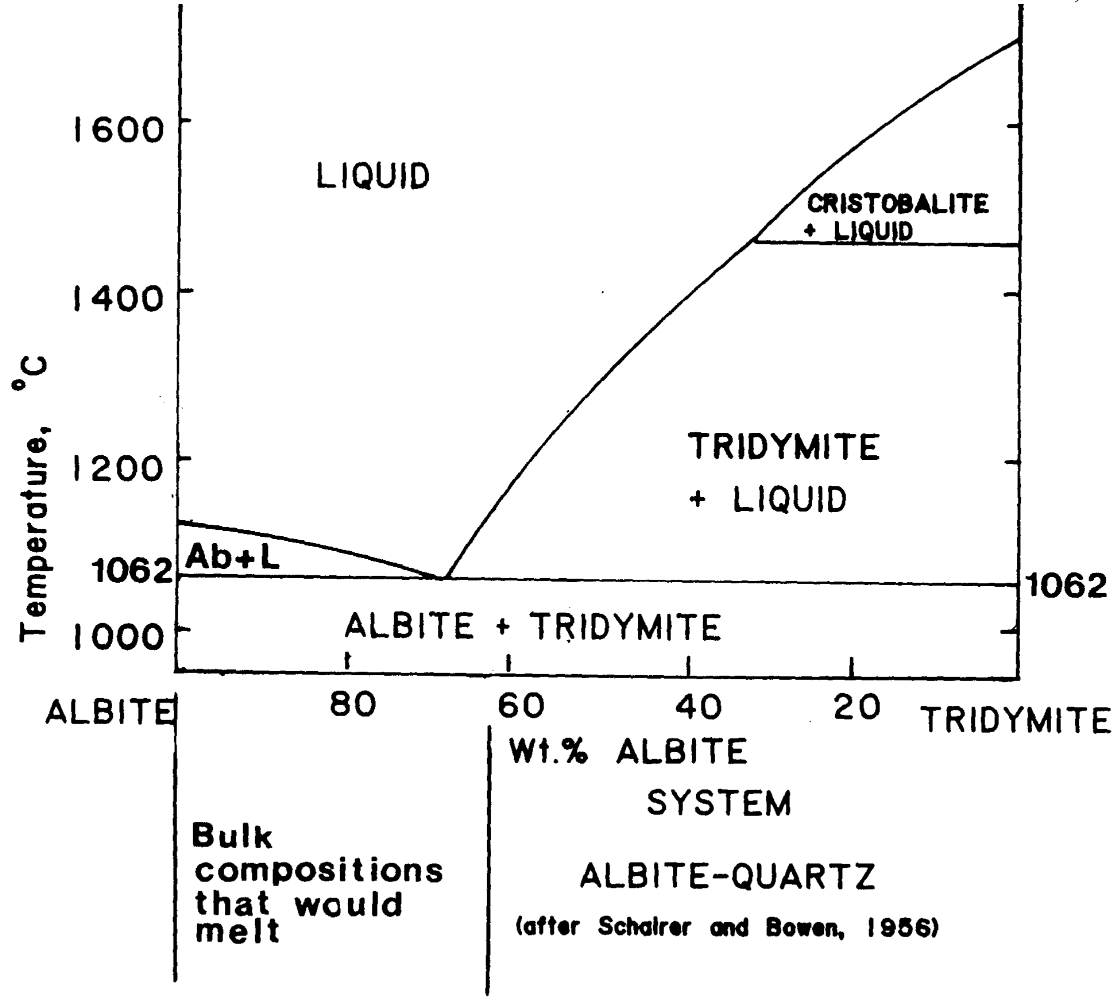

Figure 13. The dry albite-SiO_2 equilibrium diagram of Schairer and Bowen (1956)

(2) the composition of augite and plagioclase in D-1 is in accord with the above constraints and indicates the middle differentiation series of Walker 1969a;
(3) the mafic fractionation indices [(Fe) + Fe_2O_3)(100%)/(MgO +Fe_2O_3)] of W-N-60 and D-2 are respectively 66.55 and 66.92(Figure 15) , and (4) D-2 plots directly on the differentiation trend for the Palisades Sill (Figure 14). D-2 plots closer to W-N-60 than any of the Englewood Cliff specimens of Walker (1969a).

A thickness of 900 feet is assumed for the Palisades Sill at Graniteville, Staten Island in accord with a subsurface intersection of the sill revealed in drill-core at Sewaren, N.J. (Van Houten, 1969). At 525 feet above the base of the sill, a xenolith of Lockatong argillite might easily have fused.

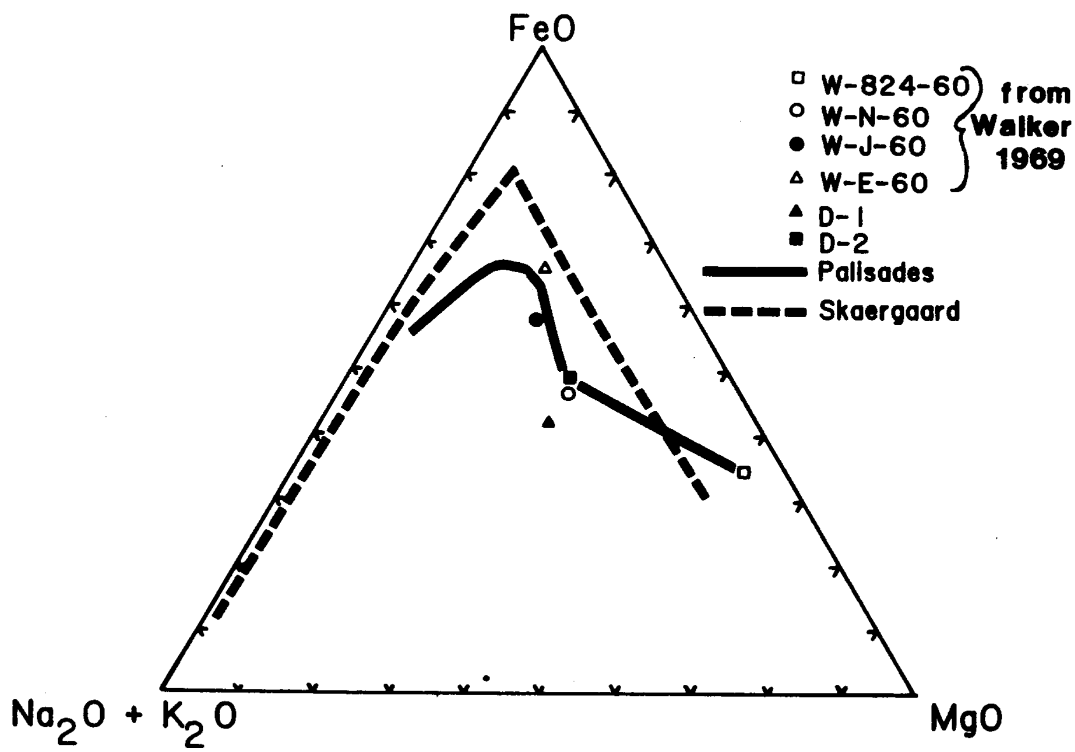

Figure 14. MgO-FeO-(Na2O+K2O) diagram which shows the composition of the Palisades Diabase from the Englewood Cliffs section of walker (1969a). D-2 falls directly on the Palisades diabase differentiation trend of walker(1969a). D-1 shows Fe -depletion and alkali enrichment. FeO represents total Fe expressed as FeO.

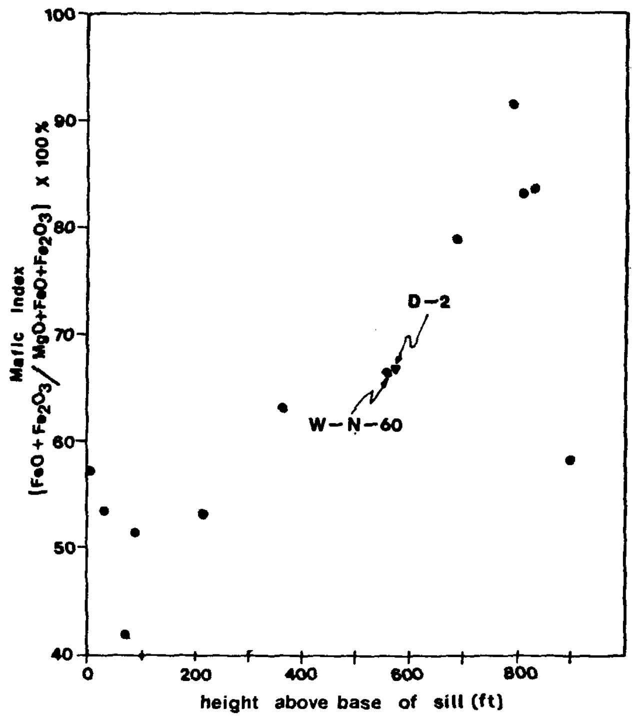

Figure 15. Plot of mafic Index vs. height above the base of the Palisades Sill. D-2 plots adjacent to W-N-60 of Walker (1969)a.

There are differences between the bulk chemical composition of the trondhjemite and that of the Lockatong hornfels (see Table 7), which are due to the paucity of the ferromagnesian component in the Lockatong hornfels. This suggests that Fe^{2+},Mg^{2+}, and some Ca^{2+} diffused from the diabase liquid into the fusion zone of the xenolith and were incorporated into the high-Ca clinopyroxene now present in the trondhjemite. This gave rise to a more complex bulk chemistry for the trondhjemite than would have been obtained solely by fusion of the xenolith. As shown in Figure 14, D-1 and D-2 differ in that D-1 is relatively enriched in sodium and relatively depleted in iron. This indicates that Na^{+} ions diffused out of the trondhjemite liquid and into the diabase magma, whereas Fe^{2+} ions diffused from the diabase magma into the trondhjemite liquid.

The identity of two contiguous magmas of diverse composition may be maintained for a limited amount of time (Yoder, 1973). The coexisting melts described in this study did not mix, although there appears to have been simultaneous diffusion of ions across the liquid-liquid interface. The coarse grain size of the trondhjemite (especially the large euhedral clinopyroxene crystals) and the evidence for chemical diffusion strongly indicate that these two melts coexisted for some time. Yoder (1973) suggests that the failure of two contrasting magmatic liquids to mix might be due to (1) immiscibility, (2) short time of contact, or (3) high viscosity occasioned by volatile loss.

Silicate liquid immiscibility involves the splitting of a homogeneous magma into two immiscible fractions upon cooling (Roedder, 1978). This occurs when $\Delta \overline{H}_{mix}$

(enthalpy of mixing), is greater than the entropy of mixing term $\mathbf{T}\Delta\overline{\mathbf{S}}_{\mathbf{mix}}$ so that $\Delta\overline{G}_{mix}$ in equation (1) is positive

$$\Delta\overline{G}_{mix} = \Delta\overline{H}_{mix} - T\Delta\overline{S}_{mix} \qquad (1)$$

(Ryerson and Hess, 1978). An upward convexity in the G-X surface of the liquid is produced such that the ΔG of the system is minimized by the liquid-liquid separation (Hess, 1977; Ryerson and Hess, 1978). This is not the case in the study presented in this paper, inasmuch as the spatial relationship of the trondhjemite to the xenolith and the diabase, and the petrochemical data reveal that the trondhjemite is a fusion product of the xenolith.

In order to determine whether or not the diabase magma and the trondhjemite magma were in an immiscible relationship, their respective positions on an FeO-(Al_2O_3 + K_2O)-SiO_2 diagram (Watson, 1976) were plotted(Figure 16). This showed that the diabase and trondhjemite compositions plot outside of the liquid-immiscible field. Also it was noted that any attempt to draw conjugate tie lines between respective diabase and trondhjemite compositions resulted in lines which were perpendicular to the conjugate tie lines within the field of liquid immiscibility. We also plotted the diabase, trondhjemite, and xenolith compositions, respectively, on the hypothetical pseudoternary phase diagram, [SiO_2] - [Na_2O-K_2O-Al_2O_3] - [FeO + TiO_2 + MnO + MgO + P_2O_5] (see Grieg, 1927; Weiblen and Roedder, 1973; McBirney, 1975), but conjugate tie lines drawn between diabase and trondhjemite compositions are still perpendicular to the tie lines shown in the field of liquid immiscibility (Figure 17). Therefore, the diabase and the trondhjemite do not appear to be in an immiscible relationship.

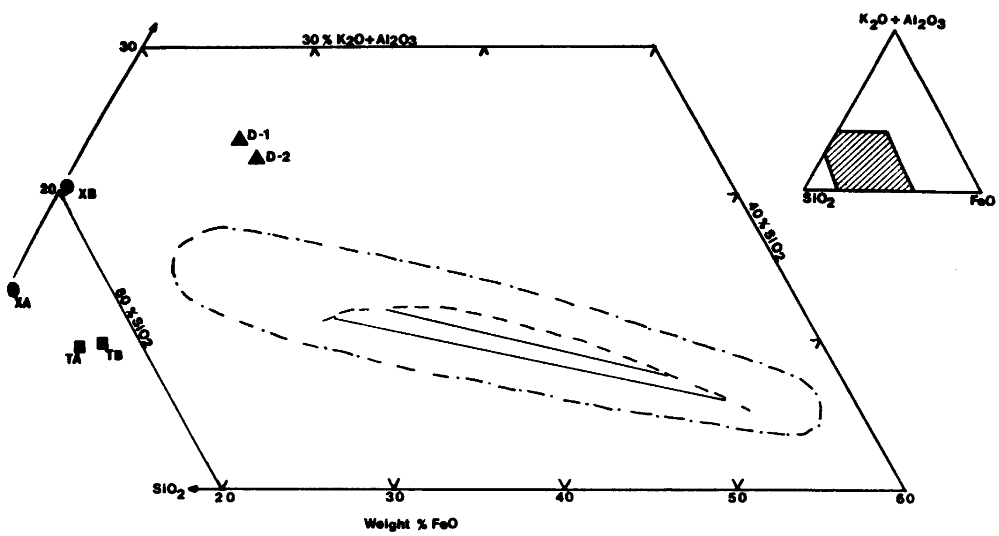

Figure 16. FeO-(Al_2O_3 + K_2O)-SiO_2 diagram (Watson, 1976).

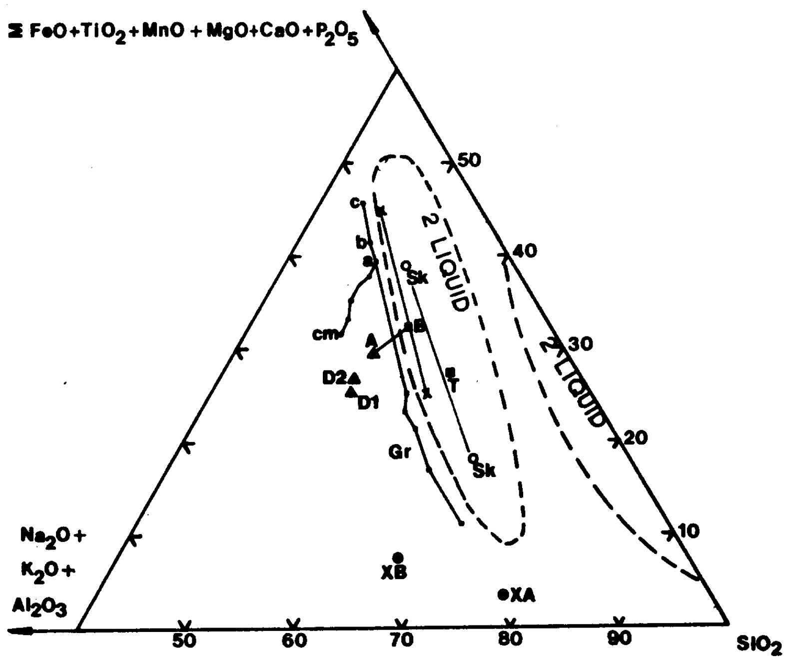

Figure 17. Diabase, trondhjemite, and xenolith compositions plotted on on the hypothetical pseudoternary phase diagram, [SiO_2] - [Na_2O-K_2O-Al_2O_3] - [FeO + TiO_2 + MnO + MgO + P_2O_5] (see Grieg, 1927).The high-temperature and low-temperature imiscibility fields are marked by dashed lines(system fayalite-leucite-silica (after Roedder, 1951). Also plotted are the compositions of (1) The Rattelsnake Hill basalt (A)and its mesostasis (B) (Philpotts, 1979, and a pair of conjugate liquids (Sk) produced experimentally from mixtures of late stage Skaergaard rocks (McBirney, 1975).The calculated Skaergaard differentiation trend is shown for comparison (cm=chilled margin;a, b, and c reprent the upper zone;Gr=melanocratic granophyres. T= specimens TA and TB; The x's represent immiscible liquids in the Rattelsnake Hill basalt (Philpotts, 1979).

It seems highly improbable, in view of the geological setting in which these contrasting liquids occur, that lack of mixing was due to the short time of contact. It also seems highly unlikely that lack of mixing was due to high viscosity occasioned by volatile loss inasmuch as the xenolith would have been devolatilized before melting of its margins occurred. The high viscosity of the trondhjemitic silicic liquid appears to be the factor responsible for the lack of mixing.

ORIGIN OF THE SPHALERITE

Sphalerite is a remarkably refractory sulfide, with a melting point in excess of 1800°C (Kullerud 1966). Nevertheless, all economic deposits of sphalerite are of hydrothermal origin, and occurrences of magmatic sphalerite are rare (Ramdohr 1980, p. 519). However, a few occurrences of magmatic sphalerite have been reported. Desborough (1963) reported the occurrence in Missouri of sphalerite of apparently magmatic origin; it occurs as disseminated grains in unaltered intrusive tabular bodies of olivine diabase, coarse ophitic gabbro, and layered gabbro. He indicated that, because of its lack of distinctive optical properties, sphalerite may easily be overlooked or mistaken for ilmenite in microscopic examination of thin and polished sections. He also reported that sphalerite may be a rare constituent of magmatic droplets of iron, nickel, and copper sulfides in mafic igneous rocks. Wilson (1953) stated that some zinc may enter the magmatic Cu-Ni-Fe sulfides, although zinc typically reaches its maximum concentration in a mafic magma at a much later stage than the formation of the magmatic sulfides. Naldrett (1989) pointed out that zinc is conspicuously absent in magmatic sulfide ore, although some of the magmatic sulfide ores of Sudbury average 100-200 ppm of zinc, and some copper-rich stringers contain up to 3700 ppm of zinc. According to Naldrett (1989, p. 52), the sulfide melt - silicate melt partition coefficient of zinc is about 1, and, consequently, zinc does not concentrate in sulfide melts. In this paper, we report evidence for the direct crystallization of magmatic sphalerite from a felsic silicate melt.

Petrographic relationships indicate that the sphalerite is part of the magmatic suite and that is probably crystallized from the silicate melt at temperatures between 1062° and 1073°C (before the albite of the quartz-albite granophyre and after the augite phenocrysts and early discrete crystals of albite). Based on the geological setting, pressure at the time of crystallization was probably less than 2 kilobars. A possible source of the zinc and sulfur was the sedimentary xenolith (now a hornfels), which contains 20-54 ppm of zinc. Another possible source of the zinc and sulfur is the surrounding uncontaminated diabase, which contains 50 ppm of zinc. Zinc and sulfide ions could have diffused from the basaltic magma across the liquid-liquid boundary between the coexisting basaltic and trondhjemite magmas.

The equilibrium boundary between sphalerite (low-temperature polymorph) and wurtzite (high-temperature polymorph) in pure ZnS occurs at 1020°C at 1 atmosphere (Kullerud 1966); the pressure-dependence of the inversion is not known. The inversion temperature is lowered at about 960°C with 15 mole % of FeS in solid solution (Kullerud 1966). These phase relationships in the ZnS-FeS system suggest that wurtzite should have nucleated as the equilibrium phase in the temperature interval 1062-1073°C and then inverted to sphalerite on cooling below the solidus, but the morphology of the ZnS crystals suggests, at first, that sphalerite was the magmatic phase. However, it is important to note that several of the polytypes of wurtzite (3R, 9R, 12R, 15R, 21R) have symmetry *R3m* (Kostov & Min_eva-Stefanova 1982). It is possible, therefore, that the magmatic ZnS crystals developed as thin tabular forms of wurtzite parallel to (0001). They would display trigonal outlines like tourmaline and appear optically isotropic regardless of whether they inverted to sphalerite. Such an occurrence of wurtzite would be compatible with the estimated temperature of crystallization and the equilibrium relationships between wurtzite and sphalerite.

Toulmin *et al.* (1991) have recently reviewed the binary system (ZnS-FeS) with specific reference to the FeS content of sphalerite in association with pyrite and pyrrhotite as a function of temperature and pressure. However, the absence of pyrite and pyrrhotite in direct association with Graniteville sphalerite precludes the application of their conclusion to this occurrence.

REE STUDIES

This occurrence constitutes an exceptional circumstance in igneous petrology in which the source rock (xenolith) and the igneous daughter product (trondhjemite) are contiguous and in which the geological, petrographical, mineralogical, and chemical evidence point unequivocally to a parent-daughter relationship. This setting provides an opportunity to test whether REE signatures reflect the source of an igneous rock. Chondrite-normalized REE plots of the xenolith, the trondhjemite, and the contiguous Palisades diabase were prepared from REE analysis obtained by INAA. The trondhjemite and the xenolith plots are characterized by a pronounced negative europium anomaly (Eu 20-30 times chrondrites), LREE concentration of 80-100 times chrondrites, and HREE concentrations of 30-50 times chrondrites. By comparison, the diabase shows a positive europium anomaly (EU 19-26 times chrondrites), LREE concentrations of 18-40 times chrondrites, and HREE concentrations of 10-17 times chrondrites. It is concluded that the REE signature of an igneous rock does indeed reflect that of the source rock. Furthermore, a plot of trondhjemite and xenolith compositions on a Sun and McDonough(1989) spider diagram(Figure 18) present compelling evidence of a parent(xenolith) - daughter(trondhjemite) relationship between the two. When we compare the chemical signatures of the three contiguous rocks namely the xenolith, the trondhjemite and the Palisades diabase we conclude that the trondhjemite was derived from the fusion of the xenolith of Lockatong argillite. The general slope of the trondhjemite curve is negative. Furthermore, the range of the trondhjemite data plots within the range of the Lockatong argillite data and both display 3 very distinct negative anomalies: potassium, phosphorous and titanium. By comparison, all Palisades diabase analyses including fractionated granophyres plot as lines with different slopes, magnitudes and do not display the 3 negative anomalies.

Our data rules out magmatic fractionation as a process capable of generating the trondhjemites of this study.

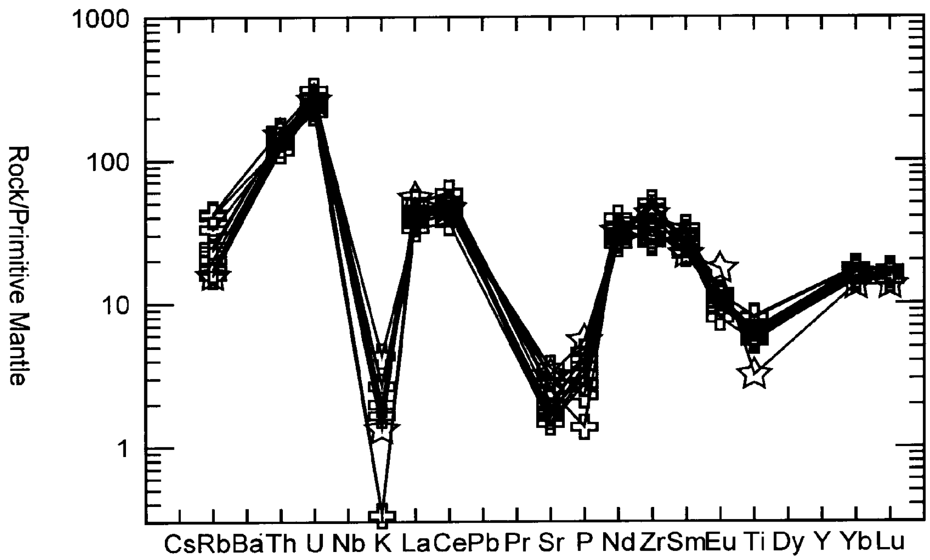

Figure 18. Sun and McDonough(1989) mantle normalized plot. Trondhjemite compositions from Graniteville represented by open plus sign; Xenolith from graniteville represented by open star.

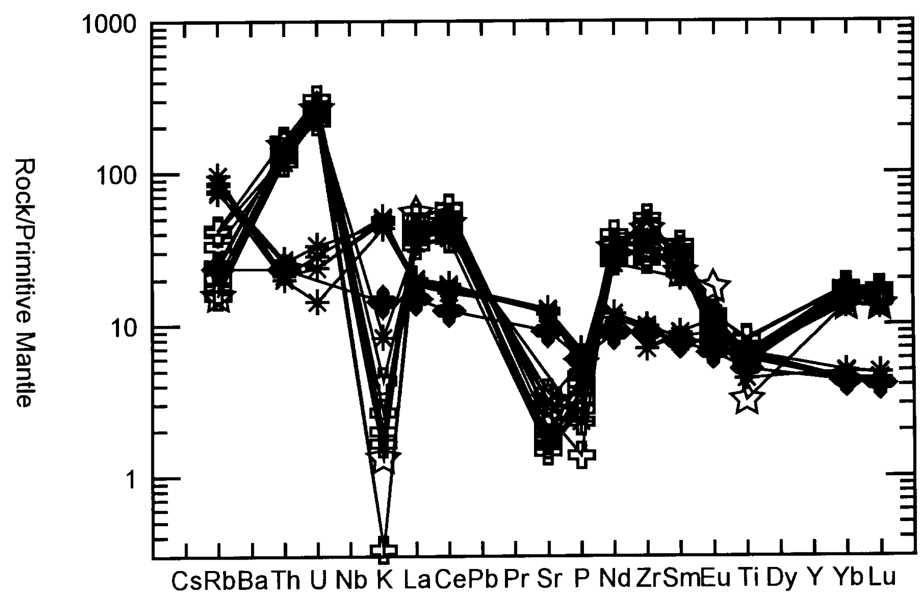

Figure 19. Sun and McDonough(1989) mantle normalized plot. Trondhjemite compositions from Graniteville open plus sign; Xenolith from Graniteville represented by open star. For comparison diabase samples from graniteville represented by an asterisk; and HTQ from Tollo and Gottfried, 1992 is represented by a closed plus sign.

DIFFUSION OF IONS ACROSS THE LIQUID-LIQUID INTERFACE

This discovery of a partly fused xenolith of Lockatong argillite in the Palisades Sill at Graniteville, Staten Island, NY provides a natural laboratory for the study of cationic diffusion profiles across a magmatic liquid-liquid interface which involves coexisting trondhjemite magma, derived by fusion of a xenolith of Lockatong argillite,and basaltic magma of the enclosing sill. Drill cores oriented normal to the liquid-liquid boundary(ie., normal to the trondhjemite-basalt contact) were cut normal to the cylindrical axis of the drill cores with a diamond wafering blade

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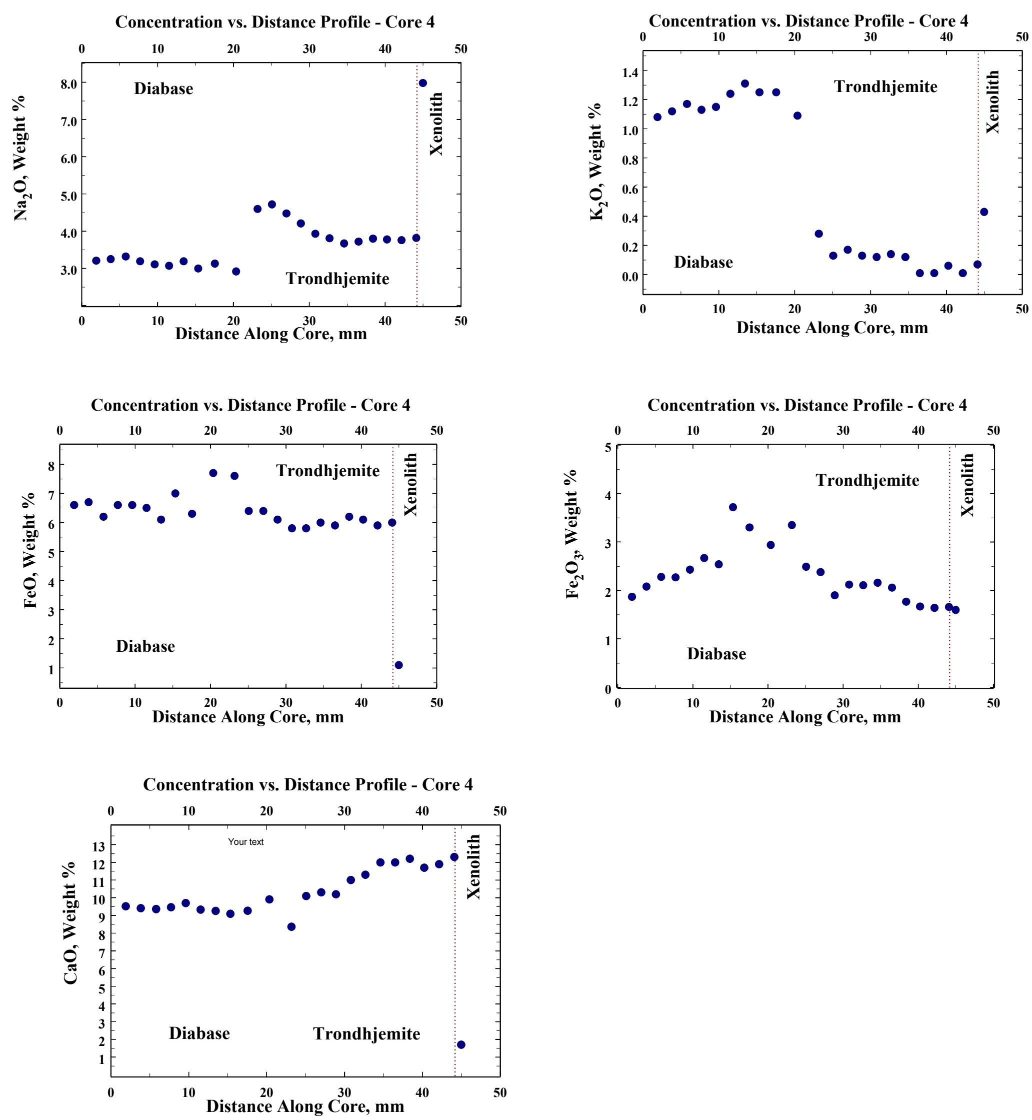
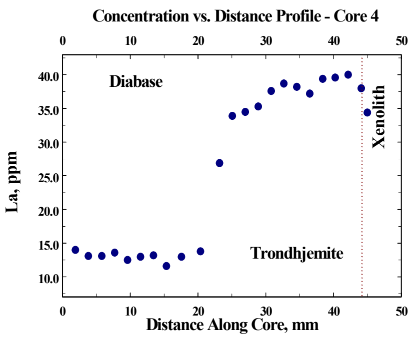
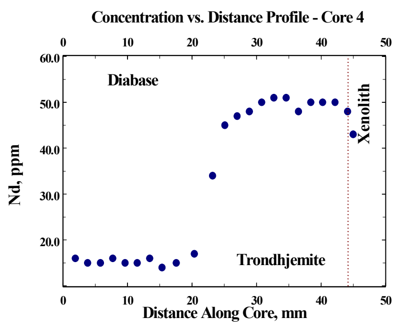
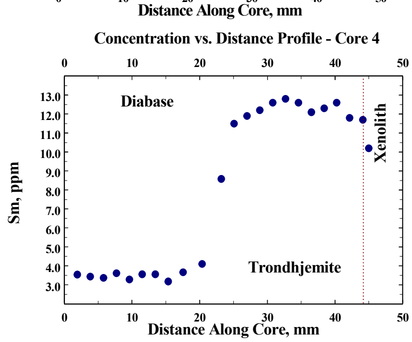
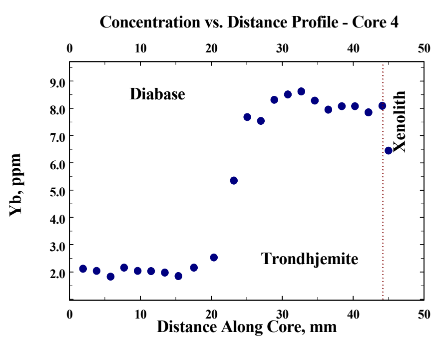


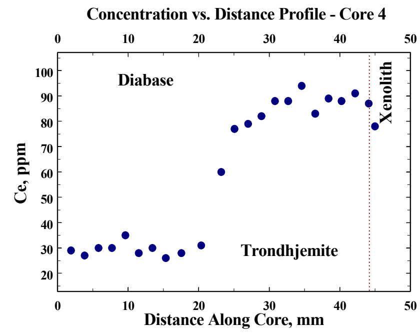
Figure 20. Concentration profiles of selected major elements **.** The Diabase-Trondhjemite contact is at 22mm.

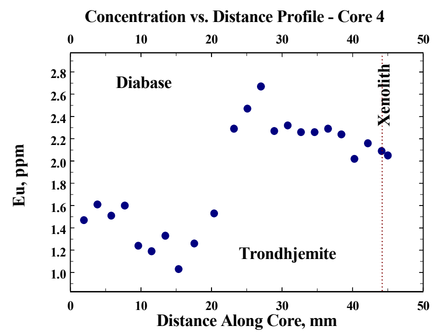

Concentration vs. Distance Profile - Core 4
La, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith

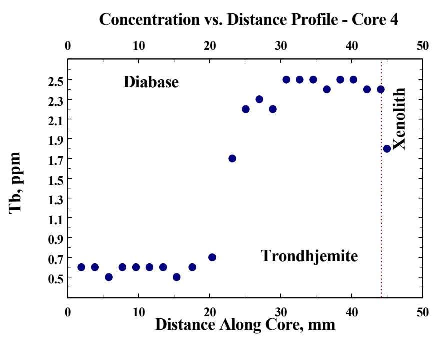

Concentration vs. Distance Profile - Core 4
Nd, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith


Concentration vs. Distance Profile - Core 4
Sm, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith


Concentration vs. Distance Profile - Core 4
Yb, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith


Concentration vs. Distance Profile - Core 4
Ce, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith


Concentration vs. Distance Profile - Core 4
Eu, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith


Concentration vs. Distance Profile - Core 4
Tb, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith

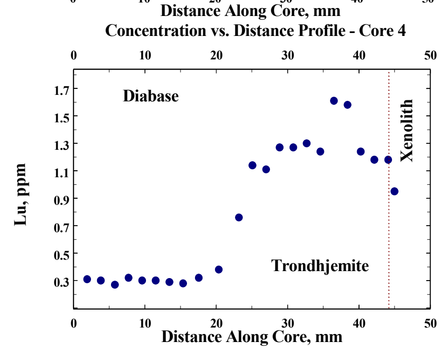

Concentration vs. Distance Profile - Core 4
Lu, ppm
Distance Along Core, mm
Diabase
Trondhjemite
Xenolith

Figure 21. Concentration profiles of selected REE's . The Diabase-Trondhjemite contact is at 22mm.

into discs approximately 1.5mm thick, and each disc was analyzed for major, minor and trace elements including REE's over a distance of 3 cm on the trondhjemite side and 1.2 cm on the diabase side of the magmatic interface. Theoretical models of the two-magma system suggest that at time equals zero the slope of the concentration-distance curve at the liquid-liquid boundary would be 90° whereas at time equals infinity the slope would be zero so that the magnitude observed of the slope between 0 and 90° would be a measure of the relative rate of diffusion for each ionic species. The relative diffusivities of selected major oxides and REE are shown in Figures 19 and 20. Despite the overprint of crystallinity and mineralogy in these holocrystalline rocks, trends in the slopes of the concentration-distance curves for ions of the same charge appear to relate the relative diffusion rates of different ionic species to ionic size and thus steric hindrance in the magma(Benimoff and Sclar, 1996).

Petrographical, mineralogical, and chemical data, plus field evidence indicate that coexisting silicic and mafic melts resulted when the margins of a xenolith of Lockatong argillite fused within the Palisades sill. There must have been diffusional interchange of ions to account for the more complex bulk chemistry of the trondhjemite as compared with the argillite protolith. This study shows that Fe^{2+}, Mg^{2+}, and Ca^{2+} diffused from the diabasic magma into the fusion zone of the xenolith, and that Na^{+} diffused from the fusion zone of the xenolith into the diabasic magma. Evidently, these two chemically divergent magmas did not physically mix.

A dry magma of the composition of the trondhjemite would have a very high viscosity compared with the diabase magma. This high viscosity apparently prevented disruption of the liquid-liquid interface and thereby minimized physical mixing of the diabase and trondhjemite magmas.

5. AT THE UPPER CONTACT OF AN EARLY JURASSIC DIABASE INTRUSION EXPOSED IN A ROCK QUARRY AT BROOKVILLE, NEW JERSEY.

Syenitic rock is exposed along the banks of a stream that flows along the upper contact of the Stockton Diabase sill (a Palisades correlative) with the Lockatong Formation (Figure 22). The syenite was first described by Ransome (l899) and Lewis (l908), then by Milton (l964), Barker and Long (l969) and Benimoff et al. (1996, 1997 and 1998). The Stockton Diabase is approximately 500 m thick and is truncated by a southeast-dipping normal fault near the syenite occurrence. The upper contact of the Stockton Diabase strikes about N. 45° E. and dips 30° NW but as observed by Barker and Long (l969) the contact is complicated by several apophyses.

The syenite is composed principally of orthoclase perthite, albite, and quartz, with minor biotite, pyroxene, and amphibole with variable concentrations of calcite.

The Lockatong meta-argillite hornfels at Brookville is composed principally of sodic plagioclase, biotite, pyroxene, amphibole and variable chlorite, calcite and muscovite. Barker and Long (l969) also report minor to trace amounts of apatite, hematite, pyrite, sphene, and fluorite. Quartz is conspicuously absent. Most of this hornfels corresponds to typical detrital cycles described by Van Houten (l965) particularly the dark gray carbonate-rich mudstone facies.

Benimoff et al. (1996, 1997 and 1998) concluded that the alkalic syenite associated with intrusive Stockton diabase at Brookville NJ is the crystallization product of a melt derived by fusion of the Na-rich nepheline normative Lockatong argillite in the thermal aureole of the intrusive Stockton diabase. They also determined that the syenite is a holocrystalline phanerite containing 15 modal % black euhedral amphibole prisms in a matrix matrix of mottled pink and white feldspar. Their electron microscopy reveals that the amphibole is zoned with a Mg -rich core and an Fe- rich rim. Based on the IMA classification of amphiboles the Mg -rich core is a sodian kaerusutite and the Fe rich rim is a sodian magnesium hastingite. Around the amphibole is an extremely fine intergrowth of oligoclase surrounded by K-feldspar,. The fine intergrowth of oligoclase surrounded by K-feldspar shows the reverse rapakivi texture. Furthermore, the presence of baddeleyite, the absence of quartz and zircon, the low tetrahedral site population of Si in the amphibole (5.824 vs. 8.00 tetrahedral sites/ formula unit), normative nepheline and no normative quartz indicate that the syenite represents a silica undersaturated melt.

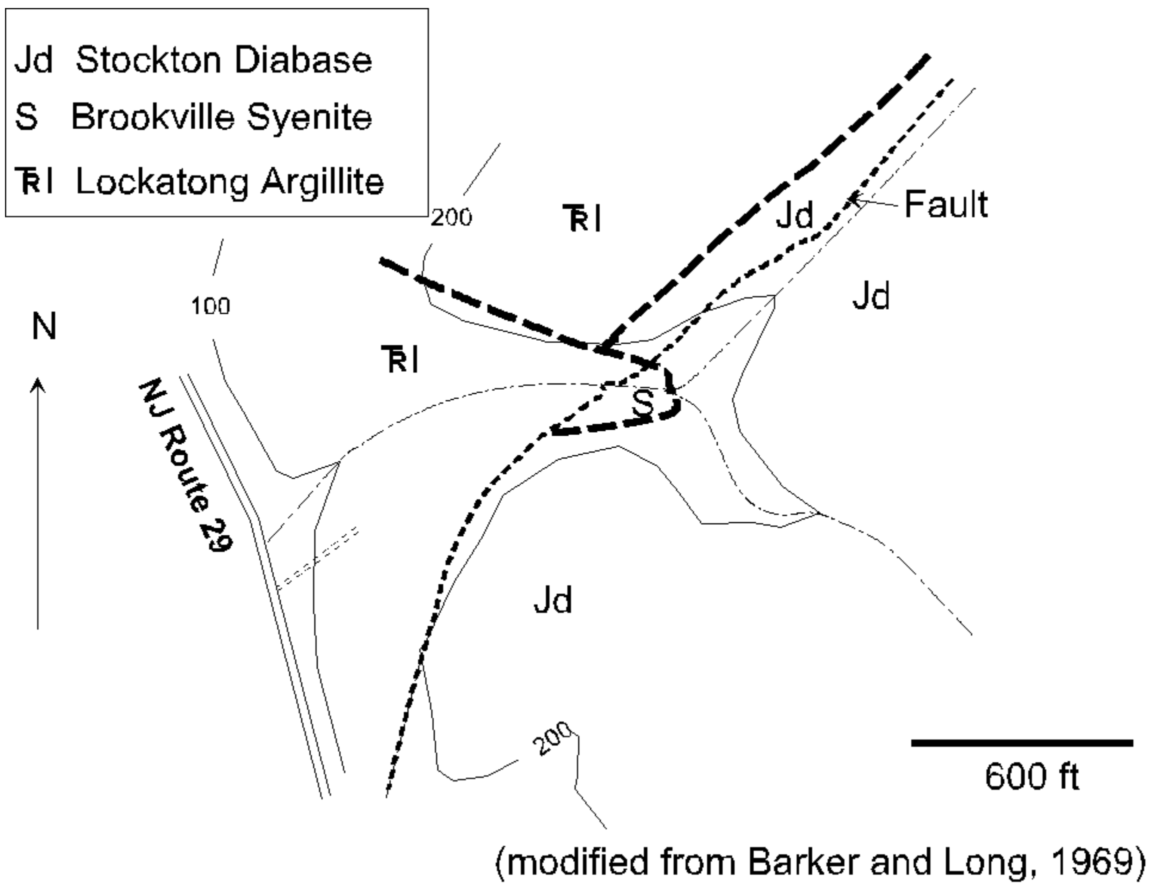


Figure 22. Geologic map from Barker and Long. 1969. Please note that this is private property and permission must be granted by, the owner of the property, Trap Rock Industries.

The chemical composition of the syenite includes about 54 percent SiO_2, 7 percent Na_2O, and 4 percent K_2O, about the same as the Lockatong hornfels, but contains less MgO and FeOt (2 and 5 vs. 5 and 7 percent). MgO and FeOt were partitioned into a black refractory residual hornfels exposed adjacent to and intergrown with the syenite resembling a migmatite. The MgO and FeOt content of the black biotite rich rock are 11 and 15 percent respectively. The build up of biotite in the black rock has displaced feldspar resulting in retention of only 0.5 percent Na_2O. Most elements in the syenite and argillite (including TiO_2 and Zr at 0.8 percent and 140 PPM) maintain approximately consistent levels indicating very high degrees of argillite fusion. The TiO_2 and Zr levels, however, differ from levels found in the Stockton Diabase and associated granophyre (about 1.4 and 120 PPM) suggesting that syenite magma mixing with diabase magma was no more than a minor factor. These data also indicate that the syenite could not be a fractionation product of Stockton Diabase in which case the TiO_2 and Zr contents would exceed Stockton Diabase levels instead of their actual lower levels.

In addition, chondrite normalized plots of the syenite, Lockatong argillite and Stockton Diabase were prepared . The syenite and Lockatong argillite plots(figure 23) are characterized by tightly grouped parallel trends and a negative europium anomaly (Eu= 20 times chondrites), LREE concentrations of about 100 times chondrites and HREE concentrations of 8 to 20 times chondrites. The similarity of REE trends for the Lockatong argillite and alkalic Brookville syenite is further evidence of a very high degree of partial melting of Lockatong argillite and the dissimilarity to the Stockton Diabase is further evidence of a low degree of mixing.

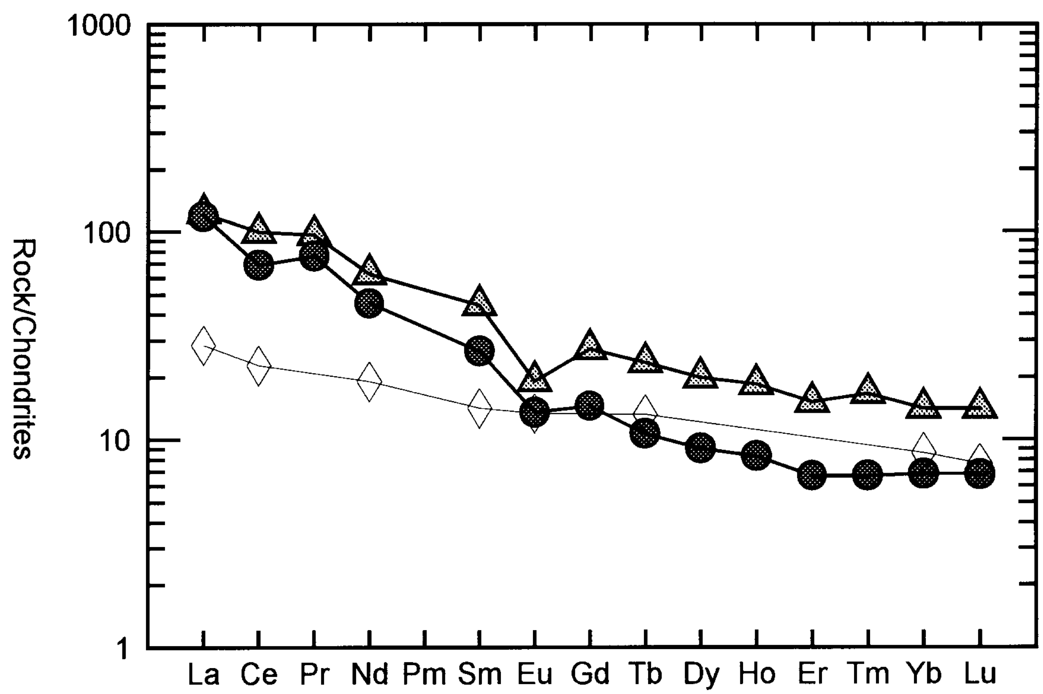


Figure 23. Chondrite normalized REE plot of Lockatong argillite(circles, Brookville Syenite(Triangles) and Stockton Diabase is represented as open diamonds(LQ1 from Husch, 1992).

Again, our complete chemical analyses of the trondhjemite when plotted on primitive mantle normalized spider diagrams(Sun and McDonough, 1989) provide compelling evidence that the syenite is genetically related to the Lockatong argillite and not genetically related to the Stockton Diabase(see Figure 24). The Brookville Syenite is an intrusion of fused Lockatong argillite. The general slope of the trondhjemite curve is negative. Furthermore, the range of the trondhjemite data plots within the range of the Lockatong argillite data and both display three very distinct negative anomalies: potassium, phosphorous and titanium. The Brookville Syenite data also plot very close to the North American Shale Composite(NASC). By comparison, all Stockton diabase analyses plot as a line with a different slope, and magnitude. This data rules out magmatic fractionation as a process capable of generating the Brookville Syenite of this study.

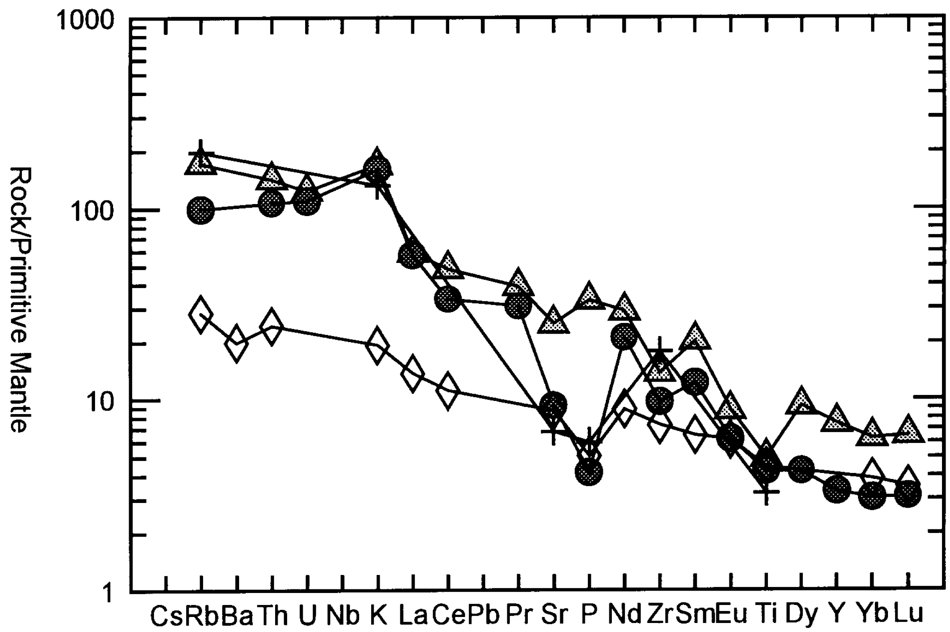


Figure 24. Sun and McDonough (1989) mantle normalized plot of Lockatong argillite (circles, Brookville Syenite (Triangles) and Stockton Diabase is represented as open diamonds (LQ1 from Husch, 1992).; + represents the North American Shale Composite (NASC)(Gromet et al., 1984). Note the similar chemical signatures of the NASC, the Brookville Syenite and the Lockatong Argillite.

CAMP

Table 8 shows average analyses of CAMP flood basalts. These data when plotted on a Sun and McDonough(1989) mantle normalized spider diagram(Figure 24) show amazing uniformity. The 200 Ma flood basalt event coincided with the extensional break-up of Pangaea and is classified as a reactivated arc-sourced, (A-Type) flood basalt, (Puffer, 2001) in contrast to a plume sourced flood basalt (P-type). The A-type classification is based on : (1) its short duration of a few thousand years, (2) its uniform composition and (3) close resemblance to basalt and basaltic andesite from known arc-related sites. The uniform chemical composition, in particular is consistent with the rapid extrusion of a limited amount of eutectic magma. Rapid extrusion and a near absence of contamination is a result of the extensional tectonic activity, which allowed unobstructed egress to the surface in contrast to the ponding , contamination and fractionation typical of magmatism in compressional settings . The reactivated arc source was enriched in water and large-ion-lithophile elements(LILE) typical of mantle metasomatic effects known to occur at subduction zones. However LIL and water enrichment into the CAMP source occurred during the Paleozoic assembly of Pangaea when subduction occurred in a compressional setting and calc-alkaline volcanism was common. When the vectors reversed from compression to extension, and Pangaea began to break up, this same arc-source underwent re-melting. Renewed melting , however, was limited by the amount of flux contained within the source and therefore, was of short duration.

The very minor role played by crustal contamination of CAMP basalt is seen in Table 8 and figure 25 with a virtual identical province-wide composition. Crustal contamination is largely limited to the margins of intrusive CAMP dikes and sills. In contrast to the flood basalts, the CAMP related dike swarms are chemically much more diverse. Table 9 lists some of the CAMP dike populations that range from mafic olivine normative(OLN) dikes common throughout the southeastern U.S. to high iron enriched quartz normative tholeiites(HFQ)(Weigand and Ragland, 1970) scattered throughout the CAMP province. The High-Titanium-Quartz-Normative population (HTQ) of dikes is chemically the same as the initial flood basalt population and is the most widespread dike population. The Low-Titanium-Quartz Normative population(LTQ) is confined to the southeastern U.S. The MgO content of the OLN population(Table 9) averages 10.52 wt. % in contrast to an average of only 5.53 wt. % for the HFQ population. In addition the TiO_2 content of the OLN population(table 9) average only 0.59 wt. % for the HFQ population. Weigand and Ragland(1970) also indicate that the mafic index for the ENA dikes range from 40 to 75.

The contrast between uniform flood basalt chemistry(table 8) and less uniform dike and sill chemistry(table 9) is probably due to a combination of factors. (1) Dikes are relatively susceptible to deuteric alteration and metasomatic effects compared with flows, (2) Dike chemistry is influenced by depth of emplacement; deep crustal dikes tend to be richer in dense phases, particularly olivine, then shallow dikes (3) Sills (such as the Palisades Sill) and even vertical intrusions (such as Snake Hill) are relatively susceptible to in-situ fractionation processes, and (4) dikes are relatively susceptible to crustal contamination and the mixing of fusion products near their outer contacts than basalt flows as this research makes clear. The kind of sediment fusion described in this study requires high sustained temperatures and insulated thermal energy that is absent in an extrusive environment. In addition, some sampling problems involving thin dikes(less than 10 m across) can add to the contrast in flood basalt vs. dike swarm chemistry. Hydrothermal alteration effects, particularly chloritization, are confined to the 10 m thick margins of Snake Hill(Puffer and Benimoff, 1997) and can generate olivine normative compositions. Any of the chloritized dikes of the southeastern U.S. OLN dike swarm that are less than 10 m thick should, therefore, be viewed with skepticism. This may, however, include the majority of the sample population. Although, mixing of the kind of fusion products described in this study is even more confined that hydrothermal alteration effects, any resulting changes in magma chemistry would, again, be confined to the intrusive rock populations.

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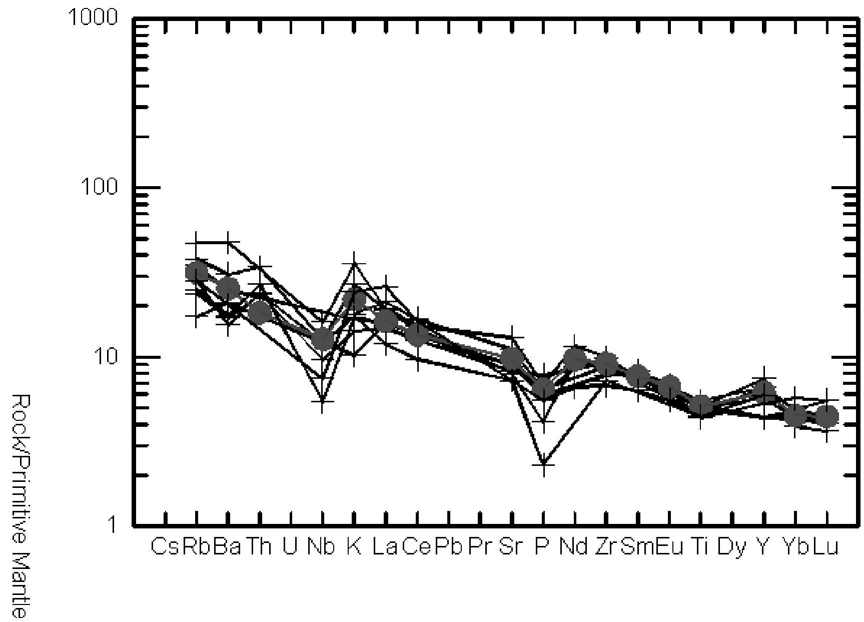

Figure 25 Sun and McDonough(1989) mantle normalized plot of CAMP extrusive rocks(table 8). The camp average is plotted as closed circles. Note the very tight clustering of CAMP extrusives in contrast to the diverse chemistry of CAMP intrusives(table 9).

TABLE 8. Average Analyses of CAMP Basalts

	A. Initial HTQ-type Flows								B. Secondary LTQ-type Flows		
Name	Orange Mt	Talcott	Mt.Zion	North Mt.	H. Atlas	Algarve	Maranhão	CAMP	Holyoke	Preakness	Sander
Location	NJ	CN	VA	N. Scotia	Morocco	Portugal	Brazil	average	CN	NJ	VA
n	11	7	7	6	15	2	16	64	28	31	16
SiO_2	52.68	51.79	52.65	53.81	52.80	51.90	52.98	52.41	53.65	52.09	52.69
TiO_2	1.12	1.07	1.09	1.08	1.19	1.01	1.10	1.11	0.97	0.95	1.09
Al_2O_3	14.25	14.25	14.62	13.79	14.38	14.44	14.87	14.50	14.67	14.33	14.22
FeO_t	10.15	10.84	9.97	9.73	9.57	10.36	9.77	10.00	11.74	12.66	12.43
MnO	0.18	0.16	0.17	0.16	0.16	0.16	0.17	0.17	0.20	0.21	0.22
MgO	7.81	7.97	7.68	8.09	8.13	7.91	7.26	7.80	5.97	6.55	5.76
CaO	10.80	11.23	10.69	10.38	10.98	11.23	10.43	10.84	9.40	9.49	9.40
Na_2O	2.45	2.06	2.68	2.06	2.10	2.31	2.21	2.37	2.61	3.11	3.44
K_2O	0.43	0.50	0.31	0.81	0.52	0.54	1.07	0.66	0.73	0.52	0.61
P_2O_5	0.13	0.13	0.14	0.09	0.17	0.14	0.15	0.14	0.05	0.12	0.14
Rb	15	22	11	24	18		30	20	18	16	
Ba	119	174	145	216	108		331	178	142	123	
Sr	192	186	191	169	173	235	275	207	153	153	
Th	2.00			2.88	2.04		0.00	1.55		2.29	
Zr	99	87	99	96	104	98	111	102	81	76	
Hf	2.40			2.90	2.64		0.00	1.97		2.02	
Nb	6.90			9.00	8.72		11.44	9.08	5.30	3.89	
Ni	99	86	79	64	89	87	76	88	36	52	
Cr	309	322	282	225	221	241	301	282	11	111	
La	10.2	11.1	10.8	13.0	10.4	14.6	13.3	11.2	18.0	8.2	
Ce	22.1	23.9	24.1	26.0	22.5	29.7	27.1	24.0	29.0	17.2	
Nd	11.8			12.8			15.8	13.1		9.0	
Sm	3.40	3.66	3.51	3.30	3.45		3.34	3.42		2.81	
Eu	1.10	1.15	1.13	1.03	1.11		1.11	1.12		0.93	
Tb	0.61	0.67	0.65	0.63	0.65		0.63	0.64		0.64	
Yb	2.10	2.35	2.30	2.13	2.41		1.91	2.22		2.81	
Lu	0.30	0.33	0.34	0.33	0.41		0.27	0.33		0.41	
Y	20			26	30		34	28	27	24	

Note all Fe is listed as FeO_t, major elements are in wt.% normalized to 100% anhydrous, trace elements are in ppm. Listed averages for Orange Mt. Basalt from *Tollo and Gottfried* [1992]; Talcott Basalt from *Puffer* [1992]; Mt Zion Church Basalt from *Puffer* [1992]; North Mt. Basalt from *Papezik et al*. [1988]; High Atlas Basalt from *Bertrand et al.* [1982]; Algarve basalt is new data; western Maranhão Basalt from *Fodor et al.* [1990]; Holyoke Basalt from *Philpotts et al.* [1996], Preakness Basalt from *Tollo and Gottfried* [1992]; and Sander Basalt from *Tollo* [1988].

Table 9. Major and minor element averages for ENA dolerites.

(from Weigand and Ragland, 1970)

	HFQ	HTQ	LTQ	OLN
SiO2	52.96	52.10	51.66	47.90
TiO2	1.14	1.12	0.76	0.59
Al2O3	14.21	14.22	14.95	15.26
FeOt	13.87	11.65	11.77	12.10
MnO	0.22	0.19	0.20	0.18
MgO	5.53	7.41	7.44	10.52
CaO	9.86	10.66	10.80	10.75
Na2O	2.51	2.12	2.23	2.00
K2O	0.64	0.66	0.48	0.29
Co	52	49	53	65
Cr	94	277	218	166
Cu	74	111	68	108
Li	18	19	16	12
Ni	34	81	48	308
Rb	22	21	15	8
Sr	178	186	127	115
Zn	99	84	86	84
Zr	94	92	60	50

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ROAD LOG
for
2001 NYSGA MEETING

The uniform composition of CAMP diabase types and the absence of wall rock contamination at Northern New Jersey and Staten Island, New York Locations

Total Mileage	Inc. Mileage	DESCRIPTION
0	0	Leave IBM Palisades Conference Center and Continue on US 9W south
11.7	11.7	Left Turn onto Main St.(Fort Lee, NJ)
13.6	1.8	Merge with County Rd. 505(River Road)

14.8	1.2	STOP 1 Caution- Dangerous Intersection! Park on left side of River Road Near Entrance to Henry Hudson Park

The Palisades diabase, exposed along its basal contact with the Lockatong Formation at Fort Lee NJ, is fine grained, massive chill-zone rock. The diabase is aphyric and subophitic and contains very few xenoliths. A few large xenoliths have been observed but they are typically spaced at least 100 meters apart. The lower contact is largely parallel to the bedding planes of the underlying metasediments but is locally discordant. In addition, there are anticlinal dome-like structures consisting of the Lockatong Formation that rise a few meters into overlying diabase. It is at these dome-like structures where most of the partial fusion of the Lockatong Formation has taken place. Three clearly exposed examples of these domes occur within 2 km of Ross Dock, where fusion has occurred resulting in syenitic rock intergrown with black laminated siltstone forming a migmatite and trondhjemite veins. The bedding at these domed structures is disrupted and may have involved movement of volatiles derived from brackish groundwater within the lacustrine Lockatong sediments. The high salt content of Lockatong Argillite may have helped flux the melting process. Both the syenite and the trondhjemite are sodic, typically containing 4 and 7.5% Na_2O respectively, but the K_2O and Rb is highly partitioned into the syenite typically containing 5 % and 125 ppm vs. only 0.5% and 25 ppm for the trondhjemite. Both the syenite and trondhjemite have similar REE contents that are comparable to the Lockatong Argillite.

Our complete chemical analyses of the syenite and trondhjemite when plotted on primitive mantle normalized spider diagrams (Sun and McDonough, 1989) provide compelling evidence that the igneous components of the migmatite are fused Lockatong argillite. For example, all three lithologies display distinctly negative Nb anomalies. However, the chemistry of these lithologies differs from that of the Palisades Diabase.

14.8	0	Head North on County Road 505
15	0.2	Bear Left at Y(Main St.)
15.2	0.2	right turn
15.4	0.2	left turn onto Bridge Plaza South
15.5	0.1	Right turn onto NJ 67
15.6	0.1	Right turn onto I-95 South, US 1 and 9 South
17.9	2.3	Bear Left onto US 1+9 South
18.2	0.3	Left turn - Follow US 1+9

		South(Broad Ave)
18.4	0.2	JCT 5 Follow US 1+9 South
18.8	0.4	JCT 93 - Follow US 1+9 South
18.9	0.1	Enter Traffic Circle Follow US 1+9 South
21.5	2.6	Pathmark supermarket on Right
22	0.5	Milepost 59
22.7	0.7	46th St.
23.6	0.9	JCT RT 3
25.5	1.9	Right Turn onto County Road(County Road 653)
26.6	1.1	Pass under I-95 NJTP
27.8	1.2	Left Turn onto New County Road (655)
28	0.2	Pass over RR tracks
28.4	0.4	Enter Hudson County Park
29.1	0.7	STOP 2 -Park car in lot

Laurel Hill(Snake Hill) is an elliptical exposure of early Jurassic diabase about 500 meters in diameter, located 2.5 km west of the underlying Palisades sill and 10 km east of the stratigraphically overlying Orange mountain Basalt. Fusion of Passaic siltstone at the southern contact has generated granitoid rocks. The granitoid rocks include trondhjemite and syenite. The potassic granitoids chemically contrast with the intrusive but closely resemble the wall rocks.

29.2	0.1	Exit on Park Road
29.8	0.6	Leave Hudson County Park
30.4	0.6	Turn Left onto County Ave.
30.7	0.3	Secaucus Diner on Right
30.9	0.2	Observe west slope of the Palisades Sill on right
32.2	1.3	Turn Right onto Paterson Plank Road
32.8	0.6	Enter NJ Turnpike South(Toll Road)
45.9	13.1	Exit at 13

46.6	0.7	Pay Toll -follow signs for Goethals Bridge -I-278 East
48.6	2	Pay Toll
48.8	0.2	Exit at Forest Avenue
48.9	0.1	Left onto Forest Ave.
50.3	1.4	STOP 3 Park on Right side between Van Name and Simonson Aves.

At this locality, we see an extraordinary example of two coexisting magmatic liquids, now represented by the diabase of the Palisades Sill and a pyroxene Trondhjemite derived by fusion of the margins of a xenolith of sodium rich Lockatong Argillite .(Benimoff and Sclar, 1978, 1980, 1984, 1988, 1992, 1996; Sclar and Benimoff,1993). The diabase is composed dominantly of plagioclase ($An_{61}Ab_{38.8}Or_{0.2}$) and augite ($En_{34-44}Fs_{17-31}Wo_{35-42}$).The trondhjemite is composed dominantly of quartz-albite granophyre in which are enclosed large discrete crystals of albite($Ab_{99}An_{0.52}Or_{0.44}$) and Ca-rich pyroxene . Minor constituents include interstitial calcite, titanite, ilmenite, optically homogeneous titanomagnetite, nickelian and cobaltian pyrrhotites, apatite, and sphalerite. The modal mineral percentages are clinopyroxene 38, albite 38, quartz 18, titanite 2.7, calcite 1.3, and opaques 2.0. The xenolith is now a hornfels and exhibits a granoblastic texture. The hornfels is composed dominantly of albite and quartz and subordinately of calcite, titanite, apatite, ilmenite, and actinolite. The modal mineral percentages are albite 66, quartz 30, titanite 2.3, calcite 0.9, apatite 0.5, and actinolite 0.3. Normative albite ranges from 56.4 to 80.2 wt.%, whereas normative quartz ranges from 7.0 to 35.4 wt.%. Chemical analyses reveal that diffusion of calcium, magnesium, iron, and sodium ions occurred across the liquid-liquid interface.

END OF FIELD TRIP

The "Age of Dinosaurs" in the Newark Basin, with Special Reference to the Lower Hudson Valley

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ABSTRACT

This field guide is intended as an introduction to the rich stratigraphic and paleontological record of the Triassic-Jurassic Newark rift basin, especially in the vicinity of the present and ancestral routes of the lower Hudson River. We will visit seven stops that illustrate this region's range of sedimentary and igneous environments and paleobiological assemblages, focusing on their significance to the understanding of global events in the early Mesozoic, in particular the beginning of the "Age of Dinosaurs".

INTRODUCTION

The Newark basin (Figure 1) is one in a remarkable series of early Mesozoic rift basins that extend from Greenland to Europe, Morocco and eastern North America, and to the Gulf of Mexico, comprising the largest known rift system. This massive set of basins - the central Atlantic margin rifts - formed during the crustal extension that led to the fragmentation of Pangea (Figure 1). The Newark basin is one of the largest segments of the outcropping, deeply eroded North American contingent of these rifts, the basin fill of which is collectively termed the Newark Supergroup (Figure 1). Continental rifting seems to have begun in eastern North America sometime in the median Permian and finished in the Early Jurassic, although the exact timing of the termination of rifting is poorly constrained. These rifts - in particular the Newark basin - also record a major tectonic paroxysm that punctuated the beginning of the Jurassic: the emplacement of basaltic intrusions and extrusions of the Central Atlantic Magmatic Province (CAMP) (Marzoli, 1999; Olsen, 1999) - the largest known igneous province (Figure 2).

A “Hot House” mode of global climate system apparently prevailed during the Triassic and Early Jurassic, with little or no convincing evidence of ice at the poles, probably due to very elevated CO_2 levels (Ekart and Cerling, 1999; Tanner and Hubert, 2001). Despite the profound difference between today’s global climate and that of the early Mesozoic, tropical climate gradients, as reflected by indicators of humidity, were evidently not much different than today – and hence quite strong (Kent and Olsen, 2000). During this time, Pangea straddled the equator, its central region drifting slowly northward through time (Figure 2). The Newark basin lay in this central region, and drifted from an equatorial position in the late Middle Triassic or earliest Late Triassic (~232 Ma) to about 7˚N by the beginning of the Jurassic (~202 Ma). As a consequence, the Newark basin itself slowly drifted from the wet tropics through a strong climate gradient towards the arid climate belt, producing a sedimentological transition that is well displayed in the basin sediments.

Within this context of the Early Mesozoic "Hot House" world, climate was far from stable in the tropics. As is true for tropical climate during the Quaternary, that of the

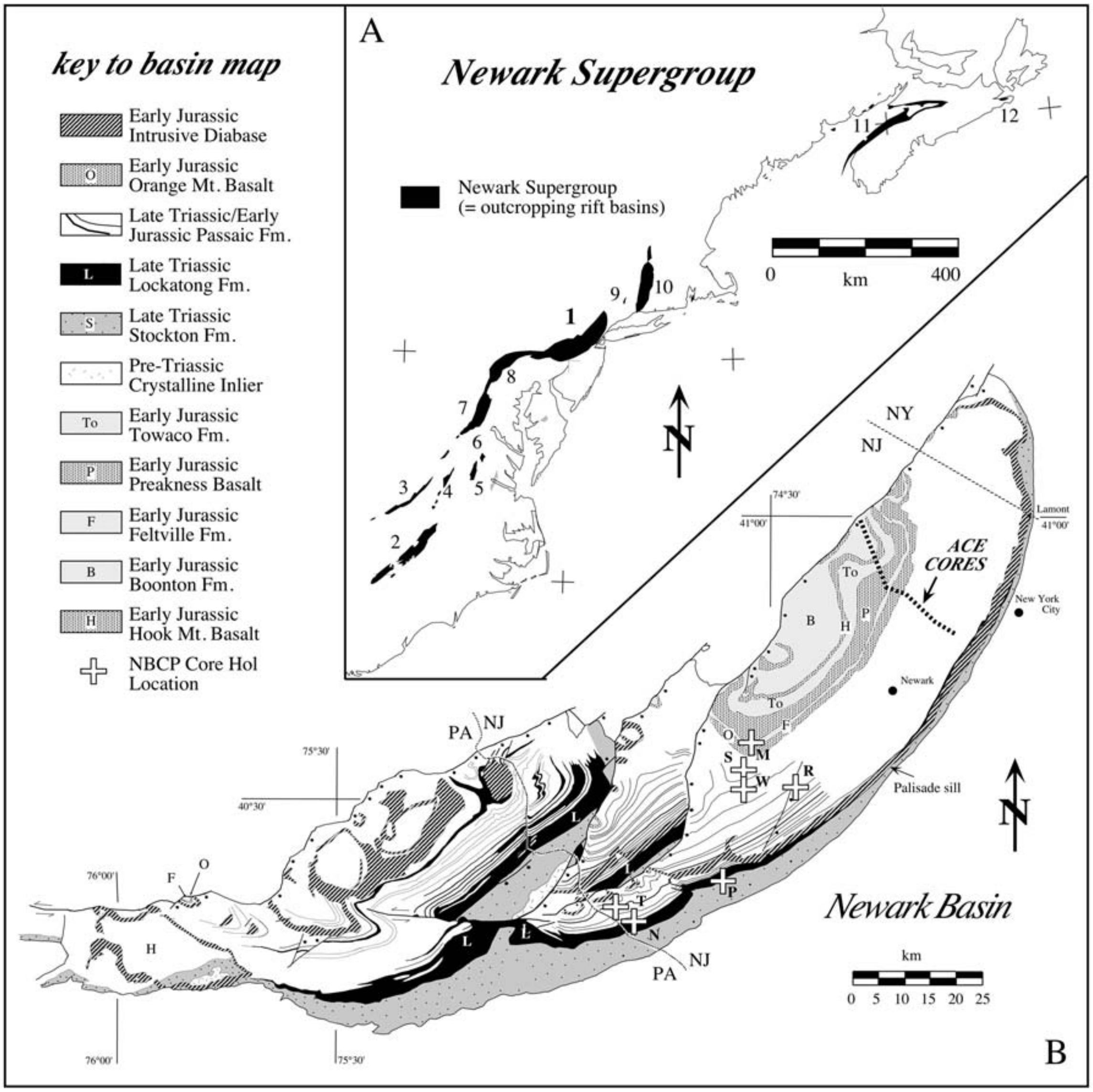

Figure 1. Central Pangean rift basins in the Early Jurassic Pangea relationship to the CAMP. **A**, rift basins of central Pangea showing the position of the Newark basin, the Long Island Platform - South Atlas fault zone (LIP-SAF) and the Minas - Gibraltar fault zone (MFZ-GFZ) (modified from Olsen, 1997). B, Pangea in the Early Jurassic showing the spatial relationship between the CAMP and the central Pangean rift zone (modified from Olsen et al., 2002a).

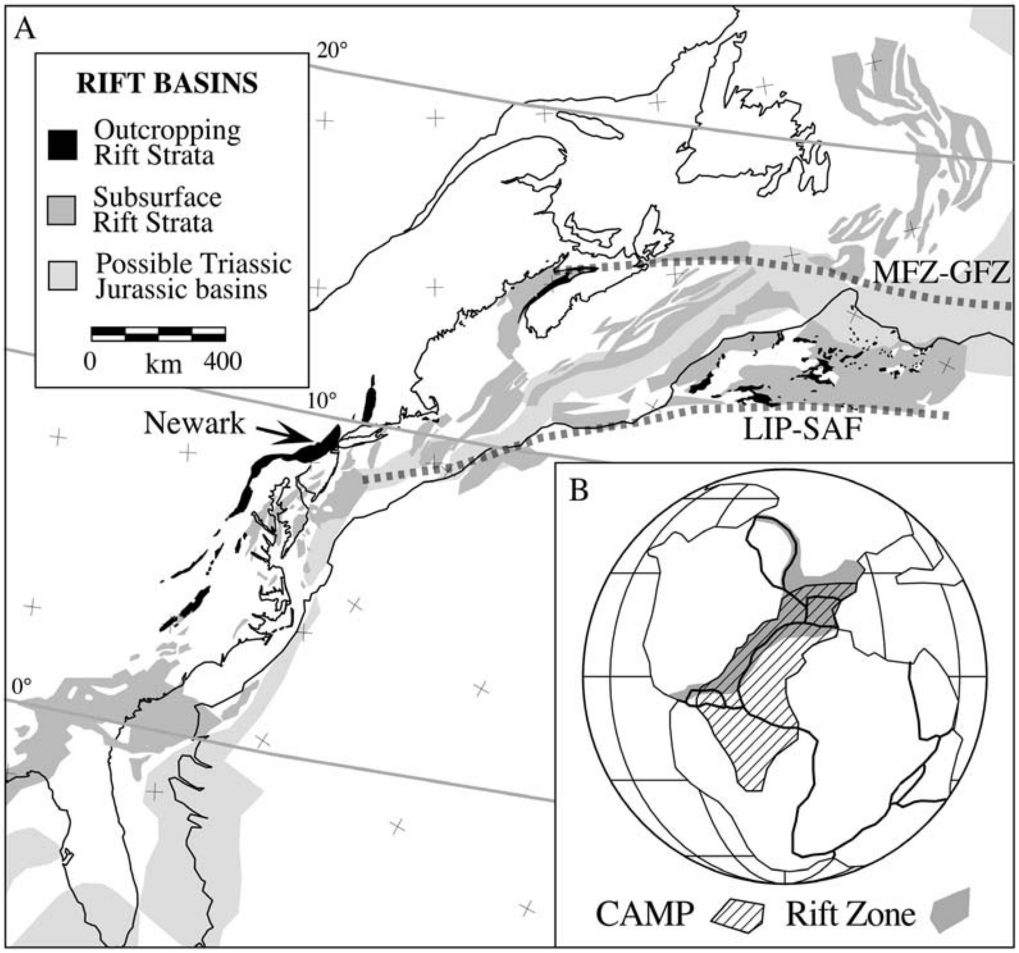

Figure 2. Newark basin section and time scale showing distribution of field stops (adapted from Olsen and Kent, 1999; Olsen et al., 2001a; Olsen et al., 2002b).

Triassic and Early Jurassic fluctuated dramatically in precipitation following Milankovitch climatic cycles, driven by variations in the Earth's orbit. These precipitation and evaporation cycles are recorded as lake level cycles of several orders of complexity in the abundant Newark Supergroup lacustrine deposits, and especially in the Newark basin (described in more detail below). In addition to providing a constantly fluctuating sedimentary environment sampling various lacustrine to fully terrestrial biological communities, these cycles provide a rigorous stratigraphic framework for the basins and a mechanism by which to calibrate the Late Triassic-Early Jurassic time scale.

Biologically, the Triassic and Early Jurassic were pivotal. The early Mesozoic opened following the largest mass-extinction of all, that of the end-Permian. However, terrestrial communities of the Early Triassic largely inherited the Paleozoic-style dominance of synapsid amniotes (the oxymoronic "mammal-like reptiles"), and very distinct northern and southern hemisphere floras. Through the Triassic, however, diapsid sauropsids (lizards, crocodilians, dinosaurs and their relatives) became progressively more abundant, diverse, and larger, so that by the beginning of the Late Triassic synapsids were rare in tropical regions, except for a narrow belt around the equator. By the early Late Triassic dinosaurs had evolved, but they did not become truly dominant until after the next great mass extinction at the Triassic-Jurassic boundary. This great transition to dinosaurian dominance is recorded in detail in Newark Supergroup and particularly Newark basin deposits, as explored in this fieldtrip guide.

STRATIGRAPHIC ARCHITECTURE, CYCLOSTRATIGRAPHIC FRAMEWORK, AND STRATIGRAPHIC NOMENCLATURE OF THE NEWARK BASIN

Based on extensive scientific and industry coring, drilling, and seismic profiles, and outcrop studies in eastern North America and Morocco, Olsen (1997) recognized four tectonostratigraphic sequences in the central Pangean rifts (Figure 3). Tectonostratigraphic sequences (TS) are similar in concept to marine sequence stratigraphic units in that they are largely unconformity-bound genetically-related packages, but differ from them in that there it is assumed they are controlled largely by tectonic events. Tectonostratigraphic sequence I (TS I) is apparently median Permian in age and is present for certain only in the Fundy basin of maritime Canada and various Moroccan basins; however, it may very well be present in the subsurface in other basins. Tectonostratigraphic sequence II (TS II) is of ?Middle (Anisian-Ladinian) Triassic to early Late Triassic (Early to early Late Carnian) age and is present in most Newark Supergroup basins, dominating the preserved record of some (e.g. Richmond basin). Tectonostratigraphic sequence III (TS III), of early Late Triassic (Late Carnian through early Late Rhaetian) age, is the most widespread of the sequences and dominates nearly all Newark Supergroup basins. Tectonostratigraphic sequence IV (TS IV) is of latest Triassic (Late Rhaetian) to Early Jurassic (Hettangian and Sinemurian) age, and contains the Triassic-Jurassic boundary, extrusive tholeiitic basalts of the CAMP, and occasionally extensive post-CAMP sedimentary strata. At the hingeward edges of the rift basins, the unconformities between the tectonostratigraphic sequences can represent large hiatuses, but may pass into correlative conformities at depth within the basins, with no break in sedimentation. The exception is the TS I – TS II boundary, which, as far as is known, always represents a hiatus of a score or so million years. TS II through IV are present in the Newark basin and represent the fundamental stratigraphic and sedimentological units of the basin sequences, transcending the traditional formational bounds.

Tectonostratigraphic sequences II – IV at least were probably initiated by major pulses of regional extension which subsequently declined, as hypothesized by the basin filling model (outlined by Schlische and Olsen, 1990, and elaborated on by Contreras et al., 1997) (Figure 3). As a consequence of the growth of the accommodation space during the extensional pulse, and disregarding climate changes, the basin depositional environments should follow a tripartite development at their depocenters, consisting of a basal fluvial sequence, succeeded by a rapidly deepening lacustrine sequence, and finally followed by slow upward shallowing. The slowing or cessation of the creation of new accommodation space would cause additional shallowing and thus a return to fluvial conditions; eventually erosion would ensue if creation of accommodation space stopped or nearly stopped. Each new pulse of extension would be expected to produce a shift of

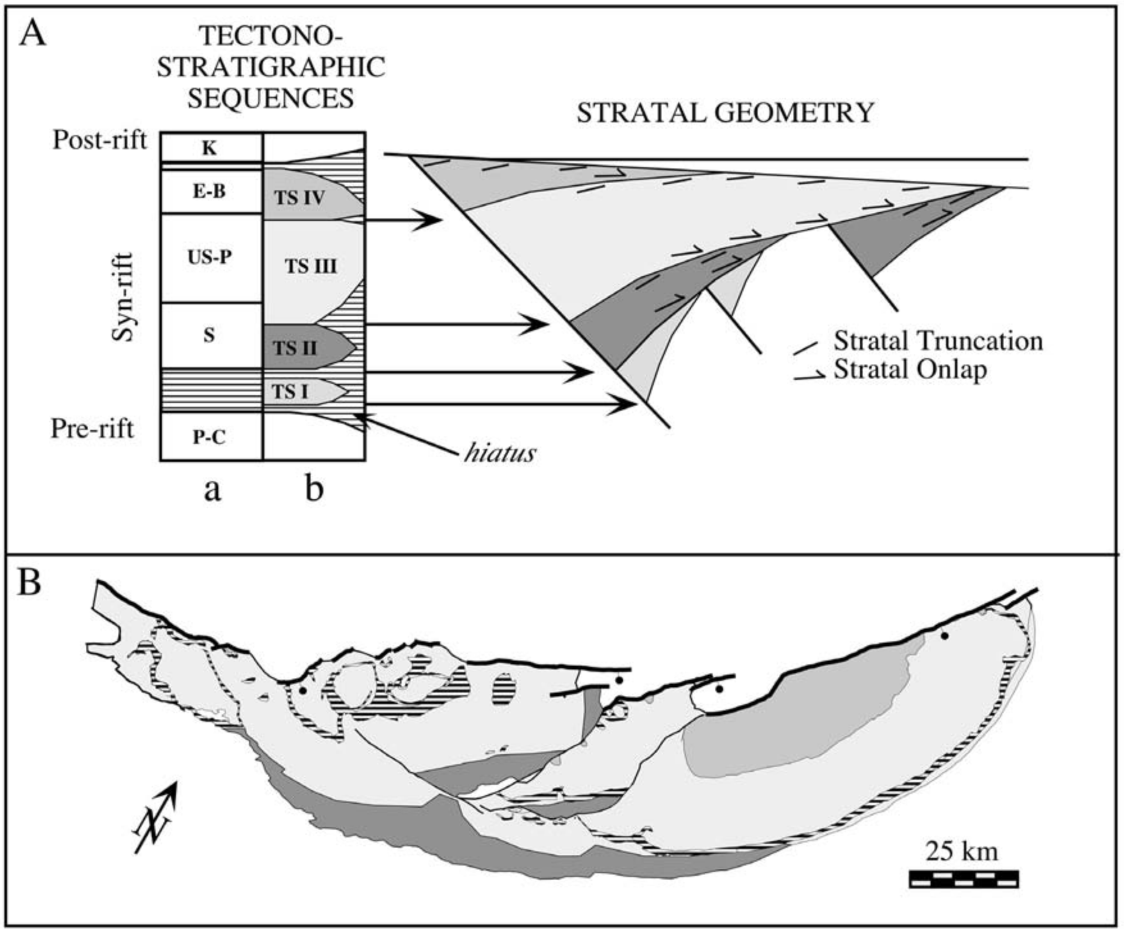

Figure 3. Tectonostratigraphic sequences of the central Pangean rifts and Newark basin. **A**, generalised tectonostratigraphic sequences of central Pangean rifts (modified from Olsen, 1997). **B**, map of the Newark basin showing extent of tectonostratigraphic sequences with the patterns corresponding to **A** above (dark horizontal ruling are intrusive diabase).

the depocenter towards the boundary fault system, accompanied by erosion of the hanging wall deposits; this would continue until the basin fill onlapped those areas of the hanging wall. We hypothesize that the hanging wall unconformities between TS II, III, and IV were each caused by a renewal of extension. This certainly is true of the TS III-IV boundary in the Newark basin, since it is actually a correlative conformity in most presently-outcropping areas. Whether or not the full basin filling sequence - termed a Schlische cycle by LeTourneau (2002) - is actually observed in outcrop, depends on the depth of erosion relative to the basin depocenter and the boundary conditions of the basin geometry and sediment input.

Cyclostratigraphy

As a result of over a century of intensive outcrop work and recent coring, drilling and seismic exploration, the Newark basin is known in more stratigraphic detail than any other central Atlantic margin rift, and arguably any rift of any age. Virtually the entire stratigraphic section of the Newark basin was cored by the US National Science Foundation-funded Newark Basin Coring Project (NBCP) (Goldberg et al., 1994; Kent et al., 1995; Olsen and Kent, 1996; Olsen et al., 1996a); the Army Corps of Engineers (ACE) cores were recovered as part of the currently-dormant Passaic River Diversionary Tunnel Project (Fedosh and Smoot, 1988; Olsen et al., 1996b). About 6.7 km of continuous core of mostly Triassic age, at a total of seven coring sites, was recovered by the NBCP, and over 10 km of mostly Jurassic core from dozens of coring sites is represented by the ACE cores. These cores have allowed the entire stratigraphy of all but the very oldest and very youngest parts of the Newark basin sequence to be recovered (Figure 4).

Based on these cores as well as outcrop studies, one of most dramatic features of the Newark basin sedimentary record is the pervasive cyclicity obvious in most of the sequence (Figure 5). This cyclicity was first described in detail, and ascribed to astronomical control of climate, by Van Houten (1962, 1964, 1969, 1980). All subsequent studies have confirmed and elaborated on these seminal works; the fundamental sedimentary cycle seen in these sequences, caused by the ~20 ky cycles of climate precession, has consequently been named the Van Houten cycle (Olsen, 1986).

The thickness of Van Houten cycles varies from about a meter to over 25 meters, depending on both the geographic and stratigraphic position in the basin, although within single formations in specific areas of the basin, the thickness tends to varies only about 25%. This cycle consists of three lithologically distinct divisions that are defined by the relative development, in comparison to surrounding units of sedimentary features and the presence of fossils indicative of submergence or exposure. These three divisions are termed 1, 2 and 3, and represent lacustrine transgressive, high stand, and regressive deposits, respectively (Figure 5). At one end of the range of expression of Van Houten cycles as seen in the Newark basin, they consist of very obvious asymmetrical sequences, that we interpret as produced by relatively wet conditions: 1, a relatively thin gray division 1, which has mud cracks or roots traces and often tetrapod footprints at the base, passing upward into oscillatory-rippled or laminated mudstone with no mudcracks, roots, or tracks; this division represents the transition from an exposed mud flat or vegetated plain to a perennial lake; 2, a moderately thick division 2, the lower part consisting of a black, microlaminated mudstone or limestone containing complete articulated fish and reptiles, produced by deposition in a deep (>80m) chemically-stratified lake; passing upward into less well laminated black or gray mudstones deposited in a perennial lake with an at least seasonally oxygenated bottom; 3, a thick division 3, marked by the presence of abundant mudcracks, and other signs of emergence, often with abundant tetrapod footprints, marking a the transition to temporally persistent playas and mudflats.

At the other end of the spectrum, even within the same more-inclusive stratigraphic sections, the Van Houten cycle can be dominated by red massive mudstone. We call this the "dry" type: 1, a red, nearly-imperceptible and abbreviated division 1, with barely-discernable mud cracks and/or roots, produced by a decrease in emergence of

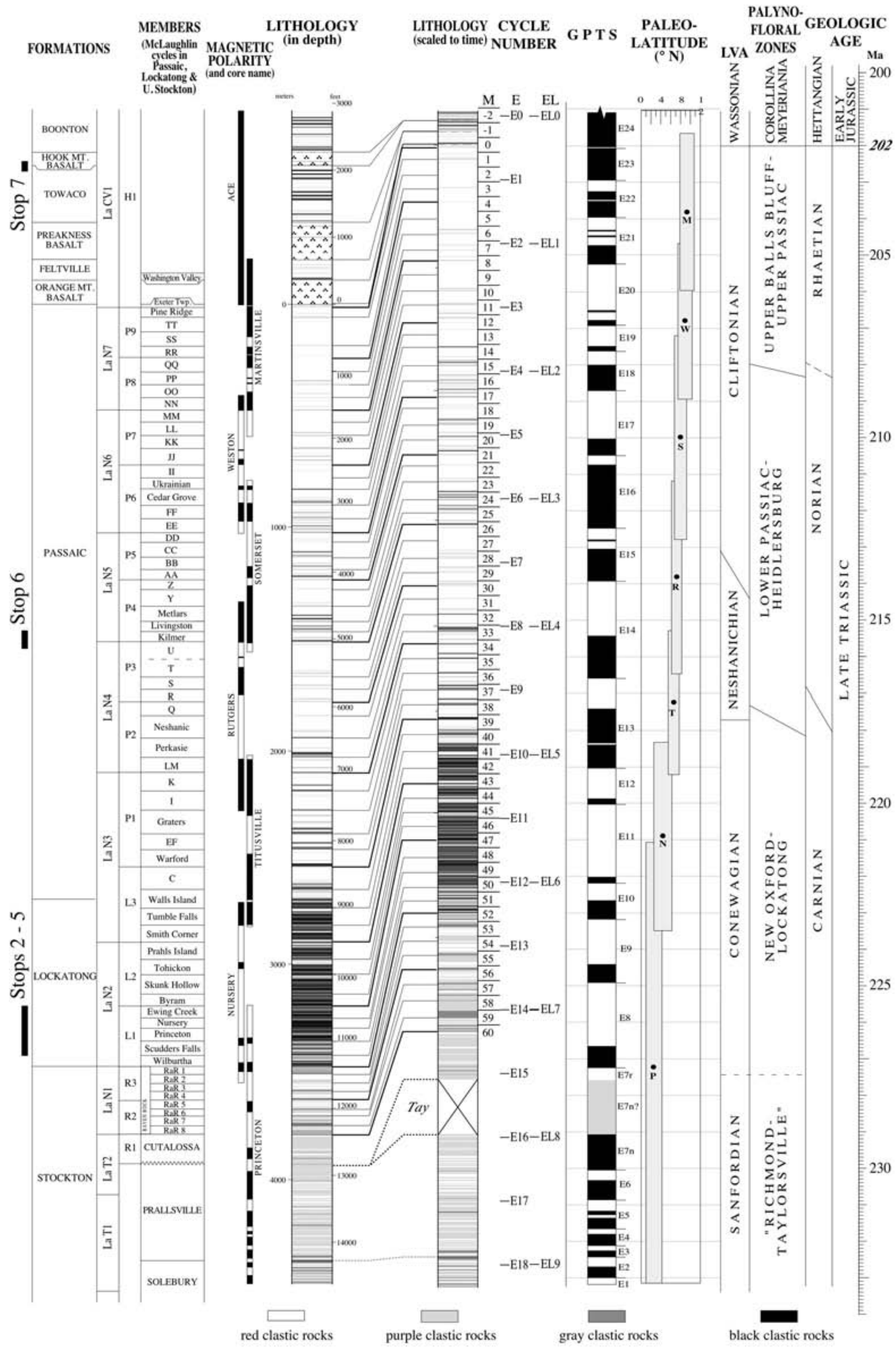

Figure 4. Newark basin section and time scale showing distribution of field stops (adapted from Olsen and Kent, 1999; Olsen et al., 2001a; Olsen et al., 2002b).

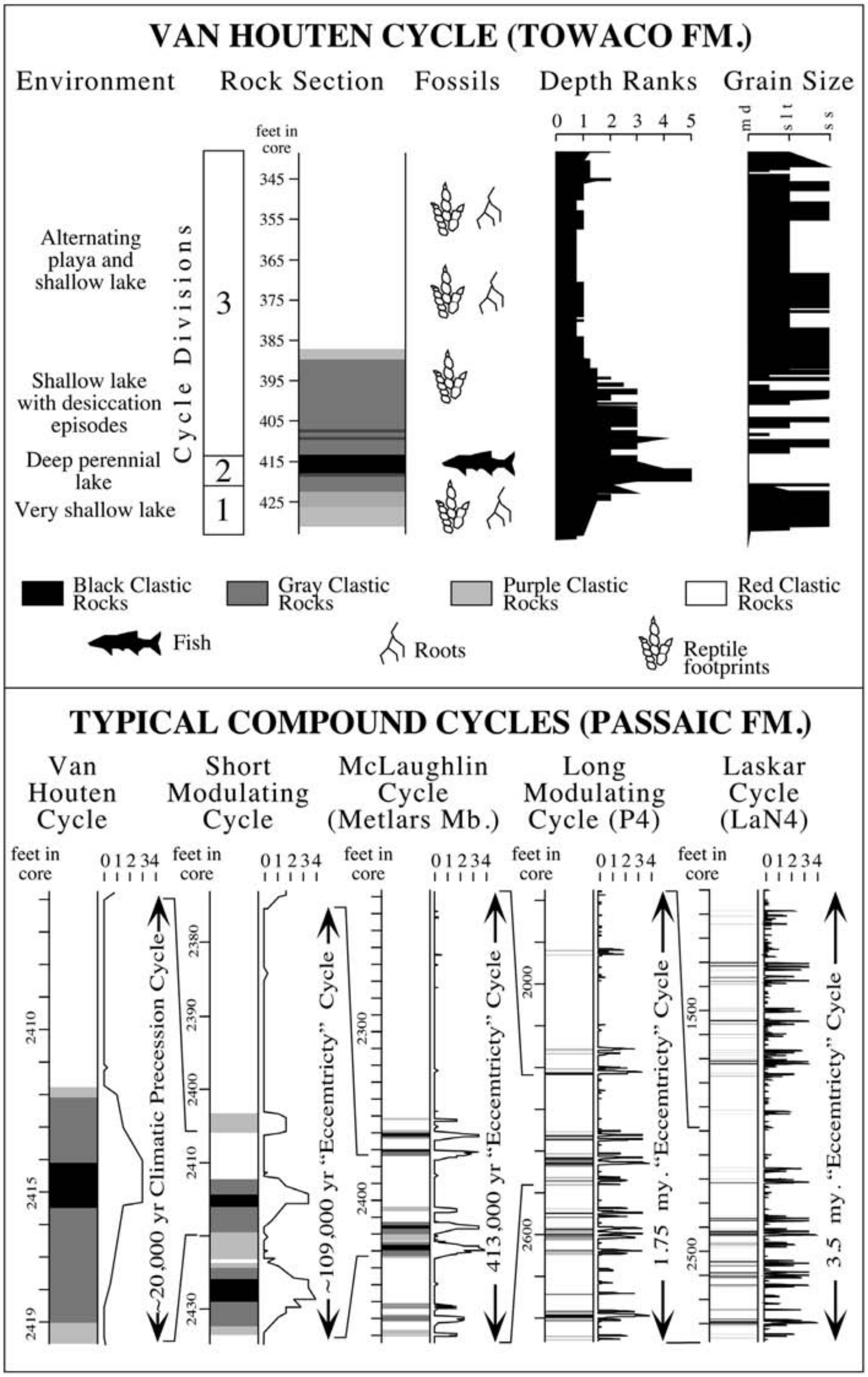

Figure 5. Van Houten and compound cycles (modified from Olsen and Kent, 1999).

the mud flat and an increase in sediment moisture; 2, a red division 2 with vague bedding and perhaps a trace of oscillatory ripples, deposited during sporadic episodes of standing water lasting years or decades; and 3, a division 3 consisting entirely of red massive mudstone, with virtually no visible structures except traces of permeating, superimposed mudcracks deposited during persistent playa conditions.

In vertical succession, the relatively wetness or dryness of Van Houten cycles is modulated by three orders of cycles, producing a characteristic and predictable pattern (Figure 5). The <u>short modulating cycle</u> consists of sequences of one to three relatively wet Van Houten cycles, followed by one to four relatively dry Van Houten cycles. The complete short modulating cycle usually has four to six Van Houten cycles; five cycles are less common. The short modulating cycle is the expression of the highest frequency of the so-called astronomical "eccentricity" cycles, which averages about 100 ky, although they are actually made up of two modes averaging 125 and 95 ky. Sequences of relatively "wet" short modulating cycles (i.e. dominated by wet types of Van Houten cycles) followed by relatively dry short modulating cycles make up the <u>McLaughlin cycle</u> (Olsen et al., 1996a), produced by the 404 ky cycle of eccentricity. This cycle is named for D.B. McLaughlin (1933, 1944, 1946), astronomer at the University of Michigan, who in his spare time mapped many of the 404 ky cycles over much of the Newark basin; ironically, there is no evidence that he ever ascribed the cyclicity to an astronomical cause. The <u>long modulating cycle</u> consists of four to five McLaughlin cycles, and following the pattern of the other modulating cycles, consists of one to three relatively wet McLaughlin cycles, followed by one to four relatively dry ones. This cycle was produced by a 1.75 my eccentricity cycle. Recent analysis reveals that alternations of relatively wet long modulating cycles with relatively dry ones marks out a 3.5 my modulating cycle (Olsen, 2001).

The two types of long modulating cycles (i.e. 1.75 and 3.5 my) find a counterpart in long time series of Neogene climatic precession where their values are 2.4 and 4.7 million years respectively. Laskar (1990, 1999) has shown that the former is caused by the interaction of the gravitational attraction of Mars and Earth, and that the value of the cycle is subject to considerable chaotic drift over time scales of hundreds of millions of years. Laskar (1990) has also shown the present 4.7 my cycle is a direct consequence of a resonance between the orbits of Earth and Mars, producing a cycle with a period twice that of the 2.4 my cycle. We ascribe the 1.75 and 3.5 million year Newark basin cycles to exactly the same mechanism, differing from the modern values because of planetary chaos, as predicted by Laskar (Laskar 1990; Olsen and Kent, 1999). We herein designate the 3.5 my long modulating cycle, as seen in the Newark lithostratigraphy, the <u>Laskar cycle</u>.

All of these astronomical cycles can be recognized by counting lithological cycles in the Newark basin record, either in outcrop or core (e.g. Van Houten, 1964). However, current quantitative determination of their periods in <u>thickness</u> relies on Fourier analysis carried out on numerically ranked sedimentary fabrics (depth ranks), color, or geophysical parameters (e.g. Olsen, 1986; Olsen and Kent, 1996, 1999; Reynolds, 1993). The periods in <u>time</u> are determined by calibration of the sedimentary record by assigning a 404 ky value to the McLaughlin cycle based on the total duration of the Newark basin Triassic section that is fully consistent with current paleontological correlations with existing radiometric time scales (Kent et al., 1995; Olsen and Kent, 1996). The 404 ky McLaughlin cycle in the Newark basin serves as a basis for an astronomically-calibrated geomagnetic polarity time scale for the Late Triassic (Olsen and Kent, 1996, 1999), which is pinned in absolute time by radiometric dates from CAMP igneous rocks (Figure

4). Use of the 404 ky cycle for time scale calibration for an interval hundreds of millions of years ago is justified because this eccentricity cycle is cause by the gravitational interaction of Jupiter and Venus, which should be stable on the scale of billions of years.

Stratigraphic Nomenclature

Traditionally, the Newark basin section has been divided into nine formal lithostratigraphic mappable formations (Figures 4) that generally do not correspond to the tectonostratigraphic divisions (Figures 3, 4) (Kümmel, 1897; Olsen, 1980a; Olsen et al., 1996a). These are, in ascending stratigraphic order: the Stockton Formation, consisting largely of fluvial tan and red sandstones and conglomerates and red mudstones, with less common gray sandstone and black and gray mudstone intervals (maximum thickness >2000 m); the Lockatong Formation, comprised of mostly cyclical gray and black mudstone, with less common red mudstones and sandstones of various colors (maximum thickness >1100 m); the Passaic Formation, made up of cyclical red, gray and black mudstones, sandstones and conglomerates, with red colors being dominant (maximum thickness >5000 m); the Orange Mountain Basalt, which consists of three major flows of high titanium quartz normative (HTQ) tholeiitic basalt (maximum thickness > +300 m) (Puffer and Lechler, 1980; Tollo and Gottfried, 1992); the Feltville Formation, a sequence with two black, gray, and red limestone cycles at its base, followed by vaguely-cyclical mostly-red mudstone, sandstone and minor conglomerate; the Preakness Basalt, made up of two major flows of high iron quartz normative (HFQ) basalt and one major flow of low titanium quartz (LTQ) normative basalt (maximum thickness > 300 m), with a minor red sedimentary interbed (Puffer and Student, 1992; Tollo and Gottfried, 1992); the Towaco Formation, comprised entirely of cyclical red, gray, and black mudstone, sandstone and conglomerate (maximum thickness >375 m); the Hook Mountain Basalt, made up of two major flows of high titanium, high iron, quartz normative (HFTQ) basalt (maximum thickness >150 m) (Puffer and Student, 1992; Tollo and Gottfried, 1992); and finally the Boonton Formation, which is closely comparable to the Towaco Formation (maximum thickness >500 m), except that there are more frequent thin gray beds, but less frequent microlaminated black mudstones.

The members of the Stockton Formation, as defined by McLaughlin, are based on grain size and very general lithologic character, while those of the Lockatong and Passaic Formation are comprised of individual McLaughlin cycles (e.g. McLaughlin, 1944; Olsen et al., 1996a) (Figure 4). The post-Passaic formations of the Newark basin have not been broken into members, with the exception of the limestone-bearing cycles of the lower Feltville Formation which are termed the Washington Valley Member (Olsen, 1980a).

In facies of the Lockatong and Passaic formations in which individual members cannot be traced with certainty, it is often possible to identify either the specific long modulating or Laskar cycles. Therefore, these have been given informal letter and number designations (Figure 4).

Definitions of the Stockton, Lockatong, and Passaic formations are based on gross lithologic features: the Stockton is characterized by the presence of abundant tan sandstones, coupled with infrequent gray and black mudstones; the transition from the Lockatong to the Passaic is based on the decrease in the frequency of gray and black mudstones in the latter. The boundaries between these formations are strongly time-transgressive along the axis of the basin, as illustrated by the fact that McLaughlin cycles of the Lockatong and Passaic Formation - which are defined as members, and which represent the 400 ky cycle - pass laterally between formations.

In the Newark basin, a high frequency of conglomerates correlates to the presence of abundant red beds. Thus the Lockatong Formation practically lacks conglomerates, while the Passaic Formation has abundant conglomerates at its northeastern and southwestern ends, including conglomerates that are laterally equivalent to the upper Lockatong Formation. In addition, parts of the lower Lockatong pass laterally into time-equivalent Stockton Formation at the northeastern and southwestern ends of the basin, especially along the Hudson River (see below), where this lithostratigraphic distinction becomes quite important when discussing the comparatively rich faunas from that area.

In contrast, the Jurassic age formations of the Newark basin are defined on the basis of separation by formations of very different lithology and origin; specifically, the alternation of basalt flow formations of distinct chemistry with intervening and overlying sedimentary formations of distinct character (Figure 4). The boundaries between these formations tend to be approximately isochronous, although there is probably some sedimentary onlap in the up-dip direction on top of each lava flow formation, as illustrated by the Feltville Formation (e.g. Olsen et al., 1996). Each of the Jurassic sedimentary formations includes some conglomerate near the border fault system and along the northeastern basin margin.

SUMMARY OF BASIN TECTONIC HISTORY

It is becoming increasingly clear that the latest Paleozoic and Mesozoic tectonic history of eastern North America, including the Newark basin, is much more complicated than is usually thought of for the type “Atlantic passive margin” (e.g. Malinconico, 2002; Olsen, 1997; Schlische, 2002; Wintsch, 1997; Withjack et al., 1995, 1998). It has long been realized that regional extension relatively quickly followed the late Paleozoic compressive and oblique assembly of Pangea. However, the discovery of median Permian post-orogenic strata (TS I) requires the initiation of deposition, probably in extensional basins, perhaps due to orogenic collapse, very soon after the Late Carboniferous-Early Permian docking of Africa with North America. A depositional hiatus of perhaps 15 million years between TS I and II suggests a temporary cessation of plate divergence. Extension began in earnest in the Middle Triassic, marked by the formation of many relatively small basins that filled with TS II strata. At about 228 Ma, a larger pulse of extension, marked by the TS II – III unconformity, coalesced and enlarged many of these smaller basins, resulting in deposition of TS III in much larger basins. This was followed by the last known major pulse of extension at about ~201 Ma, which resulted in the emplacement of the intrusives and extrusives of the CAMP, and the deposition of the sediments of TS IV. All of these NW-SE-directed extensional pulses repeatedly reactivated appropriately-oriented Paleozoic compressional faults along which the rift basins formed, following the pattern demonstrated by Ratcliffe (1971) and Ratcliffe et al. (1986).

It appears that after more than 32 million years of extension in response to divergent plate motion, compression set in coaxial with the original extension direction.

In the Newark basin, the evidence is as follows: 1, dramatically higher thermal maturities (Malinconico, 2001) and reset zircon fission-track ages (Hoek et al., 1998; Steckler et al., 1993) along the eastern margin of the presently-delineated Newark basin, suggesting significant post-rift uplift and much more erosion of the eastern relative to western basin margin; 2, much if not most of the westward tilt of the Newark basin strata occurred post-depositionally, based on a pervasive mid-Jurassic paleomagnetic overprint, which itself also suggests a major mid-Jurassic fluid-flow event (Kent et al., 1995; Witte and Kent, 1991); 3, pervasive small-scale bedding-plane faults that show consistent reverse (in present coordinate space) motion based on offset of bedding-normal joints, small scale folds and thrust faults, and slickenline orientations that appear, on the basis of our casual observations, to be coaxial with the inferred older extension direction, thus requiring post-depositional NW tilting and NW-SE shortening in an approximately horizontal plane. These observations and interpretations are consistent with the basin inversion geometry described by Withjack et al. (1998). This geometry is based on physical models that produce a major up-arching of the basin during compression, but with little reverse motion on the boundary faults and only subtle compressional deformation of the basin fill. Erosion of several km of basin section and surrounding basement rocks took place during this late Early to Middle Jurassic tectonic inversion event. This was the second of similar events through the Mesozoic and Cenozoic which progressed from south to north all along the mid- to North Atlantic margin along crustal segments defined, on the east, by major strike-slip plate boundaries inherited from the Paleozoic assembly of Pangea. The segment containing the Newark, Hartford, Fundy and most of the Moroccan rift basins is bound by the Minas Fault – Gibraltar Fault zone on the north and the Long Island Platform boundary – South Atlas fault zone on the south (Figure 2).

By the Early Cretaceous the present erosional level had been reached in the Newark basin, and marine and marginal-marine Cretaceous and Cenozoic coastal plain deposits onlapped the rift deposits. Northwest-southeast compression slowed sometime in the Cenozoic switched to the NE-SW horizontal compression that persists to the present day (Goldberg et al., 2002; Zoback, 1992; Zoback and Zoback, 1989).

TECTONOSTRATIGRAPHIC SEQUENCES OF THE NEWARK BASIN AND THEIR DEPOSITIONAL ENVIRONMENTS AND PALEONTOLOGY

Tectonostratigraphic Sequence I

There is no compelling evidence for TS I in the Newark basin, although there are hints on the NB-1 seismic line that a pre-TS II package may be present adjacent to the border fault system (Schlische, 2002).

Tectonostratigraphic Sequence II

All of the Stockton Formation below the Cutaloosa Member appears to belong in TS II (Figures 4). This is based on magnetostratigraphic and lithological correlation between the upper Stockton Formation and the Newfound Formation (Taylorsville basin) that appears to reveal a small hiatus at the base of the Cutaloosa member (LeTourneau, 1999, 2002) that we consider to be the TS II – TS III unconformity. Lithostratigraphically, TS II consists of the Prallsville Member, made up of mostly tan and pink arkosic sandstone with minor conglomerate and red bioturbated red mudstone, and the underlying Solebury Member, comprised of tan and gray sandstone with proportionally more conglomerate and gray, black, and red mudstones. These members appear, at least in outcrop and existing core, to be mostly of fluvial origin, although the thin gray and black mudstones of the Solebury Member hint at lacustrine or paludal intervals. Analysis of the NB-1 seismic line suggests that the fluvial TS II sequences at the present-day surface might pass into lacustrine sequences at depth (Reynolds, 1993).

Paleontologically, TS II is virtually unprospected. The total assemblage known so far includes root traces, some cycadeoid compressions and silicified conifer wood (McLaughlin, 1959), the extraordinarily abundant arthropod burrow *Scoyenia*, and, most tantalizingly, from the base of the Solebury Member, a fragmentary mold of a jaw of *Calamops paludosus (Sinclair, 1917),* a very large, probably cyclotosaurian, amphibian (Figure 6). Cyclotosaurian amphibians are virtually absent from the rest of tropical Pangea, except the Middle Triassic (Anisian) age Moenkopi Formation of the western US, the Economy beds of the Fundy basin, assigned a Anisian age (Baird, 1986a; Olsen et al., 1989), and the upper part of unit T4 of the Timesgadouine Formation of the Argana Basin of Morocco (Dutuit, 1976; Jalil, 1996). This suggests that the basal Solebury Member could be part of the Economian faunachron of Lucas and Huber and Lucas (1993) and thus Middle Triassic in age. The Prallsville Member of the Stockton Formation is, thus far, devoid of fossils except *Scoyenia* and roots, but it probably is of early Late Triassic age, as is the upper part of TS II in other central Pangean rifts (Olsen, 1997). Additional prospecting in the Solebury and Prallsville members could be very rewarding, as TS II formations are by far the most fossiliferous portions of the stratigraphic sections in most of the rest of the eastern North American and Moroccan basins.

There is no evidence of TS II in the northeastern part of the Newark basin, near the Hudson River. The rapid decrease in outcrop width of the Stockton from Princeton towards the Hudson is probably due to a combination of progressive onlap of younger TS II onto basement, and truncation of younger beds of TS II by the TS II – TS III unconformity (Figures 3, 7).

Tectonostratigraphic Sequence III

In the Newark basin, TS III is by far the most widespread and, at least in outcrop, the most heterogeneous portion of the basin fill, consisting of the Cutaloosa and Raven Rock members of the Stockton Formation, the entire Lockatong Formation, and nearly all of the Passaic Formation. This sequence is characterized by the dominance of cyclical lacustrine and laterally-equivalent marginal lacustrine and fluvial strata. It is richly fossiliferous, especially in the northeastern part of the basin.

The Cutaloosa and Raven Rock members are poorly known. The former appears to be the basal coarse-grained sequence of TS III. The latter consists of thick cycles of gray and tan sandstone with subordinate black and gray mudstones, overlain principally by red mudstones and tan sandstones. Relatively large-scale sets of tilted surfaces are abundant in the outcropping Raven Rock Member in the central Newark basins, and

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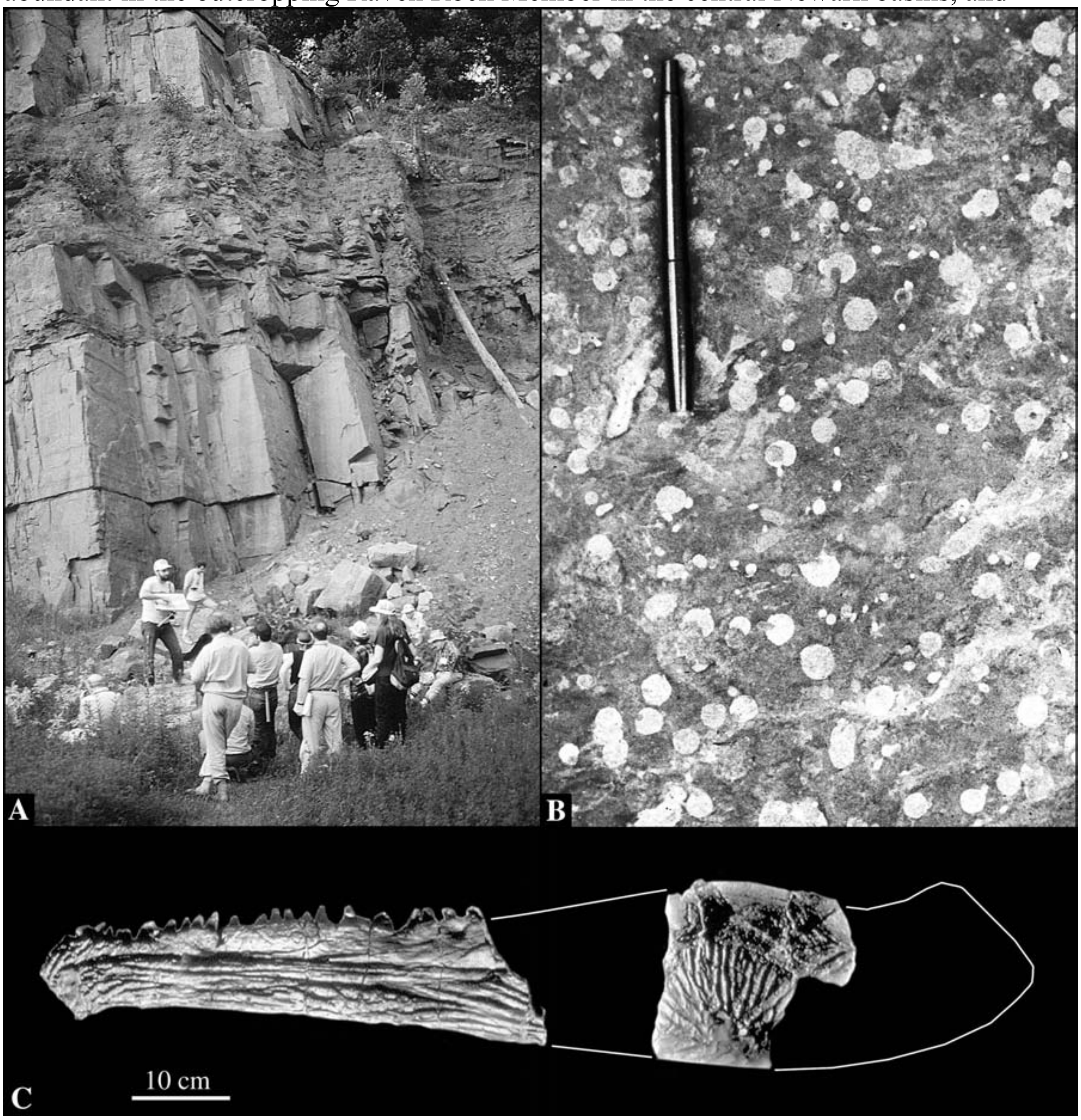

Figure 6. Facies and fossils from Tectonostratigraphic sequence II. **A**, quarry in Prallsville Member of the Stockton Formation, Prallsville, New Jersey. **B**, *Scoyenia* burrows in red-purple siltstone from same quarry as in A; **C**, rubber cast of natural mould comprising type and only specimen of the giant ?Capitosaurid amphibian *Calamops paludosis* near the base of the Solebury Member of the Stockton Formation (photo courtesy of W. Seldon and D. Baird).

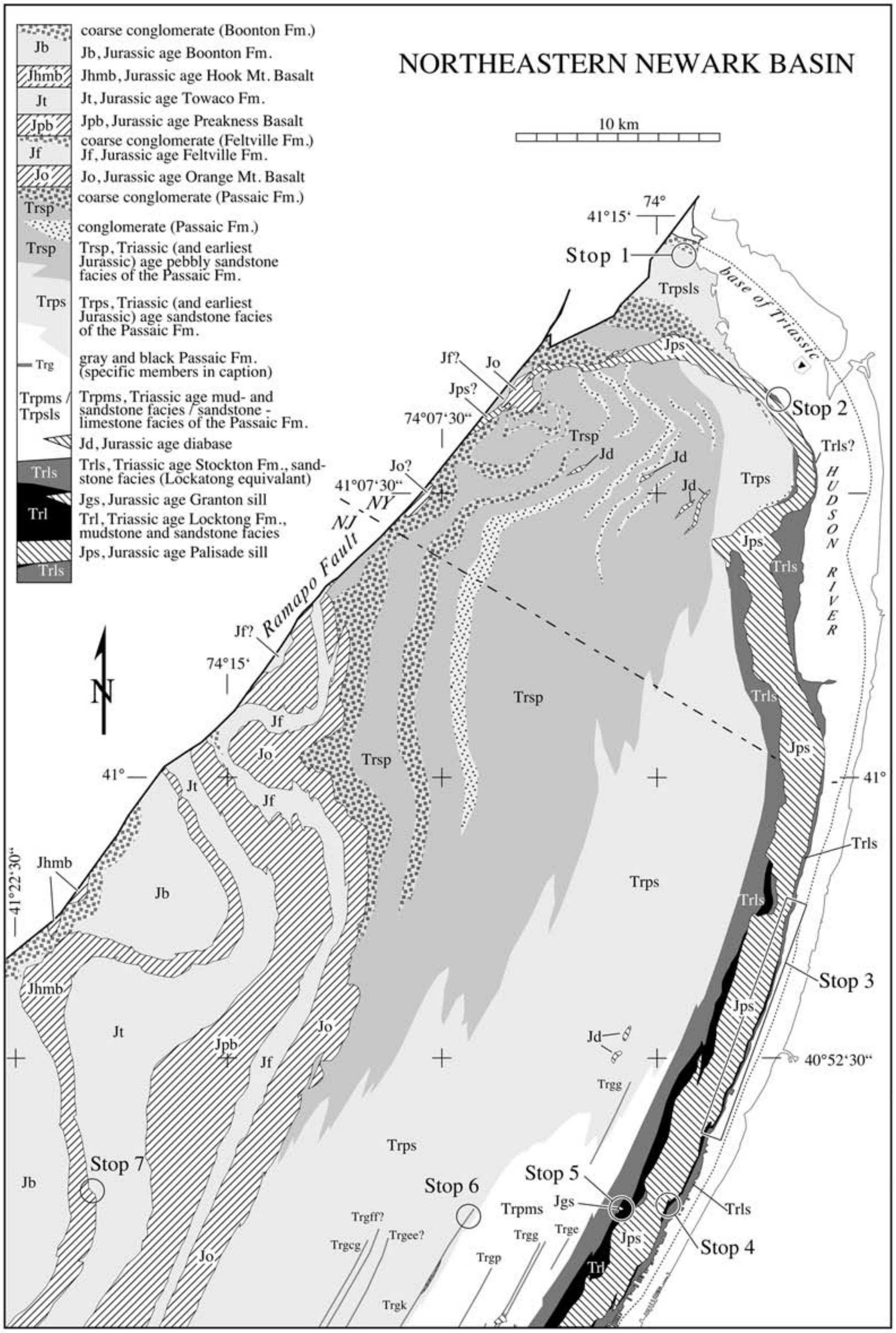

Figure 7. Bedrock geological map of the northern Newark basin showing the field stops (Based on Ratcliffe, 1999; Parker, 1993; Parker et al., 1988; and Olsen et al., 1996a).

many convincingly appear to be lacustrine deltaic sequences (Smoot, 1991; Turner-Peterson and Smoot, 1985). The large-scale cycles, tens of meters thick, appear to be a coarser, more proximal expression of McLaughlin cycles (Olsen and Kent, 1999), in which the small-scale cycles are masked by both the large size of the sedimentation units and depositional surface relief. Reynolds (1993) has shown that the Raven Rock Member passes down-dip into a seismic facies indistinguishable from the Lockatong Formation, suggesting the subsurface presence of a time-equivalent of the Raven Rock Member that is more like the Lockatong in cyclical style and sedimentary facies. This is consistent with magnetostratigraphic correlation to the more southern Newark Supergroup basins, that indicates the well-developed cyclical lacustrine strata of the lower member of the Cow Branch Formation, lower Port Royal Formation, and Cumnock Formation, are time equivalents of the lower Raven Rock Member.

Strata of the Raven Rock Member are paleontologically poorly prospected. The known assemblages consist of a single palynoflora (Cornet, in Olsen and Flynn, 1989), a relatively rich compression plant assemblages (Ash, 1986; Axsmith and Kroehler, 1989; Bock, 1952; McLaughlin, 1959), clams from several localities (Axsmith and Kroehler, 1989; Olsen and Flynn, 1989) *Scoyenia*-type burrows, conchostracans, and beetle elytra (Axsmith and Kroehler, 1989; Olsen and Flynn, 1989) (Figure 8). These assemblages are consistent with an early Late Triassic (Late Carnian) age. Again, the fossil record of age-equivalent strata in the more southern basins suggests that careful prospecting will be well rewarded.

The overlying Lockatong Formation consists almost entirely of dramatically cyclical mudstone; these cycles are the most fossiliferous in the Newark basin section (Figure 8). The full suite of lithological cycles attributed to astronomical forcing is present. In general, the most fossiliferous Van Houten cycles tend to be those in the wettest phases of the 3.5 my Laskar cycle, 1.75 my long modulating cycles, 404 ky McLaughlin cycle, and 100 ky short modulating cycles . These include the Princeton, Nursery and Ewing Creek members of the lower Lockatong, the Skunk Hollow and Tohickon members of the middle Lockatong, and the Smith Corner and Walls Island members of the upper Lockatong, although the latter have hardly been prospected.

In the central part of the Newark basin along the Delaware River, Van Houten cycles average about 5.5 m thick, short modulating cycles average about 25 m thick, McLaughlin cycles average 100 m thick, the long modulating cycles average 440 m thick, and the Laskar cycles, 880 m. In both outcrop and available core, the cycles thin away from this area, but in the subsurface to the west, they probably thicken.

Plant assemblages are surprising rare in the Lockatong; described assemblages consist of a single palynoflorule (Cornet, in Olsen and Flynn, 1989) and a few cycadeoid, conifer and equisetalian compression fossils and molds (McLaughlin, 1959; Olsen and Flynn, 1989) in gray and purple mudstones and fine sandstones. However, animal remains can be exceedingly common. Invertebrates found so far include: several types of clams (Heath and Good, 1996; McLaughlin, 1959; Olsen and Flynn, 1989) from gray platy mudstones and ripple cross-laminated fine-grained sandstones; *Scoyenia* and various other trace fossils (Metz, 1995b); burrows in red, gray and purple mudstones and sandstones; spectacularly abundant conchostracans, darwinulid ostracodes; and a single undescribed large crustacean in black to gray laminated to microlaminated mudstones. Vertebrate body fossils are very strongly dominated by a stereotyped assemblage of

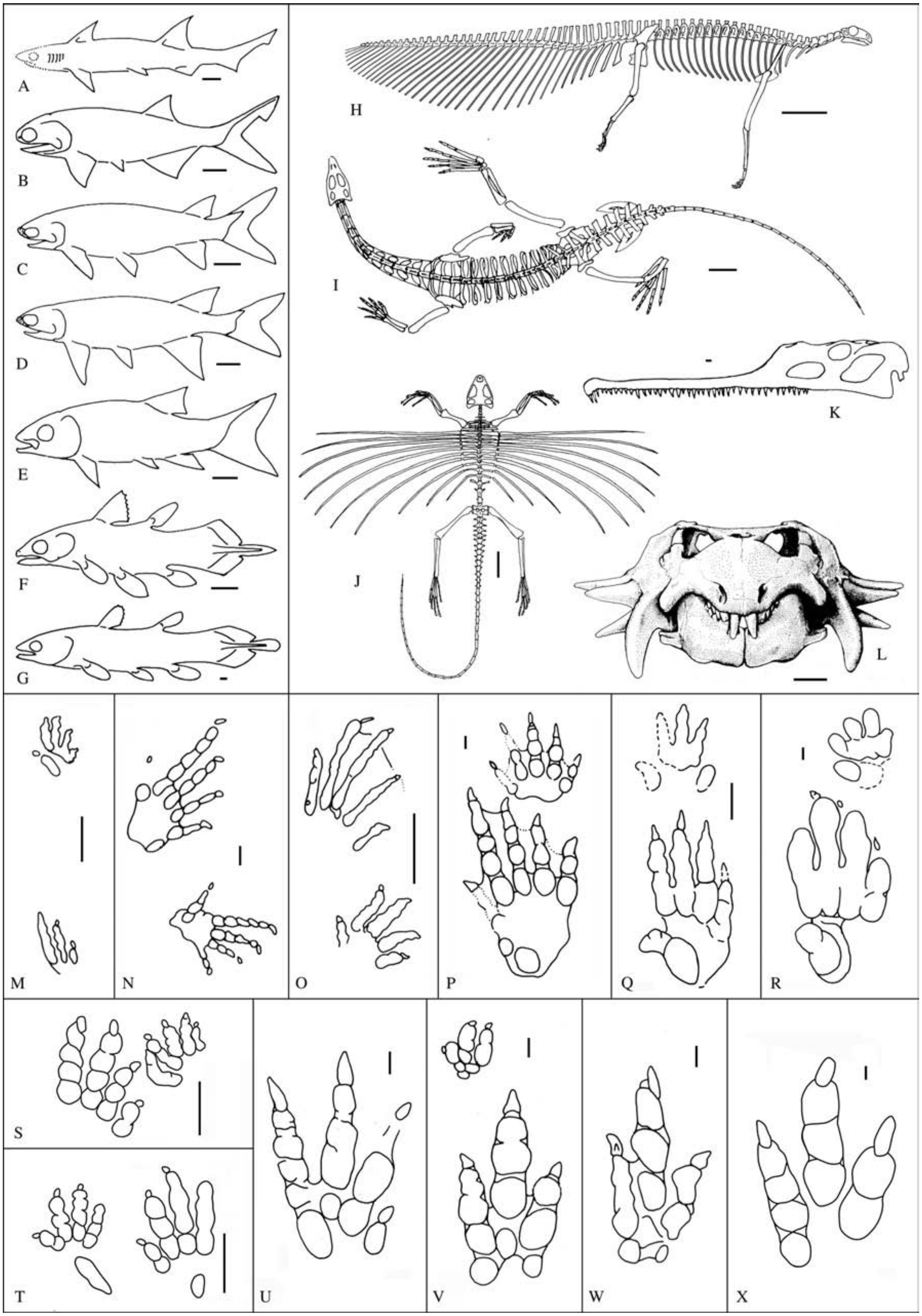
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Figure 8. Representative vertebrates from Tectonostratigraphic Sequence III of the Newark basin (modified from Olsen, 1988; Olsen and Flynn, 1989): **A**, the freshwater shark *Carinacanthus jepsoni* (Lockatong Fm.); **B**, the palaeoniscoid *Turseodus* (Lockatong Fm.); **C**, the palaeonisciform *Synorichthyes* (Lockatong and Passaic fms.), **D**, the palaeonisciform *Cionichthyes* (Lockatong and Passaic fms.); **E**, the holostean *Semionotus* (Lockatong and Passaic fms.); **F**, the small coelacanth *Osteoplurus newarki* (Lockatong Fm.); **G**, the large coelacanth cf. *Pariostegeus* (Lockatong Fm.); **H**, the drepanosaurid diapsid *Hypuronector limnaois*; (Lockatong Fm.); **I**, the tanystropheid diapsid *Tanytrachelos ahynis* (Lockatong Fm.); **J**, the lepidosaur *Icarosaurus seifkeri* (Lockatong Fm.); **K**, rutiodontid phytosaur *Rutiodon* (Lockatong and Stockton fms.); **L**, the parareptile *Hypsognathus fenneri* (Passaic Fm.); **M**, the probable lepidosauromorph track *Rhynchosauroides brunswickii* (Lockatong and Passaic fms.); **N**, the probable diapsid reptile track *Rhynchosauroides hyperbates* (Stockton, Lockatong, and Passaic fms.); **O**, the probable tanystropheid track *Gwynnedichnium* (Lockatong and Passaic fms.); **P**, the phytosaur track *Apatopus lineatus* (Stockton, Lockatong, and Passaic fms.); **Q**, the probable aetosaurid archosaur track *Chirotherium lulli* (Passaic Fm.); **R**, the probable rauisuchian archosaur *Brachychirotherium parvum* (Stockton, Lockatong and Passaic fms.); **S**, the ?sphenosuchian track "new taxon A" (Passaic Fm.); **T**, the probable crocodiliomorph track *Batrachopus bellus* (Passaic Fm.); **U**, the dinosaurian track "new genus 1" (Passaic Fm.); **V**, the ornithischian dinosaur track *Atreipus milfordensis* (Stockton, Lockatong, and Passaic fms.); **W**, the theropod dinosaur track *Grallator* (Stockton, Lockatong, and Passaic fms.); **X**, the theropod dinosaur track *Anchisauripus* (Passaic Fm.). Scale is 1 cm.

aquatic and lake margin forms, including, from microlaminated mudstones, abundant and fairly diverse articulated fish and small diapsid reptiles, and phytosaur teeth (Figure 8). Less well-laminated mudstones sometimes have disarticulated remains of the same kinds of vertebrates; occasional pockets of small tetrapod bones occur in more massive gray and red mudstones. Tetrapod footprints can be very abundant in divisions 1 and 3 of Van Houten cycles, and include one spectacularly well-preserved assemblage (Figure 8). Vertebrate coprolites are common in many facies. As a rule, vertebrate fossils are much more common outside of the microlaminated units in marginal facies of the Lockatong, especially where well-developed deltaic deposits are present in division 3 of the Van Houten cycles. The one facies of the Lockatong in which fossils of all kinds are very rare is the gray and red mudcracked massive mudstone characteristic of division 3 of the Van Houten cycles, especially in the dryer phases of the McLaughlin, long modulating, and Laskar cycles.

In the northeastern Newark basin (Figure 9), the outcrop belt of the Lockatong on both sides of the Palisade 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 (see Stops 4-6). 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. 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.

A notable feature of the microlaminated portions of division 2 of the Van Houten cycles near the Hudson River is the truly remarkable abundance of fish, especially the coelacanth *Osteopleurus newarkii*, the palaeoniscoid *Turseodus* spp. and the holostean *Semionotus braunii*, as well as two genera of small tetrapods, the tanystropheiid prolacertian archosauromorph diapsid *Tanytrachelos ahynis*, and the bizarre drepanosaurid diapsid *Hypuronector limnaios*, (Colbert and Olsen, 2001; the "deep tailed swimmer" of Olsen, 1980a) (Figure 8). Less common in the same units is a large coelacanth, probably *Pariostegeus* sp., the redfieldiid palaeonisciform fish *Synorichthyes* and *Cionichthyes*, the gliding lizard-like lepidosauromorph diapsid *Icarosaurus seifkeri*, and phytosaur teeth. More massive mudstones of division 3 of the Van Houten cycles usually have scraps and sometimes more complete remains, including a skull (Colbert, 1965), of phytosaurs; metoposaur amphibian fragments (e.g. AMNH 23579); and locally-common fish fragments and coprolites, even in mudcracked beds. Only a few poorly preserved reptile tracks, all small probable-dinosaurian forms, have been found (e.g. Gratacap, 1886).

Strata mapped as Stockton Formation along the Hudson River are almost certainly time-equivalents of the lower Lockatong Formation, and the transition from typical Lockatong (i.e. lacustrine) facies to Stockton (i.e. fluvial) facies in the Princeton and Nursery Members can be observed along the river, going north from Hudson and Bergen counties (New Jersey) into Rockland County (New York) (Figure 7). A particularly distinctive facies of the Stockton, probably the lateral equivalent of the Scudder Falls and Wilbertha members of the basal Lockatong, occurs below the Princeton Member along the Hudson River from at least Hoboken to Alpine (New Jersey), with similar facies occurring sporadically at least to Snedens Landing (New York). This facies consists of meter-scale cycles of tan cross-bedded coarse-grained or pebbly arkose passing upward into mottled or streaked purple, red and tan arkosic sandstone, and then upward into bright purplish-red massive mudstones, that often have large, widely spaced ?shrinkage cracks filled with arkose from the overlying cycle. This facies has produced one notable fossil, a partial post-cranial skeleton of a large phytosaur named *Rutiodon manhattanensis* (von Huene, 1913), found near the west abutment of the George Washington Bridge (see Stop 3g).

Passing further to the north along the Hudson into Rockland County, the Lockatong Formation disappears, having passed laterally into the Stockton Formation. From Piermont northward to Haverstraw, the Stockton consists of alternations of decameter-scale sequences of mostly red sandstones with minor purple and tan sandstones and red mudstones, and similar-scale sequences of purple, gray, and tan sandstones and conglomerates with subordinate red and purple mudstones. Fluvial, deltaic and marginal lacustrine facies appear to be present. We presume that the more red sequences represent drier phases of McLaughlin cycles, but cannot demonstrate this. Both types of sequences contain abundant fossils, including bones (see Stops 2 and 3).

Reptile tracks, *Scoyenia*, and root traces are common in the mostly-red facies, while the drab facies contains bones, tracks and root traces. So far, the osteological remains include phytosaur teeth and bone scraps, a small amphibian dermal bone, unprepared portion of a tetrapod skull or vertebra, and numerous indeterminate bone fragments. This material, while scrappy, is very important because it represents more terrestrial communities than the temporally equivalent assemblages in the Lockatong . Reptile tracks include poorly preserved trackways from Blauvelt (NY), which have been widely assigned to grallatorid ichnotaxa, and even attributed to the ceratosaurian theropod dinosaur *Coelophysis* (Fisher, 1981). However this designation is probably incorrect, and in fact the tracks more likely belong to *Atreipus* isp., which was most likely made by an ornithischian dinosaur (Olsen and Baird, 1986). Other tracks are much better preserved and include a true small grallatorid, unquestionable *Atreipus*, the (?)phytosaurian track *Apatopus*, the probable-rauisuchian archosaur track *Brachychirotherium* sp., and the lepidosaurian track *Rhynchosauroides* cf. *R. hyperbates* and related traces (Figure 8). According to Huber and Lucas (1993) and Lucas and Huber (2002), the tetrapods of the Lockatong and equivalent-age strata indicate inclusion in the Conewagian faunachron of Late Carnian age.

The Passaic Formation conformably overlies the Lockatong Formation in most of the Newark basin, and for the most part continues the cyclical pattern seen in the latter (Figures 4, 5). In the two central basin fault blocks near the Delaware River, the transition from Lockatong to Passaic occurs in precisely the same cycles (Olsen et al., 1996a). The lowermost Passaic is marked by an abrupt switch to much more abundant red massive mudstone in the upper half of the Walls Island Member, as well as an increase in Van Houten cycle thickness of about 33%, seen in both outcrop and core. As is the case of the underlying Lockatong, cycle thickness in the Passaic Formation decreases west, east, and northeast of this central area. Cycle thickness increases in outcrop and presumably in the subsurface to the west towards the border fault. However, away from the center of the basin, the transition occurs at lower stratigraphic levels, but most of the individual members are still identifiable.

Overall, the apparent wetness of the Lockatong and Passaic formations decreases cyclically upward. Part of this trend is certainly due to the northward drift of Pangea, carrying the Newark basin into more arid climes, as evidenced by the increase in evaporates and massive vesicular fabric in division 3 of the Van Houten cycles (Smoot, 1991). However, this trend is also due in part to the progressive filling and widening of the basin through the waning phases of the major extensional pulse responsible for the formation of TS III (c.f. Schlische and Olsen, 1990).

It is only recently that the paleontological richness of the Passaic Formation has become appreciated. Far from being a "monotonous sequence of red beds" devoid of fossils, the Passaic Formation is in fact very fossiliferous in the basin margin facies (Figure 8). There are many palynoflorules recovered from the TS III portion of the Passaic formation (Cornet, 1977; Cornet and Olsen, 1985; Fowell, 1993; Fowell and Olsen, 1993), several significant macrofossil assemblages (Cornet, 1977; McLaughlin, 1959), and root traces are very common in many areas and facies. Invertebrates include abundant *Scoyenia* as well as other trace fossils (Metz, 1993, 1995a, 1998), conchostracans and ostracodes in gray to black portions of division 2 of the Van Houten cycles, and from one locality, insects, apparently dipteran (fly) larvaeFish are much more rare in the Passaic than in the Lockatong; this correlates with both the much lower frequency of microlaminated strata as well as a lack of prospecting. The only articulated fish are *Semionotus*, from several localities in the black shale of division 2 of one Van Houten cycle near the base of the Warford Member (Late Carnian), the same basic facies in the Ukrainian Member (Rhaetian), and a gray laminated siltstone in Member L-M (early Norian or latest Carnian). Disarticulated fish occur at a number of localities, and include fragments of small coelacanths and redfieldiids (member I, Late Carnian or Norian age; Olsen et al., 1982) and complete *Semionotus* (Warford Member, Late Carnian age; member OO, middle Rhaetian).

Tetrapod remains are the most spectacular fossils from the Passaic Formation. Diverse and often very well preserved tetrapod footprints are very abundant at many horizons. The richest areas are the northeastern and southwestern parts of the basin, particularly the Jacksonwald syncline (Figure 8). Initially in the Passaic Formation(Carnian and early Norian), the ornithischian dinosaurian form *Atreipus* is very abundant, but its last known occurrence is in the Rhaetian (members II and JJ). The abundance and size of grallatorid theropod dinosaurian tracks, traditionally (e.g. Lull, 1953) placed in the ichnogenera *Grallator* and *Anchisauripus*, increase through the Passaic in TS II, and they are the only dinosaurian forms in Late Rhaetian aged strata. There are, however, diverse other ichnogenera in the Passaic, representing procolophonids, tanystropheids, lepidosauromorphs, and various (phytosaurian, aetosaurian and rauisuchian) archosaurs (Figures 8, 10).

Discoveries of osteological tetrapod remains are becoming increasingly common. In particular, red and gray massive root-bearing sandstones and siltstones have abundant remains locally. Thus far, skeletal material includes fragments to articulated partial skeletons of metoposaurid amphibians, the procolophonid *Hypsognathus fenneri*, phytosaurs, the aetosaur *Aetosaurus (Stegomus) arcuatus*, and the crocodylomorph *Protosuchus*, as well as unidentified forms. As an example of how rich the Passaic can be, one locality in the Jacksonwald syncline (member TT), has produced several hundred specimens of *Hypsognathus*, including five skulls and three partial skeletons.

The Van Houten cycles so obvious in some parts of the basin gradually become less well-marked as we progress into the coarser facies of the Passaic Formation in the northeastern part of the basin; therefore some important fossil localities cannot be placed in a specific McLaughlin cycle. Nonetheless, the position within the long modulating and/or Laskar cycles can still be ascertained. In the northeastern Newark basin, the TS III portion of the Passaic has yielded two palynoflorules (in the Cedar Grove Member, and the upper part of long modulating cycle P4), one compression flora assemblage (from the latter palynofloral locality), one major footprint locality (lower part of long modulating cycle P4), a minor footprint locality that nonetheless has produced significant forms (lower Laskar cycle LaN7; Baird, 1986b); and surprisingly abundant remains of *Hypsognathus fenneri* from several localities (Laskar cycles LaN5, LaN6, and LaN7; Colbert, 1946; Gilmore, 1928; Sues et al., 2000), as well as tooth and bone fragments of indeterminate tetrapods from several localities. The relatively common *Hypsognathus* and indeterminate scraps indicate that every red sandstone outcrop or temporary exposure should be carefully examined.

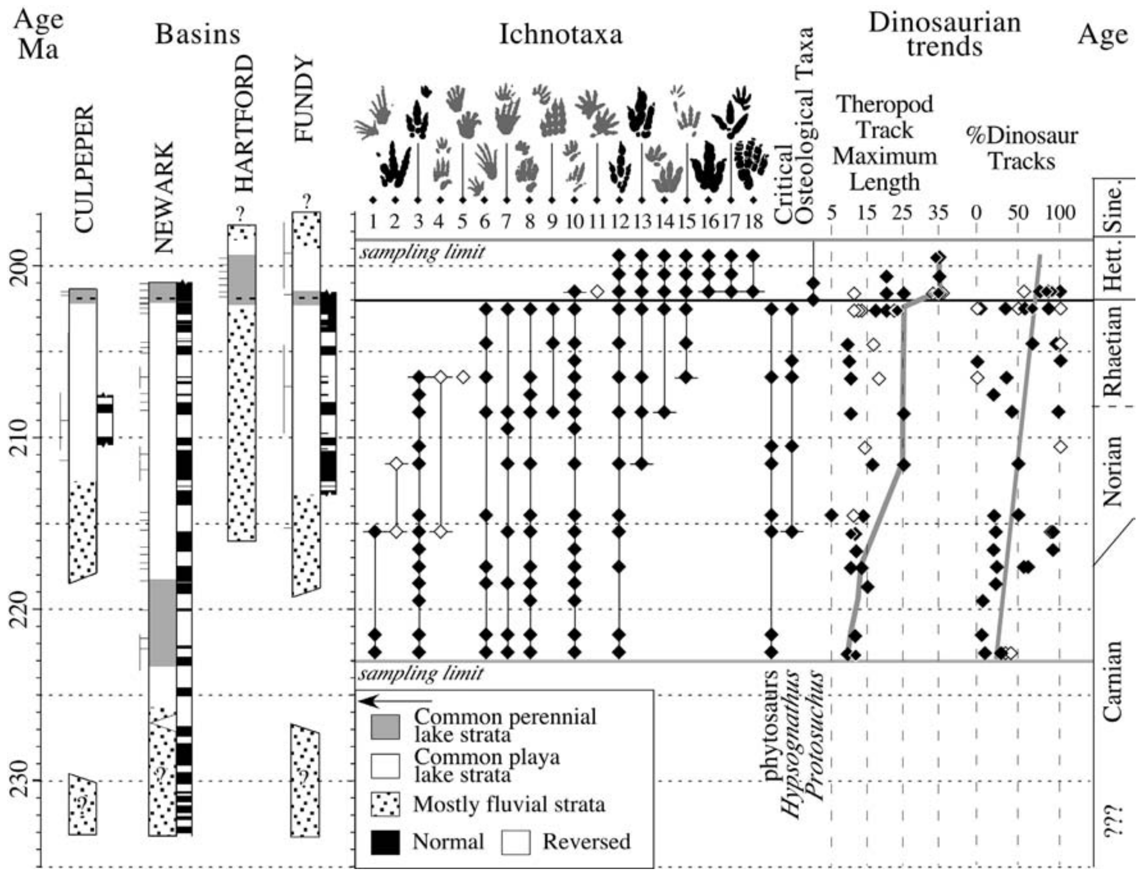

Figure 10. Correlation of four key basins of the Newark Supergroup showing the temporal ranges of footprint ichnogenera and key osteological taxa binned into 1 my intervals showing the change in maximum theropod dinosaur footprint length (line drawn through maximum) and percent of the assemblages that consist of dinosaur tracks (with linear regression line). Short, horizontal lines adjacent to stratigraphic sections show the position of assemblages and the attached vertical lines indicate the uncertainty in stratigraphic position. Ichnotaxa are: 1, *Rhynchosauroides hyperbates*; 2, new dinosaurian genus 1; 3, *Atreipus*; 4, *Chirotherium lulli*; 5, *Procolophonichnium*; 6, *Gwyneddichnium*; 7, *Apatopus*; 8, *Brachychirotherium parvum*; 9, new taxon B; 10, *Rhynchosauroides* spp.; 11, *Ameghinichnus*; 12, "*Grallator*"; 13, *Anchisauripus*; 14, *Batrachopus deweyii*; 15, "*Batrachopus*" *gracilis*; 16, *Eubrontes giganteus*; 17, *Anomoepus scambus*; 18, *Otozoum moodii*. Stratigraphic and magnetostratigraphic columns and correlations modified from (Olsen, 1997) (from Olsen et al., 2002b).

Tectonostratigraphic Sequence IV

Based on lithostratigraphic and magnetostratigraphic correlations laterally over distances greater than 100 km, the TS III – IV contact is a correlative conformity over much of the Newark basin with liittle evidence for a tectonostratigraphic sequence boundary, probably because of the very deep level of post-rift erosion. However, in the northeastern Newark basin, northeast of East Orange (NJ), there is an abrupt change in facies from coarse red sandstone, conglomerate, and massive mudstone below, to much better-bedded sandstones and mudstones above, characteristically with very abundant reptile footprints. This transition could represent the TS III - IV boundary. Further northeast near Ladentown (NY), the Passaic Formation thins dramatically below the Orange Mountain Basalt, coincident with an abrupt change in strike, which could be due to truncation by the TS III – IV unconformity (Figure 7). The evidence for a well-developed TS III – IV unconformity is, however, far from conclusive. It is very important to stress that the TS III – IV boundary and any associated unconformity occurs <u>well below</u> the palynologically-identified Triassic-Jurassic boundary. There is no evidence for - and much against - an unconformity or hiatus at the Triassic-Jurassic boundary in the Newark basin (Fowell and Olsen, 1993, 1995; van Veen, 1995).

TS IV sedimentary sequences show a development of persistent lacustrine conditions of a magnitude not seen since the Lockatong, some 16 million years earlier (Figure 4). This transition occurs at the base of the Pine Forge Member and continues in the strata between and above the extrusive basalt formations. As a consequence, sedimentary units of TS IV are very rich in floral remains, fish, and reptile footprints (Figure 9).

The floral assemblages of TS IV consist of stromatolites around trees from a single locality (middle Towaco Formation), common palynomorph assemblages from many gray units, several important compression fossil assemblages, and root traces (Figure 9). Most gray units produce palynomorph fall into two groups, that allow us to recognize the Triassic-Jurassic boundary. Palynomorph assemblages from the Pine Ridge and lower Exeter Township members are diverse and tend to have varying proportions of the Triassic taxon *Patinasporites densus*, along with other Triassic forms, and varying amounts of *Corollina* spp. (this genus never exceeding about 60% of the assemblage). Strata of the upper Exeter Township Member and the remainder of TS IV lack Triassic taxa and are overwhelmingly dominated by *Corollina* (Cornet, 1977; Fowell, 1993; Fowell and Olsen, 1993; Olsen et al., 1990), with only minor variations in the composition of palynomorph assemblages. The transition has been studied most intensively in the Jacksonwald syncline, where the two types of assemblages are separated by only 5 meters of strata (Figure 11). A thin (20-30 cm) interval, about 70 cm above the last Triassic palynoflorule, is dominated by fern spores. We regard the base of this “fern spike” unit as the Triassic-Jurassic boundary (Fowell et al., 1994; Olsen et al., 2002b,c).

Macrofloral compression assemblages occur in every formation in TS IV and tend to be in the same strata as the palynomorph assemblages. An important but structurally poorly-preserved assemblage occurs in the uppermost Passaic Formation near the Clifton-Paterson (NJ) town boundary. This assemblage is dominated by several forms of *Brachyphyllum*-type conifer shoots, conifer cone fragments, and leaf fragments of the fern *Clathropteris meniscoides*. A very poorly preserved palynoflorule from this assemblage contains only *Corollina*, suggesting a Jurassic age, probably (based on cycles) less than 10 ky younger than the Triassic-Jurassic boundary. Beautifully preserved compression assemblages in gray mudstones - particularly from several localities in the Oldwick Syncline in the Washington Valley Member - produce abundant

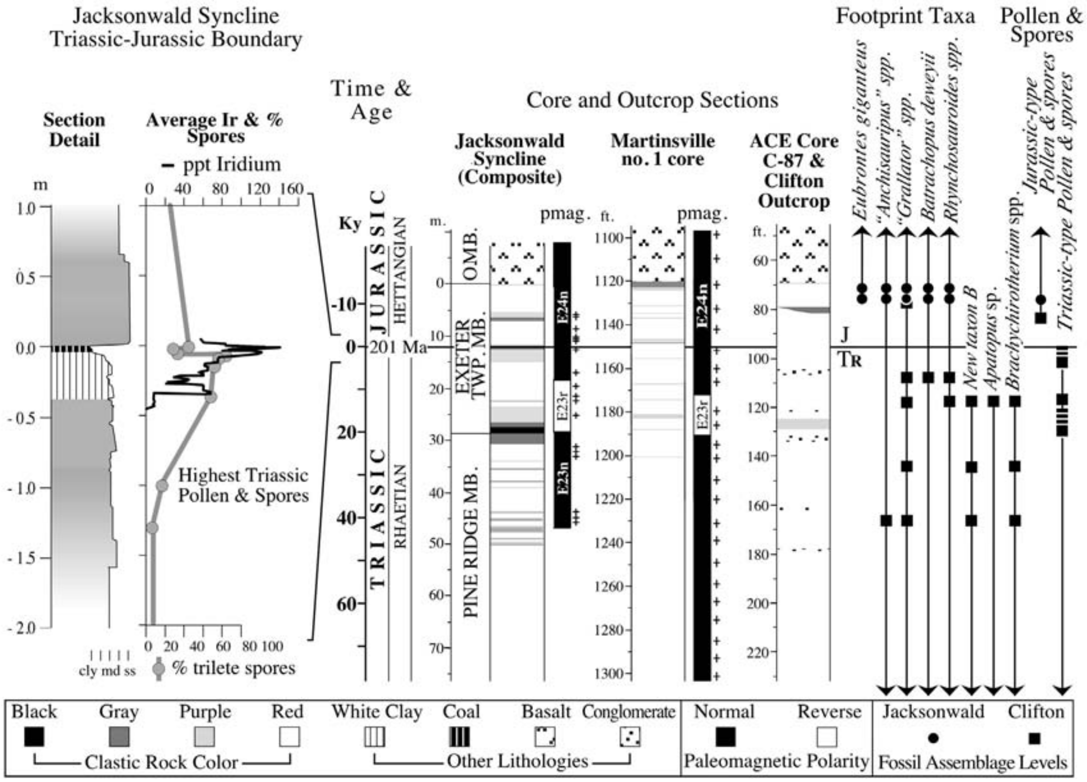

Figure 11. Fine-scale correlation between Ir anomaly, and fern spike, and footprint data from the Newark basin (from Olsen et al, 2002b). Average Ir anomaly is based on 4 localities along strike each of which have an Ir anomaly in virtually identical position (details of data and averaging in supplemental material).

Brachyphyllum shoots and associated reproductive structures . Large fronds of *Clathropteris meniscoides* in growth position have been found in gray ripple cross-laminated siltstone of this member. Conifer shoots and reproductive structures, stems of the horsetail rush *Equisetites*, and rare fragments of *Clathropteris* and the cycadeoid *Otozamites* occur in gray siltstones and claystones of the Towaco and Boonton formations (Cornet, 1977). The stereotyped nature of these assemblages stands in stark contrast to the typical Triassic assemblages from the rest of the Newark Supergroup.

Invertebrates from TS IV include various trace fossils (Metz, 1984, 1991, 1992; Boyer, 1979) with *Scoyenia* being notably rare or absent, darwinulid ostracodes and conchostracans (Pine Forge and Exeter Township members of the Passaic, and Washington Valley Member of the Feltville) (e.g. Nason, 1889; Olsen, 1980a), and an elytron of the beetle *Liasocupes* sp. (Huber et al., 2002). All of these occur in gray thin-bedded, although not microlaminated, silty claystones and limestones.

Articulated and often beautifully-preserved fish are abundant in microlaminated beds of division 2 of Van Houten cycles in the Feltville, Towaco, and Boonton Formations. These include the large coelacanth *Diplurus longicaudatus* (Boonton Formation), the palaeonisciforms *Ptycholepis marshi* (Feltville Formation) and *Ptycholepis* sp. (Boonton Formation), the redfieldiid palaeonisciform *Redfieldius* spp.

(Feltville and Boonton formations), and the holostean *Semionotus* spp. (all formations). In contrast to the Triassic examples, the *Semionotus* spp. comprise species flocks comparable to those seen in cichlid fishes in the Great Lakes of East Africa (McCune, 1987, 1996; McCune et al., 1984; Olsen, 1980a). Some of the fish from TS IV are among the best preserved early Mesozoic fish from anywhere (e.g. Olsen and McCune, 1991).

Tetrapod footprints are more common in TS IV sedimentary units than in any other part of the Newark basin section, with a few localities having produced perhaps tens of thousands of specimens. Three footprint assemblage types occur within TS IV. The oldest is restricted to TS IV strata below the palynologically-identified Triassic-Jurassic boundary (Pine Ridge and lower Exeter Township members of the Passaic Formation), and is indistinguishable from older Triassic assemblages in the basin. *Rhynchosauroides* sp., *Gwyneddichnium*-sp., *Apatopus* sp., *Brachychirotherium parvum*, *Batrachopus* cf. *B. bellus*, and *Batrachopus deweyii* occur, along with a form of probable crocodylomorph affinities referred to as New Taxon B (Silvestri and Szajna, 1993) and abundant specimens of the theropod dinosaurian forms *Grallator* spp. and *Anchisauripus* spp. (Figure 10, 11). The second type of assemblage occurs directly above the Triassic-Jurassic boundary (upper Exeter Township Member, Passaic Formation) and consists entirely of *Rhynchosauroides* n. sp., *Batrachopus deweyii*, *Grallator* spp., *Anchisauripus* spp., and for the first time, the large theropod track *Eubrontes giganteus*. All the forms typical of the Triassic are absent, even though this is one of the most heavily-sampled levels within the Newark basin (the localities occur in strata directly beneath the Orange Mountain Basalt, at several former and presently active quarries in the northeastern part of the basin). The third type of assemblage occurs in the Feltville, Towaco, and Boonton Formations. This assemblage is similar to that from just above the Triassic-Jurassic boundary, but the ornithischian dinosaur ichnite *Anomoepus scambus* is present at most localities, the mammal-like synapsid track *Ameghinichnus* n. sp. occurs at one locality (upper Towaco Formation – see Stop 7), and *Rhynchosauroides* sp. is restricted to a single specimen from the same approximate level as *Ameghinichnus* (Olsen, 1995). The tetrapod footprint assemblages thus follow the turnover pattern seen in the palynofloral assemblages.

In contrast to the exceedingly abundant and well-preserved tetrapod footprint assemblages, osteological remains from the Jurassic of the basin are virtually absent. Thus far there are only a few bone flakes in a coprolite, and a shard of a large tooth, probably of a theropod dinosaur, from the natural cast of a *Eubrontes giganteus* footprint (Olsen, 1995)!

TRIASSIC AND JURASSIC CONTINENTAL COMMUNITIES AND THE TRIASSIC-JURASSIC BOUNDARY

The superb time control and resolution provided by the astronomically-calibrated paleomagnetic polarity timescale makes the Newark Supergroup, particularly the Newark basin, one of the best venues for examining tropical continental floral and faunal change across the Triassic-Jurassic boundary (Kent and Olsen, 1999; Olsen and Kent, 1999). Its one deficit, as cited by Benton (1994), has been a lack of osteological remains of tetrapods, but this is rapidly being remedied (Carter et al., 2001; Olsen et al., 2000b, 2001b; Sues et al, 2000). Based on the Newark timescale and paleontological correlations with areas outside the central Pangean rift zone, a consistent picture emerges of the profound changes that occurred around the boundary, with some indications of what the causal mechanism for that change may have been.

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

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

The lack of time control at the appropriate level of resolution outside of the central Pangean rifting zone, in addition to significant sampling gaps, particularly in the Norian and Rhaetian, precludes detailed knowledge of how these provinces changed, although some trends are evident. It is clear that to some extent the faunas and floras tracked climate as central and southern Pangea drifted north. It is also apparent that in most areas dinosaurs became more abundant, diverse and larger through the Triassic, with the moderate- to large-sized herbivorous prosauropod dinosaurs becoming common in the later Triassic (Norian and Rhaetian) at the boundaries between the tropical and boreal regions, and perhaps at higher latitudes, but remained virtually excluded from the lower latitudes. The provinciality and within-habitat diversity led to a very high-diversity global terrestrial biota, which is only now being appreciated (Anderson et al., 1986).

The Early Jurassic global biota was much more stereotyped. Most floral provinciality was gone, with the *Dicroidium-Thinfieldia* complex being completely eliminated. Conifers, especially the now-extinct Cheirolepidiaceae (*Corollina*- producers) were extraordinarily dominant in the tropics, a pattern that would continue until the mid-Cretaceous (Watson, 1988). A northern boreal province persisted, with infrequent cheirolepidiaceous conifers, but it was dominated by different groups (e.g.

Thaumatopteris) (Harris, 1931). The boreal southern areas had much less abundant cheirolepidiaceous conifers.

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

The rate at which this change occurred can presently be assessed only in the Newark Supergroup, and most of the evidence comes from the Newark basin. In the central Atlantic margin rifts, the floral change was evidently very abrupt, estimated in the Newark, Fundy, and Argana basins to have occurred over less than 20 ky, and probably actually much less (Fowell, 1993; Fowell and Olsen, 1993; Fowell et al., 1994; Fowell and Traverse, 1995; Olsen et al., 2000a, 2002b,c), as it occurs within a single Van Houten cycle. A very similar rate of change evidently affected tetrapods, based mainly on Newark basin tetrapod footprint assemblages (as described above), although augmented with data from other Newark Supergroup basins (Figures 10, 11). This change is consistent with the much less intensely-sampled skeletal data.

In the Newark basin, the floral and faunal changes are directly associated with the fern spike and a newly-discovered iridium anomaly (Olsen et al., 2001b,c). The floral and faunal pattern (with the exception of the survival of the non-avian ornithodirans) and the associated iridium anomaly is strikingly similar to the pattern seen at the K-T boundary in the North American Western Interior (e.g. Tschudy et al., 1984), which suggests a similar cause for both extinctions – a giant asteroid impact – a suggestion which had repeatedly been made long before the new biotic and Ir data were available (Badjukov, et al., 1987; Bice et al., 1992; Dietz, 1986; Olsen et al., 1987, 1990; Rampino and Caldeira, 1993).

However, one the most striking aspects of the Triassic-Jurassic boundary in the central Atlantic margin rifts is the direct superposition of the oldest CAMP basalts on the boundary, but always with an intervening small thickness of Jurassic strata. A possible causal link is difficult to ignore, given a similar (although less precisely timed) coincidence between the Deccan Traps and the K-T boundary and the Siberian Traps and the Permo-Triassic boundary (Rampino and Caldeira, 1993). The three largest Phanerozoic mass-extinctions are evidently very close in time to the three largest Phanerozoic flood basalt provinces, and all three have at least some evidence of asteroid impacts. A mechanism linking impacts with flood basalts has been proposed (e.g. Boslough et al., 1996), but the energetics are difficult to reconcile with the observations and the models (Melosh, 2000). Nonetheless, it seems plausible that a massive impact might be able to concentrate the effusive rate of a distant flood basalt province that was close to eruption; however, this topic has yet to be explored quantitatively.

At this point we can paint a speculative picture of what the Triassic-Jurassic transition may have been like, given present data (see summary; Figure 12). Biotic diversity was rising through the Late Triassic, but this increase was terminated by the impact of one or more asteroids or comets (e.g. Spray et al., 1998). As with the K-T scenario, continental biotas were initially affected by reduced sunlight and the attendant cold for a period of months, followed, in the case of the Triassic-Jurassic boundary at least, by a significant time of elevated CO_2 (McElwain, et al., 1999), leading to a rise in global temperatures. The massive ecological disruption was initially caused by the physical effects of the impact and then the cold, but these were then exacerbated by the rise in temperature, in turn probably associated with a great increase in the incidence of lightning-induced fire. This situation could only have been made worse by the succeeding CAMP flood basalt episode. The long term disruption allowed only rapidly-growing spore-dispersed plants - largely ferns - to populate the tropical regions over the next hundreds to thousands of years. The surviving dinosaurs may have all been small forms, but within 10 ky, theropod dinosaurs became considerable larger than any that had existed during the Triassic. The massive and sustained ecological disruption led to the extinction of many tetrapod families that were presumably dinosaurian competitors, and only afterward did the familiar dinosaur-dominated communities arise that would last for the next 135 million years.

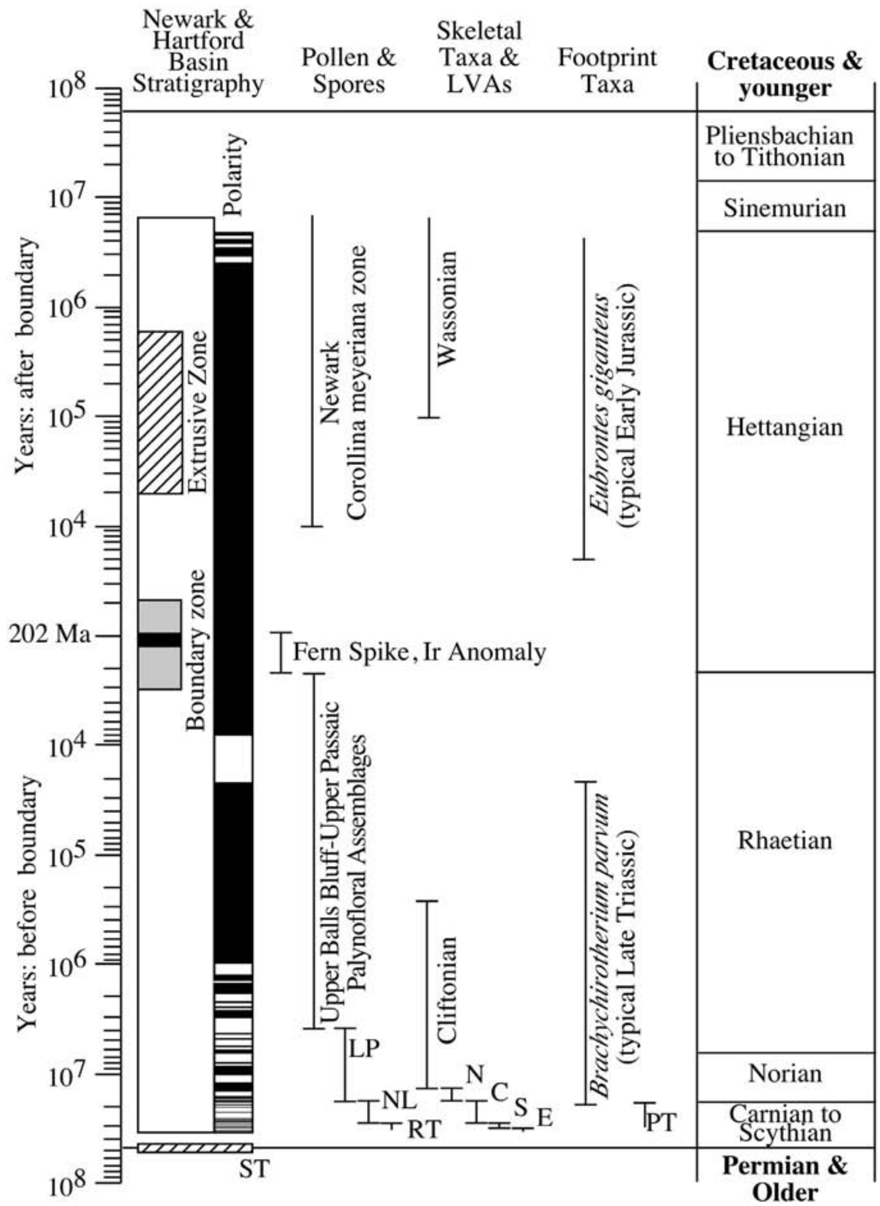

Figure 12. Summary of major physical and biotic events around the Triassic-Jurassic boundary plotted on a logarithmic scale (from Olsen et al., 2002c). LVAs are Land Vertebrate Ages of Huber and Lucas, 1997 and Lucas and Huber (2002): N, Neshanician; C, Conewagian; S, Sanfordian; and E, Economian. Pollen and spore zones are from Cornet (1977) and Cornet and Olsen (1985): LP, Lower Passaic Heidlersburg; NL, New Oxford, Lockatong; RT, Richmond Taylorsville. Footprint distribution; PT, range of the Pekin-type footprint assemblages. Note that the extrusive zone consists of lava flow formations interbedded with fossiliferous and cyclical sedimentary strata, with the latter interpreted as representing nearly all of the time shown. ST indicates the position of the Permo-Triassic Siberian Traps.

FIELD TRIP AND ROAD LOG

The field trip road log is focused around the present and ancestral route of the Hudson River (Johnson, 1931; Lovegreen, 1974; Stanford, 1993) (Figure 13) in the northeastern Newark rift basin, and present lower Hudson Valley. The road log begins and ends at Lamont Doherty Earth Observatory (LDEO), Palisades, New York, and together the 8 stops span most of the section in the northern Newark basin (Figure 14).

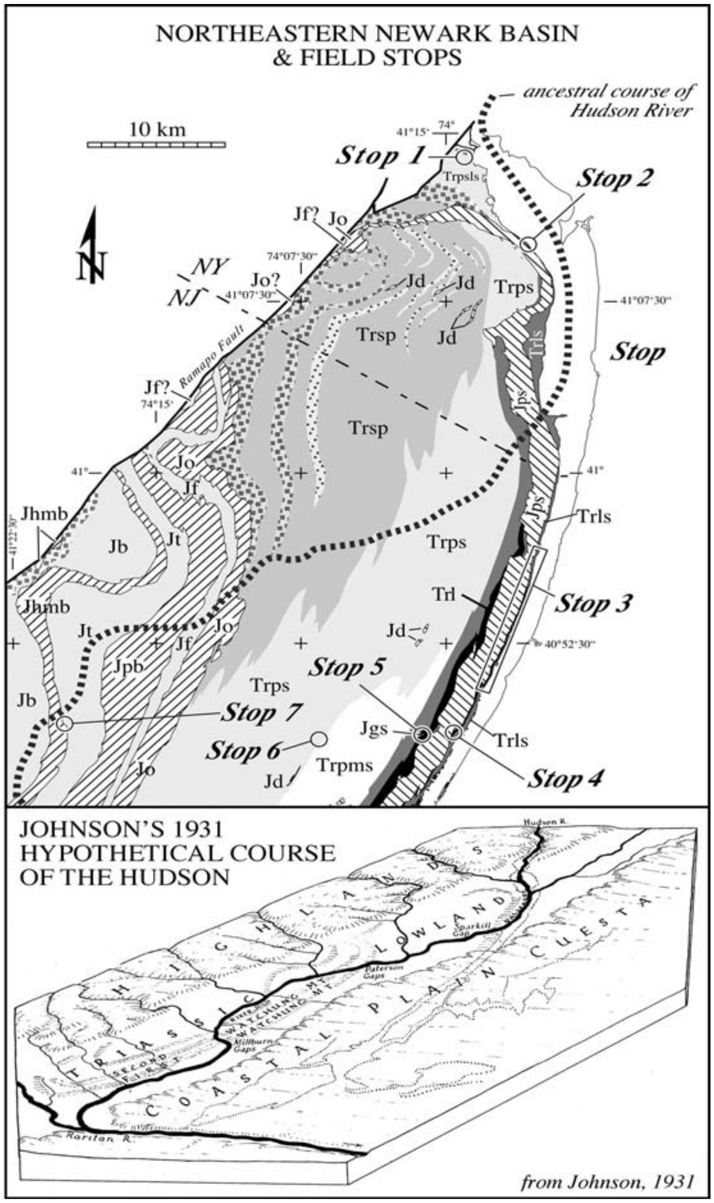

Figure 13. **A**, Northern Newark basin and ancestral course of the Hudson River (based on Johnson, 1931) with field trip route stops (see Figure 7 for key of lithostratigraphic units); **B**, late Cenozoic, pre-Holocene hypothetical course of the Hudson with assumed physiography (from Johnson, 1931).

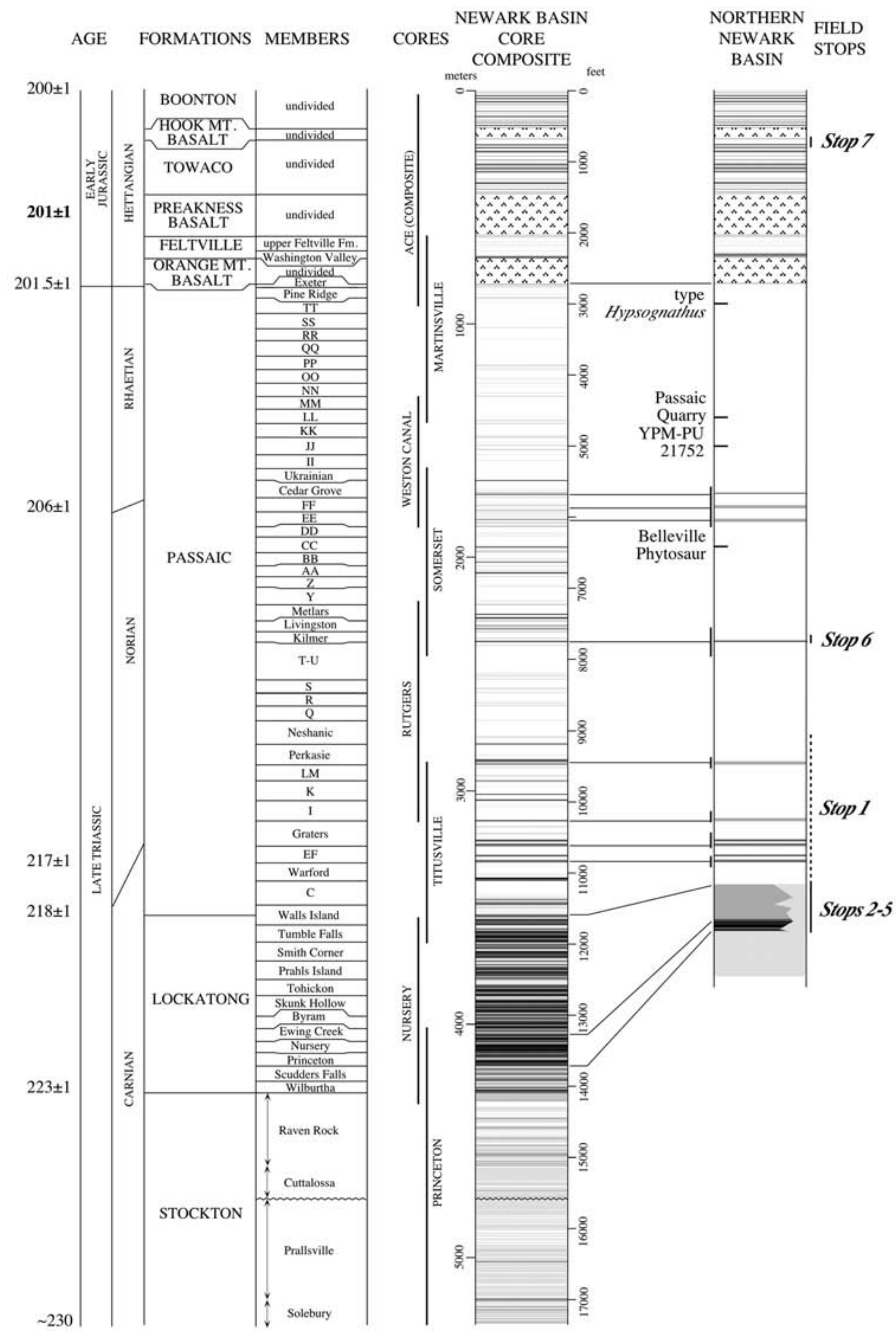

Figure 14. Newark basin section from the NBCP cores of the central Newark basin (left) compared to the stratigraphy of the northern Newark basin with field trips stops. See Figure 4 for key to lithologies.

0.0 mi. Leave entrance to LDEO, turn right and proceed north on Rt. 9W along the strike of the Palisade sill.

0.9 mi. Turn left (west) on Oak Tree Road and drive down the dip slope of the Palisade sill.

1.8 mi. Outcrops of diabase and hornfels of Lockatong-like facies.

2.3 mi. Crossing buried valley of ancestral (Miocene-Pleistocene) Hudson River.

2.6 mi. Turn right (north) onto Route 303.

7.3 mi. On right is the Blauvelt footprint site discovered in 1972 by PEO and Robert F. Salvia. Site consists of the excavated dip slope of gray, tan, pink, purple, and red mudstones and sandstones of the Stockton Formation equivalent of the Lockatong Formation.

Tracks found include a slab of tridactyl footprints often attributed (incorrectly we believe) to the ceratosaurian theropod dinosaur *Coelophysis* (e.g. Fisher, 1981), and isolated tridactyl tracks (Figures 15, 16). These tracks are rather broad and tulip-shaped suggestive of large *Atreipus* (see mileage 34.8 and Stop 6), the early ornithischian dinosaur ichnite. However definitive manus impressions are lacking, and they tend to be broader than most *Atreipus*. They could belong to a new form, but because of their overall poor preservation, they remain indeterminate.

Other footprint forms found are largely indeterminate but include indeterminate forms that are probably poorly preserved examples of small *Brachychirotherium* as well as an example of *Rhynchosauroides* cf. *R. hyperbates* (Figure 15). Indeterminate phytosaur teeth and bone scraps occur in the footprint-bearing beds.

On the south side of the exposures are greenish and tan mudstones with abundant and often large roots and tree trunk casts suggestive a strongly vegetated surface (Figure 15).

7.8 mi. Outcrop on right is Palisade sill that has moved considerably higher in the section to the Stockton-Passaic formational boundary.

8.8 mi. Pass Lone Star quarry on left in a large apophysis of the Palisade sill.

9.1 mi. Palisade sill moves lower in section in this area.

13.7 mi. Merge with Route 9 W on dip slope of Palisade sill.

14.2 mi. Long Clove Quarry in Palisade sill on left.

14.4 mi. Pass through small fault gap to the front of the questa of the Palisade sill.

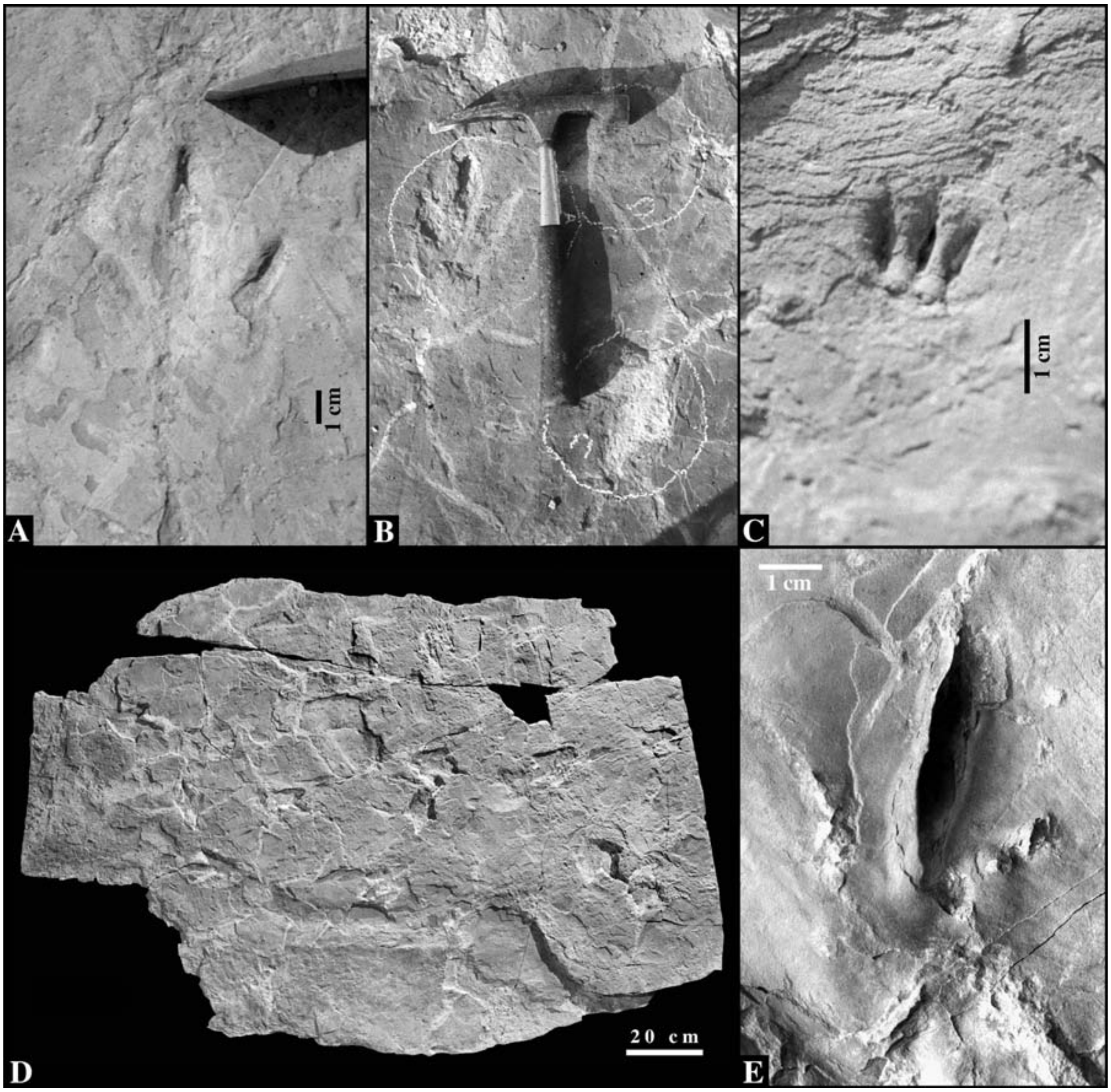

Figure 15. Footprints from the Blauvelt and Upper Nyack track localities: A, indeterminate isolated pes impression in situ (uncollected), from Blauvelt, New York; B, indeterminate tracks as found of main slab now at the New York State Museum, Albany (from Blauvelt, New York); C, leptodactylus quadrupedal track, probably *Brachychirotherium*, from Blauvelt, New York; **D**, indeterminate dinosaur tracks (possibly *Atreipus*) on main slab after collection, now at the New York State Museum, Albany (Blauvelt, New York); leptodactylus *Grallator* cf. *G. parallelus* footprint, Upper Nyack, New York.

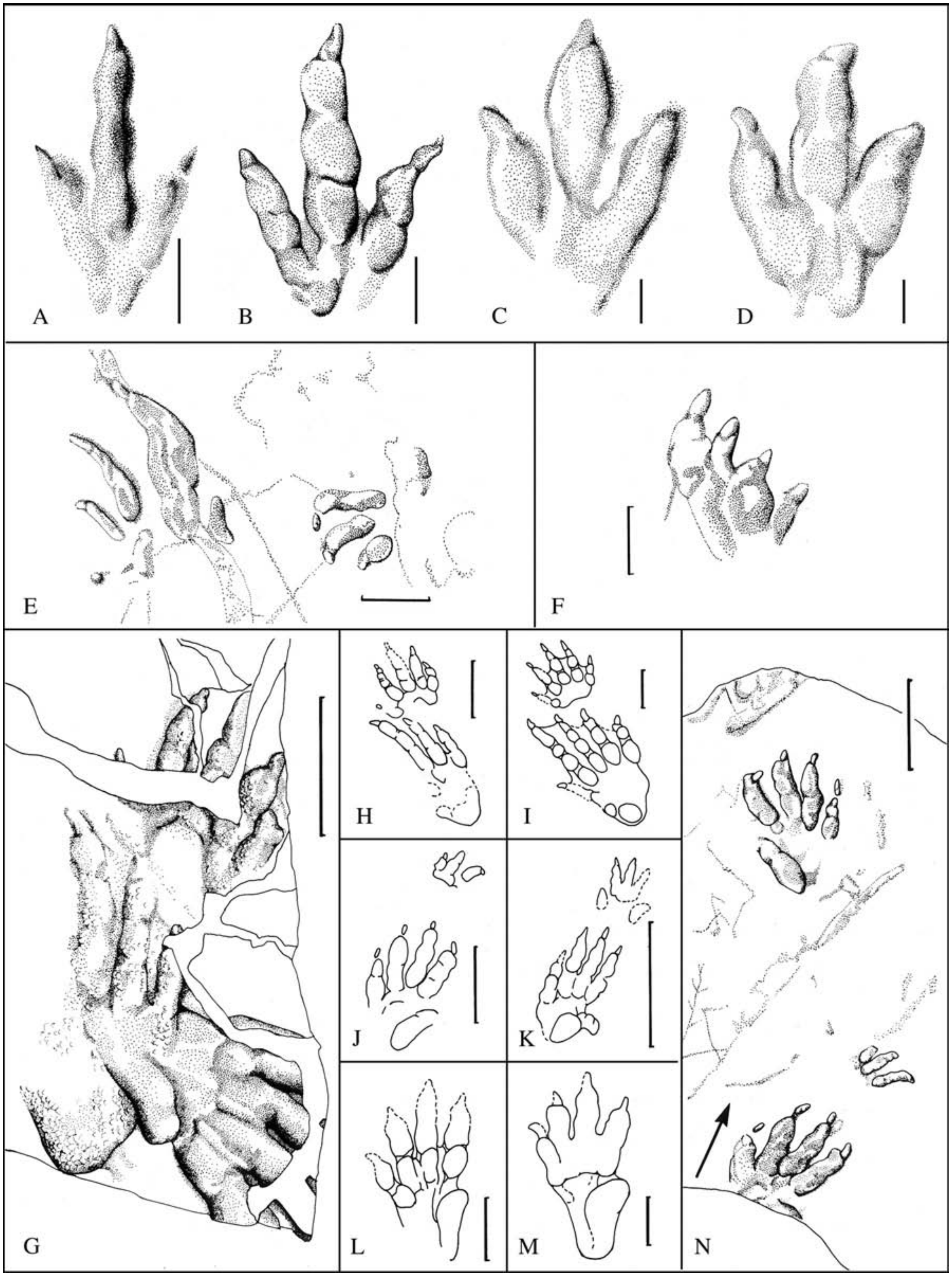
A
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Figure 16. Drawings of footprints from the Stockton Formation of New York: **A**, *Grallator* cf. *G. parallelus* from Upper Nyack (in collection of D. Fisher - see Figure 15D); **B**, Grallator sp., from Stop 2, Haverstraw, New York (specimen lost); **C-D**, indeterminate dinosauria track, possibly Atreipus on main slab after collection, now at the New York State Museum, Albany (from Blauvelt, New York); **E**, *Rhynchosauroides* cf. *R. hyperbates*, from Blauvelt, New York (YPM 7733); **F**, *Rhynchosauroides* cf. *R. hyperbates*, from Stop 2, Haverstraw, New York (YPM 8265); **G**, *Apatopus lineatus* and *Brachychirotherium* cf. *B. eyermani* from from Stop 2, Haverstraw, New York (YPM 7731); **H**, outline drawing of Apatopus lineatus shown in **G**; **I**, outline drawing of *Apatopus lineatus* from Baird (1957) see Figure 25; **J**, outline drawing of composite of manus-pes sets of *Brachychirotherium* cf. *B. eyermani* from N; **K**, *Chriotherium lulli* from Baird (1954); **L**, outline drawing of *Brachychirotherium* cf. *B. eyermani* based on **G**; **M**, *Brachychirotherium eyermani* from Baird (1957); **N**, *Brachychirotherium* cf. *B. eyermani* from Stop 2, Haverstraw, New York (YPM 8262), see Figure 25. Scale is 2 cm in A-D, 1 cm in E-F, 5 cm in G-N.

14.5 mi. Outcrops on left are of red sandstones of Stockton facies. As we travel north the contact with sill drops discordantly lower in the sedimentary section.

15.0 mi. Cliff of Palisade sill. On the back of the escarpement is the Long Clove quarry in the sill. Crushed stone passes though a tunnel in the sill to the offloading docks on the right.

17.0 mi. Strike ridge of Palisade sill curves strongly to the west.

18.9 mi. Crossing Cedar Pond Brook.

19.0 mi. Turn right on Lowland Hill Road.

19.2 mi. Turn right onto into parking lot for Lowland Town Park and park.

Stop 1: Gorge of Cedar Pond Brook, Stony Point, New York.

Latitude and Longitude: 41°13.605', 073°59.079 ' to 41°13.677', 073°59.378
Tectonostratigraphic Sequence: TS III
Stratigraphic Unit: ?lower Passaic Formation
Age: Late Carnian (early Late Triassic); ~220-223 Ma
Main Points:

1. NE terminus of Newark basin near onlap onto basement of TS III
2. Sediment-starved section with well-developed carbonate beds
3. Carbonates of ?lacustrine and pedogenic origin - local source
4. Should be ideal area for bone or egg preservation

5. Relevant to arguments about the Broad Terrane Hypothesis

Cedar Pond Brook flows from west to east cutting a gorge though a unique carbonate-rich facies of the Newark basin section at its northern terminus (Figures 17, 18). The western, coarse-grained part of the section was mentioned by Ratcliffe (1980) and described in outline (with thin sections) by Sanders (1974), but is otherwise virtually unstudied.

Conglomerates making up the basal beds of the local section outcrop to the east in a railroad cut within the Stony Point Battlefield Park (Figure 17). Basement here consists of metamorphic rocks bordering the Cortland Complex, which is locally reddened near the contact with the conglomerate. The contact itself is not exposed. According to Ratcliffe (1980), the exposed conglomerates contain clasts of Manhattan Schist and Ordovician Inwood Marble dolostone clasts, both clast types being derived very locally from the east and north. The section fines to the west, passing up-section into a cyclical sequence of red calcareous mudstones and sandstones and pink to gray limestones (Figures 19, 20).

Cycles have a mode about 1.5 m with a large amount of variation consist of a lower red silty mudstone with abundant *Scoyenia* burrows and root traces, passing upward into mudstone with carbonate nodules; these nodules often coalesce upward into a bed of pink to white massive limestone (Figure 20). The limestone contains thin (~2 mm) branching tubes that are most likely root traces, but sometimes also (apparently) ostracodes, identified in thin section by Sanders (1974). These biologic remains indicate that at least some of the limestones are of lacustrine origin, perhaps modified by pedogenic structures. Many of the mudstones and limestone beds have large (~1 m long, ~3 cm wide) root traces and most mudstones have densely branching thin root traces.

Strikingly absent from these outcrops are larger-scale fluvial units such as channels. However, also absent are any indications of Van Houten cycles, unless they are the limestone cycles themselves. This facies resembles no others in this part of the basin, although it does bear some similarity to the middle upper Passaic of the Jacksonwald syncline, where carbonate beds are well-developed (although not as well-developed as here). This suggests that this section represents a relatively dry phase of a McLaughlin cycle, probably in the lower Passaic Formation, in an area of the basin isolated from entry points of major rivers or streams.

Ratcliffe (1980) described the "Annsville Phyllite", present to the west on the hanging wall of the Ramapo fault system (here the Thiells splay). Sedimentary breccias of (?)Inwood Marble carbonate occur locally near the fault low in the section on the north side of the Cedar Pond Brook gorge. This indicates that there was indeed a local source of abundant carbonate, which probably accounts for the extreme development of limestone beds in this facies.

Recently, it has become clear that tetrapod remains are present and locally abundant, being well preserved at many levels within the Newark Supergroup in red massive sandy to silty mudstones which contain root traces and have variable development of carbonate nodules and caliches (Carter et al., 2001; Olsen et al., 2000b, 2001b; Sues et al., 2000). The abundance of carbonate helps to buffer bone dissolution in a pedogenic environment, and can allow otherwise very perishable vertebrate carbonate fossils - eggs - to be preserved. The association of pedogenic carbonate nodules (caliche or Kunkar) and tetrapod bone has been repeatedly noted elsewhere (e.g. Dodson et al.,

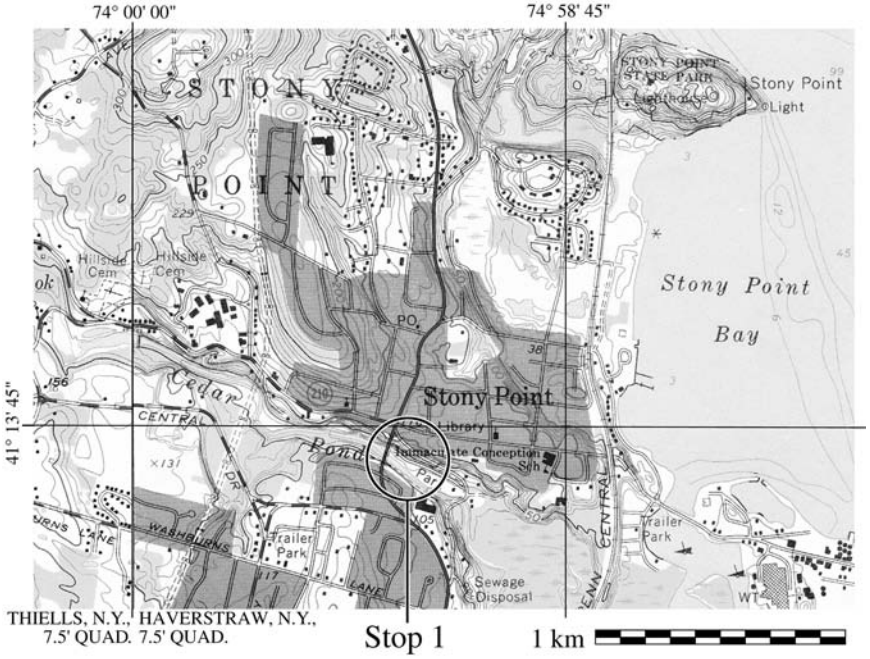

Figure 17. Location map Stop 1.

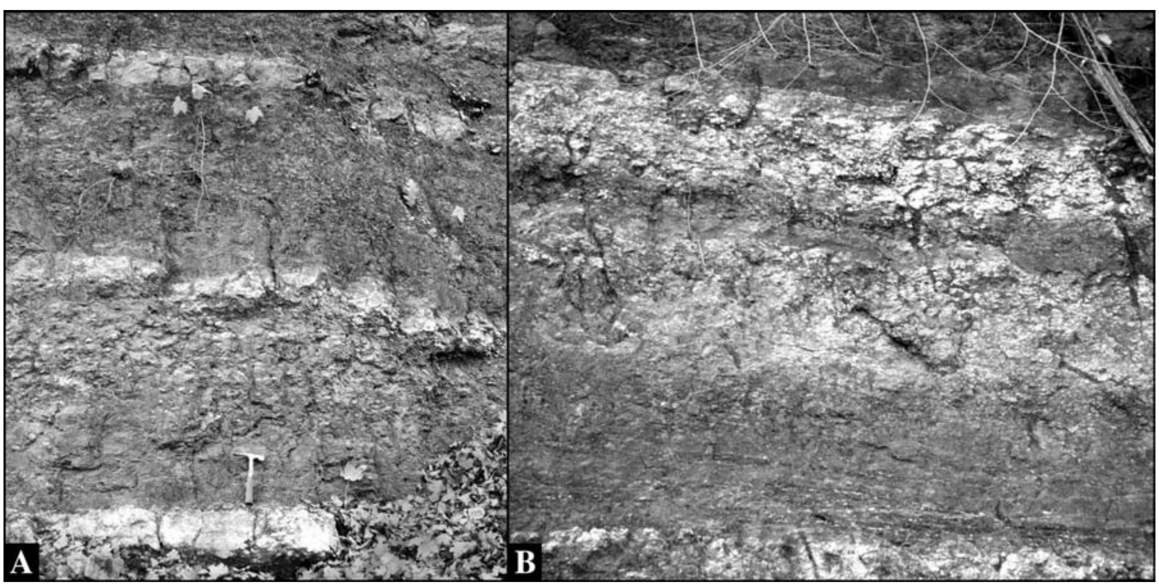

Figure 18. Photos of representative portions of the section at Stop 1: A, two complete successive carbonate-bearing cycles (from 17-20 m in Figure 19); B, two successive carbonate-bearing cycles of different thickness (25-29 m in Figure 19), note meter-scale root-traces (one very obvious at upper right).

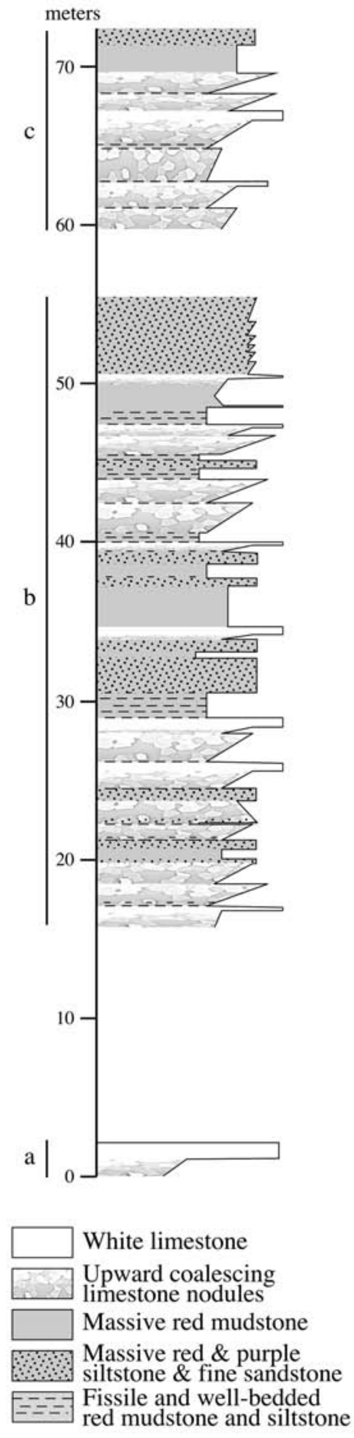

Figure 19. Stratigraphic section at Cedar Pond Brook: **a**, gully, north side of Longhill Town Park, latitude 41°13.692' N longitude 073°59.189' W; **b**, section on south bank of Cedar Pond Brook, south side of park and beneath bridge for Rt. 9W, from latitude 41°13.605' N longitude 073°59.079' W to latitude 41°13.674' N longitude 073°59.242' W; **c**, section west of bridge on slope to south of Cedar Pond Brook latitude 41°13.677' N longitude 073°59.378' W .

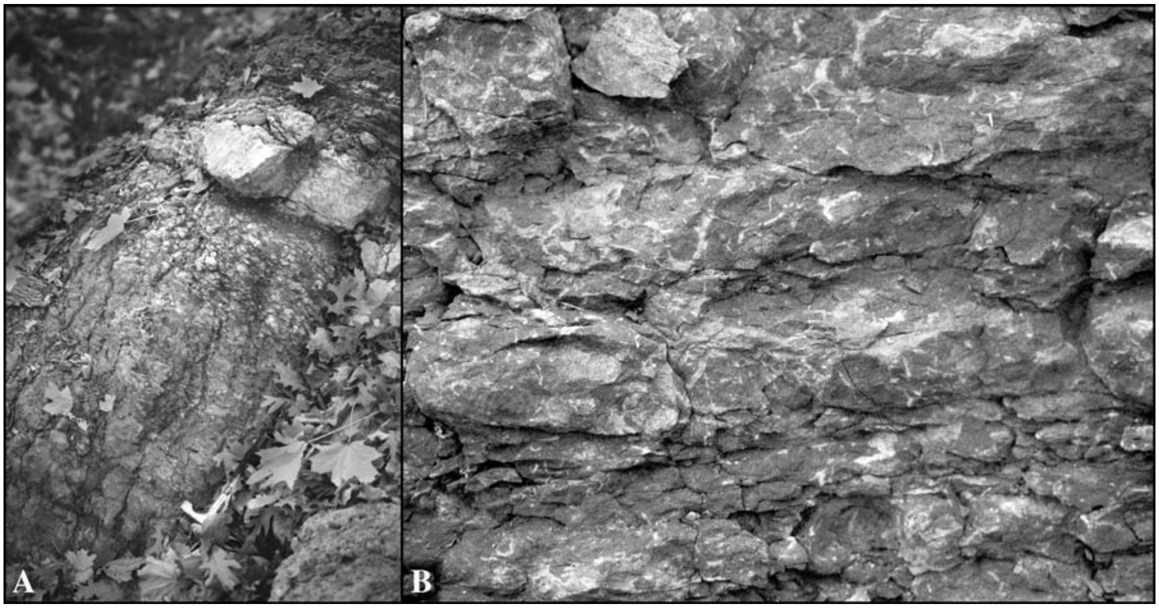

Figure 20. Carbonate cycle details: **A**, upward coalesing carbonate nodules in single cycle at 44 m in Figure 19; **B**, prevalent well-developed heavily rooted massive mudstone fabric (at 20 m in Figure19, image spans about 30 cm vertically).

1983). These outcrops should be ideal for bone preservation, but there has been little prospecting as of yet. Persistence is needed, however, because the tetrapod fossils tend to be nearly the same color as the carbonates, making it very difficult to locate.

This northeastern part of the Newark basin was crucial to Sanders (1960, 1963, 1974) arguments for the "Broad Terrane Hypothesis". This hypothesis was first proposed by I.C. Russell (1879, 1880), who argued that the Newark and Hartford half-graben basins were actually the western and eastern components of a giant full graben which had been post-depositionally uplifted in the middle and then deeply eroded (Figure 21). Sanders considered the northeastern terminus of the Newark basin to be a key area because it showed that this end of the basin was an artifact of post-deposition folding, indicated by the basement-parallel strike of the beds wrapping around to the west, towards the border fault. Sanders argued that there is no indication of onlap of progressively younger strata onto basement to the northeast; therefore the basin did not shallow in that direction, and the high between the basins did not exist at the time of deposition.

We argue that these outcrops favor exactly the opposite interpretation. There is no evidence at all that the sequence seen at Stops 1-4 is the basinward equivalent of the basal section (i.e. Stockton Formation) of the Newark basin. The facies is utterly unlike the local expression of the Stockton, and all the evidence (as seen in stops 2-4) points to a shoaling of the basin towards the north and east. In addition, the map pattern (7, 13) indicates rapid thinning of the Passaic Formation toward this area and dramatic changes in strike, both of which are consistent with the existence of a <u>syndepositional</u> basin high in the region between the Newark and Hartford basins.

However, this does not mean that there might not have been a sedimentary connection between the two basins. The Newark and Hartford basins are deeply eroded, and if TS III in eastern North America follows the pattern seen in the much less deeply eroded Moroccan basins, we should expect there to have been onlap of the younger parts of TS III over basement, eventually connecting the two basins. Strata of TS IV, particularly the basalt flows and thinned interflow units, may also have connected the basins (c.f. McHone, 1996). It is also quite likely that considerable amplification of the high between the two basins occurred postdepositionally as a consequence of late Early Jurassic to Middle Jurassic tectonic inversion. However, there is no evidence supporting the argument that the deepest part of the combined Newark-Hartford basin lay in the now-uplifted region between them, as proposed in Russell's and Sanders' versions of the Broad Terrane Hypothesis.

Go back to entrance of Lowland Town Park.

19.3 mi. Turn right back onto Lowland Hill Road.

19.5 mi. Turn left onto Rt 9W and head south.

23.2 mi. Turn left (east) onto Short Clove Road

23.4 mi. Turn right (south) onto Riverside Avenue.

24.2 mi. Park at south termination of Riverside Avenue in parking lot for Hook Mountain State Park of the Palisades Interstate Park.

Walk south along path about 1060 m looking for ruins on left.

Turn left at ruins and (carefully!) proceed down the rocky and vegetated slope to Stop 2.

Stop 2. West shore of Hudson River, Palisades Interstate Park, Haverstraw, New York.

Latitude and Longitude: 41°10.383'N. 073°56.097'W.
Tectonostratigraphic Sequence: TS III
Stratigraphic Unit: Stockton Formation
Age: Late Carnian (early Late Triassic); ?220 Ma
Main Points:

1. Stockton-like equivalent of Lockatong Formation at basin edge
2. Fluvial and ?lake margin facies of *wet* part of McLaughlin cycle
3. Multiple fossil-rich facies
4. Sampling of more terrestrial communities than usual for this time
5. Well-preserved footprints in unusual settings
6. Relatively abundant bones in tan quartz-pebble conglomerate

Abundant outcrops and exposures below the Palisade sill extend from Haverstraw, New York south to Hoboken, New Jersey allowing a close look at the facies changes along-strike progressing from mostly-fluvial environments in the northeast to the mostly lacustrine cyclical facies in the southwest. Outcrops along the west shore of the Hudson River in this segment of the Palisades Interstate Park consist mostly of various fluvial facies with some hints of lacustrine influence. This site consists of exposures in a small abandoned quarry and river-side outcrops of a ~17 m thick section of very heterogeneous clastic rocks, including tan quartz-pebble conglomerates, pink and red sandstones, and purple and red mudstones and siltstones (Figures 22, 23, 24).

Evidence for this section being the along-strike equivalent of the lower part of the Lockatong (probably the Nursery or Ewing Creek member) consists primarily of the apparent continuity of strike direction from unquestionable Lockatong outcrops to the south of this region. If there is some hanging wall onlap of younger strata onto basement in this direction - and there is a hint of this - these outcrops would still likely be part of the Lockatong rather than lower in the section. In addition, the facies change northwards along-strike from the Lockatong is slow, consistent, and significantly different from the lateral change within the Passaic Formation in the same region that trends towards exclusively red conglomerates.

The abundance of tan, pink, and purple strata at this section (Figure 24) suggests prolonged water saturation of the sediments prior to lithification, and, together with the relatively muted pedogenic influences, suggests deposition during a relatively wet episode. The thickness scale of the section showing these features suggests perhaps a wet phase of a 404 ky cycle as is also expressed in the Lockatong Formation to the south (e.g. Stops 4 – 6). However, it is important to note that we have no direct knowledge of the scale of depositional relief (e.g. of marginal deltas), nor is there a clear indication of a hierarchy of cycles, so this interpretation has to be regarded as tentative.

What makes this site particularly interesting is the relative abundance of tetrapod fossils in virtually all lithologies. Most of the fossils are found right at the Hudson shoreline, where rubble piles from the old quarry are eroding into the river, providing clean and fresh rock faces for examination. It is well worth pulling material directly from the eroding rubble piles, as well as turning over slabs in the river itself.

The lithology most consistently producing bones is a tan quartz-pebble conglomerate. Bones and teeth are white in this matrix and comparatively easy to see. Blocks of this material should be broken up to hand-size pieces to reveal fresh faces before being discarded. Thus far, a small partial (?)skull (as yet unprepared; Figure 25), an amphibian dermal bone, numerous (?)phytosaur teeth, and many unidentified bone and teeth fragments have been found. Red and purple siltstones and sandstones have also yielded white bone and tooth scraps, so these lithologies should also be carefully examined.

Reptile footprints are surprising common at this site and occur in a variety of lithologies, showing nearly the full range of footprint preservational styles (Figure 26). Perhaps the most unusual style of preservation is in a red to dark reddish-purple medium-grained sandstone, in which the tracks are preserved as sometimes high-relief natural casts on the underside of beds, frequently preserving skin impressions and other exceptional details. The tracks are actually impressions in a thin massive silty claystone which crumbles when the overlying sandstone is separated. Tracks are also found as

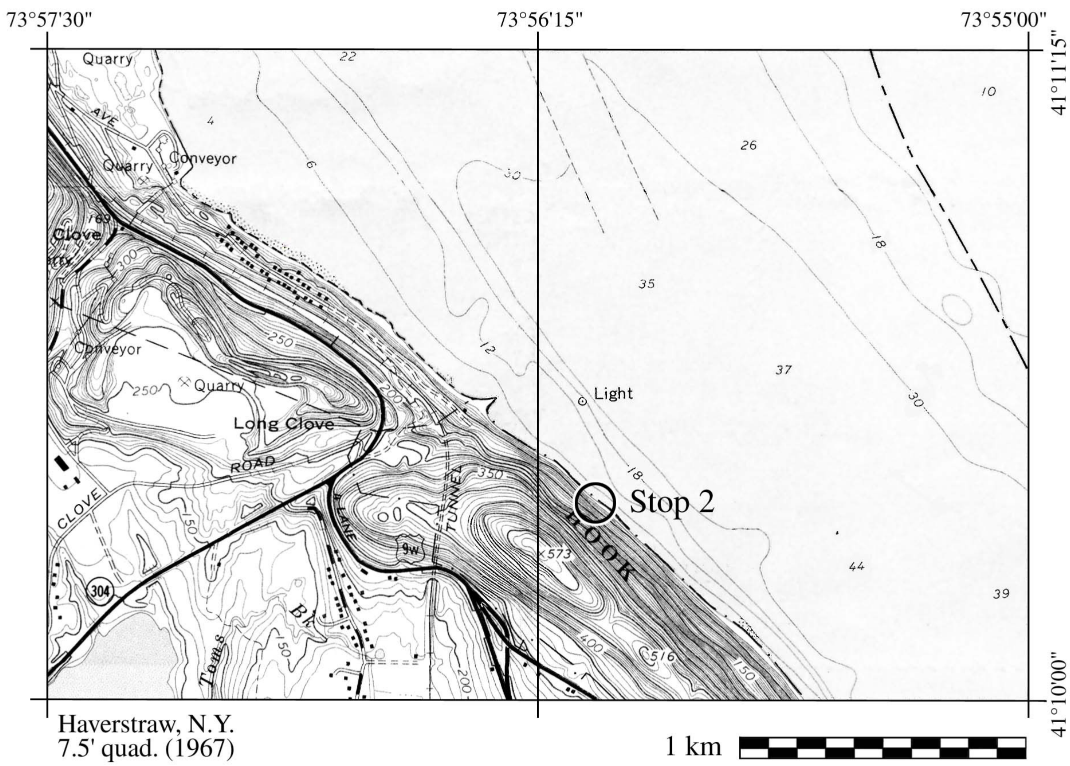

Figure 22. Stop 2 site map.

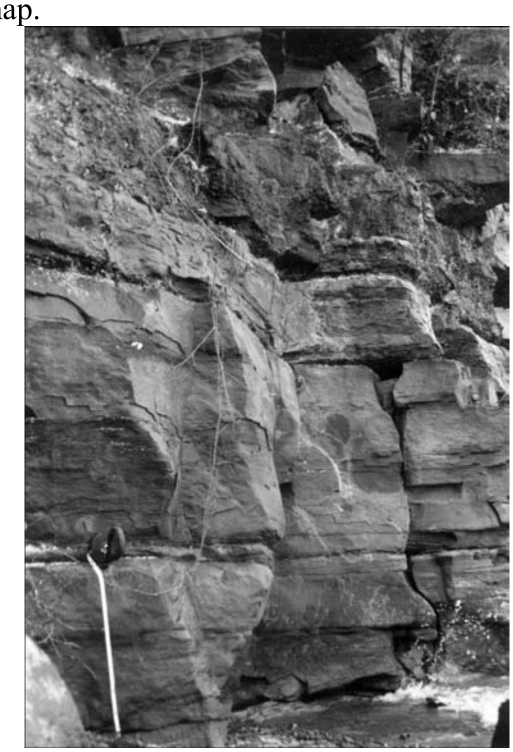

Figure 23. Photograph of outcrops looking north at waters edge at Stop 2 (latitude 41°10.394' N longitude 073°56.088 W). Length of exposed tape measure is 1.5 m.

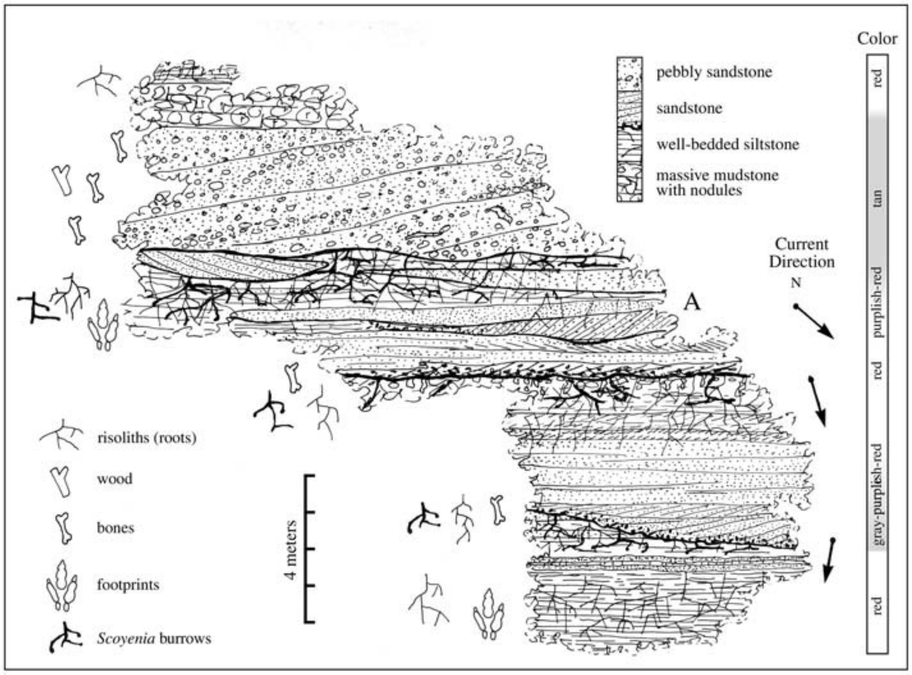

Figure 24. Measured section at Stop 2, Haverstraw, New York from the old quarry to the river (from latitude 41°10.379' N longitude 073°56.097' W to latitude 41°10.396' N longitude 073°56.088 W). Vertical exageration about 3 x.

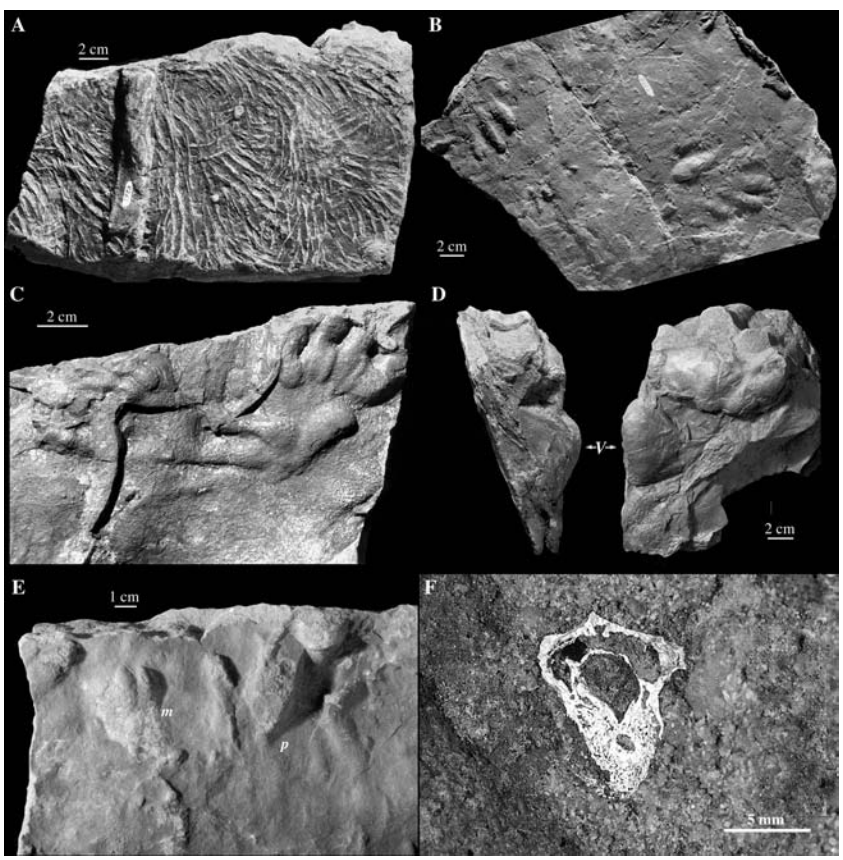

Figure 25. Fossils from Stop 2, Haverstraw, New York: **A**, natural cast of scratch marks probably made by the animal that produced *Rhynchosauroides hyperbates* (YPM 8263); **B**, natural cast of partial trackway of *Brachychirotherium* cf. *B*. *eyermani* (YPM 8262); **C**, natual cast of manus-pes 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 indicate digit V (five); **E**, natural cast of poor manus (m) and pes (p) set of ?*Atreipus* sp. (YPM 8553); F, partial tetrapod skull (AMNH uncataloged).

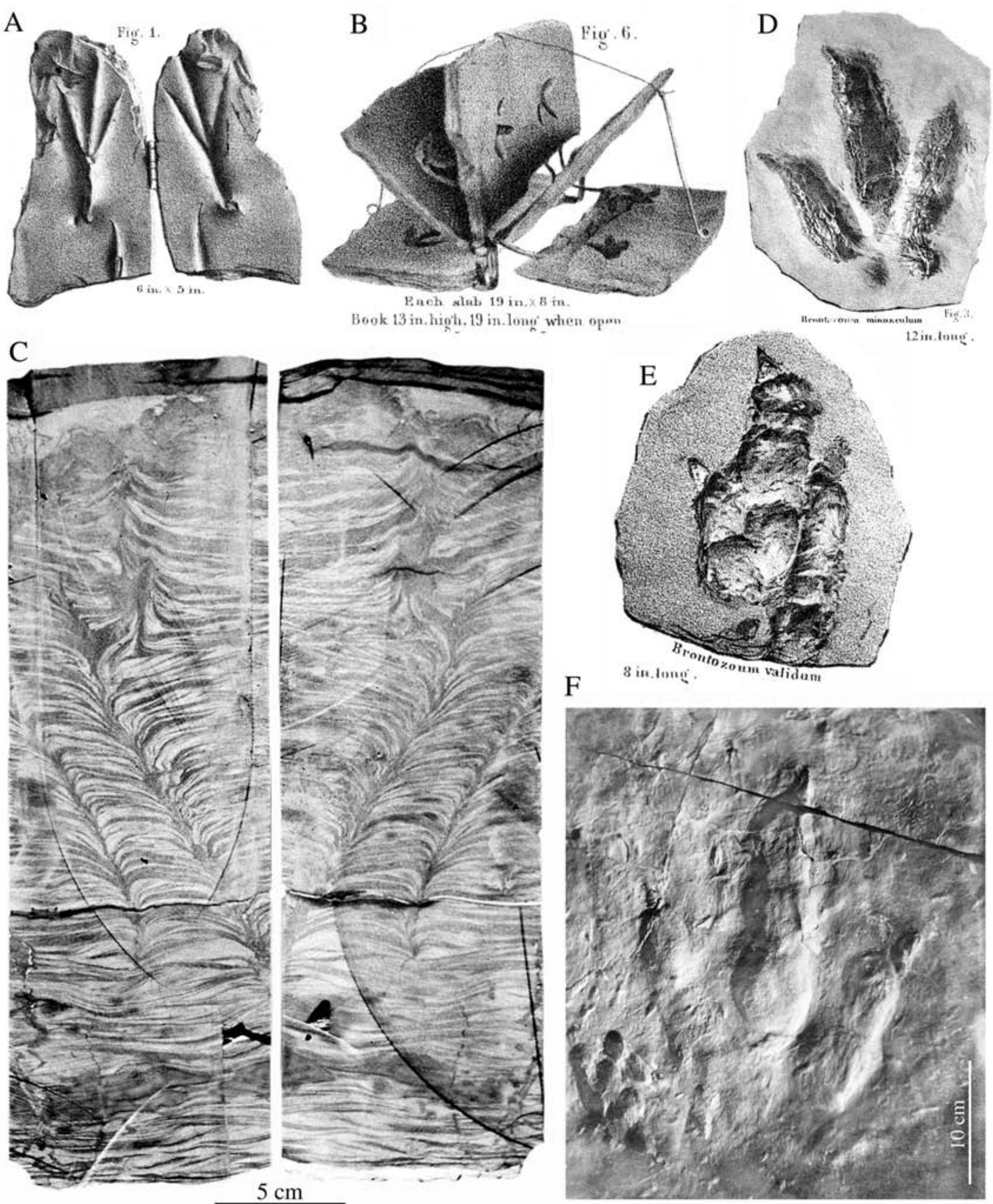

Figure 26. Footprint preservational styles: **A**, leptodactylus form labeled *Triaenopus leptodactylus* (AC 27/9 from Hitchcock, 1956, Plate LII, fig. 2); **B**, another classic leptodactylus form making up a "stony volume" in which the track is impressed though many layers given the name of *Platypterna varica* (AC 27/4 from Hitchcock, 1956, Plate LII, fig. 6); **C**, cross section of leptodactylus track from a core (Park River core FD 12T, 44 ft) from the Portland Formation, Hartford, Connecticut (downwardly merging laminae were produced as the foot was implanted and upwardly merging laminae were produced as the foot was withdrawn); **D**, natural cast of an underprint type fooptrint in which the actual surface in contact with the foot was a centimeter or so above the layer that spit open, labelled *Brontozoum minusculum* (from Hitchcock, 1956, Plate LVII, fig. 2); **E**, natural cast of a normal track in which the surface that visible is the surface that was in contact with the foot, labelled *Brontozoum validum* (from Hitchcock, 1956, Plate LVII, fig. 3); **F**, natural mould of a normal track in which the surface that visible is the surface that was in contact with the foot as is made obvious by the presence of the impression of skin on parts of the track, labeled *Brontozoum minusculum* (from Dean, 1861, Plate 16).

lower-relief natural casts and impressions in flaggy sandstones and siltstones, as faint underprints, and as leptodactylous forms (Figures 25, 26). The group name *Leptodactyli* ("thin-toed") was introduced by Edward Hitchcock (1836) for what he regarded as a major subdivision of tracks. In fact such tracks are extremely interesting traces representing one end-member of preservational style, that record the implantation and extraction of a foot in deep (relative to the foot size), soft, usually ripple cross-laminated silt or very fine sandstone. The substrate must have been water-saturated at the time of impression, and probably not bound by micro-organisms such as algae. Gatesy et al. (1999) was (to our knowledge) the first to recognize the true nature of these traces and to extract useful information about trackmaker locomotion from them. The hallmarks of the leptodactylous style of impression is that the “soles” of the toes appear sharply pointed or creased in cross-section, the impression passes through many layers, and on each layer the track has a different shape. This style of track can be extremely abundant where ripple cross-laminated siltstones are present, and while they have great potential to understand locomotory mechanics, they provide little information on the identity of the trackmaker, and sometimes lead to spurious interpretations (e.g. the supposed presence of feathers at the back of the feet).

Thus far, *Rhynchosauroides* cf. *R. hyperbates*, *Apatopus lineatus*, *Brachychirotherium* cf. *B. eyermani*, and cf. *Atreipus* and other indeterminate dinosaurian tracks have been found at this locality, which is one of the oldest footprint assemblages in the Newark basin. Associated with the *Rhynchosauroides* cf. *R. hyperbates* are crisscrossing scratch marks of the appropriate size to have been made by the same trackmaker (Figure 25). Similar scratch marks arranged in crescent-shaped arrays occur associated with *Rhynchosauroides* at other localities (Olsen et al., 1989), and have even been interpreted to be the tracks of a hairy cynodont (Ellenberger, 1976). These scratch traces are almost certainly all foraging traces of reptiles. A striking feature of this and all of the older track localities in the Newark Supergroup (see mileage 34.8 and Stop 6) is that dinosaurian tracks are not common and all are relatively small.

Lockatong Formation fossil assemblages tend to be very stereotyped, almost entirely biased toward preservation of aquatic forms (unsurprising given the dominance of lacustrine facies). Thus this type of assemblage, from fluvial facies (i.e. Stockton Formation) of Lockatong age , is very important, in that it provides a sampling of a more terrestrial assemblage for Lockatong time. Careful further prospecting at this site and others along the Hudson, should provide much more information.

Retrace route back to 9W.

25.3 mi. Turn left onto Route 9W, going south.

26.8 mi. Intersection with Rt. 303, proceed on 9W south.

27.2 mi. Tunnel for Penn Central Railroad. Western portal has outcrops of tans sandstones and dark gray hornfels. Contact with sill is strongly discordant and was figured by Kümmel (1899).

30.8 mi. Cut through discordant limb of Palisade sill.

31.3 mi. Exposures behind Volkswagen repair shop are of tan, gray, and red sandstone. Reptile tracks occur in well-bedded gray siltstones. Tracks include *Grallator* cf. *G. paralellum* (sensu Olsen et al, 1999) and *Brachychiroptherium* sp. (Figure 15). Tan sandstone beds with shale chips and carbonate bleb lags have scales of *Turseodus* and indeterminate small tetrapod bones and teeth.

31.5 mi. Turn left onto Old Mountain Road.

32.2 mi. Turn left onto Castle Heights Road,

32.5 mi. Turn right onto Broadway, heading south. To north is Hook Mountain State Park with outcrops of the Palisade sill and underlying Stockton Formation briefly described by Sanders (1974).

33.3 mi. Turn left onto Depew Road.

33.35 mi. Turn right onto Piermont Avenue and head south.

34.4 mi. Overpass for New York State Thruway and west portal of the Tappan Zee bridge. Borings for the bridge identified the Stockton-basement contact in the Hudson (Figures 7, 13) (Sanders, 1974).

34.8 mi. Brownstone Quarries, Grandview, New York

A string of small- to medium-sized quarries were developed during the 19th and early 20th centuries in red clastics below the Palisade sill, from just south of the present position of the west ramp for the Tappan Zee Bridge in South Nyack, to northern Piermont. These quarries all seem to have exploited the same facies of red fluvial sandstone and mudstone. Much of this distinctive bright red-orange-brown sandstone was used for the construction of buildings in Nyack and homes in Grandview. Most of these quarries now contain private homes.

The sections exposed in these quarries consists of fluvial sequences, much of which are heavily bioturbated by roots and burrows. Several moderate-sized channels are exposed. The abundance of pedogenic structures, and the lack of drab colored beds indicative of long-term water saturation, suggest that this facies is within a dry phase of a 404 ky (or longer) cycle. In support of this, the belt of red clastics progressively rises in elevation to the south and is underlain by a bench of drab coarser-grained clastics that also rises in elevation southwards to central Piermont, where it is exposed along River Road and Ash Street. However, the lower contact of the Palisade sill does <u>not</u> change elevation significantly over the same area. Therefore the contact between the sill and the country rock becomes stratigraphically lower to the south: the sill rests above a thick section of red clastics at Grandview (this Stop), but rests on the afore-mentioned tan sandstones at Piermont. This relationship could also be taken as evidence of onlap of younger strata to the north, but it might instead be due to a slight rise in the position of the Palisade sill, shifting the topographic high to the west.

According to Lawrence Blackbeer (pers. comm., 2001) tracks were recovered in these quarries by a quarry operator. They were sold to Blackbeer, and subsequently sold at Phillips auction house in New York City. ShayMaria Silvestri (Rutgers University) was kindly allowed to photograph these specimens; her photographs are reproduced here (Figure 27). The slabs reveal well-preserved unquestionable examples of large *Atreipus milfordensis* and *Brachychirotherium* cf. *B. eyermani*, the latter preserving skin impressions. The former are remarkably similar to specimens from the middle Passaic at Stop 6 of this field trip, also in the Blackbeer collection. *Atreipus* has also been described from an *in situ* occurrence in the Lower Lockatong Formation at Arcola, Pennsylvania (Olsen and Baird, 1986). The track-bearing layers have not as yet been identified at the present locality. As far as we know, these quarries have never been carefully prospected; additional searching could reveal important new material. *Because all of these sites are privately owned and residential, permission to prospect must be sought in every case, with due care given to private property and privacy.*

36.7 mi. Outcrop of tan arkose and gray and purple mudstone on right.

36.75 mi. Outcrops of tan arkose and gray mudstone and contact with Palisade sill are on northwest corner of Ash and Kinney streets. The red clastics seen in the Grandview quarries to the north are evidently at a higher stratigraphic level.

37.0 mi. Cross Sparkill Creek and turn right onto Ferdon Avenue. .Sparkill Creek flows through Sparkill Gap, which is where the ancestral Hudson River deviated from its present course and headed south west though the Palisade cuesta.

37.8 mi. Turn left onto Valentine Avenue.

37.9 mi. Turn left up dip slope of Palisade sill.continuing on Valentine Avenue to access Route 9W.

38.0 mi. Turn right onto 9W south floowing the strike ridge of the Palisade sill.

40.3 mi. Pass Lamont-Doherty Earth Observatory and New Jersey - New York state line.

43.7 mi. Turn left onto Alpine approach road.

43.8 mi. Continue straight ahead onto Henry Hudson Drive and Enter Palisade Interstate Park.

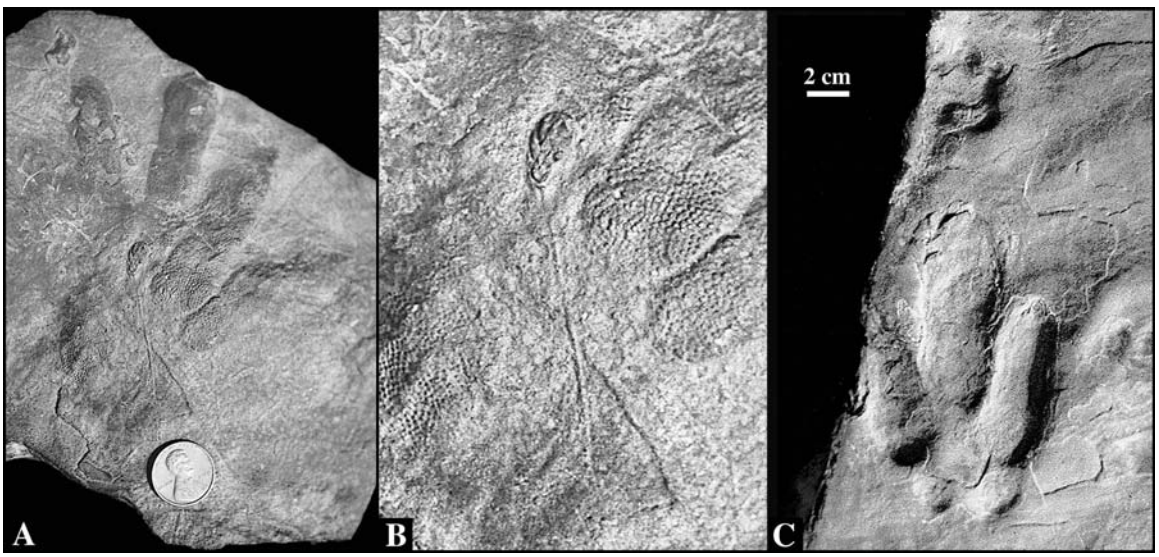

Figure 27. Footprints (all natural casts) from Grandview quarries: **A**, *Brachychirotherium* sp. with clear skin impressions; **B**, detail of skin impressions; **C**, *Atreipus milfordensis*. Specimens formerly in collection of Lawrence Blackbeer sold at auction.

Stop 3. Along-strike transect, Palisades Interstate Park, Alpine to Edgewater, New Jersey.

Latitude and Longitude: 40°56.393'N 073°55.285'W to 40°50.798'N 073°57.882'W

Tectonostratigraphic Sequence: TS III
Stratigraphic Unit: Lockatong and Stockton formations
Age: Late Carnian (Early Late Triassic); 225-226 Ma
Main Points:

1. Lateral transition from Stockton to Lockatong formations at basin edge
2. Marginal to deep-water lacustrine environments
3. Well-developed Van Houten cycles appear
4. Van Houten cycles very thin at hinge margin
5. Two cycles traced through basin with distinctive faunal patterns
6. Palisade sill as part of the CAMP event
7. Intrusion-related structures and sill contact
8. Distinctive facies of Stockton Formation with bones

Lockatong and Stockton sediments and their contacts with the Palisade sill are exposed at numerous places along the Palisade escarpment from Hoboken, New Jersey to Haverstraw, New, and permit a cycle-by-cycle correlation of the Lockatong for at least 15 km of this distance. The individual cycles were informally designated a series of letters and numbers (Olsen, 1980; Figure 28). Exposures in the Palisades Interstate Park, along Henry Hudson Drive and the Hudson River shore, from the Alpine boat basin to the park's entrance off of River Road in Edgewater, provide one of the best places to examine the lateral facies changes in the lower Lockatong and Stockton formations in northeastern New Jersey. In this along-strike transect we will have three mini-stops and five more significant stops (Figures 29, 30).

44.8 mi. Pull off on right side of road.

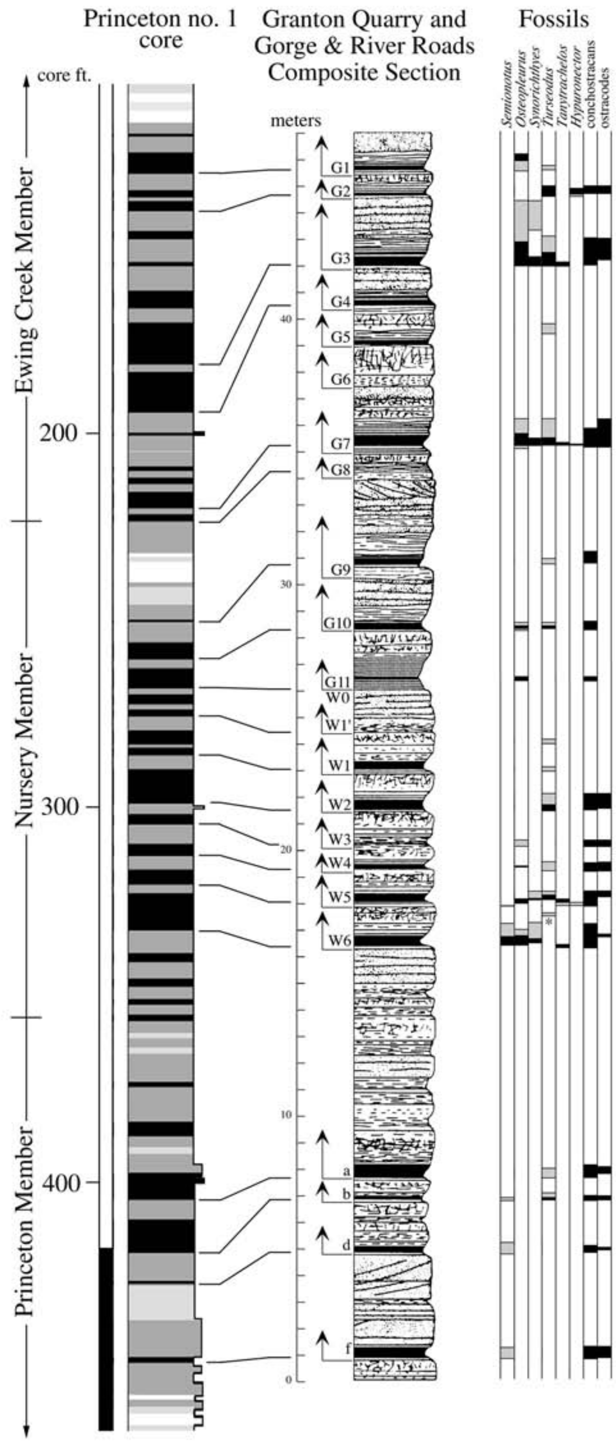

Figure 28. Composite section of Lockatong in vicinity of the Hudson River with the distribution of fossils and correlation to Newark basin coring project core, Princeton no. 1. Key for lithologies of the Princeton no. 1 core in Figure 4 and for northeast composite in Figure 34.

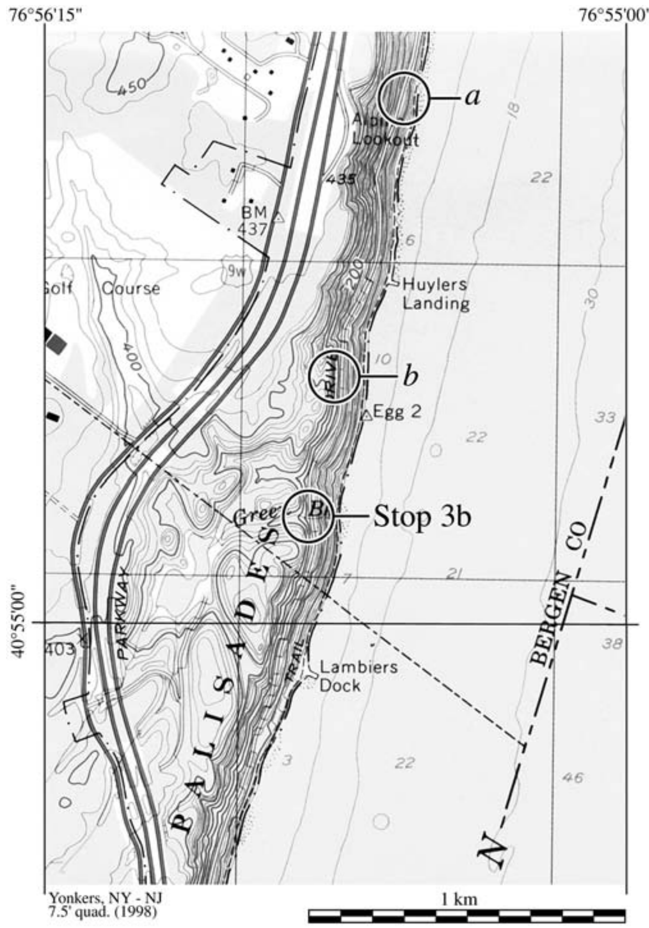

Figure 29. Location map for Stop 3a.

Ministop 3a: Stockton Formation entrance to Alpine Boat basin: **latitude 40°56.393'N, longitude 073°55.285'W.**

Tan and purple arkose with large-scale cross-bedding, and red and dark purple mudstone is exposed along the west side of Henry Hudson Drive and the traffic circle at the south end of the Alpine boat basin. Trough cross-bedding seems to dominate, with paleocurrents heading generally south and west. Given that this outcrop is near river level, it is more likely to be the lateral equivalent of those parts of the Stockton Formation fluvial sequence seen at Stop 3g (see below), rather than a lateral equivalent of the Lockatong Formation as seen at higher elevations (Stops 3b-3f). However, the strata also resemble the outcrop passed at Ash Street in Piermont (NY), which is probably the lateral equivalent to some of the Lockatong levels at Stop 3g.

To the north of the circle is the Alpine Boat Basin and the Blackledge-Kearny House. From the Blackledge-Kearny House (often reported as the headquarters of Cornwallis' assult on Washington's garrison), north there are excellent outcrops of the Stockton Formation very similar in facies to what is seen to the south, notably at Stop 3g. To our knowlege these have not been prospected for vertebrates, or studied.

Head south on Henry Hudson Drive.

45.4 mi. Cross bedding in Lockatong-Stockton lateral transitional facies along road; latitude 40°55.910'N, longitude 073°55.468'W. Currents heading south.

Another small outcrop of cross bedded tan arkose of the Lockatong-Stockton lateral transitional facies; latitude 40°55.432'N, longitude 073°55.607'W.

46.3 mi. Pull off on right side of road.

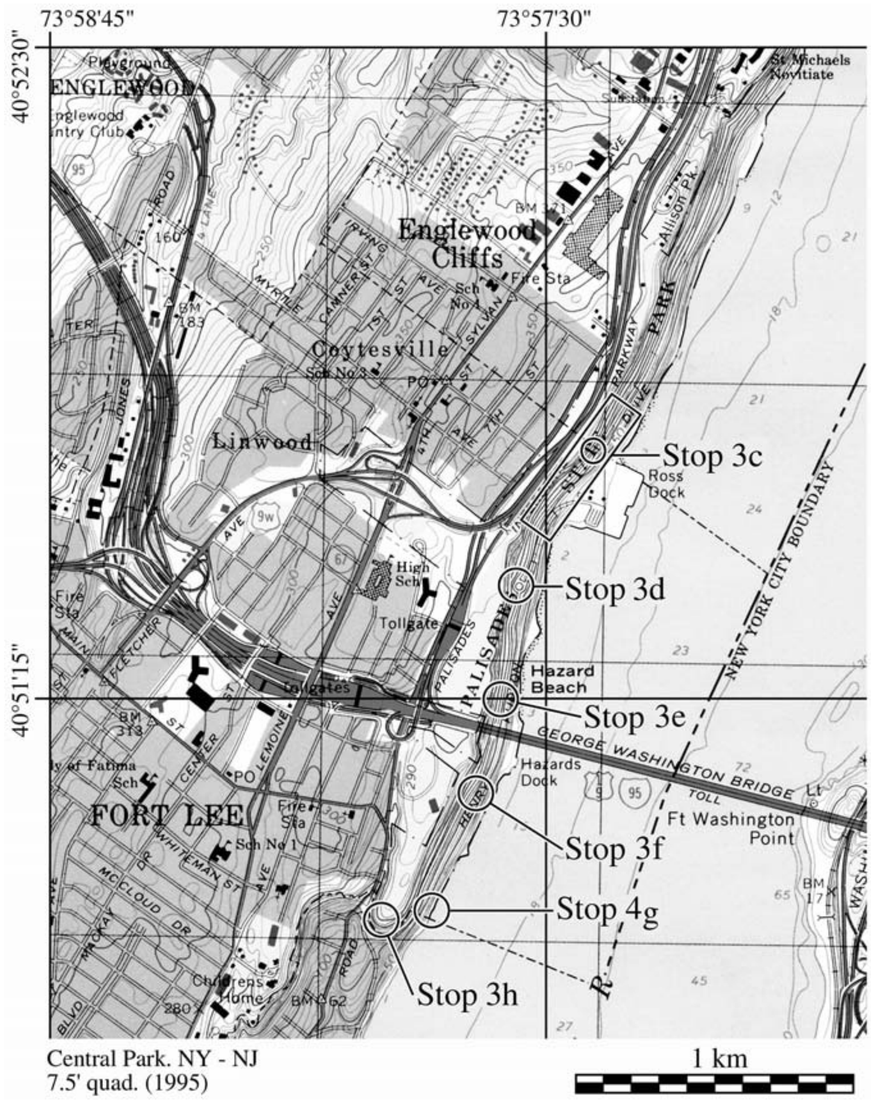

Figure 30. Location map for Stops 3b-h.

Stop 3b: Lockatong-Stockton lateral and vertical transitional, Greenbrook Falls, Alpine, New Jersey: **latitude 40°55.189'N, longitude 073°55.696'W.**

A shear cliff at the waterfall for Green Brook Creek and adjacent exposures on the west side of River Road reveal a long vertical section of the lower Lockatong and underlying Stockton formations. This is the only outcrop of this vertical transition known. Unfortunately, apart from the exposures along the road, the waterfall outcrops are virtually inaccessible and very dangerous. *Note that climbing is NOT permitted in the park.*

The roadside exposures show a small anticline in the still-cyclical lower Lockatong Formation (Figure 31). Here the cycles are considerably thicker than further south along the Hudson, and it is possible that these beds are part of a deltaic complex, changing laterally into the Stockton Formation. Here, each cycle averages about 4 m thick (N=2: Figure 32), consists of a gray massive to faintly parallel-bedded mudstone, overlain by crudely-bedded, locally cross-bedded and oscillatory-rippled tan arkose. These cycles do not show any hint of microlamination and vertebrate fossils have yet to be found. If the gray mudstones were slightly thinner and the irregularity of the bedding increased, this outcrop would be indistinguishable from some of the nearby units mapped as Stockton Formation. These units are probably the lateral equivalents of part of the Nursery or Princeton members at Stop 3c (below).

The anticline could be due to deformation caused by emplacement of the Palisade sill. More likely, it represents bedding distortion caused by modest gravity sliding along depositional relief at the lateral terminus of the Lockatong lake system.

Head south on Henry Hudson Drive.

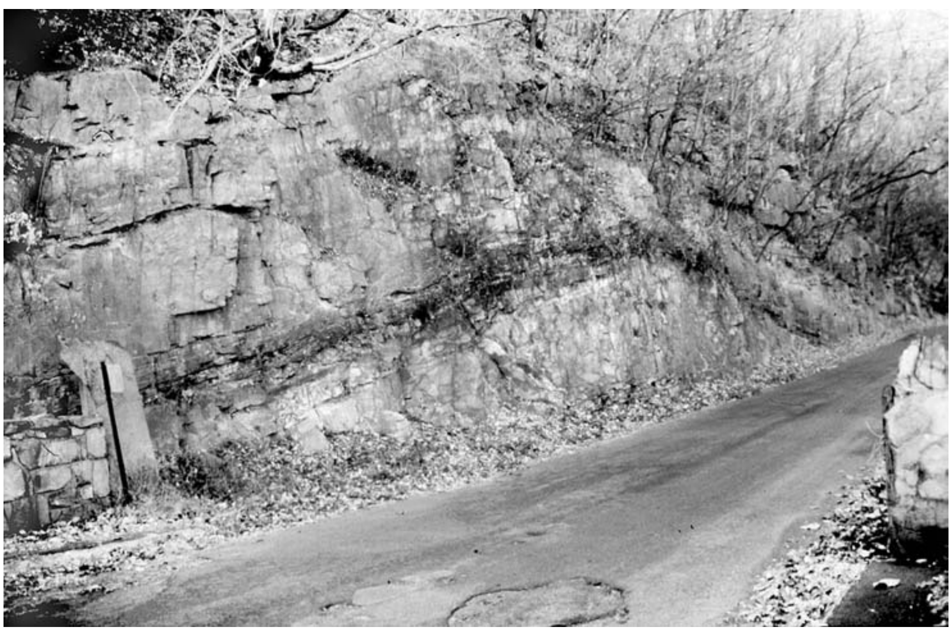

Figure 31. Anticline at Greenbrook Falls, Stop 3b. Prominent dark band is gray mudstone at 4 m in section in Figure 32.

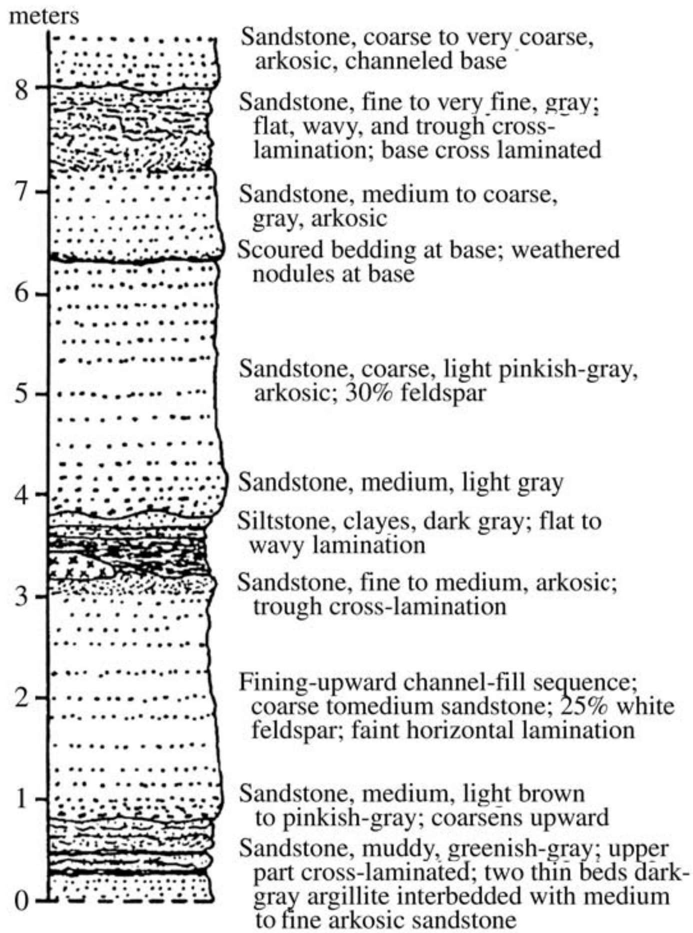

Figure 32. Measured section at Greenbrook Falls, Stop 3b (modified from Parker et al., 1988).

48.8 mi. Exposures of dark gray hornfels of laminated mudstone that appears to be part of cycle W6; 40°52.969, 073°56.705. At this outcrop the burrow *Scoyenia* penetrates the mudstone, consistent with a stronger surface modification and a nearshore environment.

50.6 mi. Pull off on right side of road.

Stop 3c: Van Houten cycles of lower Lockatong Formation, Henry Hudson Drive near Ross Dock: from 40°51.602'N, 073, 073°57.495'W to 40°51.777'N, 073°57.295'W.

These exposures comprise a long section though most of the Nursery and Princeton members at the northernmost outcrops at which these members can be unambiguously identified (Figures 29). This site has been previously described by Olsen (1980) and Olsen et al. (1989). We will begin at the north end of the exposure and proceed south, going up-section (Figure 33).

The series of Van Houten cycles in the Nursery and Princeton members of the Lockatong Formation are among the most distinctive in the entire Newark basin (Figure 34). Not only do they present a specific sequence of distinctive lithologies, but the faunal content of the cycles and the vertical changes in faunal composition is distinctive and persists laterally for at least 150 km (Figure 35). Two cycles in particular, - cycles W5 and W6 - are the lynchpin for the lateral correlation. These two cycles were quarried extensively at outcrops in Weehawken, New Jersey by a team from Yale University in 1979 and 1980 (Olsen, 1980) (Figure 36). More than 3000 specimens of fish and reptiles were recovered, resulting in a detailed sampling of faunal and taphonomic change through the two cycles (Figure 37, 38, 39). At Weehawken, the lower cycle (W6) has a very fine-grained clay-rich partially-microlaminated division 2. Articulated specimens of the tanystropheid *Tanytrachelos ahynis* are present in the transitional beds leading into the microlaminated portion of this division. However, most of the microlaminated portion of this cycle is dominated by the holostean fish *Semionotus braunii*, which was originally described by Gratacap (1886) and Newberry (1888) from a small exposure to the north of the Yale Weehawken excavations, although other fish are present - notably articulated specimens of the small coelacanth *Osteopleurus newarki*, which are relatively common in the upper part of the microlaminated interval. Division 3 is distinctive in that the lower part has a distinct interbed that marks a return to perennial lake conditions. This makes W6 appear as though it has a double division 2. This is characteristic of Van Houten cycles near the paleoequator, because there are two insolation maxima per 20 ky near the equator, rather than one, due to the twice-annual passage of the sun over the equator (Crowley et al., 1992; Olsen and Kent, 2000). Many Newark basin Van Houten cycles show hints of this pattern consistent with the basin’s near-equatorial position, but none show it as strongly as cycle W6. Interestingly, this deeper water interbed has a different dominant fish - the palaeoniscoid *Turseodus* - represented, however only by disarticulated, albeit distinctive, elements.

The succeeding cycle, W5, has disarticulated rare *Semionotus* at its very base, associated with abundant disarticulated elements of the bizarre drepanosaurid diapsid *Hypuronector limnaios* (Colbert and Olsen, 2001) (Figures 8, 39), Above this, articulated *Osteopleurus* occur, but higher up, articulated *Turseodus* dominate the microlaminated interval. The microlaminated interval contains metamorphic minerals (e.g. diopside) that show that the laminite was much more calcareous than that of cycle W6. Virtually all of these features persist laterally from Weehawken northward to this stop (Figure 37).

Cycles W5 and W6 have also been identified in the NBCP cores and in outcrops in the southwestern part of the basin. Because of the distinctiveness of these two cycles, the correlation of the adjacent cycles is certain, and the pattern of cycles as well as their constituent fauna is maintained across much of the basin (Olsen at al., 1996a).

In contrast to sections further south that contain these cycles, this section contains the largest proportion of tan arkose, most notably in many of the shallow-water portions of the Van Houten and short modulating cycles. Much of this tan arkose exhibits oscillatory ripples, especially in division 1 of the cycles. Compared to sections further south, it is clear that the dry phases of the short modulating and McLaughlin cycles are abbreviated, with some parts even omitted (Figure 35). In addition, the degree of lamination is generally less. The average thickness of the Van Houten cycles at this stop

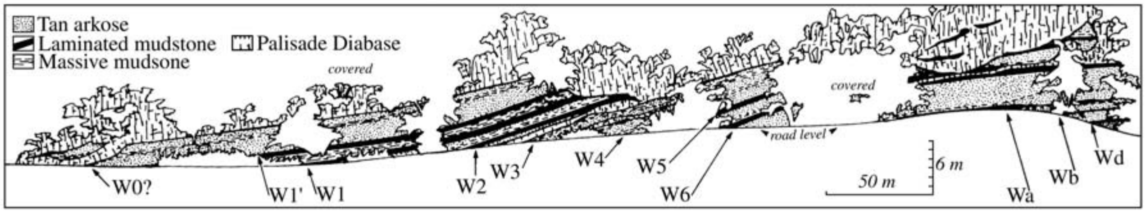

Figure 33. Panel diagram of the section along Henry Hudson Drive near Ross Dock (Stop 3c) showing the various cycles and the stratigraphically down stepping to the north of the Palisade sill (north is on the right). From Olsen 1980b.

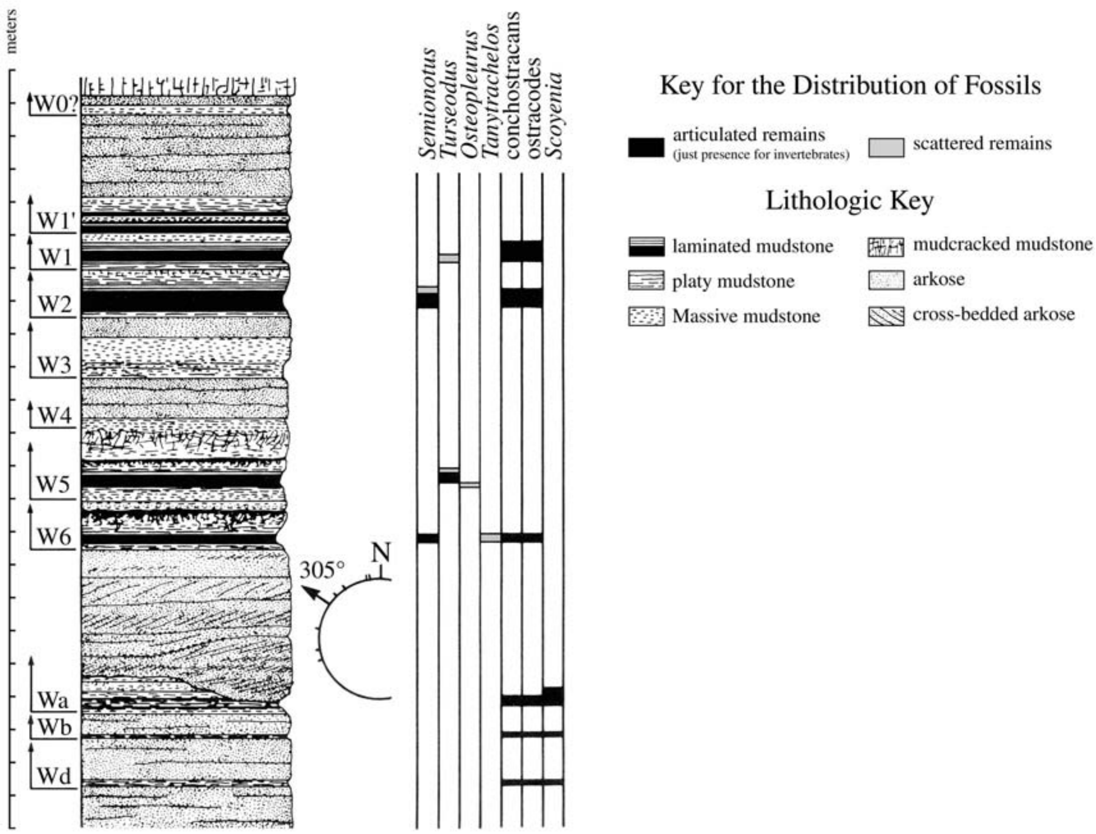

Figure 34. Composite stratigraphic section at Ross Dock with distribution of fossils. From Olsen (1980b).

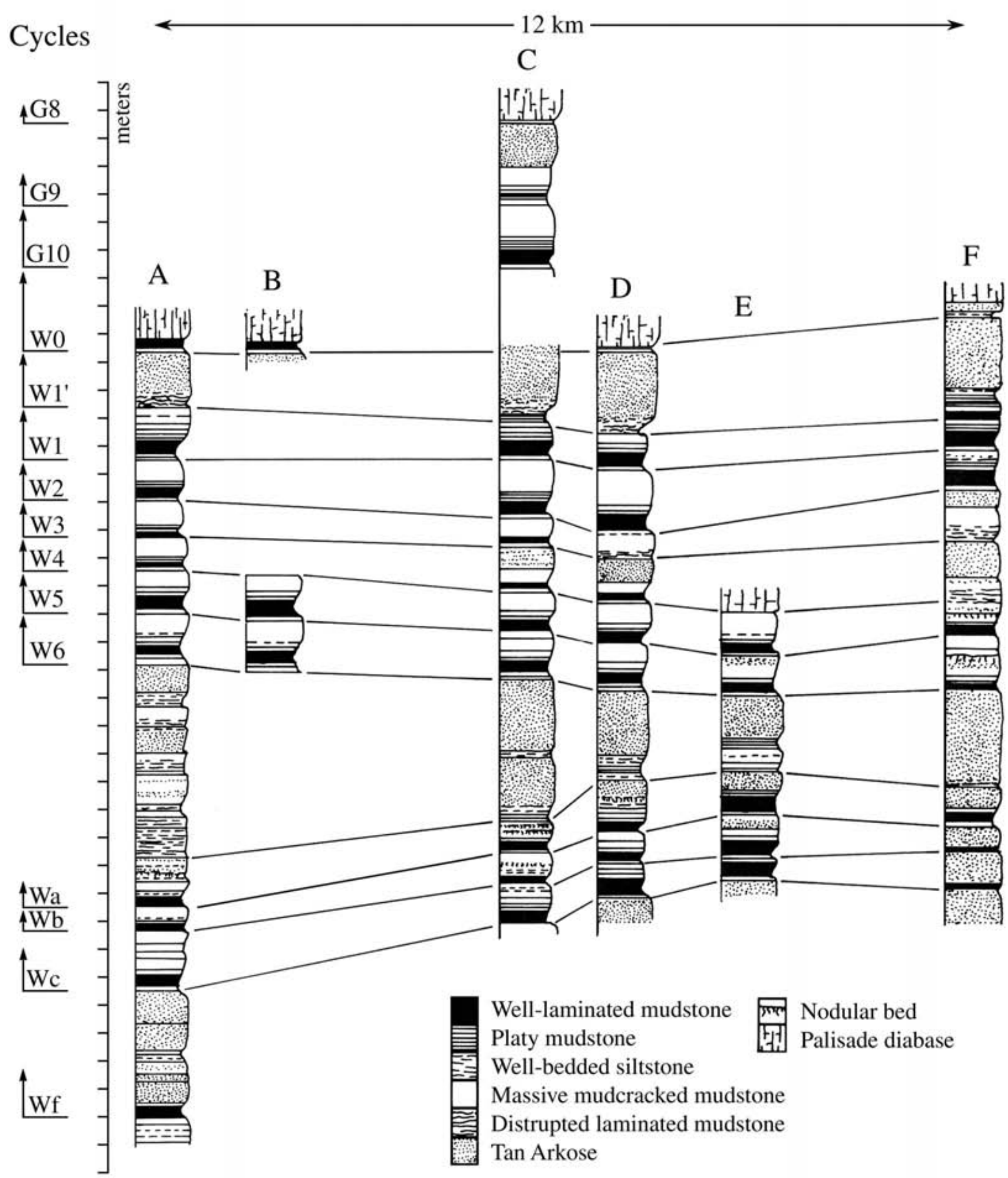

Figure 35. Lateral correlation of cycles from Kings Bluff (Yale quarry), Weehawken to Ross Dock area, Fort Lee (Stop 3c) (modified from Olsen 1980b): A, Kings Bluff exposure; B, Gratacap's (1886) Weehawken locality; C, George and River roads (Stop 4). Edgewater; D, east portal for old New York, Susquahanna Southwestern Railroad; E, "old trolley route" below former site of Palisades Amusement Park, Edgewater, New Jersey; F, Ross Dock area (Stop 3c), Fort Lee. Exposure A and F are 12 km apart; the other sections are positioned to scale.

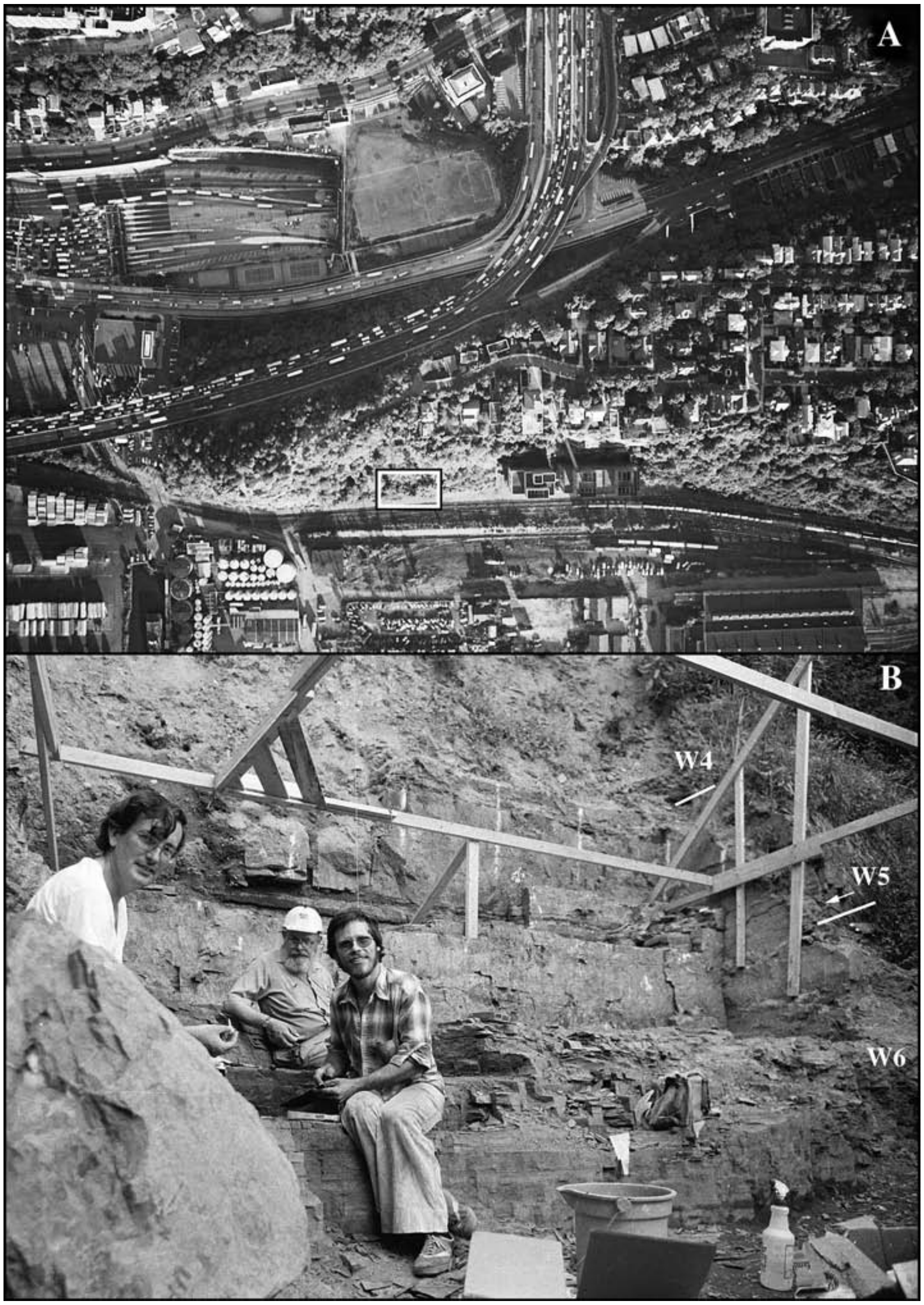

Figure 36. **A**, Aerial view of Weehawken Yale quarry location (box) at Kings Bluff, just south of the ventilation buildings for the Lincoln Tunnel (1979). Major highway is Interstate 495 feeding into the Lincoln Tunnel toll plaza (center, upper left). North is on right. Photo by William K. Sacco. **B**, Yale quarry, Weehawken, New Jersey showing the main cycles excavated at the site, 1979: from left to right, Keith Stewart Thomson, Donald Baird, Paul E. Olsen. Wooden frame was for identifying 3 dimensional position of all specimens recovered.

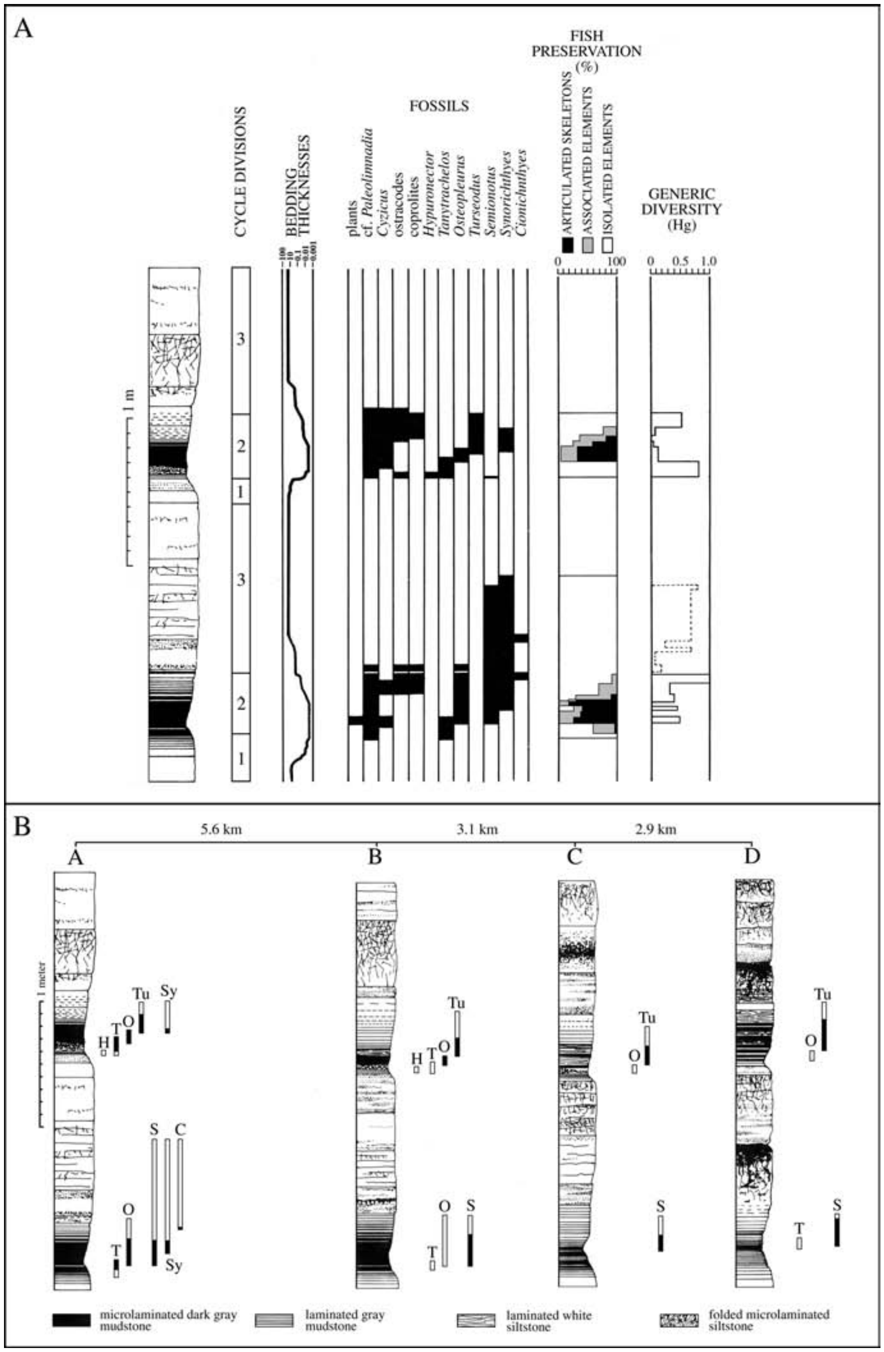

Figure 37. A, Microstratigraphy and biostratonomy of W5 and W6: key to lithologies as in Figure 34 except as shown. **A**, biostratonomy of W5 and W6 at the Yale quarry at Weehawken. Generic Diversity (Hg) is the Shannon-Weaver (1949) information index. **B**, Comparision of sections of cycles W5 and W6 along strike showing distribution and preservation style of fish: A, Yale Quarry, Kings Bluff, Weehawken; B, Gorge and River Roads, Edgewater (Stop 4); C, "Old Trolley Route", Edgewater; D, Ross Dock area (Stop 3c); C, *Cionichthyes*; H, *Hypuronector*; O, *Osteopleurus*; *Semionotus*; Sy, *Synorichthyes*; T, *Tanytrachelos*; Tu, *Turseodus*.

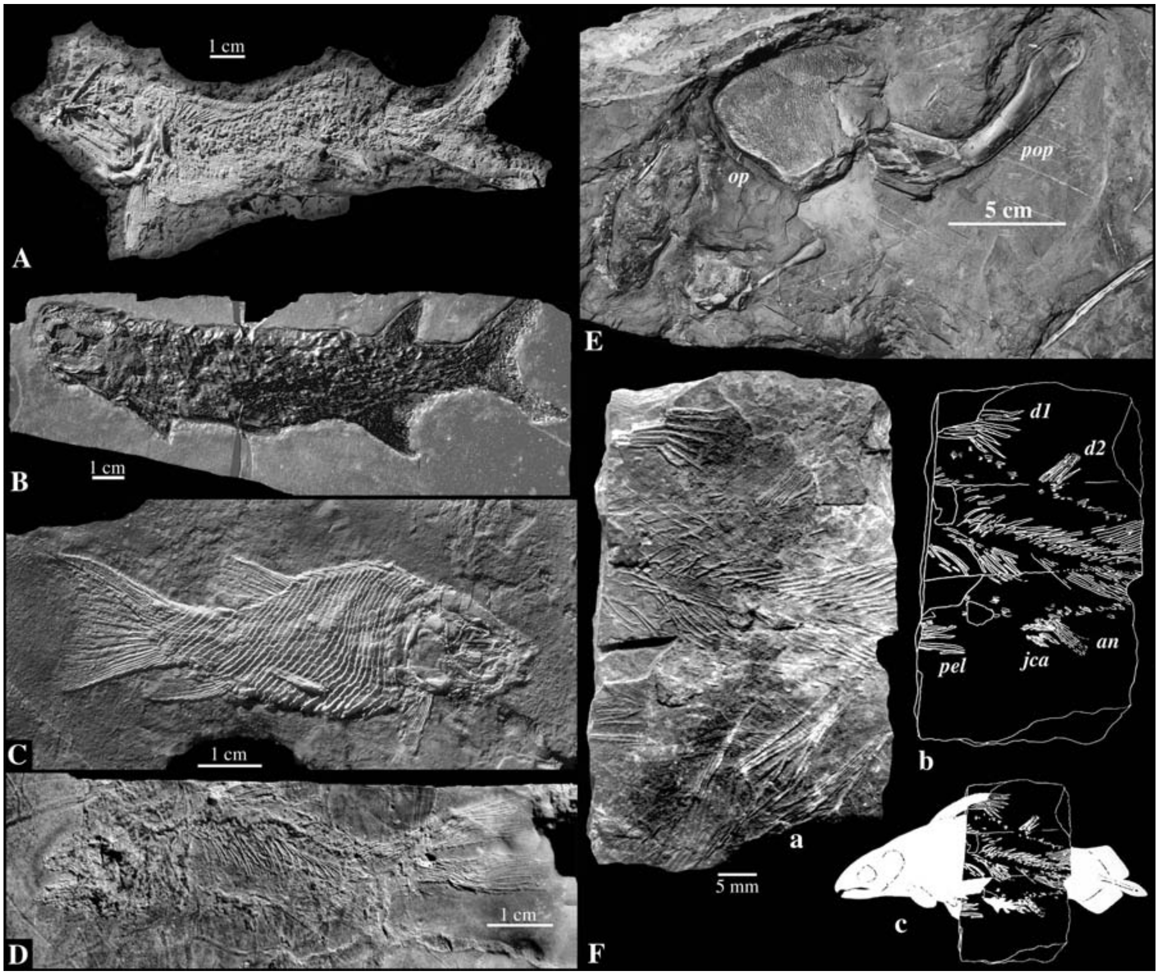

Figure 38. Examples of fish from Lockatong Formation. **A**, polysufide cast of the palaeoniscoid *Turseodus* sp. from cycle W-5, Yale Weehawken quarry (YPM field collection number W5-663). **B**, the refieldiid palaeonisciform *Synorichthyes* sp. from Gwynedd, Pennsylvania (YPM 8863). **C**, the holostean *Semionotus braunii*, from cycle W-6, W-5, Yale Weehawken quarry (YPM xxxx). **D**, the small coelacanth *Osteopleurus newarki* from cycle G-7, Granton Quarry (Stop 5). **E**, disarticulated partial skull of the large coelacanth cf. *Pariostegeus* sp. from float at Granton Quarry (Stop 5) (NJSM 16697: see Rizzo, 1999a): *pop*, preopercular; *op*, opercular. **F**, fragmentary, articulated specimen of *Osteopleurus newarki* from float at Granton Quarry (Stop 5) that seems to have been buried in the process of giving birth to a baby (see Rizzo, 1999b) (NJSM 15819): a, slab preserving mid section of probable adult and juvenile; b, drawing of same (*d*1, anterior dorsal fin; *d2*, posterior dorsal fin; *pcl*, pelvic fins; *an*, anal fin; *jca*, juvenile (caudal fin) in position of cloaca, just in front of anal fin); c, orientation of slab relative to outline of complete fish.

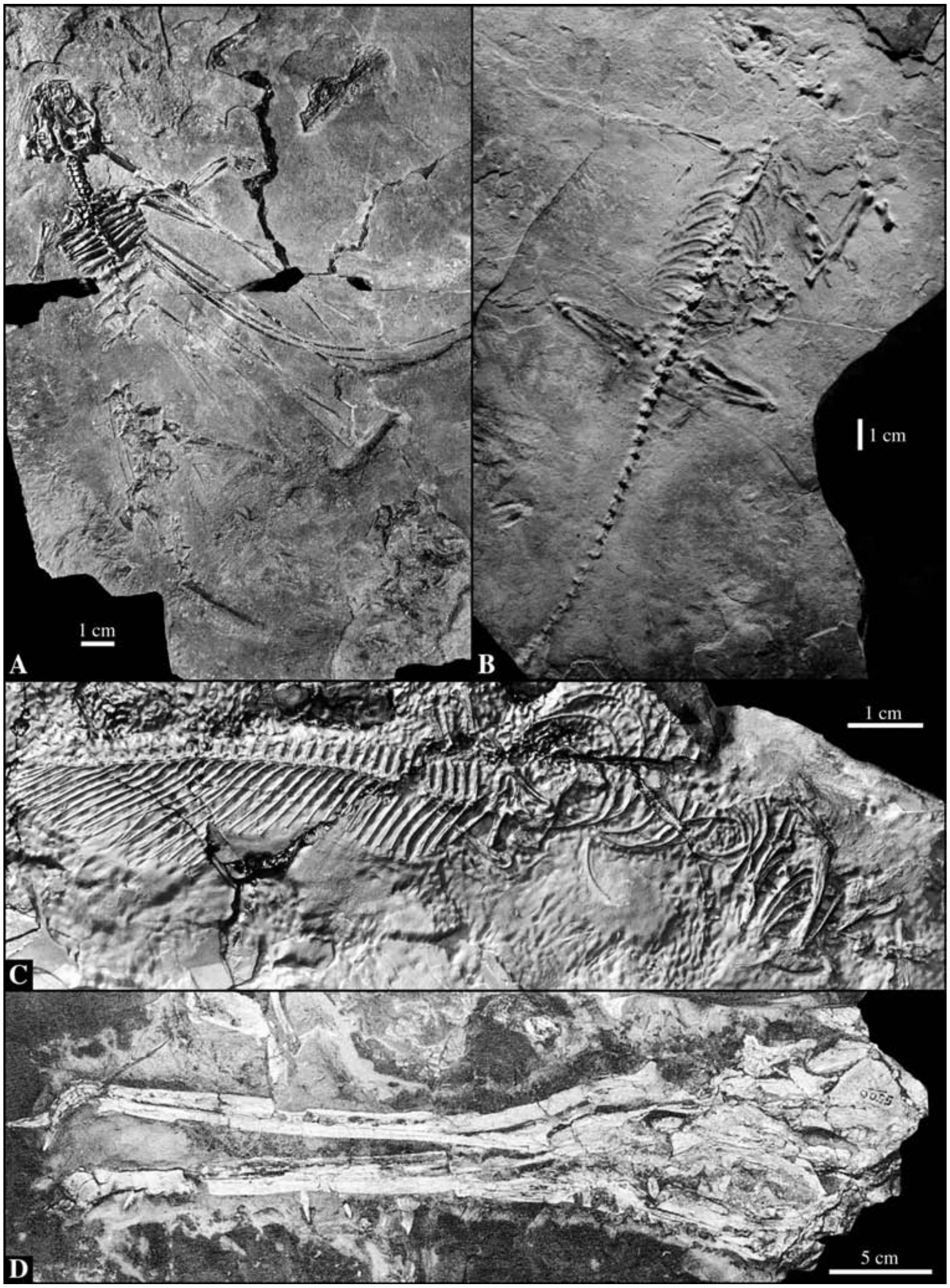

Figure 39. Reptiles from the Lockatong Formation at Granton Quarry (Stop 5): **A**, type specimen of *Icarosaurus seifkeri* from cycle G-3 (uncataloged AMNH specimen) (from Colbert, 1966; with permission of the American Museum of Natural History); **B**, female *Tanytrachelos ahynis* found by Steven Stelz, Trinny Stelz and James Leonard; **C**, type 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).

is about 1.5 m. This is only about 25% of their thickness in the central Newark basin, and is consistent with the position of this section on the shoaling, hinge side of the basin.

On the whole, the distribution of vertebrate remains is consistent with stratigraphic equivalents further south, although the degree of articulation is less at this stop. In all cases at these exposures, bone and scale material is preserved as a translucent milky or pinkish phosphate, making it difficult to see. Combined with the extreme hardness of the metamorphosed sediment, this makes finding vertebrate material very difficult at this locality - although good specimens are present and can be found.

The underlying cycle, Wa is penetrated by numerous *Scoyenia* burrows (which were definitely not present at Stop 4) as are cycles Wb and Wc. We are clearly in the shallow-water facies of division 2 of cycles Wa-c at this point, but only just leaving the deep-water facies of division 2 in cycles W2, W5, and W6. Cycle Wa shows well-developed fracture cleavage in division 1. This cleavage dips 25° - 30° and strikes S78°W. It is strata-bound but discontinuous, passing laterally into breccia or non-cleaved beds. What is the significance of these structures?

The tongue of buff arkose between cycles W6 and Wa is thinner than at Stop 4 and displays unidirectional, oscillatory, and possibly hummocky cross-stratification, suggesting wave reworking of sheet deltas. Division 2 of cycle Wa is cut out by the arkose beds of this sequence by a channel-fill sandstone with mudcracked mudstone interbeds. Mean paleocurrent direction for these cross-beds is N59°W (based on 8 readings) (Figure 34).

In cycles W5 and W6 the degree of microlamination is significantly less than in these cycles further to the south, consistent with the rest of the section, significantly so in W5. The sequence of vertebrate taxa is still consistent with the pattern seen at Weehawken, but the degree of articulation has decreased. However, articulated *Tanytrachelos* and *Semionotus* are still present in division 2 of cycle W6. The upper part of division 1 of W6 has produced a partial arthropod, of uncertain relationships, about 20 cm long (Olsen, 1980). The most dramatic change from outcrops to the south, however, is that in this area, nodules, probably originally of carbonate, increase upward from the microlaminated portion of division 2 of cycle W6, coalescing into a nodular bed in the lower part of division 3, just below the more-laminated bed within division 3 that is such a distinctive feature of this cycle (Figures 34, 40). We interpret this nodular ?former-carbonate level as a caliche, developed in the dry lakebed of the lower parts of cycle W6. Hints of this caliche begin to appear in outcrops to the south (Figure 37).

Cycles 3 and 4 are evidently replaced by wave-influenced buff, cross-bedded arkose. Cycle 2 is very well exposed and contains the fish *Turseodus* and *Diplurus*. Cycle 1' is present, overlain by 4 m of arkose, but there is no sign of cycle 0, which presumably has pinched out or was cut out south of here. Cycle 1 is present but poorly exposed.

The facies trend in the Lockatong Formation from Stops 3 to 5 is from a basin-margin facies to a more central basin facies. The lateral heterogeneity seen toward the basin's hinge margin [Stops 3-5] gives way to monotony in horizontal continuity to the south. Those cycles with the best developed microlaminae and the best preserved fish at this stop are also those which persist the furthest laterally with the least change.

Figure 40. Exposures along Henry Hudson Drive, Stop 3c near Ross Dock showing cycles W5 and W6: *a*, nodular probable caliche horizon.

Continue south along Henry Hudson Drive.

50.8 mi. Pull off on right side of road.

Ministop 3d: Abandoned quarry in Palisade sill: latitude 40°51.460'N, longitude 73°57.534'W.

This old quarry, at the traffic circle at the ramp leading to Ross Dock, exposes the lower half of the Palisade sill (Figure 41). As noted by Walker (1969), the olivine zone of the sill produces an obvious bench along the escarpment to the immediate south, and elsewhere in the area, essentially paralleling the lower contact of the sill. In the cliff face, the olivine zone is marked by a zone of deflected columns. Flow banding is present within the olivine zone (Naslund, 1998).

Proceed south along Henry Hudson Drive.

50.9 mi. Pull off on right side of road.

Stop 3e: Concordant contact between Palisade sill and Lockatong Formation: latitude 40°51.253'N, longitude 073°57.587'W.

This comprises what is probably the most spectacular exposure anywhere in the basin of the Palisade sill with the underlying sedimentary rock (Figure 42). The contact is exposed for more than 50 m along strike, and reveals the intimate structure of the Lockatong-sill contact. The sill itself is part of a series of probably once-continuous sills and plutons that extended over almost the entire Newark basin, making this component of the CAMP one of the most extensive sills in the world.

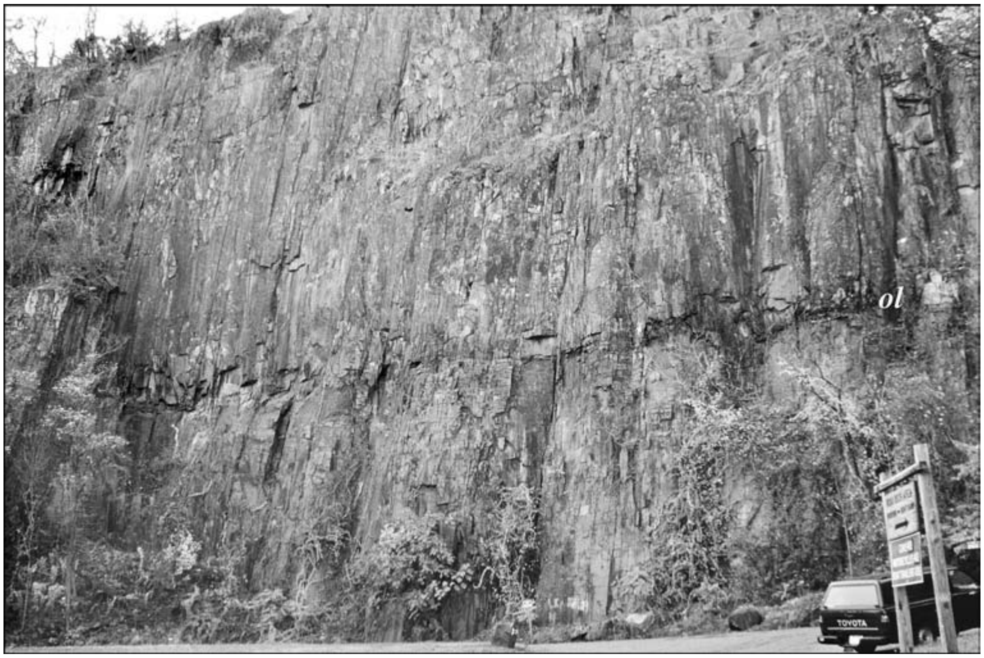

Figure 41. Quarry face at Stop 3d, with olivine zone (*ol*) in zone of deflected columns.

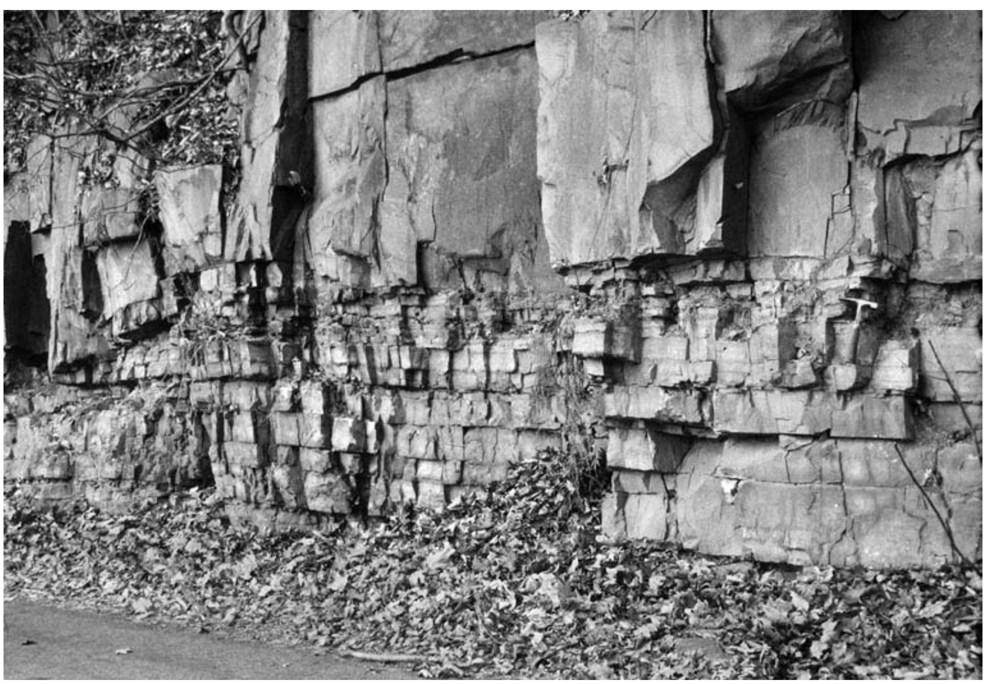

Figure 42. Mostly concordant contact of Palisade sill with mudstone and arkose of Lockatong Formation just north of the George Washington Bridge at Stop 3e.

Note that the contact is extremely sharp and that there is virtually no evidence of assimilation of sediment into the sill. Here and there the sediment-sill contact jumps a few tens of centimeters up or down, but on the whole it is remarkably concordant. Because so little thickness of Lockatong Formation is exposed here, we don't know what cycle is represented by this exposure.

Proceed south along Henry Hudson Drive.

51.2 mi. Pull off on right side of road.

Ministop 3f: Discordant contact between Palisade sill and Lockatong Formation: latitude 40°51.063N, longitude 073°5.672'W.

This exposure shows the contact between the Palisade sill and Lockatong Formation jumping first up several meters and then back down (Figure 43). The more mud-rich intervals appear to behave as if brittle, with essentially no assimilation into the sill, while the tan arkose appears to have flowed at the sill contact, with considerable chaotic mixing into the sill.

Proceed south along Henry Hudson Drive.

51.4 mi. At the middle of the sharp right turn to the west. cross over rock wall onto path, and walk down to the river's edge.

Stop 3g: Outcrops of Stockton Formation along Hudson River shore: 40°50.886'N, 073°57.727'W to 40°50.802, 073°812'W.

From just south of here to beneath the George Washington Bridge there are scattered outcrops of variegated strata of the Stockton Formation; their facies are characteristic of the units immediately below the Lockatong in this region. Tan, gray, and purple trough cross-bedded arkose and pebbly arkose grade upwards, through a series of irregular beds, into bright purplish-red massive mudstone, at meter-scale repetitions (Figure 44).

A partial disarticulated postcranial skeleton of a large phytosaur was recovered on private land just south of where the path comes down to the water's edge (Figure 44). It appears to have come from the transition between the arkose and red mudstone. The specimen was discovered in the summer 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. They contacted Barnum Brown and W. D. Mathew at the American Museum of Natural History (AMNH) in New York City, and 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 mis-identified in most published reports, and is usually cited as being "a half-mile south of the George Washington Bridge, opposite 155th St. [ref. and exact quote], even though 155th St is closer to 1 mile south of the bridge! However Mathew (1910) states that the specimen is from opposite <u>160th</u> St., which <u>is</u> approximately 1/2 mile south of the bridge; The specimen is also commonly referred to as the "Fort Lee phytosaur" (e.g. ref.),

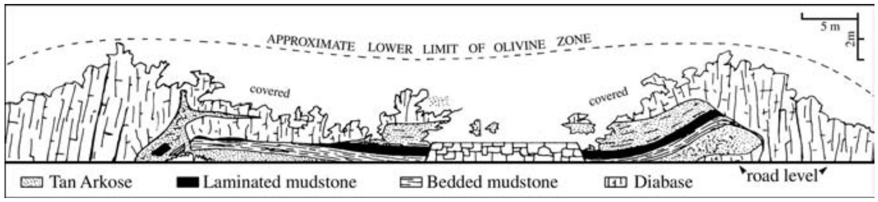

Figure 43. Sketch of discordant contact of Palisade sill and Lockatong Formation along Henry Hudson Drive just south of the George Washington Bridge at Stop 3f (from Olsen, 1980).

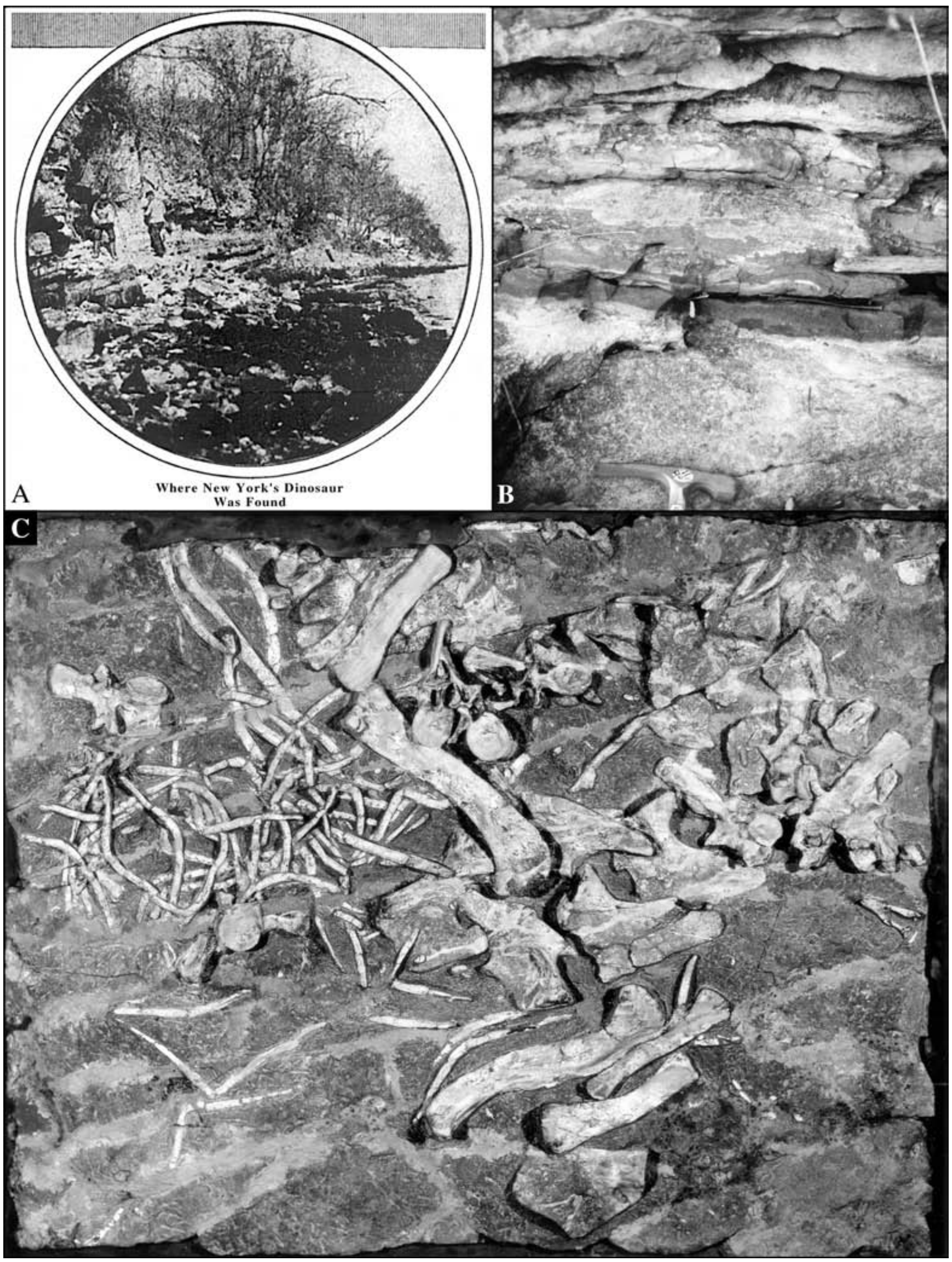

Figure 44. *Rutiodon manhattenesis* and outcrops at Stop 3g, Stockton Formation: **A**, photograph from the front page of the magazine section of the New York Times for December 25, 1910, showing the location of the phytosaur skeleton just south of the boundary with the Palisades Interstate Park (Stop 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).

although it is actually from Edgewater. The specimen (AMNH 4991; Figure 44) 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 should prove fruitful. AMNH 4991 is currently on exhibit in the Hall of Vertebrate Origins at the American Museum of Natural History in New York.

Walk back up hill to Henry Hudson Drive and proceed west toward intersection with Main Street (Fort Lee) entrance to the park.

51.6 mi. Stop at outcrops of diabase on north side of road.

Mini Stop 3h: Rotten olivine zone: **latitude 40°50.798'N, longitude 073°57.882'W.**

Here weathering profile of the olivine zone can be clearly seen. The olivine zone is weathered to a very crumbly diabase, that according to Naslund (1998) still has much fresh-looking olivine crystals.

51.8 mi. Leave park and turn left onto Main Street in Fort Lee and proceed south. Main Street becomes River Road once in Edgewater.

53.0 mi. Undercliff Road parallels escarpment on right. Stop 6.7 in Olsen et al. (1989) is along old trolley cut for Palisades Amusement park (Van Houten, 1969). Cut exposes upper Stockton and lower Lockatong Formation including cycles W5, W6, and Wa-c, as well as the contact with the Palisade sill including a large xenolith of cycle W5 within the sill.

54.4 mi. Turn right onto Old River Road.

54.5 mi. Exposures in old quarry in back of buildings exposing cycles W1-6.

54.6 mi. Driving over east portal of the tunnel for the former New York Susquehanna and Western Railroad. Cut below bridge comprises excellent exposures W1-6 and cycle Wa-c, (Olsen et al., 1989). Cycle W2 is very fossiliferous here with excellent articulated *Turseodus* and large disarticulated coelacanths (cf. *Pariostegus*).

55.2 mi. Park near truncation of Old River Road.

Stop 4. Gorge and River Road Section, Edgewater, New Jersey

Latitude and Longitude: 40° 48.458'N, 073° 59.534'W
Tectonostratigraphic Sequence: TS III
Stratigraphic Unit: Lockatong Formation
Age: Late Carnian (Early Late Triassic); 222 Ma
Main Points:

1. Long section of typical facies at hinge margin of Lockatong
2. Marginal to deep-water lacustrine environments
3. Well-developed Van Houten cycles in more basinward environment
4. Same cycles as at Stop 3c
5. Metamorphic mineral suite

A long section, recently expanded, is exposed along Gorge Road from its intersection with Old River Road to the contact with the Palisade sill at the Legend Hills Condominium development (Figures 45, 46). The same cycles seen at Stop 3c (from W6 upwards) are exposed here; however the section continues up into cycles not seen elsewhere along the east side of the Palisade sill, overlapping into the section exposed at Granton Quarry (Stop 5).

The most striking differences between this stop and the cycles seen at Stop 3c, are the increase in the degree of lamination in all cycles, and a decrease in the proportion of the cycles containing arkose. This is coupled with an increase in the abundance of fossils, especially fish and reptiles. The biostratonomy (*sensu* Schafer, 1972) of the cycles are now closely comparable to what is seen at the Yale Quarry at Weehawken, including such details as the presence of isolated bones of *Hypuronector* in the thin (~2 cm) massive gray mud at the base of division 2 of cycle W5. In addition the paleosol caliche in cycle W6 has disappeared and is replaced by massive mudstone.

We will begin at the north termination of Old River Road, on the south side of Gorge Road adjacent to 180 Old River Road. Cycles W5 and W6 and the underlying arkosic sequence are exposed near road level, but only W6 and the arkose are easily reached. The 5.5 m of arkose between cycles W6 and cycle Wa has some unidirectional cross-bedding indicating transport to the west and southwest (Van Houten, 1969), along with some wave-influenced sandstones, including some very handsome oscillatory ripples. (J.P. Smoot, pers. comm.). The arkose tongue and thin siltstones present in this interval are the lateral equivalents of relatively more shallow-water cycles present in the central Newark basin. Van Houten (1969) described the metamorphic minerals at this stop (Figure 45). Overall, from Stop 3c to this stop the facies trend, in terms of both lithology and fossils, is consistent with moving further away from the basin edge.

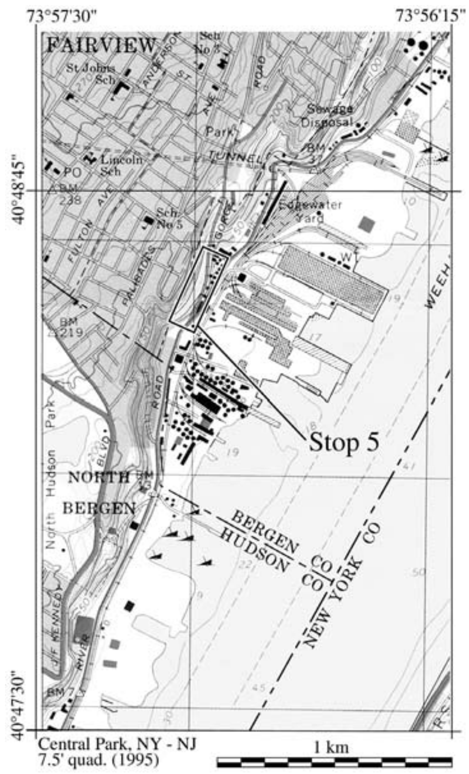

Figure 45. Map for Stop 4, Gorge and River Roads.

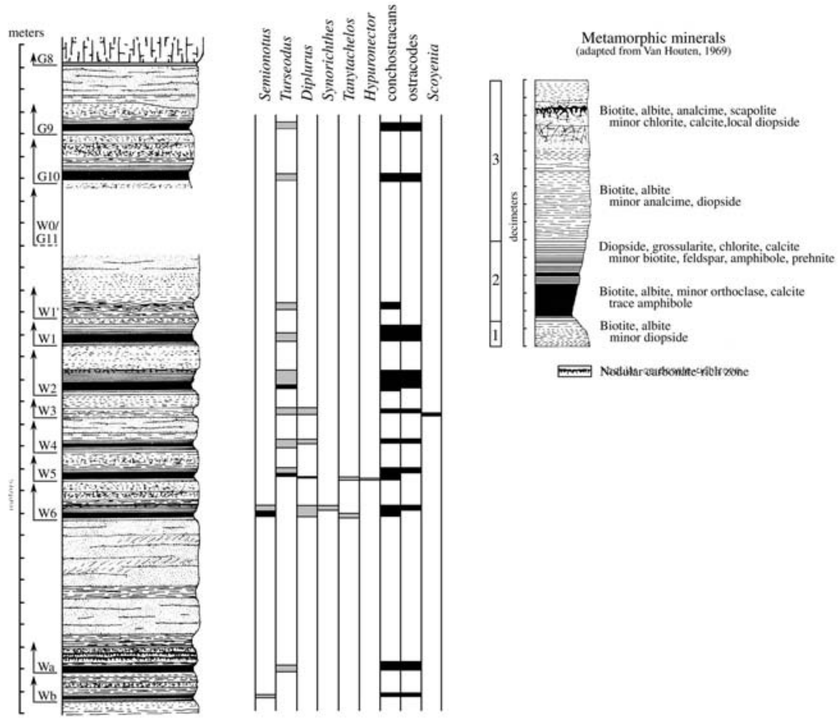

Figure 46. Section and fossils at Gorge and River Roads, Stop 4 (modified from Olsen, 1980b).

Proceed (with caution) east across Gorge Road, walk north along Gorge Road on the sidewalk, and then cross back to the exposures on the north side of the entrance road for the Legend Hills Condominiums. Lying in superposition on top of the sections exposed in 1980 are two more cycles separated by a gap of non exposure. The gap has to contain cycle W0 as seen at other nearby exposures. The two cycles are separated from the beautifully-exposed contact with the Palisade sill by a tan arkose sequence. Lithologically the two newly-exposed cycles resemble cycles W2 and W1, but are different enough that we can be sure they are indeed not simply structural repetitions of these cycles, but rather they are nearly identical to cycles G9 and G10 at Granton Quarry (Stop 5). Recognition of the identity of these two cycles at Granton Quarry is very important because it allows the sections on the east and west sides of the Palisade sill to be concatenated (Figure 28).

The arkosic sequence at the top of the section locally cuts deeply down into cycle G9; this is visible on both the south and east faces of the cut at the entrance to the Legend Hills Condominiums, and there is cross bedding visible within the the arkose. Just at the contact with the diabase, the arkose becomes very well-bedded and is capped by mudstone. This evidently marks the base of cycle G8 as seen at Granton Quarry.

Note that Gorge Road follows a fault-line ravine north of the Lockatong exposures; the eastern block is downthrown about 53 m (Van Houten, 1969).

Turn around, heading north on Old River Road.

55.2 mi. Turn right onto unnamed road.

55.25 mi. Turn right onto River Road.

55.28 mi. Turn right onto Gorge Road.

56.2 mi. Turn left onto Edgewater Road.

57.4 mi. Turn left onto Shaler Avenue.

57.5 mi. Cut in back of Shaler Avenue School is remnant of old quarry. Exposures consist of about 10 m of tan arkose and minor gray and purple mudstone of the upper Lockatong Formation.

58.0 mi. Turn left (south) onto Broadway (Routes 1 and 9).

58.7 mi. Cut for west portal of the tunnel for the former New York Susquehanna and Western Railroad exposes tan arkose (Parker, 1993).

59.3 mi. Turn right onto 83 Street.

59.5 mi. Turn left into back entrance for Tonnelle Plaza and drive up ramp to back of plaza complex.

59.8 mi. Park.

Stop 5. Granton Quarry, North Bergen, New Jersey.

Latitude and Longitude: 40° 48.431'N, 074 01.071'W
Tectonostratigraphic Sequence: TS III
Stratigraphic Unit: Lockatong Formation
Age: Late Carnian (Early Late Triassic); 222 Ma
Main Points:

1. The classic locality for Lockatong fossils
2. All reptile and fish taxa represented
3. Well-developed Van Houten cycles
4. Granton CAMP sill
5. Bedding plane faults as evidence of inversion

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 (Figure 47). 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 (Figure 39). 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 (Figure 48), three additional cycles are exposed on the east-acing exposure; and all 11 cycles are exposed on the north-facing exposure, which is where we will examine them. The base of the section appears to be 38-46 m above the contact with the Palisade 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 (1980), 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 recrystalized 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 (Figure 48).

Cycles G3 and G7 (Figure 48) have produced representatives of all the known skeletal remains of Lockatong vertebrates except the holostean *Semionotus*. The basal portions of division 2 of both of these cycles have extremely high densities of fossil fish, 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;

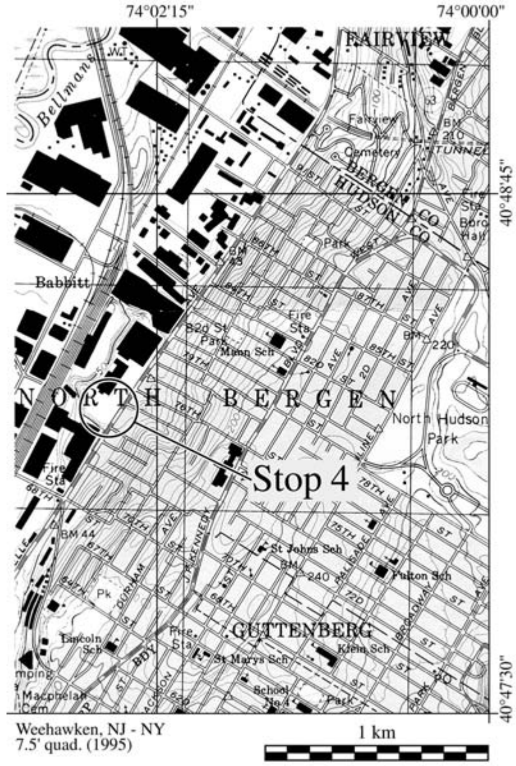

Figure 47. Map for Stop 5, Granton Quarry.

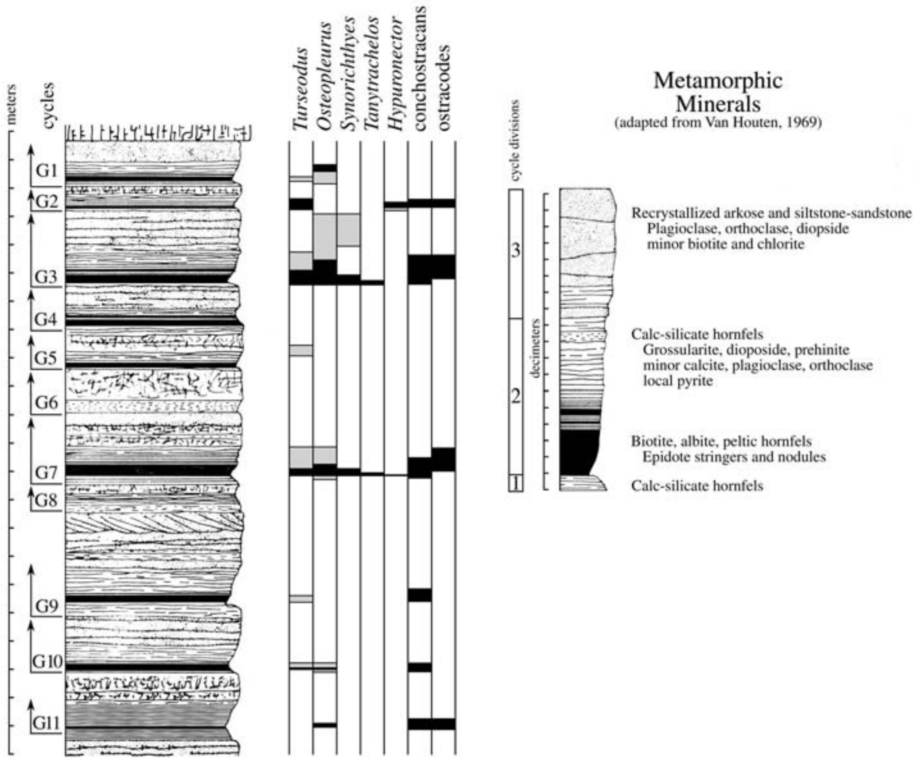

Figure 48. Section and fossils at Granton Quarry, Stop 5 (modified from Olsen, 1980b).

Schaeffer and Mangus, 1971). Without a doubt the three most spectacular skeletons of small reptiles found in the Lockatong come from the Granton Quarry. These include the type specimen of the bizarre "deep-tailed swimmer", *Hypuronector limnaios* (Colbert and Olsen, 2001), the peculiarly-abundant tanystropheid *Tanytrachelos ahynis* (Olsen, 1979), and the gliding lepidosauromorph *Icarosaurus seifkeri* (Colbert, 1966) (Figure 39). Larger remains occur as well, of which the most spectacular is the skull of a juvenile rutiodontine phytosaur (Figure 39), but isolated phytosaur bones and teeth are fairly common and isolated vertebrae of a metoposaur amphibian also been found.

Cycles G8 through G11 overlap with the section on the east side of the Palisade 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 eastern most 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 Palisade sill combined, it is now possible to look at trends in 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) (Figure 28).

Although each cycle has its unique properties, there are prominent general paleontological patterns repeated in most cycles, which are well shown in cycles G7 and G3, that are common in Van Houten cycles in general (Olsen et al., 1989). The most obvious and least surprising pattern is seen in the correlation between the degree of fish preservation and the degree of lamination of the sediments. Microlaminated beds tend to preserve beautifully articulated fish, laminated mudstones produce disarticulated but still associated fish, and mudcracked mudstones have only dissociated scales and skull bones. This correlation almost certainly reflects the often-quoted dependence of fish preservation on a lack of oxygen, bioturbation, macro-scavenging, and physical disturbance.

Inversely correlated with this fish-preservation trend is one which at first appears very peculiar: a trend to <u>lower</u> fish diversity in the beds with the best fish preservation, and vice versa. This observation is based on the results from the Yale Quarry in Weehawken (Olsen et al., 1989) (Figure 37). This trend is all the more surprising because many more fish (individuals) are identifiable from the beds producing the best-preserved fish. The explanation seems to be that the highest diversity of lake environments tends to be near the shores, whereas the deeper-water zones tend toward low diversity (this is especially true for lakes with anoxic bottom waters, because they lack benthic forms). Thus, the cycle in fish diversity is a consequence of proximity to the shore which varies as a function of depth, which in turn controls the degree of lamination and absence of bioturbation.

The taphonomic pattern seen in the microlaminated division 2 of Van Houten cycles fits a chemically-stratified lake model (Bradley, 1929, 1963; Ludlam, 1969), in which bioturbation is perennially absent from the deeper parts of the lake bottom because the bottom waters lack oxygen, which is required by almost all macroscopic benthic organisms. Chemical stratification (meromixis) can arise by a number of mechanisms, but the main physical principle involved is the exclusion of turbulence from the lower reaches of a water column. This tremendously decreases the rate at which oxygen diffuses down from the surface waters, and retards the upward movement of other substances. The main source of water turbulence is wind-driven wave mixing. This turbulence usually extends down about one-half the wavelength of surface wind waves, which depends on the fetch of the lake, wind speed, and wind duration. If the lake is deeper than the depth of the turbulent zone, the lake becomes stratified with a lower non-turbulent zone and an upper, turbulently-mixed zone. The thickness of the upper mixed zone is also dependent on density differences between the upper waters (epilimnion) and lower waters (hypolimnion), which can be set up by salinity differences (saline meromixis) or by temperature differences, as seen in many temperate lakes. In the absence of saline or temperature stratification, chemical stratification can still arise in a deep lake with relatively high levels of organic productivity. Because oxygen is supplied slowly by diffusion, consumption by bacteria of abundant organic matter sinking into the hypolimnion plus oxidation of bacterial by-products eliminates oxygen from the hypolimnion. Lakes Tanganyika and Malawi in East Africa are excellent examples of very deep lakes in which there is very little temperature or density difference between the epilimnion and hypolimnion, but still chemical stratification occurs with the exclusion of oxygen below 200 m. Such a pattern is common in deep tropical lakes. The preservation of microlaminae and fossils in Van Houten cycles may have been a function of great water depth relative to a small surface area of the lake, which in the case of Lockatong lakes was none the less huge (in excess of 10,000 km^2); the depth, based on the area of the lake that must have been below the turbulent zone, was a minimum of 80 m during the deposition of the microlaminated beds (Olsen, 1990).

Thus, Van Houten cycles with a microlaminated division 2, such as G3 and G7 at Granton Quarry, and W5 and W6 at Stop 4, reflect the alternation of shallow, ephemeral lakes or subaerial flats with deep perennial lakes with an anoxic hypolimnion set up by turbulent stratification under conditions of relatively high primary productivity and low organic consumption (e.g. low ecosystem efficiency). The low organic content of divisions 1 and 3 of Van Houten cycles in general probably reflects higher ecosystem efficiency caused by shallow water depths, rather than lower total organic productivity.

This model also accounts for an exceptionally useful (for collecting) property of those cycles with a microlaminated division 2. Articulated small reptiles, such as *Tanytrachelos*, are found with predictable regularity in the basal few millimeters of the microlaminated unit. This pattern was first noticed in 1977 in Van Houten cycles in the upper member of the Cow Branch Formation of the Dan River basin (North Carolina and Virginia) (Olsen et al., 1978). The discovery of *Tanytrachelos* in the base of microlaminated units in this southern basin prompted a concerted search for reptiles in the homologous position in Van Houten cycles 500 km further north, in northeastern New Jersey. It took less than an hour for PEO to find the first skeleton, and that was at the locality described by Gratacap (1886) in Weehawken. Although Gratacap collected several hundred fish specimens from this site, no reptiles were found. That is, because without an appropriate model of <u>why</u> extra effort should be expended in those specific units, they get short shrift from the collector due to the presence of abundant fish in other parts of the unit. After the discovery of *Tanytrachelos* at Gratacap's locality, PEO informed Steven Steltz and James Leonard (two dedicated amateur collectors) about their predictable pattern of occurrence. The next time they visited Granton Quarry they found a complete *Tanytrachelos* (Figure 39) as well as pieces of other *Tanytrachelos* individuals, exactly where predicted. Prior to this, articulated *Tanytrachelos* had not been found at

Granton Quarry despite the fact the site had been a famous fossil locality for several decades.

With the sections on the east and west sides of the Palisade sill now combined, several larger-scale patterns emerge that are reinforced by data from elsewhere in the Newark basin. Looking at the distribution of fossil fish taxa through several cycles of different scales, a hierarchy of self-similar ecological patterns from the scale of the Van Houten cycle to the long modulating cycle is revealed. The full basic pattern of occurrence is, from oldest to youngest, *Semionotus*; *Semionotus* + *Osteopleurus*; *Osteopleurus* + redfieldiids (especially *Synorichthyes*); *Osteopleurus* + redfieldiids + *Turseodus*; *Turseodus* + redfieldiids, and finally just *Turseodus*. Most of this sequence can be seen in cycles W6 and W5 (Figure 37) at the Yale quarry in Weehawken, supplemented by data from the Eureka Quarry (Eureka, Pennsylvania). The pattern can also be seen within the short modulating cycle, e.g. N2 (which contains cycles W6 - W3). Thus, *Semionotus* is dominant in cycle W6, and *Osteopleurus* and *Turseodus* are dominant in cycles W5-W3. The basic pattern can be seen again in the McLaughlin cycle, such as the Nursery Member. *Semionotus* is abundant in short modulating cycle N2, near the bottom of the member, and *Osteopleurus* and *Turseodus* are dominant in short modulating cycles N3-N4. Finally, sequences of McLaughlin cycles, making up long modulating cycles, also show the pattern, with *Semionotus* being more common in the Princeton Member than in the Nursery Member, while *Semionotus* is virtually absent from the Ewing Creek Member. The distribution of fish taxa largely tracks the lithology of the fish-bearing units. More clastic units tend to be dominated by *Semionotus*, while more calcareous (or formerly calcareous) units tend to be dominated by *Turseodus*. In turn, the lithology is a function of paleoclimate, with more calcareous units being deposited higher in the cycles by more concentrated lake waters in drier times.

Interestingly, a very similar pattern is evident in Jurassic age strata of the Hartford basin (the Shuttle Meadow and East Berlin formations), Deerfield basin (the Turners Falls Formation, temporally equivalent to the East Berlin Formation), and probably also in the less well known Newark basin (Feltville, Towaco and Boonton formations). In these Jurassic age strata, *Osteopleurus* and *Turseodus* are absent, with *Redfieldius* being the only redfieldiid present. Again, *Semionotus* tends to occur low in the Van Houten, short modulating, and McLaughlin cycles.

The stratigraphic sequence in the Hackensack Meadowlands, underlying the Granton sill to the west of it'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 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. Notice the absence of prominent folding at the diabase-sediment contact. Because this sill is thin, and the

Palisade sill fairly remote, much of the sedimentary rock is not as metamorphosed as at previous stops; some cycles still have considerable organic matter.

One or two bedding plane thrust faults, always thrusting to the east, are present in division 2 of nearly every cycle at Granton Quarry. Slickensides are usually present and indicate that movement occurred parallel to dip. All the joint sets are cut by these thrusts, their displacement indicating that each fault has a net slip of 0.5 to 1.5 cm. This type of minor thrust fault is evident in virtually all Newark Supergroup lacustrine cycles and can be seen at every stop of this trip. The fact that all of these faults are thrusts requires post-depositional northwest-southeast shortening, a steepening of dip, and a $\sigma 3$ that would have been vertical. This is completely incompatible with the extension that produced the basins; thus we take these faults as evidence for structural inversion.

Continue east.

59.9 mi. Turn right onto Routes 1 and 9 south.

60.5 mi. Dip slope of Palisade sill mantled by northwesterly dipping Lockatong Formation. Contact appears concordant from here to Granton Quarry.

60.8 mi. Contact of uppermost Stockton with Palisade Sill directly on left (east). Stockton dips 15 ° NW and the contact dips 45° to 80° NW. Recrystallized arkose of Stockton Formation is sheared and dragged upwards close to contact with slikensides indicating down-dropping to the west. There are no indications of movement in the diabase, however. Basal Lockatong hornfels 100 m to the north, dip 15° NW and lie concordantly on the Palisade Sill. A possible interpretation of these exposures might be that the apparent movement in the Stockton occurred during intrusion of the sill.

61.5 mi. On left is open cut in Stockton Formation and west portal of tunnel for Penn Central Railroad. Cut exposes upper Stockton beds described by Darton (1883) and Olsen (1980b). At east end of cut at tunnel is excellent exposure of the contact between the Stockton Formation and the Palisade Diabase. Stockton beds dip at 15° NW while the irregular contact dips at 60° NW. This contact is locally welded according to Darton (1890) and Lewis (1908). This appears to be a continuation of the contact surface exposed at the Penn Central tunnel described above and this is another exposure where the Palisade sill is described as having a dike-like appearance. Actually, these exposures and similar ones near by are perhaps better explained by a local stepping up of the Palisade sill as described by Olsen (1980b).

62.5 mi. Turn right onto entrance ramp for New Jersey Route 3 (and New Jersey Turnpike) and proceed straight ahead onto New Jersey Route 3 west..

64.3 mi. Ridge underlain by more resistant strata in proximity to the Graters Member. Outcrops of this member occur further south at abandoned diabase quarry at Laurel Hill.

64.9 mi. Crossing over Hackensack River. At one stage or another the Hudson probably flowed along this route (Lovegreen, 1974).

66.9 mi. Exit right for service road.

67.1 mi. Turn right onto Orient Way: becomes Rutherford Avenue.

68.5 mi. Turn left (south) onto Polito Avenue.

68.8 mi. Turn right into parking lot for Ellen Tracy outlet store.

Stop 6. Bluff and Copper Prospect at Lyndhurst, New Jersey

Latitude and Longitude: 40°48.434'N, 074°06.506'W
Tectonostratigraphic Sequence: TS III
Stratigraphic Unit: Passaic Formation
Age: Late Norian (middle Late Triassic); 212 Ma
Main Points:

1. Part of ridge representing wet phase of long modulating cycle P4
2. Wetter facies better cemented
3. Kilmer Member of Passaic, marginal lacustrine facies
4. Vague pattern of Lockatong cycles
5. Accumulation rate similar to cored areas
6. Copper hosted in sandstones
7. Drier facies similar to more basinward facies with gypsum nodules
8. Footprints abundant in lake margin facies
9. Dinosaurs becoming larger and more abundant
10. Crurotarsans still very abundant

The Meadowlands are bordered on the west by a distinct ridge that extends from Hackensack to Kearny (NJ). For the most part this ridge is characterized by a heterogeneous assemblage of red mudstones and sandstones. However, there are a few purple and gray units present; based on their stratigraphic position they are part of long modulating cycle P4. The most eastern gray and purple unit probably is the lower part of the Kilmer Member (Figure 14).

Stop 6 is located on the eastern side of this ridge (Figure 49) and exposes the uppermost part of member T-U and the lower half of the Kilmer Member (Figure 50). 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, so much so that during mapping of this member its surface trace became known as the "dead zone" because plants have trouble growing on the copper-rich regolith. There is surface evidence of copper prospect pits where and unnamed brook crosses the Kilmer Member in Piscataway (NJ), but these are not mentioned by either Lewis (1907) or Woodward (1944).

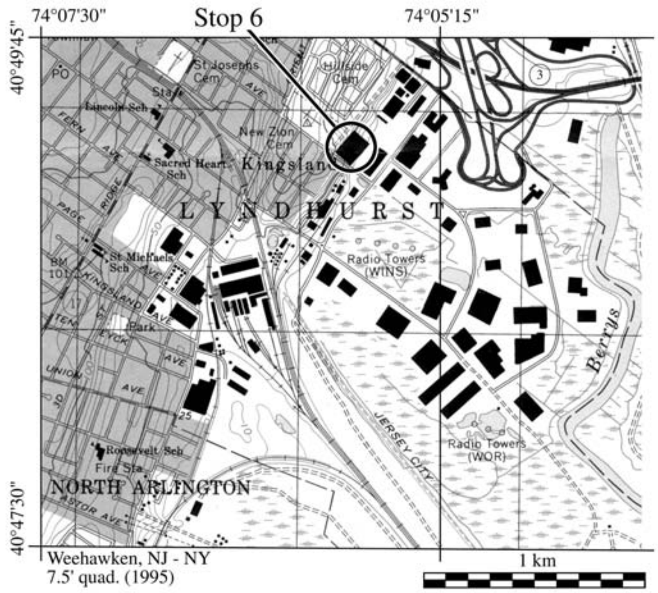

Figure 49. Map for Stop 6, Lyndhurst, New Jersey.

Figure 50. Exposures of copper-bearing white and tan sandstone surrounding purple siltstone of lower ?Kilmer Member at Stop 6, Lyndhurst, New Jersey.

In the 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 km 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). According to Woodward (1944), the Schuyler copper was discovered a few years prior to 1719 and an extensive mine was developed there, remaining profitable on and off until nearly 1830. In addition to copper it also produced a small amount of silver and trace amounts of gold. Possibly the first steam engine to arrive in North America (1753) was used to drain the mine, but the machine was damaged by fire in 1765, 1768, and 1773 (Woodward, 1944). There were several attempts at reopening the mine through the latter part of the 19th and earliest 20th centuries, but by 1903 the mine had been abandoned because of various engineering problems, and no doubt the much greater profitability of the copper deposits in the central and western United States.

The exposures at this stop may be the prospect mentioned by Woodward (1944) on the Kingsland estate, 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) (Figure 50).

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 T-U and the Kilmer member in the NBCP cores.

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). Although the exact location was not recorded, the lithology of the footprint slabs is consistent with the local expression of the Kilmer Member. The assemblage is distinguished in the Newark basin by the presence of relatively large grallatorid footprints, up to the size of *Anchisauripus tuberosus*; this is the oldest level in the basin with such tracks (Figure 51). The nomenclatural problems associated with these kinds of footprints are discussed in the text for Stop 7. Relatively large examples of the ornithischian dinosaur track, *Atreipus milfo*rdensis, are present, along with a new dinosaurian ichnogenus ("*Coelurosaurichnus*" sp. of Olsen and Flynn, 1989), as well as the non-dinosaurian *Brachychirotherium parvum*, and *Rhynchosauroides* sp.

Rhynchosauroides brunswickii and *Grallator* sp. were found specifically at this site by PEO during the early 1970s. PEO also found *Kouphichnium* sp., made by horseshoe crabs, as well as *Scoyenia* burrows. These trace fossils were all found in the red siltstones immeadiately above the gray sandstones.

Further southwest, (0.3) mi are additional exposures along former Erie Lackawanna Railroad tracks showing an unusual reverse fault dipping to the west and downthrown on the east. Slickensides confirm the dipslip nature of the fault.

About 2.9 km west of here on the west bank of the Passaic River are what were the Avondale and Belleville quarries that produced a large amount of building stone ("brownstone") to the region (based on location of quarries shown by Darton et al (1908). These quarries are now the location of Father Glotzbach and Monsignor Owens parks in Nutley (Avondale) New Jersey. These quarries produced a fragmentary phytosaur skull (Edwards, 1895; Lull, 1953; Baird, 1986b) as well as dinosaur footprints (Woodworth, 1895) and fragmentary plant remains (Nason, 1889). Unfortunately the whereabouts of the dinosaur tracks are unknown. The stratigraphic level of the quarries appears to be close to the Cedar Grove Member and almost certainly with in long modulating cycle P6.

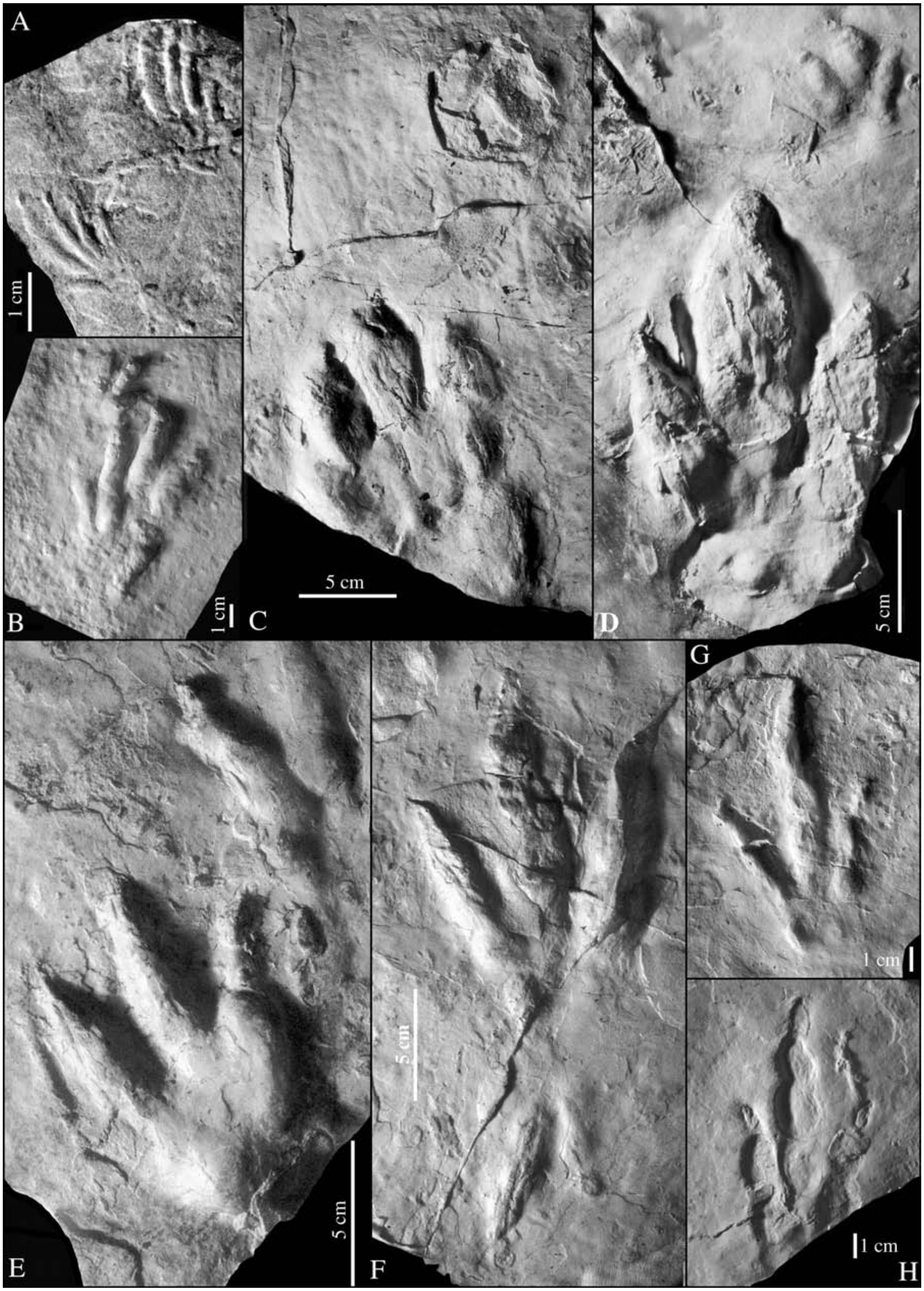

Figure 51. Footprints from Lyndhurst or vicinity, Lawrence Blackbeer collection (all polysulfide casts): **A**, lepidosauromorph track *Rhynchosauroides* sp; **B**, probable rauisuchian track *Brachychirotherium* sp.; **C**, probable rauisuchian track *Brachychirotherium parvum*; **D**, ornithischian dinosaurian track *Atreipus milfordensis*; **E**, ?saurischian dinosaurian track "new genus 1"; **F**, theropod dinosaur tracks *Anchisauripus* sp. (above) and *Grallator* sp. (below); **G**, theropod dinosaur track *Grallator* sp.; **H**, theropod dinosaur track *Grallator* cf. *G. parallelus*.

Return to Polito Avenue and turn left, heading north.

69.2 mi. Turn right onto Rutherford Avenue.

69.5 mi. Turn left onto ramp for Route 3 east (and New Jersey Turnpike).

69.9 mi. Enter onto Route 3.

70.6 mi. Exit for New Jersey Turnpike. Take New Jersey Turnpike (Western Spur) south.

75.4 mi. Exit 15W for Interstate Route 280. Take Route 280 west.

78.5 mi. Crossing Passaic River.

78.5 mi. Underpass for North 6th Street. Outcrop on right is a gray sandstone of the Cedar Grove Member. According to Neal K. Resch (quoted in Cornet (1977) The gray zone is roughly 4.8 m thick, the basal beds of which consists of 90 cm of black laminated shale passing upward into gray shale. This is followed by 60 cm of tan sandstone and 3 m of thinly bedded gray sandy shale and sandstone, grading upward into red sandstone. During the construction of Route 280, Resch collected conchostracans and leafy shoots of *Brachyphyllum* from the shale and various reptile footprints, including *Apatopus* sp.. *Brachychirotherium* sp., *Grallator* sp., and Anchisauripus sp. from the well bedded siltstone. Cornet (1977) described a palynoflorule from gray shale (sample SPPS-23A) indicating a probable early Rhaetian age. This unit was originally thought to be part of the Ukrainian Member, but Witte et al. (1991) show that the red beds directly overlying this gray sequence are of reversed polarity indicating correlation with the Cedar Grove Member.

80.6 mi. Exposures of Passaic Formation in cuts for connections to the Garden State Parkway.

83.4 mi. Exposures on left have produced a partial trackway of *Apatopus* sp., a phytosaurian trackway figured in Baird (1986b). This is amongst the youngest *Apatopus* know dating from about 500 ky before the Triassic-Jurassic boundary.

84.1 mi. Latitude 40°47.483'N, longitude 074°14.923'W. Huge cut in Orange Mountain Basalt - type section of the formation (Olsen 1980a) (modified from Olsen and Schlische in Olsen et al. (1989).

When this road was constructed in 1969, this 33 m high roadcut was the deepest federally financed highway cut east of the Mississippi River (Manspeizer, 1980). A complete section of the lower flow unit of the Orange Mountain Basalt is exposed (Figure 52) (Olsen, 1980a,c; Manspeizer, 1980; Olsen et al., 1989). Unusual curved patterns of columnar joints are present in the cut, described as chevrons, oblique and reverse fans, and rosettes. This flow overlies red beds of the Passaic Formation which contains the Triassic-Jurassic boundary 10 m or so below the basalt contact. During the construction of this cut in 1969, PEO found several examples of *Batrachopus* sp. just beneath the contact with the Orange Mountain basalt.

At this outcrop of the type section, the west-dipping lower flow of the Orange Mountain Basalt is almost completely exposed (Olsen, 1980a; Puffer, 1987). The Orange Mountain Basalt is an HTQ basalt (Puffer and Lechler, 1980) which at this exposure shows a complete, beautifully displayed Tomkeieff (1940) sequence (Olsen, 1980a; Puffer, 1987) almost exactly comparable to Long and Woods (1986); Type III flows. The thin (6 m) lower colonnade is fine-grained with large columns, the entablature is thick (35 m) with very well-developed curvicolumnar jointing, and the upper colonnade (10 m) is massive with poorly-developed columns.

According to Lyle (2000) the curvicolumnar jointing of the type seen at this outcrop are a direct result of ponded water on top of the cooling flows. The curved columns result from joints forming normal to large widely-spaced vertical master fractures in a hexagonal array that form very early in the cooling of the basalt flow. Water percolates down the mater fractures and provides a surface cooling faster than adjacent basalt resulting in surface-normal hexagonal fractures that propagate away from the master fractures and curving towards the upper cooling colonnade of the flow. The master fractures can be clearly seen separating bowl-shaped fans of the curvicolumnar jointing. Additional bowl-shaped fractures, as seen here may also have helped to control the radiating pattern of columns.

The largest fault in this exposure strikes N05°E, dips 80°E, and does not visibly offset subhorizontal cooling joints, suggesting that it is a predominantly strike-slip fault (Schlische, 1985). Another fault with a 25-cm-wide breccia and gouge zone strikes N30°E and dips 70°NW. Although slickensides suggest that the last slip on the fault was predominantly dip-slip, this fault may have originated as a normal fault during NW-SE extension. Several minor slickensided surfaces, mostly reactivated cooling joints in the basalt, show evidence of an earlier period of predominantly normal-slip and a later period of predominantly strike-slip (Schlische, 1985)

The Orange Mountain Basalt was probably fed by the Palisade sill (Ratcliffe, 1988) and it (and correlative units in other basins) represents the oldest of the known CAMP lava flows, flowing out less than 20 ky after the Triassic-Jurassic boundary (Olsen et al., 1996b).

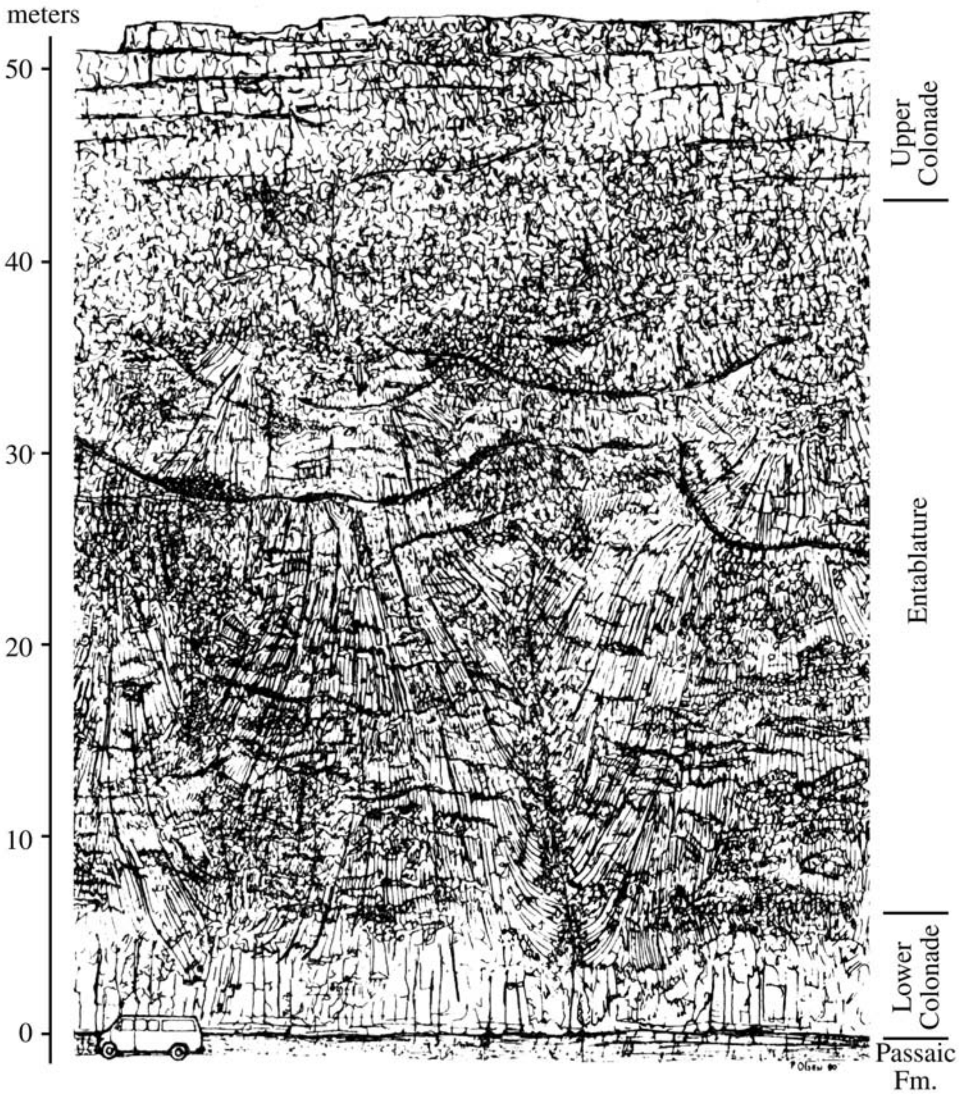

Figure 52. Composite drawing of the type section of the Orange Mountain Basalt along Interstate Route 280 in East Orange, New Jersey. Traced from a composite of a continuous series of photographs with the dip removed and compiled vertically (Olsen, 1980c).

84.9 mi. Exposures of ?second flow of Orange Mountain Basalt.

85.2 mi. Unexposed contact between Feltville Formation and underlying Orange Mountain Basalt below this point.

85.6 mi. Latitude 40°47.946'N, longitude 074°16.302'W. Contact between Preakness Basalt and Feltville Formation poorly exposed on south side of road and type section of the Preakness Basalt in deep open cut for Route 280 (modified from Olsen and Schlische in Olsen et al. (1989).

This type section (Olsen, 1980a.c) exposes about 100 m of the lowest flow of the Preakness Basalt, the thickest extrusive multiple flow unit in the Newark basin section (Figure 53). It is a high-Ti, high-Fe quartz-normative tholeiite indistinguishable from the Holyoke Basalt of the Hartford basin. The lowest flow is characterized by a thin lower colonnade; a massive, thick, and coarse-grained entablature with very characteristic splintery columns; a massive upper colonnade; and a comparatively thin vesicular flow top. It does not closely resemble any of the flow types of Long and Wood (1986), but it is similar to the second flow of the Sander Basalt of the Culpeper basin, the Holyoke Basalt of the Hartford basin, and some parts of the North Mountain Basalt of the Fundy basin.

The splintery columns are defined by what Faust (1977) calls a platy prismatic joint system, and it is characteristic of the lower flow throughout the areal extent of the Preakness Basalt. The joint system is not hexagonal and does not appear to be a faulting phenomena (although faulting does exaggerate it). Its origin remains a mystery.

Several faults cut the rocks at this exposure (Schlische, 1985) but most are now covered with gunnite for the safety of traffic. The largest fault zone strikes 005°, dips 85°E and is marked by a gully 1.5 m wide. The zone of brecciation varies in width from 40 to 60 cm. Slickenlines are curved from more steeply-plunging to more shallow-plunging attitudes. Another major fault strikes 035° and dips 83°NE and is marked by a breccia zone (with clasts 2 to 5 cm in size) 35 to 40 cm wide. Slickenlines rake 35°N. Three other macroscopic faults have orientations 010°, 82°NE; 040°W, 82°NE; and 020°, 83°NE; brecciated zones are 10 to 15 cm wide. A systematic study of minor fault populations showed that, although most of the faults followed pre-existing joints within the basalt, the slickenside striae strongly indicated strike-slip on NW-striking planes (Schlische, 1985), consistent with the larger faults at this exposure.

At this locality the contact (now overgrown) with the underlying Feltville Formation is simple, with massive basalt of the lower colonnade in direct contact with sedimentary rocks. At other localities thin flow units, pillow lavas or rubble flows are present (Olsen, 1980a).

89.0 mi. Take exit 4A onto Eisenhower Parkway south paralleling to the south the ancient course of the Hudson. Entering former course of Hudson River (Johnson, 1931).

89.6 mi. Intersection with Eagle Rock Avenue.

89.9 mi. Intersection with Nob Hill Road. Outcrop on left is of the two flows of the Hook Mountain Basalt and underlying upper Towaco Formation described in Olsen et al. (1989, Stop 6.3).

90.1 mi. Turn left onto Beaufort Avenue.

90.5 mi. Turn left onto access road for the Essex County Art Park and drive up the dip slope of the Hook Mountain Basalt..

91.0 Park in uppermost lot of Essex County Art Park. Follow path down to Walter Kidde Dinosaur Park.

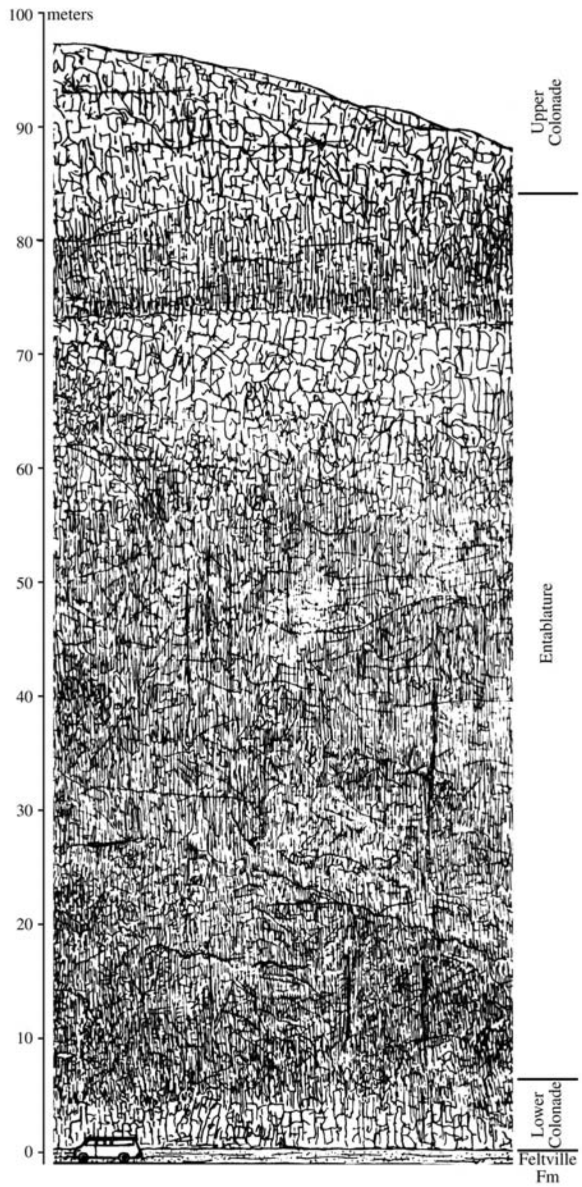

Figure 53. Composite drawing of the type section of the Preakness Basalt along Interstate Route 280 about 2.25 km west of the type section of the Orange Mountain basalt (Figure 52). Traced from a composite of a continuous series of photographs with the dip removed and compiled vertically (Olsen, 1980c).

Stop 7. Walter Kidde Dinosaur Park, Riker Hill, Roseland (New Jersey)

Latitude and Longitude: 40°48.920'N, -074°19.550'W
Tectonostratigraphic Sequence: TS IV
Stratigraphic Unit: Towaco Formation and Hook Mountain Basalt
Age: middle Hettangian (Early Jurassic); ~202 Ma
Main Points:

1. Long section of upper Towaco Formation: type section
2. Very thick Van Houten cycles
3. Peak wet phase of long modulating cycle H1 and Laskar cycle LaCV1.
4. But dry phase of 404 ky cycle
5. Footprints very common in lake margin sheet deltas
6. Typical earliest Jurassic assemblage
8. Extraordinary preservation
7. Includes level of Dinosaur State Park in Hartford basin
8. Level of unique assemblage of track types, including synapsids
9. Association with waning phases of the CAMP

The discovery of dinosaur footprints in the "Riker Hill" (aka "Roseland") quarry - part of which is now the Walter Kidde Dinosaur Park (Figure 54), was first reported in the local newspapers of Livingston and Roseland, NJ about 1968 (Figure 55). The quarry occupied a 55-acre tract on the northeast side of Riker Hill in Roseland, and was owned by the Walter Kidde Company, Inc. Over the next few years the Riker Hill quarry became locally very well known for its abundant reptile footprints, and in 1971 the owners agreed to give the most fossil-rich portion of the tract to the Essex County Department of Parks and Recreation. The resulting publicity made the site internationally famous. In 1977 the 17 acres of the present Walter Kidde Dinosaur Park was formally donated to Essex County (Figure 54) and today the park remains one of the premier sites for Jurassic age fossils in eastern North America. In this paper the 55-acre tract will be referred to as the Riker Hill quarry, and the term Walter Kidde Dinosaur Park will be used for the 17-acre portion of the Riker Hill quarry that is now the county park. This stop has been described in part by Olsen (1980b) and Olsen et al. (1989), and described in detail by Olsen (1995).

Although thousands of footprints have been found in the Riker Hill quarry, only a tiny fraction have made it into museum collections. Presently, the Walter Kidde Dinosaur Park is administered by the Center for Environmental Studies of the Essex County Department of Parks, Recreation and Cultural Affairs *Prospecting and collecting are no longer allowed at the Dinosaur Park without permission from the Center for Environmental Studies.*

The Towaco Formation at the Riker Hill quarry (the type section; Olsen. 1980a, c) consists of relatively fine-grained red, gray, and black units, mostly mudstone and fine sandstone. This facies represents some of the more basinward deposits of the Newark basin Jurassic, although the sequences deposited near the geographic center of the basin have been lost to erosion. The Hook Mountain Basalt as seen in the park is representative of most of its preserved extent.

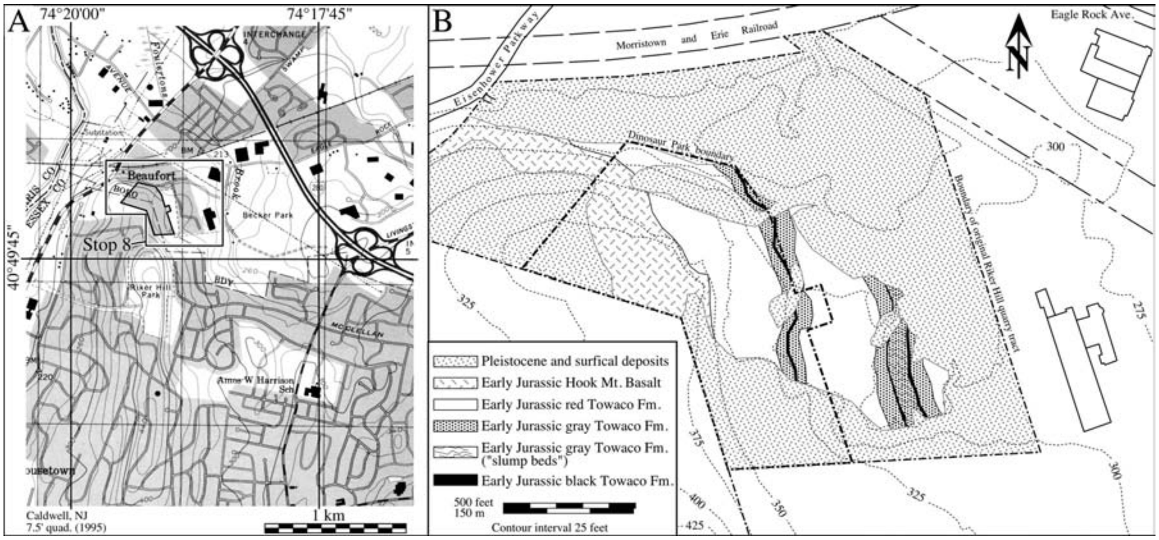

Figure 54. Walter Kidde Dinosaur Park, Stop 7, site maps: A, topographic map showing vicinity of park; B, geological map of park and surrounding area (from Olsen, 1985).

Prior to the development of the Nob Hill complex on what was the east side of the Riker Hill quarry, the exposed section below the Hook Mountain Basalt consisted of the uppermost red beds of one Van Houten cycle (RVH-1), two complete gray and black shale-bearing Van Houten cycles (RVH-2 and RVH-3), and the lower part of an entirely red fourth cycle (RVH-4) (Figure 56). Together, these cycles constitute most of a short modulating cycle (~100 ky duration). Presently, only the uppermost beds of RVH-3 and RVH-4 are exposed, representing less than ~40 ky of Jurassic sedimentation in the drying phase of a 100 ky short modulating cycle, which is in turn the drying phase of a 404 ky McLaughlin cycle, itself in the wet phase of a 1.75 my long modulating cycle (Figure 4). In the following description of the paleontology, all of the fossils will be keyed into the section shown in Figure 56 so their positions within the pattern of cyclically-shifting climate can be seen.

Varied assemblages of plant and animal remains have been found in the Riker Hill quarry. Most famous are the reptile - notably dinosaur - footprints, but well-preserved plants, fish and even insect body fossils have been found as well. As in the Towaco Formation in general, trace fossils (mostly tracks and burrows) are abundant in the red and gray beds of divisions 1 and 3 of the Van Houten cycles, while insects, fish, plants, and pollen and spores are restricted to the gray and black beds of division 2. Apart from a single tooth fragment and a coprolite, all of the fossils of tetrapods from the Walter Kidde Dinosaur Park are trace fossils.

Pollen assemblages recovered from units 9-10are, as is typical for the Early Jurassic of tropical Pangea, dominated by the extinct conifer genus *Corollina* (*Classopolis*), accompanied by various other conifer, cycadophyte, and fern spores (Figure 57) (Cornet, 1977). Remains of cheirolepidiaceous conifers are the most common macroscopic plant fossils in all facies at Riker Hill. Compressed wood and fossil charcoal almost certainly belonging to these plants are found in all of the gray and black beds at the, and large roots are present in the gray siltstones and sandstones of unit 16 of division 1 of cycle RVH-3. Leaf and shoot compressions and cone fragments are similarly present in all the gray beds, and well preserved material occurs in units 9 and 10 of division 2 of cycle RVH-3, which is still exposed. Impressions of leafy shoots and clay casts of roots and stems are common in the red units, sometimes on the same surfaces as footprints (Figure 57). *Imponoglyphus torquendus* was originally described from Late Triassic age strata of the former Soviet Union and an example of this form taxon has been described from Walter Kidde Dinosaur Park by Metz (1984). This form species consists of impressions resembling truncated cones fitted into one another. *Imponoglyphus torquendus* is almost certainly an impression of a conifer shoot similar to that shown in Figure 57. From the extreme dominance of the remains of cheirolepidiaceous conifers in all facies at the, both as macro- and microfossils, it is clear that the woodlands and scrub lands of the Newark basin in Towaco time were strongly dominated in biomass by the cheirolepidiaceous conifers. However, it is not yet possible to tell how many biological species of cheirolepidiaceous conifers are represented, and the species diversity could be quite low.

A rich invertebrate trace fossil assemblage has been recovered over the years from the Riker Hill quarry, mostly from strata exposed in Walter Kidde Dinosaur Park (Figure 58). As described by Metz (1992), *Cochlichnus anguineus* consists of smooth, narrow

Figure 55. Riker Hill Quarry (Stop 7) ca. 1969: a, contact between Hook Mountain Basalt (above) and Towaco Formation (type section); b, top of gray bed, unit 6; c, base of gray bed unit 15-16 (Figure 56).

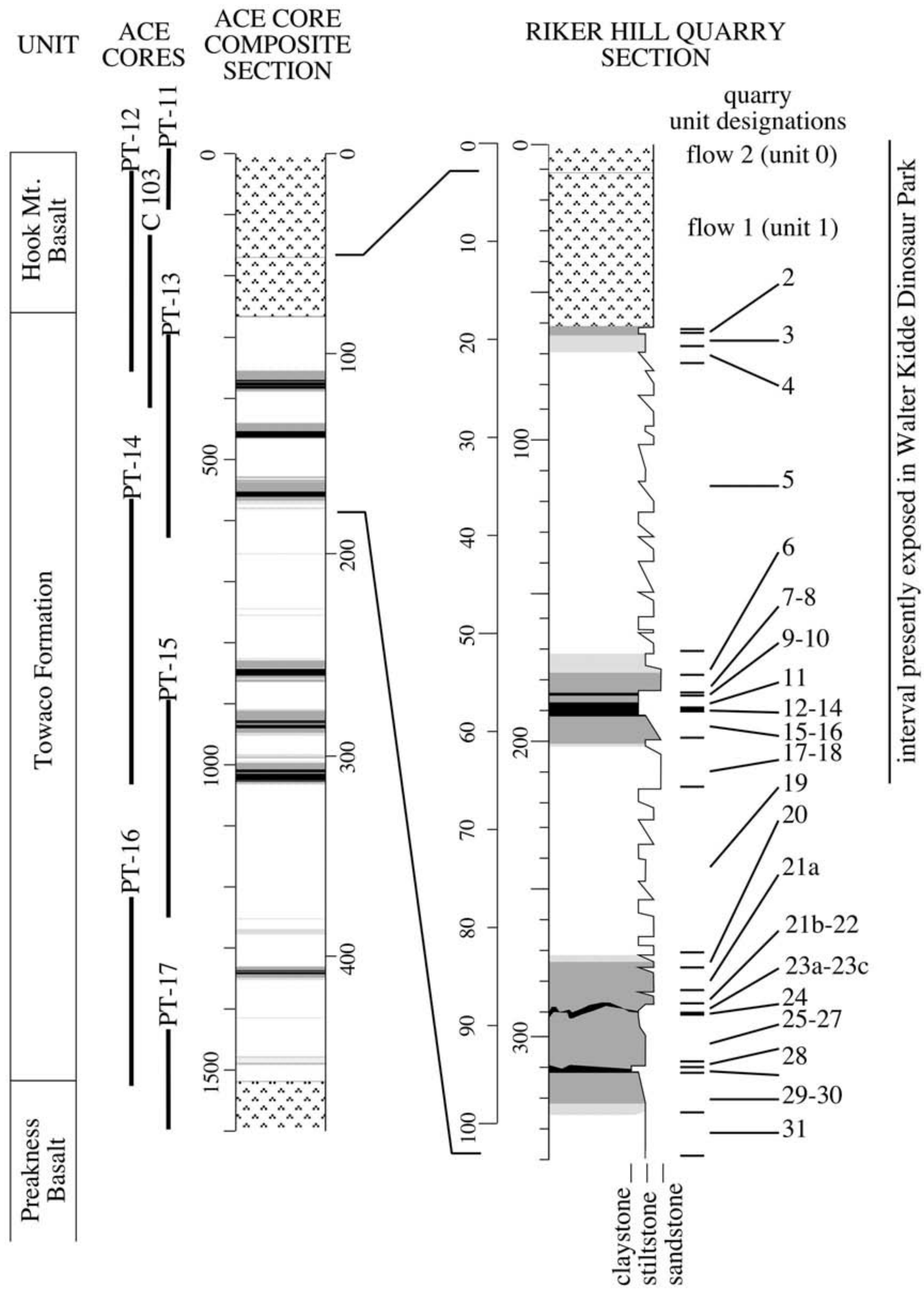

Figure 56. Figure 4. Section in old Riker Hill quarry compared to reference section of the Towaco Formation from the Army Corps of Engineers cores (Olsen et al., 1996b). Unit designations are referred to in text. Key to lithologies as in Figure 3. From Olsen (1995).

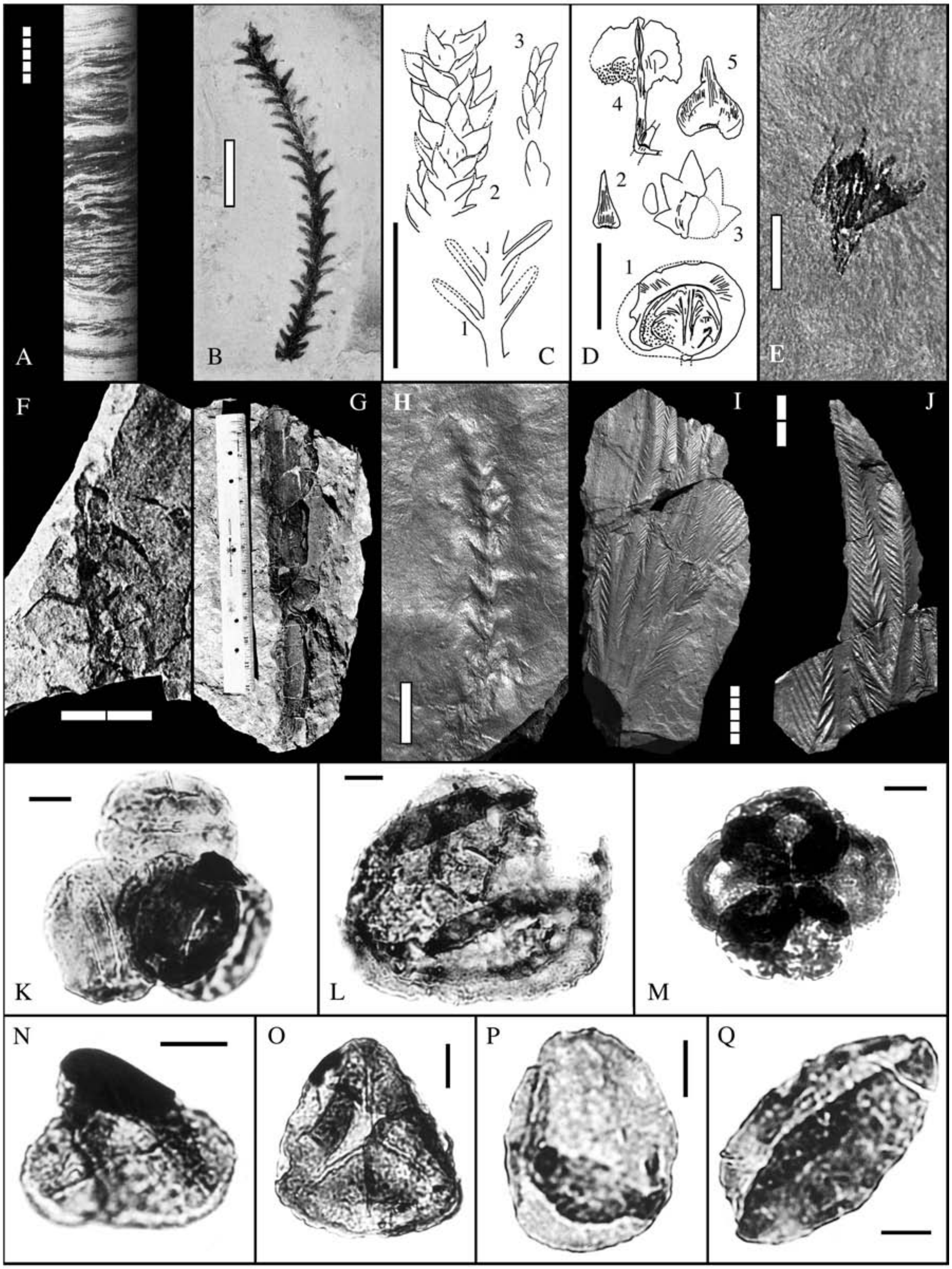
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Figure 57. Figure 6. Molecular and plant body (organ) fossils from the Riker Hill Quarry. **A**, Portion of Army Corps of Engineers core of the upper Towaco Formation showing oil staining in a pale gray fine rippled sandstone (molecular fossil) (scale, 5 cm). **B**. *Pagiophyllum* sp.- example of compression from Portland Formation; similar examples of conifer shoots have been found at the Dinosaur Park, but none have been photographed or archived in museum collections (scale, 1 cm). **C**, Conifer shoots from the Towaco Formation (adapted from Olsen and others (1989) (scale, 1 cm): 1, *Pagiophyllum* sp. 7a, from the Dinosaur Park (unit 10); 2, *Pagiophyllum* sp. 8p from the middle Towaco Formation of Pompton, NJ; 3, *Pagiophyllum* sp. 5p, equivalent unit to gray part of cycle RVH-3, from Toms Point, Lincoln Park, NJ. **D**, Conifer Cone parts (adapted from Cornet, 1977) (scale, 1 cm): 1, cone scale seed complex from gray part of cycle equivalent to RVH- 3 in Hartford basin, Rt. 9/91 road cut, Cromwell, CT; 2, Cone scale bract, unit 10, Dinosaur Park; 3, partly reconstructed cone scale bearing ovule, unit 10, Dinosaur Park; 4, cone scale bract and seed(?), unit 10, Dinosaur Park; 5, cone bract, unit ?10, Dinosaur Park (scale, 1 cm). **E**, partial cone scale bract complex, unit 10 (scale, 2 m), Riker Hill quarry; **F**, fragmentary cone axis, unit 10, Dinosaur Park (scale, 2 cm).. **G**, small compressed log, gray part of cycle RVH-2, Riker Hill Quarry (ruler is 1 ft). **H**, impression of *Brachyphyllum* shoot, upper unit 5, Dinosaur Park (scale, 1 cm). **I**, large set of rill marks (not a fossil) (scale, 5 cm); **J**, small piece of rill marks showing fine detail (scale, 2 cm). **K-Q**, pollen and spores from unit 9-10, courtesy of Bruce Cornet (pers. com., 1994), (scale, 10 microns) Dinosaur Park: **K**, cheirolepidaceous conifer pollen, *Corollina meyeriana*, tetrad; L, possible araucarian pollen, *Araucariacites australis*; M, cheirolepidaceous conifer pollen, *Corollina meyeriana*, tetrad, **N**, spore of fern *Clathropteris*, *Converrucosisporites cameronii*; **O** and **P**, fern spore, *Dictyophyllidites* sp.; **Q**, cycadophyte pollen, *Cycadopites* sp. From Olsen (1995).

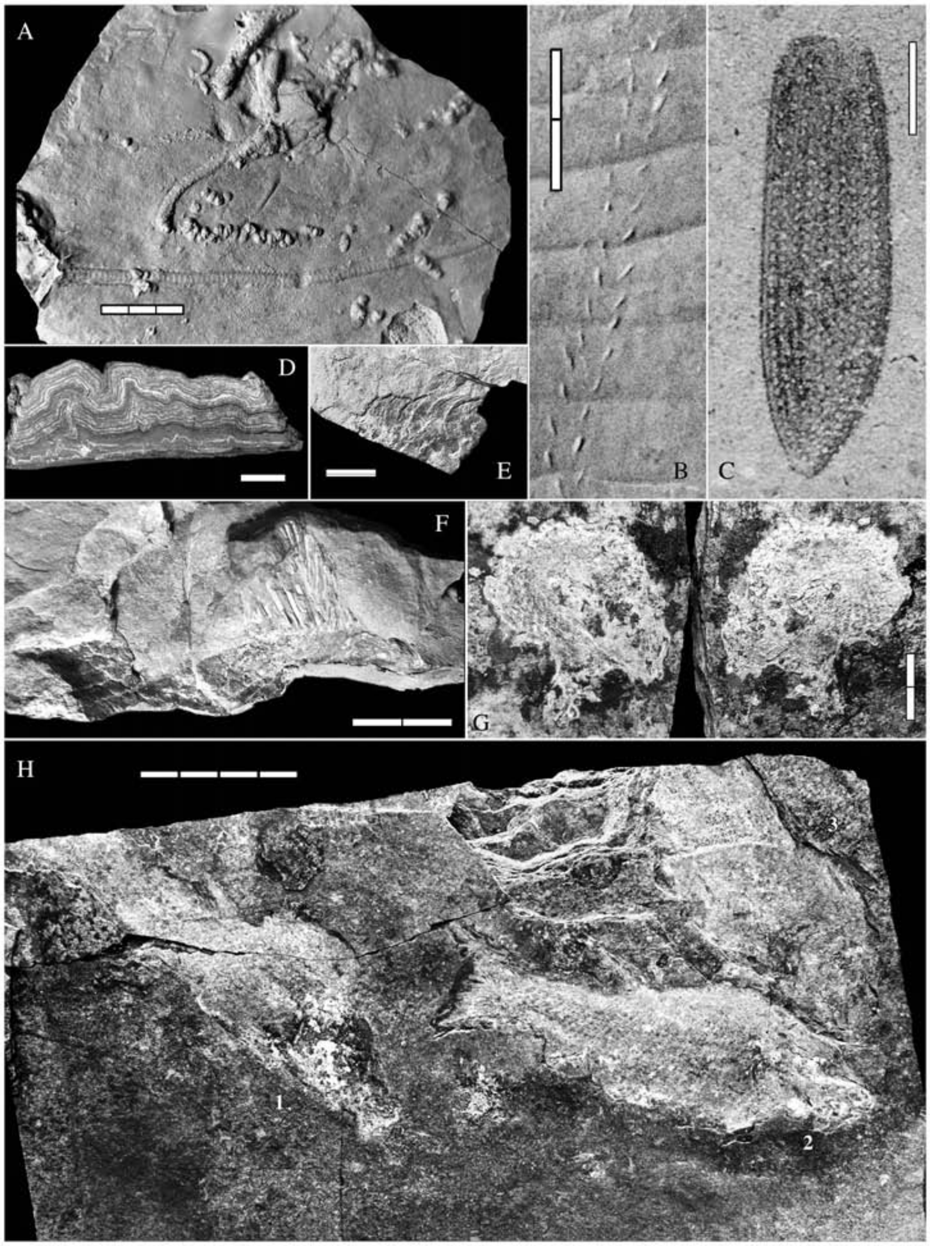

Figure 58. Invertebrates and fish from the Riker Hill Quarry. **A**, unusual, unidentified burrow (scale, 3 cm) from unit 5, Dinosaur Park; **B**, insect walking trace, *Acanthichnus* sp., from low in unit 5, Dinosaur Park; **C**, beetle elytron (wing cover), *Liassocupes* sp., unit 10, Dinosaur Park (specimen in YPM collection) (scale, 4 mm); **D**, Folded microlaminated D, folded microlaminated -carotane-bearing black shale of unit 13 (scale, 1 cm); **E**, fragment of back of fish with dorsal ridge scales of *Semionotus tenuiceps* group semionotid (scale, 1 cm); **F**, Three dimensional example of indeterminate *Semionotus* sp., from unit 23b, Riker Hill quarry (scale, 2 cm). **G**, Curled up. part and counterpart of Semionotus sp. from unit 23c (YPM 6472). **H**, three *Semionotus* from unit 23b, Riker Hill quarry; one on left is of the *Semionotus tenuiceps* group, while the other two are indeterminate.

(1.5-2 mm), sinusoidal, unlined, and unbranching horizontal burrows possibly made by nematodes or perhaps fly larvae. *Helminthopsis* sp. (Metz, 1991) is a smooth, straight to gently winding burrow of constant width that does not show sediment layer crossings. It may possibly have been produced by a worm-like form. Metz (1992) has described two species of *Planolites*, a form genus comprised of small, horizontal or inclined filled burrows lacking exterior ornament or interior structure. *Planolites montanus* is a very small form (1-1.5 mm) having occasionally-branching burrows, often curved, filled with material coarser than the matrix, and with crossovers and interpenetrations (Metz, 1991). *Planolites beverleyensis* is a larger (5-6 mm) burrow that is similar in form and filling to *P. montanus*, but shows discontinuous rings where the burrow tapers (Metz, 1991). A perhaps similar form is shown in Figure 58; in this example a burrow which shows distinct annuli is present, which apparently was broken up at one end, releasing pellets which were scattered by a weak current. In general, however, *Planolites* is a catch-all taxon with few defining characters, that could have been made by variety of worms or even arthropods. *Trepyichnus bifircus*, as described by Metz (1991), consists of a "... straight to curved trace (1 mm in diameter), with short extensions (1 mm - 2 mm) possessing slightly thickened terminations projecting from junctures between longer segments, creating a zigzag pattern". The originator of this kind of trace is unknown. *Biformites* sp. and *Fustiglyphus roselandensis* are possibly related to *Trepyichnus* described by Boyer (1979) from the Riker Hill quarry. *Fustiglyphus* consists of two kinds of linked trails: a thin (0.4-0.6 mm) part 4-7 mm long with distinct annulae and a thicker (2-2.5 mm) part 2 to 3 mm long with faint annulae . There is a faint groove running down the middle of the trace. *Biformites* is a tapering trace with annulae and a faint longitudinal groove. According to Boyer (1979), the Roseland *Fustiglyphus* is a succession of repeated *Biformites*-like traces, probably produced by a small arthropod seeking refuge in a deteriorating environment. Metz (1992) has described *Scoyenia gracilis* from Walter Kidde Dinosaur Park, where it is not common. *Scoyenia* is a lined burrow with a meniscate filling and distinct rice-grain-like prod marks on the outside surface. This form genus is the most common trace fossil in deposits of Triassic age in the Newark basin (and Newark Supergroup). It is markedly more rare in the Newark basin Jurassic, notably so in the Towaco Formation. There is little consensus on the makers of *Scoyenia* with opinions ranging from polychaete worms (D'Alessandro et al., 1987), to insects (Frey and others, 1984), and crayfish (Olsen, 1988). This taxon is very badly in need of detailed study.

Insect body fossils are represented by a single beetle elytron (wing cove) from unit 10. The narrow rows of punctures between ridges and the general shape of the elytron distinguishes the beetle family Cupedidae, hence the common name "reticulated" beetles. The elytron from Walter Kidde Dinosaur Park most closely resembles the Early Jurassic genus *Liassocupes* Whalley 1985 (Figure 58, Huber et al., 2002). The cupedids are often regarded as the most primitive of the beetle families. The family is extant; both the larvae and adults feed on rotting wood. *Cupes concolor* is the most common living member of the family in the United States and is very similar to the Walter Kidde Dinosaur Park fossil. This isolated elytron is illustrative of how incomplete our sampling of Early Jurassic life is. Then as now, beetles were probably the most diverse insect group, and insects the most diverse animal group. The lack of insect fossils is probably

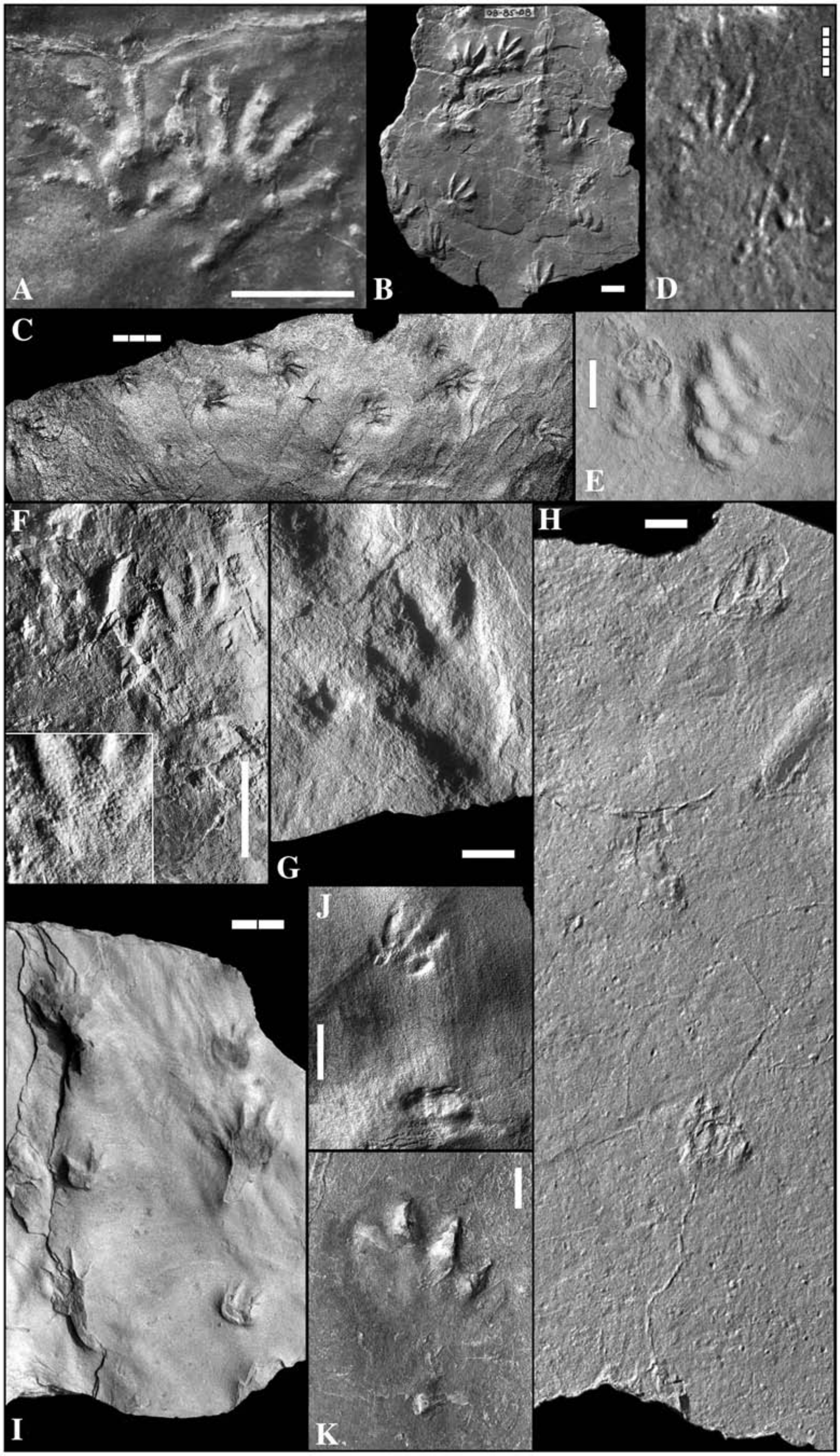

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Figure 59 Quadrupedal tracks from the Riker Hill quarry. **A**, Right manus-pes set of *Ameghinichnus* n. sp. from upper unit 5, Dinosaur Park (AMNH 29298, collected by Larry Felder, 1978). Manus is on left and pes is on right. **B**, Drawing of trackway of *Ameghinichnus* n. sp. Arrow shows manus pes set in A (above). **C**, Leptodactylus trackway of *Ameghinichnus* from upper unit 5, Dinosaur Park (Charles Rizzo collector, Charles Rizzo collection) (scale, 3 cm). **D**, Natural cast of deep underprints of at least one trackway of *Ameghinichnus* from upper unit 5, Dinosaur Park (Charles Rizzo collector, Charles Rizzo collection). **E**, Right manus-pes set of *Rhynchosauroides* n. sp., from uppermost Towaco Formation, Lincoln Park, NJ (John Colegrande collector, PU 18563) (scale, 5 mm). Manus is very faint, below and to right of hand-like pes. E, manus pes set of Batrachopus deweyii, xxxx collection. **F**, Natural cast of right manus-pes set. with at least one other additional superimposed pedes, of *Batrachopus* sp., unit 5, Dinosaur Park (Charles Rizzo collector, Charles Rizzo collection) (scale, 1 cm). Inset shows detail of scale impression on manus. **G**, Underprint of manus-pes set of *Batrachopus* sp., unit 5, Dinosaur Park (Charles Rizzo collector, Charles Rizzo collection) (scale 1 cm). Manus impression is incomplete. **H**, Trackway of *Batrachopus* from same slab as Figure 62K, Riker Hill quarry, collector and disposition unknown (scale 1 cm). **J**, Manus and partial pes impression of *Batrachopus* sp. from unit 5, Dinosaur Park (Charles Rizzo collector, Charles Rizzo collection) (scale 1 cm). **I,** Very deep underprint of trackway of ?*Batrachopus* sp. from rubble, Riker Hill quarry (specimen lost). Note deep impression of heel area. **K**, *Batrachopus deweyii* pes from the Vreeland Quarry, stratigraphic equivalent to unit 5 at Riker Hill (Rutgers New Brunswick Museum main display slab) (scale 1 cm). Adapted from Olsen (1995).

due to both a real bias against fossilization as well as a collection bias. Recent years have seen a strong increase in the number of insect body fossil occurrences in the Newark Supergroup (Fraser et al., 1996; Huber et al., 2002; Olsen, 1988) and there is no reason not to expect more finds at Walter Kidde Dinosaur Park (especially with intense collecting of unit 10).

Several trackways attributable to *Acanthichnus* (Hitchcock, 1858) have been found at Walter Kidde Dinosaur Park (Figure 59). This ichnogenus is distinguished by two rows of thin impressions. These kinds of tracks could be made by any of a number of types of walking insects. The well-defined trackways shown in Figure 59 come from the lower part of unit 5, which has very small oscillatory ripples characteristic of very shallow water.

The only fish genus found thus far at the Walter Kidde Dinosaur Park is the holostean *Semionotus*, the most abundant fish throughout the Newark Supergroup Jurassic. *Semionotus* has been found in four units in the Riker Hill section (Fig. 4). Cycle RVH-3 has produced articulated fish in the upper microlaminated zone (unit 23c) and the overlying platy fine sandstone (unit 23b) of division 2. Fish fragments have been found in cycle RVH-2 in the gray claystone that produced the beetle elytron (unit 10) in division 2, and in a coprolite in the lower part of division 3 (lower part of unit 5).

In the entire Newark basin, the Towaco Formation has produced only *Semionotus*. It is very unusual for lakes, especially large ones, to have only one genus, and other genera are probably present but in low abundance. Studies of thousands of well-preserved *Semionotus* from the Towaco Formation of Pompton (NJ) by McCune (1987, 1996) show that although the genus-level diversity is very low, the species-level diversity may be very high. McCune has identified over 30 species of *Semionotus* in the laminated division 2 of a single Van Houten cycle. This type of high species diversity in one genus in a geographically circumscribed area is termed a "species flock". Closely analogous species flocks of cichlid fishes occur today in the great lakes of Africa, and species flocks of the fruit fly *Drosophila* occur in Hawaii. In these modern cases, the high species diversity correlates with a local deficit in the generic diversity due to geographic isolation.

At the Riker Hill quarry, at least three species of *Semionotus* appear to be present, although the preservation is too incomplete for certain identification (Figure 58). These are *Semionotus tenuiceps,* a small thin bodied form with small dorsal ridge scales, and a large form. *Semionotus tenuiceps* has a distinct hump at the back of the head and has expanded shield-like dorsal ridge scales. The two other forms are much too poorly preserved to be assigned to known species.

Preservation of the *Semionotus* at the park is variable (Figure 58). Fish from unit 24 in division 2 of cycle RVH-2 are preserved as flat films. No bone appears to be preserved, although an organic matrix outlines the gut and eye regions. The mineral matter of bone is a form of calcium phosphate (hydroxylapatite). Generally, in the process of fossilization, the cellular spaces within the bone become filled with minerals - often calcite - introduced by groundwater; the original mineral matter of the bone becomes somewhat altered to carbonate fluorapatite (i.e. francolite) (Shemesh, 1990). In anoxic environments, degradation products of the organic matrix of the bone also remain, coloring the bone black. In the case of the *Semionotus* in unit 23c, the phosphatic mineral matter of the bone has been dissolved away, leaving an outline marked by the residuum of the organic matrix of the bone and organic matter in the gut. This dephosphatization has been noted elsewhere in the Newark Supergroup and is generally more prevalent in the portions of lacustrine strata farther from the basin edge (McDonald and LeTourneau, 1989). A completely different style of preservation is represented by the fish from unit 23b, in which siltstone and fine sandstone have preserved the fish as a natural mold in high relief (Figure 58). In this case the bone tissue has been dissolved by recent near-surface weathering. Bone is preserved (along with the decay products of the organic matrix) in the *Semionotus* fragments from unit 10 in division 2 of cycle RVH-3. Bone is also preserved in the fish fragments in a coprolite from the lower part of unit 5 (in the lower part of division 3 of cycle RVH-3); however, the organic matrix is not preserved in the red mudstones, and hence the bone is white. The coprolite itself may be the excrement of a small theropod dinosaur.

By far the most spectacular fossils found at the Riker Hill quarry are tetrapod footprints. Thus far, excellent examples of *Ameghinichnus*, *Batrachopus*, *Grallator*, *Anchisauripus*, *Eubrontes*, and *Anomoepus* have been found. Only one other ichnogenus is known from elsewhere in the Newark basin Jurassic - the lepidosauromorph track *Rhynchosauroides*.

Ameghinichnus was first found by Larry Felder in 1978, in the upper beds of unit 5 of Walter Kidde Dinosaur Park. The ichnogenus was established by Casamiquela in

1964 for small five-toed quadrupedal tracks from the Late Jurassic Matilde Formation of northern Santa Cruz province in Argentina. The Towaco form differs somewhat from *A. patagonicus* (Casamiquela, 1964) and we consider it to be a new (although unnamed) species. The genus is characterized by a pentadactyl manus and pes of equal size, with nearly symmetrically-disposed digits of subequal length (Figure 59). Although the inferred structure of the manus and pes are consistent with mammals in both the type and new species, this arrangement appears phylogenetically well below the base of the Mammalia. In fact, such tracks could have been made by any of a variety of advanced therapsids, including the tritylodonts (which were contemporaneous with both *Ameghinichnus* ichnospecies) or trithelodonts (which were contemporaries of at least the Towaco ichnospecies). The size of the Towaco form is more consistent with trithelodonts or the largest of the Early Mesozoic mammals. Trithelodonts (e.g. *Pachygenelus monus*) have been found in abundance in the earliest Jurassic McCoy Brook Formation of the Fundy basin in Nova Scotia, in strata very close in age to the Towaco Formation (Olsen et al., 1987; Shubin et al., 1991). We therefore favor trithelodonts as the makers of the Riker Hill species of *Ameghinichnus*, although we cannot exclude other therapsids (including mammals) on the basis of existing evidence. Since the discovery by Mr. Felder, several more specimens of *Ameghinichnus* have been recovered from closely adjacent beds, although all of these are leptodactylous forms (Figures 26, 59). A complete treatment of this new species will be given elsewhere.

A single trackway found by John Colagrande in the uppermost Towaco Formation of Towaco, New Jersey represents the sole known post-Passaic Formation occurrence of *Rhynchosauroides* in the Newark Supergroup (Figure 59). It is described here because it should also occur at Riker Hill. During the Triassic, there were a very wide range of lepidosauromorph reptiles that could have made *Rhynchosauroides*-type footprints including the Trilophosauria, Rhynchosauria, Protorosauria, and the Lepidosauria, but by the Early Jurassic only the Lepidosauria remained. The Lepidosauria include the Rhynchocephalia (Sphenodontia; which includes the living *Tuatara*) and the Squamata (lizards and snakes). Small members of either group could have made these Jurassic *Rhynchosauroides*, although we note that an appropriate sized sphenodontian, *Clevosaurus bairdi*, is one of the most abundant skeletal forms known from the earliest Jurassic of Nova Scotia (Sues et al., 1994).

The probable-crocodylomorph track *Batrachopus deweyii* is the most common quadrupedal ichnite in the Newark Supergroup Jurassic, and is common at Walter Kidde Dinosaur Park as well (Figure 59). The form genus is diagnosed as a small quadrupedal form with a five-toed manus and a functionally four-toed digitigrade pes. Skeletal examples of crocodyliforms have been found in the Early Jurassic Portland Formation of Massachusetts (*Stegomosuchus longipes*; Walker, 1968) and the McCoy Brook Formation of Nova Scotia (*Protosuchus micmac*; Sues et al., 1996). The crocodyliforms of the earliest Jurassic, such as *Protosuchus*, were rather different in their overall appearance from living crocodilians (crocodiles and alligators). They were small, slender, short-snouted, and lightly armored, with no obvious aquatic adaptations. Their skeletons had elongate limbs, which, based on *Batrachopus*, appear to have carried the body in a high walk, with the legs more or less under the body. Similarly, they were digitigrade most of the time, while modern crocodilians walk plantigrade nearly all the time. In contrast to the large lunging semi-aquatic modern crocodilians, the makers of *Batrachopus* were small, fully terrestrial, active, fast predators.

By far the most abundant dinosaur tracks at the Riker Hill quarry are bipedal three toed forms (Figure 60) that never have manus impressions. The smallest ones (1.5-15 cm long) tend to be very narrow, with a distinctly elongate middle digit (III); the largest ones (20-30 cm long) tend to be broad, with a relatively short digit III. Only very rarely is there an imprint of the tip of digit I (the hallux). The in-between sized forms are intermediate in all proportions. These types of tracks have been traditionally called *Grallator* (the smallest forms), *Anchisauripus* (the intermediate-sized forms), and *Eubrontes* (the largest forms) (e.g. Lull, 1953). The phalangeal formula and general proportions are consistent with small to medium-large theropod dinosaurs. One would think that because these kinds of tracks are very common, they must be well known and understood; unfortunately the reality is a nomenclatural quagmire badly in need of revision.

This nomenclatural mess has two origins. First, the history of the nomenclature is sloppy and in desperate need of revision: most of what are proffered in the literature as the type specimens are not, and virtually every named taxon has a tortured and confused history (e.g. Olsen et al., 1998); even the commonly-used names are not necessarily valid by strict application of nomenclatural rules!. Second, organisms change in shape as they grow. This is termed "allometry" and is caused by different growth rates in different parts of the body, and we have argued that much of the variation in shape in these footprints can be explained by growth alone.

For over 90 years the standard references for Newark Supergroup tracks of Early Jurassic age have been the works of Lull (1904, 1915, 1953), essentially revisions of Hitchcock's monographs (1858, 1865). As defined by Lull's concept of their type species, the major differences between the genera *Grallator*, *Anchisauripus*, and *Eubrontes*, apart from size, are the ratio of length to width, the relative projection and length of digit III, and the angle of divarication between digits II and IV. When considering the type specimens (as identified in Lull, 1953) alone the genera appear morphologically quite

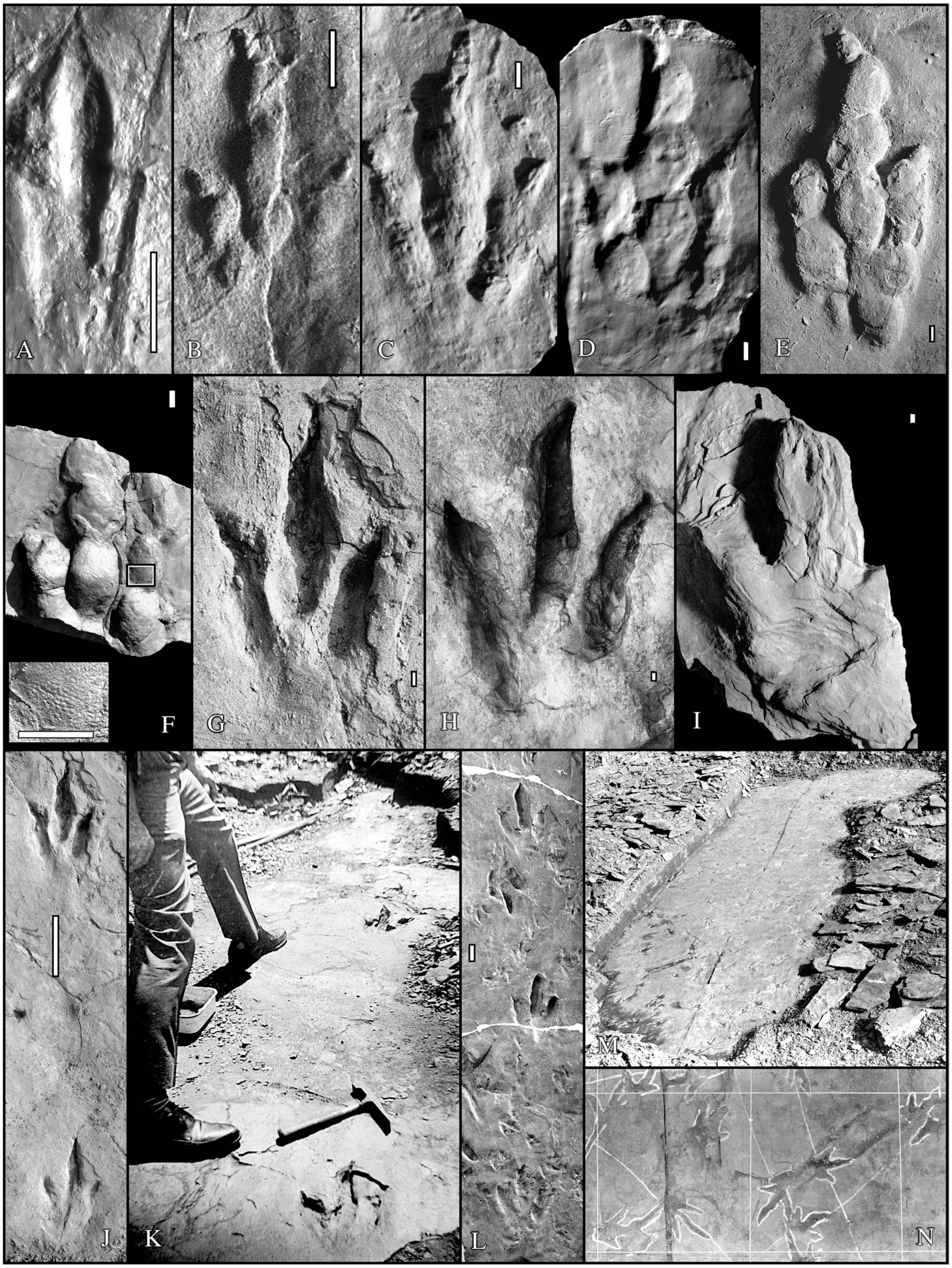
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Figure 60. Theropod dinosaur footprints from the Towaco Formation (mostly of the Riker Hill quarry). A, Natural cast of left pes of very small, Grallator (*Grallator*) sp., upper unit 18, Walter Kidde Dinosaur Park, John Colegrande collector, John Colegrande collection (same as Figure 10Aa and same slab as Figure 61C). **C**, Natural cast of left pes of *Grallator* sp., Riker Hill Quarry, ECPC 21; *Anchisauripus* sp., Riker Hill Quarry (Robert Salkin collector, Robert Salkin collection). **D**, Right pes of *Anchisauripus* sp., plaster cast of specimen, Riker Hill Quarry, AMNH 29299. **E**, Natural cast of left pes of *Anchisauripus* sp., upper unit 18, Walter Kidde Dinosaur Park, lost specimen. **F**, Natural cast of right pes of *Anchisauripus* sp., same trackway as E (above); note scale impressions in inset (specimen lost). **G**, Left pes of *Anchisauripus* sp., upper unit 18, Dinosaur Park, specimen not collected; same track as second in trackway in J (below). **H**, Left pes of *Anchisauripus* sp., upper unit 18, Dinosaur Park, specimen lost; same track as first in K (below). Scale for A-I is 1 cm. **I**, Partial natural cast of *Grallator* (*Eubrontes*) *giganteus* , upper unit 18, Dinosaur Park, specimen lost. **J**, *In situ* trackway of *Anchisauripus* sp., upper unit 18, Dinosaur Park, specimens not collected). **K**, *In situ* trackway of small *Eubrontes giganteus* sp., upper unit 18, Dinosaur Park, second track in series is ECPC 9. **L**, Trackway of *Eubrontes giganteus* and many other footprints, Vreeland Quarry, Towaco, NJ, Rutgers University Geology Museum main display slab. **M**, In situ *Grallator* footprint bearing layer of unit 23c, Riker Hill quarry. **N**, *Grallator* footprints on surface of unit 23c from slab shown in M (above).

distinct. However, when we attempt to place other specimens in these taxa we find that there are a multitude of intermediate forms. As suggested by Olsen (1980b) and Olsen et al. (1998), all of the forms lie on a morphological trend varying in a consistent way with size. It is not at all apparent to us that it is possible to objectively isolate portions of this trend as separate genera. As the size of the footprint increases, the relative width of the footprint increases, as does the divarication between digits II and IV; there is thus a decrease in the projection of digit III beyond the outer two toes. The same general proportional changes can be seen between the skeletons of the small ceratosaurian theropod *Coelophysis* (Colbert, 1989) and the much larger ceratosaur *Dilophosaurus* (Welles, 1984). The larger specimens of *Coelophysis* fit the proportions of large *Grallator* or small *Anchisauripus*. If these footprints cannot be objectively split up into several genera, they should all be give the earliest proposed valid name, which should be *Eubrontes* (Ornithichnites Hitchcock 1836) (Olsen et al, 1998). But, pending a revision of these genera (underway by ECR), we refer to them the same way as did Lull (1953): as *Grallator*, *Anchisauripus*, and *Eubrontes*, and collectively as grallatorids.

Among the many grallatorid tracks from Riker Hill are some very small examples (Figures 60). These tracks are among the smallest dinosaur footprints known from anywhere geographically and temporally. A nice gradational series of tracks intermediate between the smallest grallatorids and the largest *Eubrontes* have been found at the site (Figure 60).

Remains of small theropods have been found in Jurassic Newark Supergroup strata in the Portland Formation of Connecticut and Massachusetts and the McCoy Brook

Formation of Nova Scotia (Colbert and Baird, 1958; Olsen et al., 1987; Sues and others, 1987; Talbot, 1911). A single large tooth fragment found within a *Eubrontes* track at Walter Kidde Dinosaur Park (unit 18) is the only skeletal material of theropods yet found in the Newark basin (Olsen, 1995).

The ichnogenus *Anomoepus* (Hitchcock, 1848) is the only other common dinosaurian track form that has been positively identified from the Riker Hill quarry. This genus is distinguished in bipedal walking tracks by having the metatarsal-phalangeal pad of digit IV nearly in line with digit III, having a short digit III for the size of the foot, and more divaricate pedal digits than similar-sized grallatorids. Sitting traces are fairly common, in which the entire pes including the metatarsus is impressed, as is the five-toed manus. The wide range of sizes of *Anomoepus* present at Roseland demonstrate the range of variation in the genus, although there is far less variability than is seen in the grallatorid forms described above. We have not, however, found criteria for recognizing more than one footprint species, and therefore we synonymize all *Anomoepus* species from the Newark Supergroup with *Anomoepus scambus*, the type species of the genus from the Turners Falls Formation of Massachusetts (Olsen and Rainforth, 2002). Also unlike *Grallator*, there seems to be little change in shape with size, except perhaps for a slight relative shortening of digit III with increasing foot length.

The structure of the inferred pedal skeleton is compatible with either a theropod or an ornithischian dinosaur; however, the hand flatly rules out a theropod trackmaker, because all five digits remain well developed, albeit short. In contrast to theropods (in which manual digits I-III are dominant, IV and V eventually being lost), manual digits II-IV are dominant in *Anomoepus* - the beginnings of a trend seen clearly in more advanced ornithischians (e.g. hadrosaurs), in which the outer digits are reduced and finally lost. *Anomoepus* was therefore probably made by a small, herbivorous, relatively primitive ornithischian dinosaur, similar to *Lesothosaurus* or some other "fabrosaur". Scrappy bones and isolated teeth of "fabrosaurs" have been found in the McCoy Brook Formation of Nova Scotia (Olsen et al., 1987; Sues et al., 1987).

Anomoepus is characterized by a less well-developed pad structure in the pes than grallatorids. Although the skin texture is very similar to grallatorids, the pads tend to be separated by much smaller narrower pads, suggesting that some flexing of the foot was common. The pads on the hand show this very well, and in fact appear to show creases or grooves over the articulations, which is to be expected in a hand used more for grasping than walking. Thus, it seems likely that *Anomoepus* used its hands, and sometimes its feet, to grasp things - probably branches; a reasonable scenario for a small herbivorous dinosaur.

One of the most unusual aspects of the Riker Hill assemblages is the unusually large number of presumably-juvenile *Anomoepus* tracks. One layer in particular in the upper part of unit 18 was covered in many small and a few larger *Anomoepus*. The association of the uncommon larger forms with the much more abundant smaller forms suggests, but of course does not demonstrate, herding of young. The smallest *Anomoepus* tracks are, like the tiny *Grallator*, among the smallest dinosaur footprints known. Unfortunately, also like the tiny *Grallator*, these tracks lack pads and thus the assignment to *Anomoepus* can only be tentative. One of the minute *Grallator* trackways is from the same bedding plane as the abundant "baby *Anomoepus*". The meaning of the diminutive carnivore among the baby herbivores is unknown.

The track assemblage at Walter Kidde Dinosaur is typical for the Towaco Formation, and is very similar to all of the Newark Supergroup assemblages that postdate the Triassic-Jurassic boundary. They are the post-catastrophe assemblages, largely consisting of survivors of the end-Triassic mass extinction.

Return to Beaufort Avenue.

91.4 mi. Turn right onto Beaufort Avenue.

91.8 mi. Turn right onto Eisenhower Parkway.

92.4 mi. Turn left onto Eagle Rock Avenue.

93.1 mi. Crossing Passaic River.

93.8 mi. Turn right onto Ridgedale Road. This area is often flooded by waters from the Passaic River. Control of this flooding was the goal of the proposed Passaic River diversionary tunnel, the planning stages of which resulted in the collection of over 10 k of short cores (each ~<100 m) spanning the upper Passaic Formation and most of the succeeding Jurassic Formation (Fedosh and Smoot, 1988; Olsen et al., 1996b) collected by the Army Corps of Engineers (ACE cores).

95.2 mi. Turn right onto New Road.

96.1 mi. Turn right onto Bloomfield Avenue.

96.5 mi. Cut in Hook Mountain Basalt. Immediately north in the same ridge is the cut for Interstate Route 80 that comprises the type section of the Hook Mountain Basalt (Olsen, 1980c).

96.6 mi. Turn left onto Mountain Road.

96.7 mi. Turn right onto Maple Avenue

97.0 mi. Turn left onto entrance for Interstate Route 80 East.

102.9 mi. Gap in ridge for Preakness Basalt through which the Hudson River flowed.

104.8 mi. Crossing Passaic River.

106.1 mi. Gap on left in Orange Mountain Basalt for Hudson River.

106.3 mi. On right are outcrops of the Orange Mountain Basalt and underlying uppermost Passaic Formation. This is section E of the type section of the Passaic Formation of Olsen (1980c). A series of faults cut the Orange Mountain Basalt here, some of which are visible in the cut on the left, just west of the Passaic - Orange Mountain Basalt contact. Triassic-Jurassic boundary is somewhere within a few meters below contact.

107.1 mi. Garrett Mountain visible on right (south), Passaic Falls is on the left (north). The upper Passaic Formation of Rhaetian age (latest Triassic) has produced near here a series of well preserved skeletons of the highly specialized procolophonid reptile Hypsognathus (Colbert, 1946) including the type specimen, which was found just south of here (Gilmore, 1928). About one skeleton or skull is found in the northern Newark basin per decade.

108.8 mi. Section D of type section of Passaic Formation (Olsen, 1980a,c).

112.6 mi. Section C of type section of Passaic Formation (Olsen, 1980a,c).

113.7 mi. Section B of type section of Passaic Formation (Olsen, 1980a,c).

116.8 mi. Beginning of type section of Passaic Formation (section A, Olsen of Olsen, 1980a,c) in open cuts for Route 80.

116.1 mi. Merge with Interstate Route 95.

118.7 mi. Open cut in Palisade sill and Lockatong hornfels. According to Van Houten (1969), hornfels include grosularite, andradite, prehnite, and diopside varieties. Lockatong cycles fossiliferous, as usual, and these cycles may tie in with Granton Quarry cycles (Stop 4). Geochemistry of the sill at this cut described by Naslund (2000).

119.3 mi. Exit left for US Route 9W and Palisades Interstate Parkway.

119.6 mi. Turn left onto US Route 9W north and follow along the creast of the strikeridge of the Palisade sill.

120.1 mi. Keep left on US Route 9W north.

130.6 mi. Lamont-Doherty Earth Observatory. End of Field Trip. End of Fieldtrip.

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Bedrock geology, geochemistry and geochronology of the Grenville Province in the western Hudson Highlands, New York

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Abstract

New integrated field, regional, geochemical and geochronological studies in the western Hudson Highlands yield a detailed history of the building of the supercontinent of Rodinia during the Middle Proterozoic. An early volcano-plutonic complex formed in a ca. 1.2 to 1.1 Ga island arc or oceanic magmatic arc. The arc was rimmed by aprons of volcaniclastic detritus in uncertain relation with euxinic clastic>carbonate sedimentary basins. Collision likely with the proto-South American continent peaked at ca. 1020 Ma and resulted in granulite facies metamorphism, local felsic plutonism, pervasive gneissic foliation, and westward to northwestward directed fold nappe emplacement. A dioritic to bimodal plutonic event closely followed the peak conditions at 1,008 Ma. An extensive dextral strike-slip event then overprinted the area forming a 35 km-wide anastomosing system of mylonite zones. These mylonites display abundant kinematic indicators and are fringed by mesoscopic sheath folds and asymmetric boudins and well as transpressional folds and drag folds. Late in the deformational history, the zones became dilational and hydrothermal fluids mineralized them with local magnetite deposits within a locally buffered gangue assemblage. This strike-slip system has a synchronous conjugate system of the same magnitude in the Adirondacks of New York. Together they form a deep seated syntaxis analogous to that in the Himalayan orogen of southern Asia.

Introduction

The Reading Prong is a Grenville massif that links the Blue Ridge and the Green Mountain provinces to form the spine of the U.S. Appalachians (Figure 1). The Hudson Highlands comprises the northern extent of the Reading Prong (Figure 2). The details of the Grenvillian orogeny in the Appalachians are difficult to decipher because of complex structural relations, lack of exposure, and pervasive granulite facies metamorphism, as well as extensive overprinting during the Paleozoic and Mesozoic locally (Ratcliffe *et al.*, 1972; Bartholomew and Lewis, 1988; Krol et al., 1992; Krol and Zeitler, 1994). Any

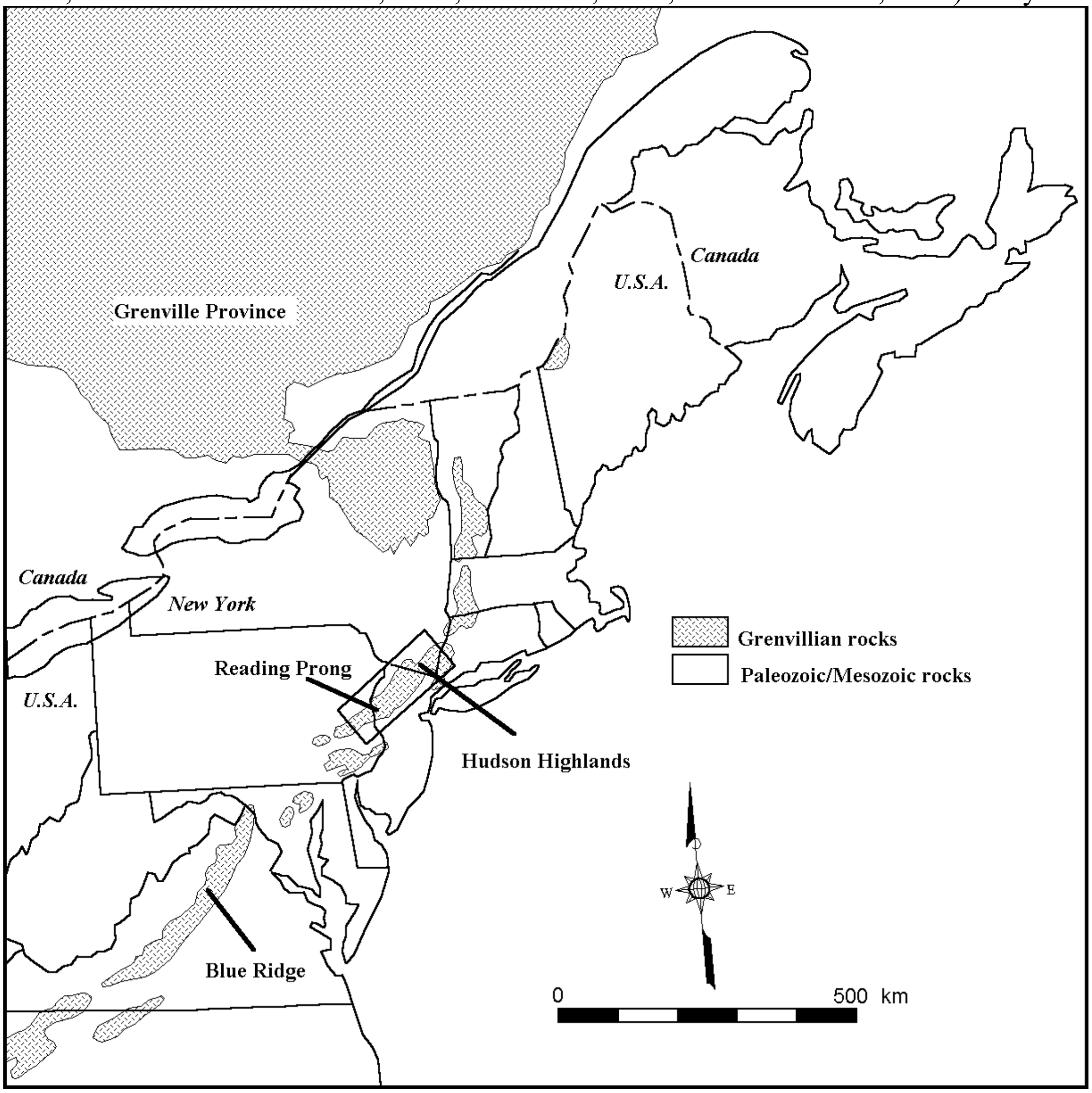

Figure 1. Regional map of eastern United States and Canada showing the geographic distribution of Grenville rocks. The area of Figure 2 is outlined by a rectangle.

granulite crystalline massifs, or with an isotopic age between ca. 1,300 Ma (Mose, 1982) and 893 Ma (Ratcliffe *et al.*, 1972) age in the Reading Prong of the north-central Appalachians (Figure 1) was assigned to the Grenville orogeny. Middle Proterozoic tectonism in the Canadian and Adirondack Grenville rocks have been subdivided into three to four orogenic events (Easton, 1986). McLelland (1986) suggested that the older Grenvillian ages are related to anorogenic plutonism in the Adirondacks of New York. McLelland and Isachsen (1980) and Whitney (1983) proposed a three-stage tectonic model for the deformation and metamorphism in the Adirondacks. Virtually all studies concur that the culmination of the Grenvillian event occurred about 1,100 - 1,000 Ma,

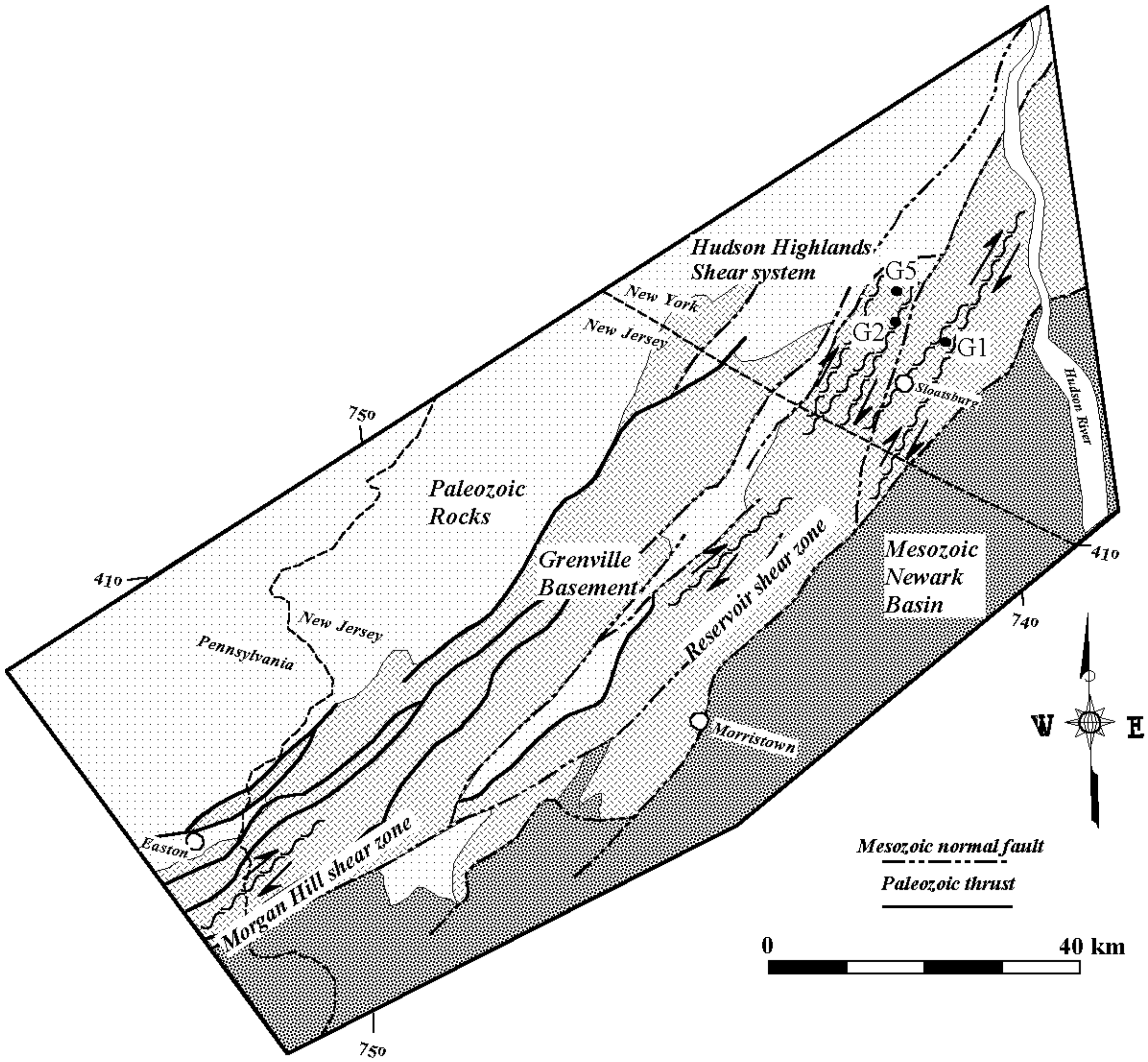

Figure 2. General geologic map of the Reading Prong and Hudson Highlands. The study area, in southern New York State, is outlined.

approximately equivalent to the Ottawan orogeny (Easton, 1986). The thermal and deformational peak in the Hudson Highlands and the entire Reading Prong occurred around 1,150-1,050 Ma (Silver, 1969; Ratcliffe *et al.*, 1972; Mose, 1982; Weiner *et al.*, 1984; Drake, 1984).

Gates (1995) and Gates and Costa (1999) proposed a major late Grenvillian dextral strike-slip shearing event in the Reading Prong. This shearing was constrained to discrete faults, such as the Ramapo and Reservoir Faults (Figure 2), which were active well after peak Grenville tectonism and to much lower temperatures. A Middle Proterozoic escape tectonic (Tapponnier *et al.*, 1982) event in the central Appalachians resulting from accretion to the north is interpreted to have produced this deformation.

Stratigraphy

The field trip area is located in parts of the Sloatsburg, Thiells, Monroe, and Popolopen Lake quadrangles west of the Hudson River within the central Hudson Highlands, NY (Figures 2 and 3). Previous mapping in this area, divided the units by rock types (Dodd, 1965; Dallmeyer, 1974). Considering that about 80% of the rocks are quartz-feldspar gneisses, this system is useful for geologic maps but not for purposes of tectonic reconstructions. Gundersen (1986) suggested that lithologic and stratigraphic associations and sequences should be grouped as units in a kind of sequence stratigraphy for metamorphic rocks. This system of grouping lithologies is adopted for this field guide.

Metasedimentary Lithofacies

Throughout the western Hudson Higlands there are belts of rock considered to have sedimentary protoliths including pelitic-, psammitic-, calcsilicate-gneisses, quartzite and marble. Belts of rock upward of a few kilometers wide may contain all or some of these lithologies interlayered at the scale of meters to 100's of meters. These rocks have been included in the metasedimentary lithofacies (Figure 3). The metapelite consists of interlayered biotite-garnet gneiss with medium to coarse quartz, plagioclase, K-spar and local sillimanite, and cordierite with quartzofeldspathic layers. Within the metapelite are zones of graphite-pyrite-garnet gneiss with biotite, quartz, K-spar, plagioclase, and sillimanite locally. Quartzite layers of 10-50cm thickness also occur within this unit as do rare and discontinuous layers of diopside and diopside-garnet marble to calcsilicate of 10 cm to 2 m thickness. The calc-silicate is quartzofeldspathic with salite, apatite, sphene, scapolite, and hornblende. It is commonly migmatitic. There is also a rare quartz-garnet granofels. Common intrafolial pegmatites exhibit rootless isoclinal folding. The contacts with the quartzofeldspathic gneiss and rocks of the metavolcanic lithofacies are usually gradational.

Metavolcanic Lithofacies

Sequences of strongly banded interlayered gneisses with mafic, intermediate and felsic compositions are interpreted to represent rocks with volcanic protoliths. The mafic gneiss domains are medium to coarse grained and aligned hornblende, plagioclase, clinopyroxene and hypersthene with local concentrations of magnetite. The intermediate and felsic gneiss bands contain medium to coarse-grained plagioclase, quartz, and minor hornblende, biotite, clinopyroxene and hypersthene. Banding ranges in thickness from 5 cm to 5m with varying proportions of each rock type. There are local interlayers of quartzite and calcsilicate gneiss. The contacts with the quartzofeldspathic gneiss and rocks of the metasedimentary lithofacies are gradational. Migmatites occur locally in this lithofacies (Figure 3).

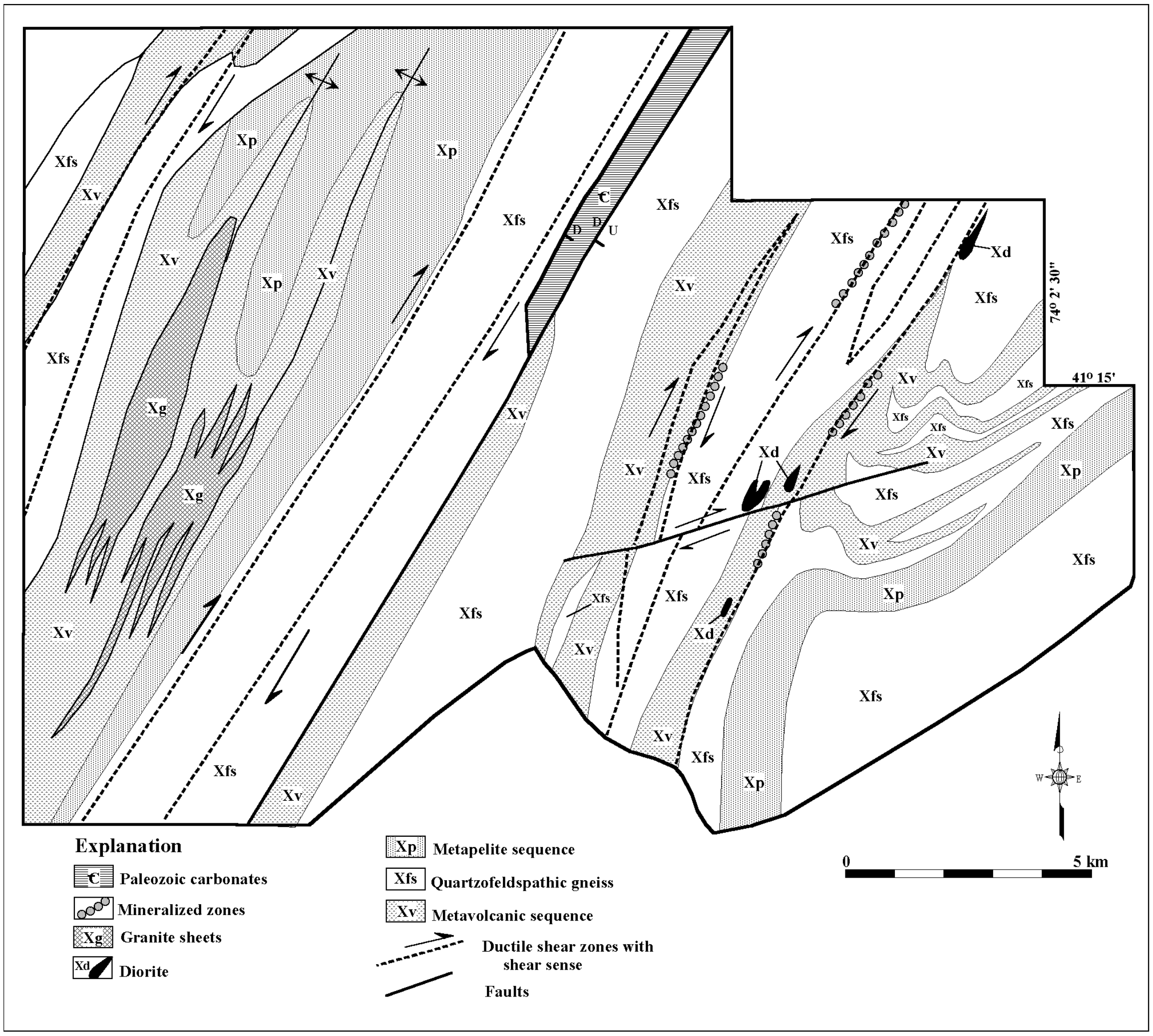

Figure 3. Geologic map of the area of Harriman State Park and Sterling State Forest.

Quartzofeldspathic Gneiss

The quartzofeldspathic gneiss ranges from massive to layered quartz-plagioclase gneiss and quartz-k-feldspar-plagioclase gneiss with minor amounts of clinopyroxene, hypersthene, hornblende and/or biotite. Locally, this unit contains magnetite or garnet in trace amounts. Compositional layering is defined by the proportion and type of the mafic mineral component. Locally, this unit contains apparent fining upward sequences by an increase in the amount of mica and decrease in layer spacing with sharp contacts between sequences. However, such relict sequences in granulite terranes are difficult to interpret. It is locally interlayered with quartzite and with mafic gneiss at the contact with the metavolcanic lithofacies. The gradational contacts with the metavocanic and metasedimentary lithofacies,and the internal compositional layering suggests the quartzofeldspathic units represent a volcaniclastic sequence. However, the occurrence of relict plagioclase grains in the main body west of the New York State Thruway suggests an intrusive origin for some of the rocks (Figure 3).

Granite Sheets

A series of granite sheets intruded rocks of the metavolanic and metasedimentary lithoficies and the quartzofeldspathic gneiss. The sheets range in thickness from 5 to 200 m and are laterally continuous for several kilometers. The granite is typically medium to coarse grained, locally megacrystaic, and lacking foliations. However, locally granite sheets are foliated where intersected by later ductile shear zones. The granite sheets are leucocratic with K-spar, quartz, plagioclase, minor biotite, apatite, and titanite. The texture is equigranular with subhedral to anhedral interlocking grains, and locally they contain xenoliths of country rock. Where the granite sheets are mylonitic, the contact with the quartzofeldspathic gneiss is very difficult to determine.

East of the New York Thruway there are isolated occurrences of granite sheets. One of the more famous outcrops at Claudia Smith's Den immediately east of the New York Thruway. In the central part of the Sterling Forest there are numerous granite sheets that strike northest-southwest and dip moderately southeast (Figure 3). As previously stated, sheet thickness varies considerably, however, the granite sheets in Sterling Forest extend out from two tabular shaped bodies of granite. One is central to Hogback Mountain and the other is located in Bare and Tiger Mountains. These two bodies of granite and sheet appendages occur within parallel antiforms that can be traced from the Monroe quadrangle southward.

Diorite

Coarse- to very coarse-grained black and white speckled diorite contains plagioclase, pyroxene, hornblende, and biotite locally. The diorite grades to lower pyroxene, anorthositic compositions locally. Texture ranges from granoblastic to foliated and mylonitic with S-C fabric. The diorite locally contains xenoliths of country rock with ductile contacts that are partially melted to form a rind of coarse to pegmatitic granite around them and filling fractures in the diorite (Figure 3).

Pegmatite Dikes

There are two generations of pegmatite dikes. Early dikes are white and contain K-spar, quartz, muscovite, and garnet locally. They are largely parallel to subparallel gneissic foliation (concordant), commonly boudinaged, and contain internal foliation and deformed grains. Thickness ranges from 10cm to 1m. Many are associated with granite sheets. The late dikes are pink, and very coarse grained with K-spar, quartz, and locally muscovite, hornblende, magnetite, pyroxene, titanite, and/or garnet depending upon the rock intruded. They are highly discordant, commonly within brittle faults, and contain xenoliths of fault rocks. They exhibit no deformational fabric. Thickness ranges from 1m to 10m. They are locally associated with small granite bodies.

Mineralized Zones

Late stage concordant to slightly discordant brittle fracture zones occur within several of the mylonite zones and are mineralized. They contain randomly oriented, coarse to megacrystic intergrowths of scapolite, salite and phlogopite followed by magnetite and cemented by calcite in areas of marble. Other zones contain hornblende and clinopyroxene followed by magnetite that are contained within massive quartz. Zones that connect the magnetite deposits are thinner and typically composed of randomly oriented to aligned clinopyroxene with only minor magnetite, phlogopite and/or quartz. They are commonly intruded by late pegmatites that contain mineralized rock as xenoliths. Thicknesses of the zones range from 2m to 15m.

Deformation

There are at least two major Precambrian deformational events recorded in the crystalline rocks of the western Hudson Highlands. The structural features produced during these events are described and placed in the context of the metamorphic conditions.

First Deformation Event

The first deformational event is regional in extent, penetrative and found in most rock units. A pervasive gneissosity formed during this event in every unit except the diorite and granite sheets. This gneissosity is defined by virtually all minerals but especially by platy and elongate minerals. Biotite, amphibole, sillimanite, and pyroxene are aligned in the strongly foliated quartz-feldspar matrix. Additionally, aggregates of quartz and feldspar define layering in some lithologies. Pegmatites are commonly parallel to subparallel to the gneissosity and exhibit well developed pinch and swell. These pinch and swell pegmatites are locally asymmetric indicating a component of simple shear in their formation. Amphibole and pyroxene clots show similar rotation textures forming δ porphyroclasts (Passchier and Simpson, 1986). Some pelitic rocks contain garnet-fish structures, and locally, some rocks contain intrafolial asymmetric isoclinal folds 5 to 20 cm thick though larger map scale folds may also exist.

Mesoscopic and megascopic folds produced during this event are recumbent to shallowly reclined. They are tight to isoclinal and commonly asymmetric with the lower limbs sheared out. This asymmetry consistently indicates northwestward transport. The weak and sparse kinematic indicators described above support this shear sense. Thinner layers in these folds contain mesoscopic parasitic folds that are especially well developed on the upper limb. The occurrences of the granite sheets is in the hinge region of two of the map-scale isoclinal folds. Undeformed diorite contains xenoliths of deformed gneiss constraining the age of the first deformational event to pre-diorite intrusion.

Second Deformation Event

The second deformational event is characterized by a group of anastomosing shear zones across the area (Figures 2 and 3). These shear zones overprint the features of the first deformational event. The shear zones strike northeast and are either vertical or steeply northwest to southeast dipping. They range from 0.5 to 2 km in thickness though the boundaries are diffuse and difficult to determine in some areas. The shear zones are marked by well-developed type II S-C mylonite (Lister and Snoke, 1986) with shallowly northeast plunging mineral lineations. The dominant lithology within the mylonite is quartzofeldspathic gneiss but some rocks of the metavolcanic and metasedimentary lithofacies are also sheared. Diorite also locally forms S-C mylonite constraining the time of emplacement to pre-kinematic with regard to this event. Kinematic indicators within the mylonite include C/S fabric, rotated porphyroclasts, shear bands, asymmetric boudins and flattened asymmetric intrafolial folds. There are well-developed mesoscopic sheath folds with shallow northeast plunge, and megascopic drag folds adjacent to the main shear zone around Little Long Pond and Lake Tiorati respectively. All kinematic indicators show a consistent dextral strike-slip sense of shear. Minerals within the sheared rocks include amphibole and biotite as well as quartz and feldspar, all of which show plastic deformation but with full recovery. By texture and mechanical response of the minerals (Simpson, 1985), metamorphic conditions must have been upper amphibolite to granulite facies.

Late in the movement history, the shear zones became dilational. One to 6 km long mineralized veins occur along the shear zones (Figure 2). The veins parallel the zones but clearly cut the mylonitic foliation with ragged to planar contacts. The veins were progressively filled with salite and scapolite locally followed by magnetite as described earlier.

Mesoscopic gentle to open upright folds also occur adjacent to the shear zones locally. These folds plunge gently from due north to north-northeast. The folds occur in well-layered metavolcanic sequences and within 150 m of the shear zone boundary. They locally appear en echelon.

A pervasive steeply northwest-dipping crenulation cleavage occurs throughout the area. It is best developed in the metapelitic and thin layered metavolcanic units. Intersection lineations with the gneissic foliation produced in the early event are generally parallel to the stretching lineations in the mylonite.

Thermochronology

Ar/Ar thermochronology

Ar/Ar thermochronology was performed on hornblende and biotite samples from the area around the Hogencamp mine (Gates and Krol, 1998). Samples were collected along the southeastern margin of a major dextral strike-slip shear zone. Mineral separates were prepared and analyzed at the Ar/Ar thermochronology lab at Massachusetts Institute of Technology using standard incremental heating procedures. All uncertainties in the ages are quoted at the 1 sigma-level and include the error associated with the J-value.

Hornblende from the gangue minerals in the Hogencamp mine HSP-2A and HSP-2B yield ages of 914 $\pm$ 3.6 Ma and 922 $\pm$ 3.4 Ma respectively. Hornblende from an undeformed granitic pegmatite that intruded mineralized vein material yielded an age of 923 $\pm$ 2.8 Ma. Biotite from the gangue minerals (HSP-2A) yields an age of 840 $\pm$ 5.0 Ma whereas biotite from the pegmatite yielded 794 $\pm$ 3.0 Ma.

The closure temperature for argon diffusion in hornblende is 500 - 550^{o} C depending upon the cooling rate. The ages obtained for hornblende from the pegmatite and the veins may represent either the initial emplacement and crystallization or cooling ages. The relations among the rocks and their ages presents a complex picture. Generally, it is clear that the mineralization and pegmatite intrusions occurred at about the same time. The pegmatite must have closely followed the mineralization where they intrude the veins. The source of fluids for mineralization may have been related to the pegmatite magmatism.

The Reservoir fault, a related dextral strike-slip shear zone, was active until 876 $\pm$ 5 Ma (Gates and Krol, 1998). Therefore, the fault could have been active after pegmatite intrusion and the 914 Ma age is explainable as crystallization or deformation. In some areas, the vein rock is sheared.

The closure temperature for biotite is about 300 ^{o} C. Therefore, cooling through 200-250 ^{o} C took at least 74 Ma and up to 134 Ma. This duration for a small change in temperature represents very slow cooling on the order of 1.5-3.5^{o} C/m.y. Slow cooling rates suggest that the interval between 920 and 786 Ma does not represent a major period of crustal thickening but instead there was slow unroofing and minor lateral movement.

U-Pb Geochronology

Zircons from three samples of gneiss from the field area analyzed at the SHRIMP lab at the Geological Survey of Canada to obtain U-Pb ages. The first sample was from a semi-pelitic gneiss layer within the metasedimentary lithofacies, the second sample was from the quartzofeldspathic gneiss body located west of the New York Thruway, and the third sample was collected from a small diorite body (Lake Tiorati Diorite). All uncertainties in the ages are quoted at the 1 sigma-level.

The zircons from the semi-pelitic gneiss are strongly zoned with distinct cores and rims (Figure 4a). We interpret the cores of these zircons to be detrital in origin wheras the clear rims are probably associated with the high-grade metamorphism. Numerous analyses from zircon cores and rims are shown on Figures 5a and 5b. Zircon cores yielded a range of ages from 1200 to 2000 Ma, whereas the rims yielded ages from 1000 to 1030 Ma with the bulk of the analyses centered on 1020 Ma.

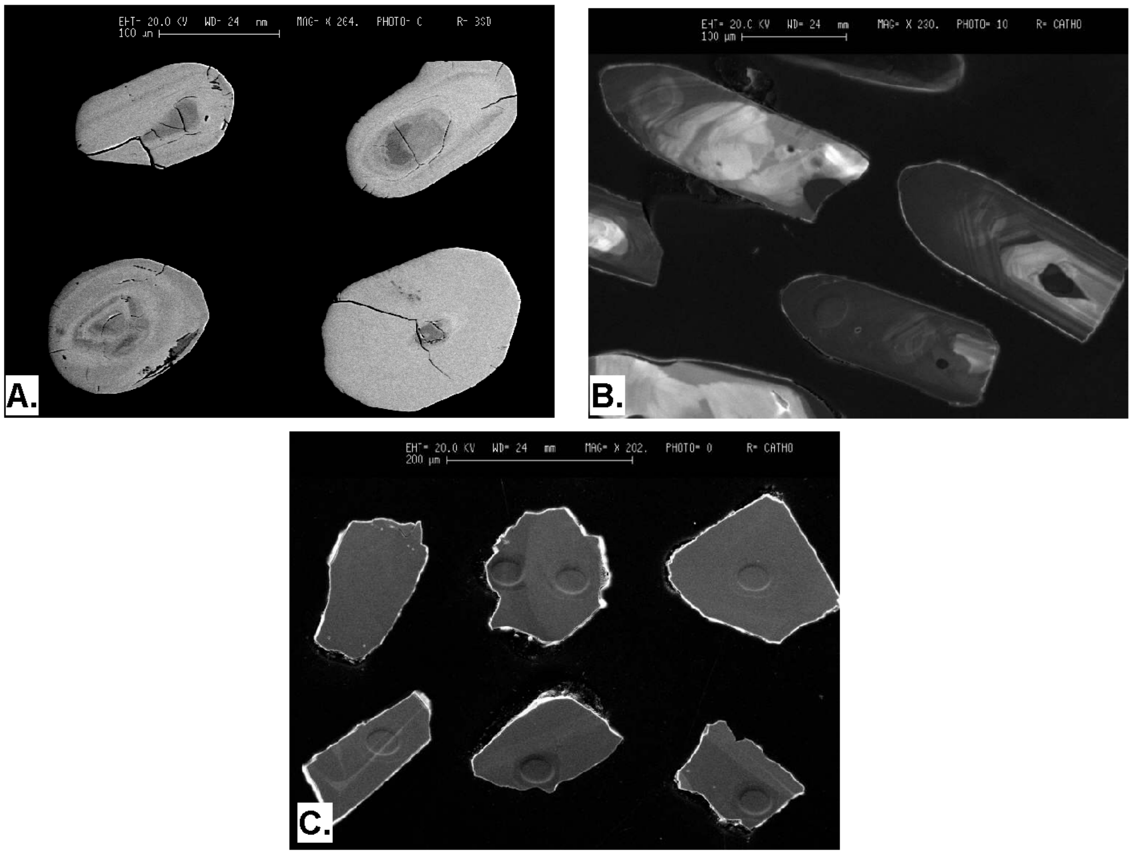

Figure 4. Representative cathodoluminescence images of zircons that were analyzed using the SHRIMP. A. Zircons from metasedimentary rocks showing complexly zoned cores and metamorphic overgrowths. B. Zircons from metavolcanic rocks showing igneous cores and metamorphic overgrowths. C. Zircons from diorite with no zonation.

The quartzofeldspatic gneiss produced zircons with rhythmically zoned cores and clear rims (Figure 4b). Analyses of the cores and rims produced two clusters of concordant ages (Figure 5c). The cores range from 1160 to 1220 Ma, and the rims exhibit a range of ages from 1000 to 1080 Ma. We interpret the zoned cores and associated ages to represent the original igneous history of this rock, and the rims to represent the regional metamorphic overprint.

The Lake Tioroti diorite body produced small subhedral zircons with minimal zoning (Figure 4c). Analyses of these zircons yielded a cluster of concordant ages averaging 1008 +/- 4 Ma (Figure 5d). Because this pluton is partially deformed in a dextral strike-slip shear zone, this age provides an upper limit to the local strike-slip event.

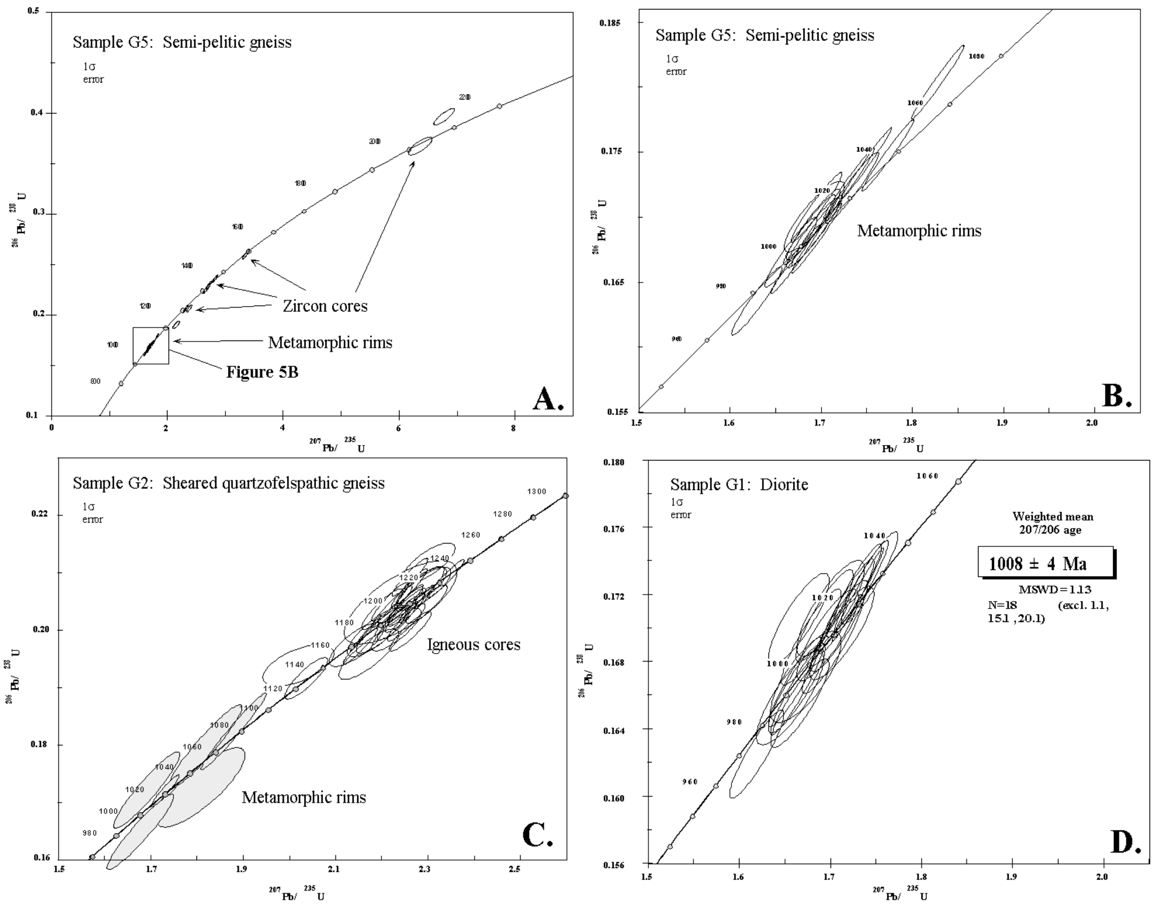

Figure 5. Concordia plots for SHRIMP analyses completed on zircons from Hudson Highlands rocks. A. and B. from semi-pelitic gneiss (B is a detailed view of A); C. from sheared quartzofeldspathic gneiss; and D. Lake Tioroti diorite body.

Geochemistry

Recent geochemical, structural, and geochronological investigations in the southwestern Hudson Highlands, NY has identified at least four discrete tectonomagmatic events. The earliest igneous events are represented by a sequence of metavolcanic (mafic and intermediate, quartz-plagioclase gneisses) and quartzofeldspathic gneisses (meta-plutonic and/or metavolcanoclastic). The metavolcanic unit consists of interlayered mafic (amphibolites) and intermediate to felsic (quartz-plagioclase) gneisses that are variably HFSE-depleted and LREE-enriched and are interpreted to have erupted in volcanic arc and back-arc setting. The second event (examples of which will not be displayed on this trip) is represented by metaplutonic, hornblende granite with A-type chemistry that is correlated with the Byram Intrusive Suite (~1095 Ma, Drake et al., 1991; ~1100 Ma, Volkert et al., 2000). The third magmatic event generated a suite of syn- to late-orogenic alaskite sheets (granite sheets on Figure 3) that have syn-COLG granite signatures and depleted HREE contents indicative of deep crustal melting in a thickened crust. The fourth magmatic event in this region is represented by the small, diorite pluton. These rocks have strong depletions in HFSE and high HREE and Y contents indicating shallow partial melting (<65 km) of asthenospheric or arc-modified lithospheric mantle with subsequent crustal contamination.

Metavolcanic lithofacies and the quartzofeldspathic gneiss

Mafic gneisses have major and trace element compositions broadly similar to tholeiitic to calc-alkaline basalts and thus, support a mafic volcanic protolith for these rocks. This is illustrated on a classification diagram based on High Field Strength Elements (HFSE), where mafic gneisses plot within and overlap the fields for tholeiitic and calc-alkaline basalts (Figure 6A). Rare Earth Element (REE) patterns are quite variable (Figure 6B); however, most samples have slightly LREE-enriched (La/Yb_N = 1.5 to 2) to almost MORB-like, LREE-depleted (La/Yb_N = 0.8) patterns with minor negative Eu anomalies (Eu/Eu* = 0.9 to 1.0) (Figure 7). Sample LT-5 is an exception with a distinct LREE-enriched pattern (La/Yb_N = 10) with a significant negative Eu anomaly (Eu/Eu* = 0.7). The lack of strong LREE/HREE fractionation and relatively high concentrations of HREE (~8-12x chondrite) in all samples indicates melt generation occurred at relatively shallow mantle depths above stability field of garnet peridotite (e.g. <60 km depth). These rocks also show variable HFSE depletions and on tectonic discrimination diagrams they consistently plot in overlapping fields defined by volcanic arc basalts and/or MORB (Figures 6C-D). Based on this data, we interpret these rocks to have erupted in an oceanic island arc/ back-arc setting or perhaps a continental arc built on attenuated crust. Mafic gneisses of similar chemistry have also been reported in the Central Metasedimentary Belt of the Grenville Province in SE Ontario (Tudor Volcanics, Turriff Volcanics, and Belmont Lake Volcanics; Smith and Holm, 1990; Harnois and Moore, 1991; Smith et al., 1996) and have similar tectonic interpretations.

Geochemical data for intermediate and felsic gneisses are also supportive of a volcanic protolith for this unit. Based on major elements, these rocks are most similar to calc-alkaline, low-K rhyodacites. Mineralogically, these rocks are tonalites (Figure 6A). They have moderately LREE-enriched patterns (La/Yb_N = 5 to 7) with relatively small negative Eu anomaly (Eu/Eu* = 0.7 to 0.9) for rocks of this silica content (68 to 70%) (Figure 6E). Similar to the mafic gneisses of this unit, intermediate and felsic gneisses lack strong LREE/HREE fractionation and have relatively high HREE abundances (8-10x chondrite) indicating crustal melting in the absence of residual garnet in crustal source rocks. HFSE depletion in these rocks is variable, but overall most samples have strong depletions in Nb, Ta, P, Hf, and Ti that are characteristic of calc-alkaline, volcanic arc rocks. They are mineralogically and chemically very similar to other tonalitic to trondhjemitic gneisses found in the New Jersey Highlands (Losee Metamorphic Suite; Volkert and Drake, 1999). Similar rocks also occur in the Greens Mtns, VT (Mount Holly Complex; Ratcliffe et al., 1991) and in the Adirondacks (McLelland and Chiarenzelli, 1990) that have U-Pb zircon ages between 1300 and 1350 Ma. The data on these rocks are consistent with the interlayered mafic gneisses and again, suggest an island arc and/or continental arc on attenuated crust tectonic setting.

Lake Tioroti diorite

Major element chemistry of coarse-grained, relatively undeformed samples of the Lake Tiorati Diorite indicate they are uniformly mafic plutonic rocks that have moderate to strong calc-alkaline geochemical signatures (Figure 6A). REE patterns of most samples are weak to moderately LREE-enriched (La/Yb_N = 1.5 to 5) and have slightly concave upward or "dished", MREE-depleted patterns (Figure 6F). They also have variable negative Eu anomalies (Eu/Eu* = 0.6 to 1.0). The mafic, calc-alkaline composition, relative strong negative Eu anomalies and slight MREE depletions in some samples suggests that significant plagioclase ± hornblende crystallization was important in the petrogenesis of these rocks before final emplacement. The lack of strong HREE and Y depletions relative to other trace elements indicates mantle melting occurred at relatively shallow depths above the garnet stability field (e.g. <65 km). All samples have very strong HFSE depletions and on plot well within volcanic arc fields on tectonic discrimination diagrams characteristic of calc-alkaline rocks associated with subduction zones (Figures 6C-D). These rocks were emplaced synchronously with a major right-lateral, ductile shearing event and thus, the strong calc-alkaline, arc-like signatures are somewhat enigmatic. We interpret the arc signature in these rocks to have been inherited from lithospheric mantle sources that had been metasomatized by prior subduction events and/or extensive crustal contamination during emplacement in the crust.

Quartzofeldspathic gneiss

The exact protolith for the quartzofeldspathic rocks is somewhat controversial. At least parts of this unit have good textural and field evidence for being meta-plutonic rocks (e.g., feldspar augen; mafic gneiss xenoliths), while other parts are interpreted as metamorphosed metavolcanoclastics. It is likely, that this unit contains rocks of intrusive and extrusive origin. Mineralogically and chemically, the quartzofeldspathic gneisses can be characterized as K_2O-rich, metaluminous (ASI ~ 0.9) hornblende-biotite granites (Figure 7A). They have distinctly A-type granite chemical characteristics defined by high K_2O/Na_2O (~2), Ba/Sr (~6), Fe/(Fe+Mg) (~0.90), total REE (~500-600 ppm), Ba (~500 ppm), Zr (400-500 ppm), Nb (20-30 ppm), Y (100-150 ppm), and low Sr, (~100 ppm), MgO (<0.5%), CaO (<2%), Cr and Ni (<5 ppm). On tectonic discrimination diagrams, they form tight clusters well within the within-plate granite (WPG) field (Figures 7E-F). REE patterns are LREE-enriched (La/Yb_N = 10), but flat through the MREE and HREE, and with strong negative Eu anomalies (Eu/Eu* ~0.30) (Figure 7B). Total REE content is very high with LREE at ~300x chondrite and HREE at ~30-40x chondrite. The lack of strong HREE depletion relative to LREE and the strong negative Eu anomalies are consistent with melting of plagioclase-bearing, garnet-free, mafic source rocks. These rocks are strikingly similar (essentially identical) in terms of mineralogy and chemistry to less deformed, A-type hornblende granites and granitic gneisses exposed 5-10 km to the west in the Greenwood Lake Quadrangle (Sonzogni et al., 2001), to the Byram Granite of the northern NJ Highlands (Volkert et al. 2000), and to the Storm King Granite in the northeastern Hudson Highlands (Rankin et al., 1993). These rocks are also chemically similar to mildy A-type granite gneisses of the AMCG and Hawkeye suites of the Adirondacks (McLelland and Whitney, 1986). The A-type affinity and similarity to AMCG suites suggests a similar origin by shallow crustal heating during syn- and post-Elzevirian lithospheric delamination and orogenic collapse.

Granite sheets

The granite sheets are high SiO_2 (~75%), leucocratic, hornblende-bearing granitoids with <5% modal mafic minerals (Figure 7A). They are metaluminous to slightly peraluminous (ASI = 0.95 to 1.1) and have highly variable K_2O/Na_2O (0.3 to 3.3) reflecting variability in the modal abundance of K-feldspar and/or Na-plagioclase as the dominant feldspar. Trace element chemistry of these rocks are distinctive from the granitic gneisses of the quartzofeldspathic unit in that the have overall much lower concentrations of most trace elements (e.g., total REE = 25-100; Y = 2-30 ppm; Zr = <125 ppm; Nb <3 ppm). This difference is also reflected on REE diagrams (Figures 7C-D) and tectonic discrimination diagrams, where the granite sheets plot scattered along the boundary between fields for syn-collisional (syn-COLG) and volcanic arc (VAG) granitoids (Figures 7E-F). These rocks are divided into two chemically distinct groups based on REE patterns. The first group has higher concentrations of total REE's, modest negative Eu anomalies (Eu/Eu* = 0.35 to 1) and either LREE enriched patterns or concave upward, "dished" MREE-depleted patterns (Figure 7C). The second group is defined by very low total REE's, strong LREE enrichment, depleted and flat MREE to HREE, and extreme positive Eu anomalies (Eu/Eu* up to 3.5) (Figure 7D). The sheets exposed at Stop 6 are of the latter type and are best interpreted as partial melts of plagioclase-free source rocks with abundant residual amphibole ± garnet coupled with fractional crystallization of quartz + feldspars ± trace element-rich accessory phases (e.g., zircon, apatite, monazite, allanite). The garnet-bearing, plagioclase-free source mineralogy implies melt generation probably occurred at deep crustal levels. In comparison, granite sheets with distinctly "dished" REE patterns clearly were generated by partial melting of garnet-free source rocks and hence melt generation probably occurred at shallower crustal levels. These two groups of granite sheets have similar field relations and appear to be part of the same magmatic event, thus crustal melting apparently occurred at various crustal levels. Based on these geochemical data, and on based on field and textural relations, we tentatively interpret these as a syn- to post-Ottawan (~1050 Ma) magmatic event related to a Himalayan-type continent-continent collision.

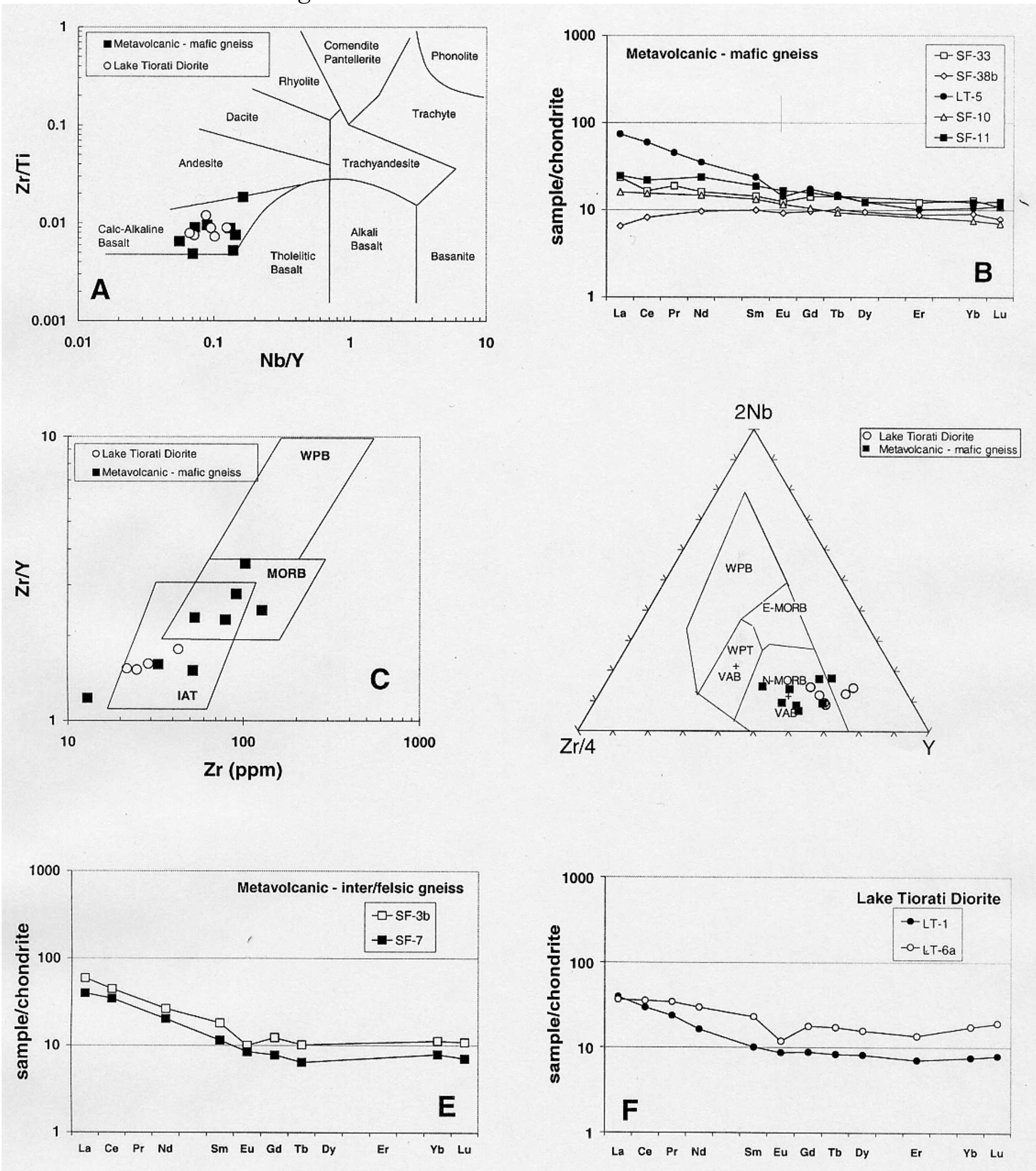

Figure 6 (A-F). Geochemical plots for amphibolites and quartz-plagioclase gneiss within the Metavolcanic Unit and diorite.

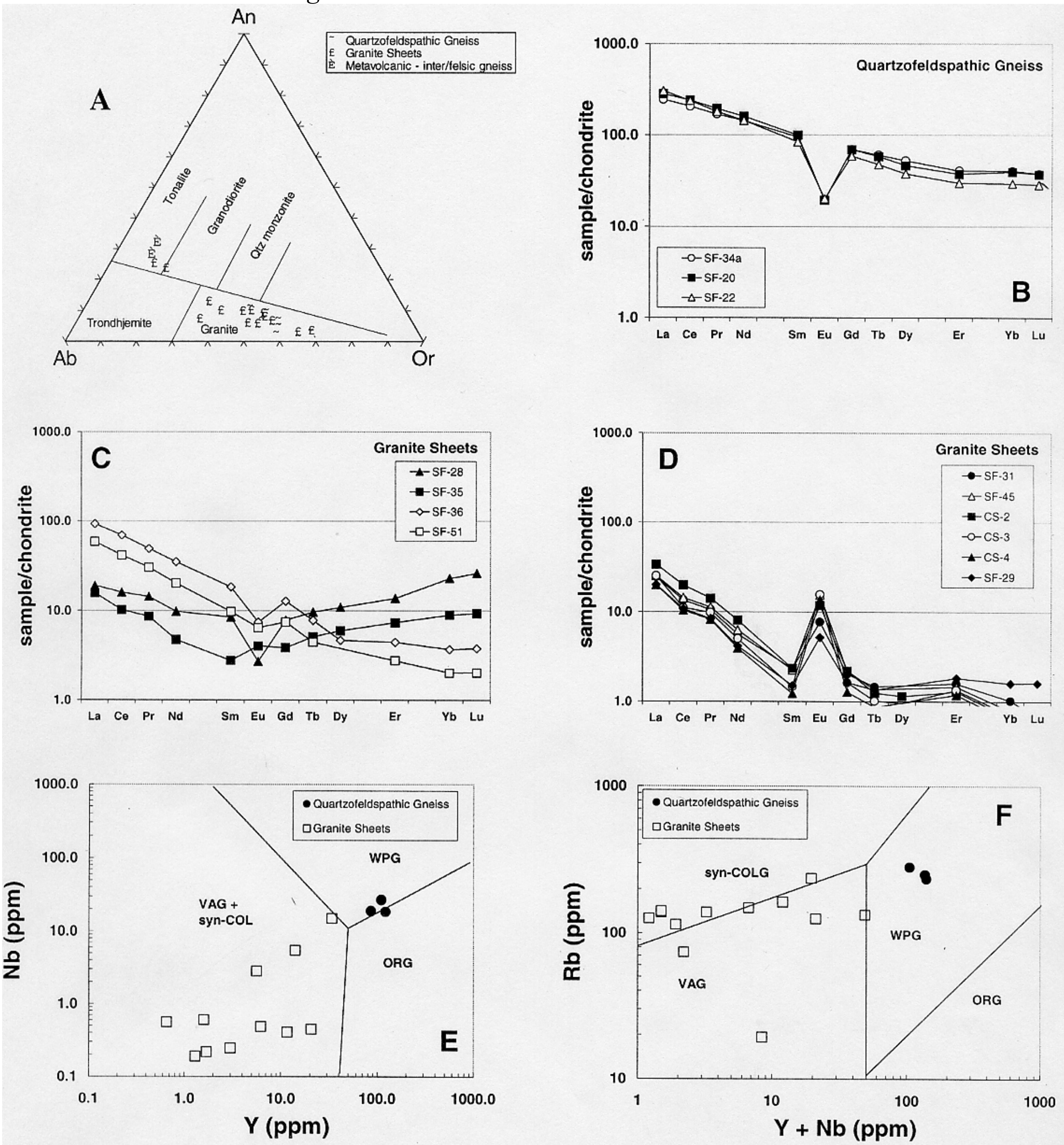

Figure 7. Geochemical plots for granite sheets and quartzofeldspathic gneiss.

Tectonic History

A subduction zone developed in the current study area about 1.2 Ga. A volcanic pile was formed in an island arc or marine magmatic arc consisting of interlayered mafic and intermediate rocks as well as associated plutons. Submarine aprons of volcaniclastic material formed along the volcanic islands and were interlayered with the volcanic rocks. These units taper away from the source. The coarse-grained volcaniclastic rocks varied in composition depending upon proximity to the volcanic source and the amount of volcanic material subject to erosion. The pelitic and calcareous rocks are in uncertain relation with the other units. By virtue of the abundance of graphite-sulfide rocks, they are interpreted to reflect a restricted and euxinic marine basin. Whether they formed during periods of volcanic dormancy, were synchronous but distant or represent a temporally separate sequence, is not clear. They are commonly interlayered with metamafic rocks. It is possible that they are back-arc basin deposits. Zircons from some of these sediments contain detrital cores with a variety of ages including possible Pinwarian (1.45 Ga) and trans-Amazonian (2.05 Ga) affinities.

The subduction sequence terminated in a collision between the volcanic or magmatic arc and another continent, likely proto-South America (Dalziel, 1991), about 1020 Ma. This collision is the Grenville orogeny (Ottawan phase) and was a Himalayan type event (Windley, 1985). It also produced severe deformation and heating of the rocks at temperatures in excess of 700^{o} C and pressures in excess of 6.5 kbars (Young and Cuthbertson, 1994). Anatexis occured locally producing the migmatites, granite sheets and the early pegmatite dikes that occur throughout the area. The intense orogenesis caused the rocks to become gneissic and produced the recumbent folds. These mesoscopic folds reflect large-scale fold nappes that were emplaced westward across the area.

Subsequent to the intense tectonism, there was intrusion of dioritic melts. Exposed in and around the study area, these bodies are dikes and small stocks. The diorite locally grades into anorthosite. SHRIMP U/Pb age determinations of zircons indicate that this event occurred at 1,008 $\pm$ 4 Ma. Geochemical data are consistent with this magmatic activity between the events having resulted from mantle delamination at the termination of the first event or the early dilational stages of the later strike-slip event.

The second event is characterized by dextral strike-slip movement during a period of rapid uplift and unroofing at approximately 1,008 Ma to 924 Ma in the study area (Gates and Krol, 1998). Thick zones of ductile deformation formed during this event and overprinted all previous features to varying degrees. Judging by the number and thickness of shear zones and the drag of some units into one of the zones, offset was significant (100s of kilometers). Late in their history, these faults were mineralized with magnetite and related minerals (uranium minerals, scapolite, pyroxene). Sheath folds within the fault and open upright folds adjacent to the fault are associated with this event. The entire area was intruded by granitic pegmatites as fault activity waned. The pegmatites are concentrated along the faults suggesting a genetic relationship. The early folds in contrast to the late dilation in the fault zones may indicate a transition from transpression to transtension during the event. A component of gravitational collapse is also possible. This second event could reflect another accretionary event but far to the north of the Hudson Highlands. A collision in the area of the Canadian Appalachians and Scandinavia may have generated tectonic escape (Tapponnier *et al.*, 1982; Burke and Sengor, 1986) of eastern Laurentia to the south along large dextral strike-slip faults that are well displayed in the Hudson Highlands (Gates, 1995). It could also just be tectonic escape as a second phase of the continental collision with proto South America similar to the scenario in the modern Himalayas. There, strike-slip overprints the contractional features produced at the onset of the collision.

Early and Middle Proterozoic were times of compressional tectonics on a global scale (Hoffman, 1988; Dalziel, 1991; Borg and DePaolo, 1994). As the Proterozoic supercontinent of Rodinia was built by the accretion of continental fragments, contractional orogens were built all along the margins and then nested into the interiors. Each accretion event reactivated adjacent old contractional zones of weakness as strike-slip faults. In this way, lithotectonic terranes escaped in directions away from the collision zones. Such extensive escape tectonism similarly occurred during the building of Pangea during the Late Paleozoic as well as in the Alpine-Zagros-Himalayan orogeny today (Burke and Sengor, 1986). If the strike-slip event in the Reading Prong-Hudson Highlands formed by tectonic escape, the locus of a major continental collision would have taken place somewhere to the present northeast of the study area. This collision would likely be the Ottawan event, a Himalayan-type collision (Windley, 1986). All terranes to the east of the dextral strike-slip faults therefore escaped to the south. A conjugate E-W striking left-lateral shear system appears to have formed synchronously in the Adirondacks of northern New York. We propose that these two systems may form a deep crustal syntaxis analogous to that which is observed in the Himalayan orogen (Gates et. al, 2001).

Conclusions

The Grenville event in the Hudson Highlands of the north-central Appalachians was formed in a four-fold tectonic scenario.

1) Deposition of volcanic, volcaniclastic sediment, and pelite-carbonate sediment within a subduction zone complex about 1.2 Ga.
2) Continental collision of the arc with another continent to the east during the building of the Rodinian supercontinent occured at about 1020 Ma. Granulite facies metamorphism, extensive pegmatite intrusion, and westward directed fold nappe emplacement accompanied this event.
3) After orogenesis ceased, dioritic to anorthositic melts intruded the area, about 1,008 Ma. Their origin is unclear but an accompanying period of extension or mantle delamination would be consistent.
4) Strike-slip shearing resulting from tectonic escape probably lasted from 1,008 to 900 Ma. There was a rapid decrease in temperature during this event resulting in the shear zones crossing the brittle-ductile transition and becoming dilational. Extensive mineralization occured within these dilational fractures.

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Road Log

0.0	Start from Parking Lot at Lamont-Doherty Earth Observatory Palisades, NY
0.1	Left on Route 9W
0.3	Right on Palisades Interstate Parkway Northbound
8.7	Exit 13 on New York State Thruway North
19.4	Exit 15A on Route 17 North. Pass through town of Sloatsburg to second stop light.
21.5	Right on Seven Lakes Drive
22.5	Right into Parking Lot at Reeves Meadow Visitors Center. Starting point for field trip. Permission for trip is required from Palisades Interstate Park Commission. See website http://harrimanrocks.rutgers.edu for a permit and a virtual field trip of these rocks.

0.0	Turn Right on Seven Lakes Drive North.
1.8	Large cut on both sides of road. Camp road at north end of outcrop (left) and spillway for Lake Sebago. Park along roadside.

STOP 1: Rocks of the metasedimentary lithofacies

The rocks at this stop include sillimanite-garnet gneiss, cordierite-sillimanite gneiss, garnet-biotite gneiss, all of these are locally migmatitic, garnet-quartz granofels, graphite-pyrite or marcasite gneiss, and quartzofeldspathic gneiss. The sulfide bearing rocks weather to a red rust color on the surface. The deformation state ranges from somewhat randomly oriented grains to mylonitic. Late warping to gentle folding can be seen on the layer surfaces. They have shallowly northeast-plunging fold axes that parallel the mineral lineation.

These rocks are representative of the low energy deposits of the sequence. They are interpreted to have formed in a restricted marine basin that was likely euxinic and with a significant volcanic input. In other areas, these rocks can contain biotite gneiss with 55% garnets, thin marble lenses, and layers of pyroxene-plagioclase gneiss that are interpreted to be of volcanic origin.

1.8	Continue North on Seven Lakes Drive.
9.1	Park on roadside just past Cedar Pond campground on right. Lake Tiorati is on the right. There is a small rock island directly across from outcrop which is on the left (west) side of the road.


Mountain
Lake Stahahe
Stahahe High Peak
Carr Pond Mtn
NURIAN
NEW YORK
SOUTHERN
STATE
Arden
Elk Pen
ARDEN SUREBRIDGE
Valley of Boulders
ISLAND POND ROAD
Green Pond Mtn
APPALACHIAN
Clove Furnace
Boston Mine
Garfield Mine
Island Pond
Black Rock
Black Rock Mtn
RAMAPO
DUNDERBERG
WHITE
BAR
ARDEN-SUREBRIDGE
CROOKED RD
ARDEN
VALLEY
ROAD
Harris Mine
Echo Lake
Lemon Squeezer
Island Pond Mtn
Echo Mtn
Goldthwaite Memorial
Surebridge
Bald Rocks
LICHEN
Ship Rock
LONG PATH
Hogencamp Mtn
Bowling Rocks
DUNNING
Little Long Pond
BOTTLE CAP
Mountain
Surebridge Swamp
3
4
Hogencamp Mine
Picnic Area
Kanawauke Circle
Lake Kanawauke
Cape Horn
Pine Swamp Mine
Pothole
Surebridge Mine
Times Square
MINE RD
Greenwood Mine
Lake Cohasset
Pine Swamp Mtn
Pine Swamp
SEVEN
LAKES
DRIVE
Fingerboard
TELEPHONE
Rock House
Lake Askoti
HARRIMAN
BEECH
RED CROSS
Camp Thendara
Cedar Pond (Group Camping)
2
Rockhouse Mtn
R-D Picnic Area
Lake Tiorati

STOP 2: Diorite Intrusion

Pluton of coarse to very coarse-grained black and white speckled diorite. On the south side of the outcrop, the diorite is equigranular in texture with random grain orientation. It contains a roof pendant of well-foliated biotite quartzofeldspathic gneiss that exhibits crenulation cleavage. The xenolith contains drag folds along it's contact with the diorite. It also contains a rim of granitic pegmatite that connects to pegmatite and quartz veins within the diorite. The diorite contains plagioclase and hornblende and clinopyroxene but with brown cores of orthopyroxene. Other phases include magnetite and ilmenite. In the northern part of the exposure, the diorite is crossed my anastomosing mylonite bands. The mylonite strikes northeast and is near vertical. Lineations plunge shallowly to the northeast. Kinematic indicators include rotated porphyroclasts and S-C fabric. Where it can be determined, shear sense is consistently dextral.

Subsequent to the first tectonic event which included the nappe emplacement and granulite facies metamorphism, there was a period of intermediate plutonism. The xenolith was deformed and metamorphosed prior to intrusion. The xenolith became more ductile as a result of the heat of the pluton. Thus drag folds formed along its edges as it fell into the magma. The magma was hot enough to cause partial melting of the rim of the xenolith, producing a granitic melt. The diorite crystallized at higher temperature than the granitic melt. Fractures opened in the newly crystallized rock and the remaining granitic melt squeezed into them forming the veins. Later deformation produced the mylonitic fabric in the diorite. This outcrop is at the eastern edge of a large dextral strike-slip shear zone with similar orientation. The dioritic Canopus pluton in the eastern Hudson Highlands is proposed to be synchronous with dextral strike-slip movement. It is dated at 1065 Ma by Rb/Sr whole rock methods (Ratcliffe et al., 1972).

9.2	Turn around at the maintenance office 200 m to the North and drive south on Seven Lakes Drive.
11.7	Drive ¼ way around Kanawauke Circle to first right, Rt. 106 west.
12.4	Park on left (south) side of road on small pull off and walk west about 100 m. Outcrop is on right (north) side of road just past bridge over the neck between Lake Kanawauke (east) and Little Long Pond (west).

STOP 3: Rocks of the metavolcanic lithofacies

Black and white, strongly interlayered mafic and intermediate gneiss with migmatitic veins. The mafic layers in the melanosome are composed of clinopyroxene, hornblende, plagioclase, magnetite, sphene and apatite. The intermediate layers are dominantly plagioclase with minor quartz, K-spar locally, apatite, hornblende and biotite. The leucosome is composed of coarse interlocking plagioclase, quartz, and K-spar and form net veins and clots. Minerals are aligned in the gneiss and granular in the leucosome.

The interlayered mafic-intermediate gneiss are interpreted as metavolcanics of island arc affinity. During the nappe emplacement event, metamorphism achieved granulite facies. Locally, the gneiss underwent anatexis and formed migmatite. Note that this rock still preserves the evidence of the first tectonic event with no overprinting.

12.4 Continue west on Route 106.
12.6 Find Parking along roadside and hike up paved service road through gate (wire). After 0.8 miles, woodland road (no unpaved) will join with Dunning Trail with yellow trail markers (turn right). After 0.2 miles mine workings will be on the left and tailings pile on the right. Mine workings extend for several hundred yards.

STOP 4: Hogencamp Mine

The Hogencamp mine was active in the 18th and 19th centuries. Magnetite was mined from the mineralized veins. The vein that hosts the Hogencamp deposit is about 6 km long and ranges in thickness from about 2 to 15 m. The wall rock is mylonitic and in this area, it is composed of quartzofeldspathic calcsilicate, amphibole-pyroxene gneiss (metavolcanic), and diopside marble. The contact of the vein with the wall rock is sharp and generally parallel to mylonitic foliation. On the small scale, however, it crosses foliation and generally the vein appears to eat into the wall rock. There is a bleached zone in the wall rock at the contact with the vein. In quartzofeldspathic rock, the bleached zone is marked by retrogression of feldspar to mica and pyroxene to amphibole. It also contains scapolite, calcite locally and apatite. The vein is composed of distinct compositional band characterized by mineral assemblages. Nearest the wall rock, there is pargasite, scapolite, K-spar, and phlogopitic biotite. The next zone in contains mainly biotite pargasite and salite. The next band is salite and pargasite. Minerals in interior zones are salite, magnetite, and calcite in that order. The salite and locally magnetite crystallized in cavities because they are euhedral and locally form doubly terminated crystals. The bulk composition of the salite and pargasite rich zones is identical to an ultramafic rock. These are metamorphically produced ultramafic rocks. The mineralized veins are intruded by very coarse grained pegmatites which locally contain xenoliths of vein material. Ar/Ar dating of the hornblende in these deposits yields 924 Ma.

The veins are interpreted to have formed in dilational joints and fractures during the waning stages of dextral strike-slip shearing. Metamorphic fluids flushed through these fractures and reacted with the wall rock. The fluids mobilized elements from the reactions with the wall rock. These reactions buffered the composition of the fluids. When these fluids encountered the right conditions either physically or chemically, they deposited the ore and gaunge minerals. With the banding of different assemblages and compositions reflects the changing chemistries of the fluids. These changes may reflect changes in flux, fluid source, or physical conditions. The pegmatites may have intruded along the same pathways as the fluids.

12.6 Return to vehicles. Continue west on Route 106.
18.1 Route 106 changes to 4-lane highway Route 17A across Route 17. Stop at the west end of first outcrop with rocks on both the median and westbound lanes of the highway. Park to the right along Route 17A.

STOP 5: Sheared quartzofeldspathic gneiss

The rock is a quartzofeldspathic mylonitic gneiss with interlayers of biotite gneiss locally. The mylonite is well foliated and lineated and composed of plagioclase, quartz, K-spar, and biotite. The biotite gneiss is composed of biotite, quartz, plagioclase, magnetite, and hornblende locally. It is well foliated and commonly folded into open to tight shallowly northeast-plunging asymmetric folds. There are pegmatite dikes that are parallel to mylonitic foliation and which commonly displays pinch and swell. There are also late pegmatites that form in "gaps" in the mylonite. At this locality there is a dike of coarse grained granite (few meters thick) with large crystals of hornblende that for radiating and linear aggretates.

This mylonite exhibits well developed kinematic indicators including S-C fabric, reverse shear cleavage (RSC), rotated porphyroclasts, and shear bands. These kinematic indicators show a consistent dextral shear sense. The width of the zone and low S-C angle indicate significant offset. Locally there are small sinistral shear zones that crosscut the main foliation and are interpreted to be conjugate. The sheared quartzofeldspathic rocks at this locality occur within the Indian Hill shear zone (Figure 3), which is only one of the zones in the anatomizing system of ductile shear zones that occurs in the Hudson Highlands.

18.3	Immediately get into the left lane of Route 17A westbound and turn left onto Eagle Valley Road from the left turn lane.
19.5	After two ponds on right (west) side of road, the first large outcrop is a road cut on the west side of the road (faces an excavated area on the east side). Park along the road.

STOP 6: Granite Sheet

Concordant leucogranite sheets intruded the quartzofeldspathic gneiss, and the rocks of the metasedimentary facies and metavolcanic lithofacies. The granite sheets range from a few meter up to 100 meters thick and contains K-spar, quartz and plagioclase, with minor hornblende, biotite and muscovite locally. Most exhibit interlocking subhedral to anhedral grains, and they are only locally foliated where the granite sheets occur near dextral shear zones. The granite at this locality is not foliated.

Most granite sheets occur within two domains inside the area of Sterling Forest. These two domains correspond to the hinge regions of map-scale upright antiforms (Figure 3). Some small granite sheets occur east of the NY Thruway.

19.5	Turn around and drive north on Eagle Valley Road.
20.7	Return to Route 17A and turn right (east), returning the way you came.
22.1	Turn left at Route 17 junction onto the service road to Route 17 North and continue onto Route 17.
23.4	Turn left onto Orange Turnpike (gas station and deli). Road will split after about one mile (large furnace on right) where the left (west) fork is Lake Mombasha Road. Remain on Orange Turnpike to right (east).

25.9 Large fresh roadcut on right (east) side of road on a right (east)-curving decline in the road.

STOP 7: Calcsilicate rocks of the metasedimentary lithofacies

Melanosome of calc-silicate gneiss is of diopside, plagioclase, quartz, K-spar and phlogopite and leucosome of pink granite veins. The rock is strongly foliated and the leucosome veins are parallel to the foliation. Later deformation is minimal. The rock is interpreted to have been a very dirty carbonate mud within the metasedimentary sequence. That sequence is dominated by metapelite. It was metamorphosed to the point of anatexis during the first tectonic event, hence the foliation in this rock is associated with the first deformation event.

25.9 Turn around and return to Route 17 South to end the trip.