

# **Neoproterozoic, Paleozoic, and Mesozoic Intrusive Rocks of Northern New Jersey and Southeastern New York**

**Field Guide and Proceedings**

**Edited by  
John H. Puffer  
Rutgers University  
and  
Richard A. Volkert  
NJ Geological Survey**

**Twenty-First Annual Meeting  
Geological Association of New Jersey**

**October 16-17, 2004  
The Ramapo College of New Jersey  
Mahwah, New Jersey**



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# **PROTEROZOIC, PALEOZOIC, AND MESOZOIC MAFIC INTRUSIONS OF NORTHERN NEW JERSEY AND SOUTHERN NEW YORK, Geological Association of New Jersey, Field Guide and Proceedings, Volume 21**

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# **Contrasting Geochemistries, Tectonic Settings, and Environmental Impacts of the Neoproterozoic, Paleozoic, and Mesozoic dike Swarms of Northern New Jersey and their Correlatives**

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## **ABSTRACT**

Three mafic magmatic events have affected northern New Jersey, each in profoundly different ways. Latest Neoproterozoic magmatism was part of a huge superplume event centered near Quebec City, Quebec that coincided with the break-up of Rodinia. Most of the Neoproterozoic magmatism is alkalic and viscous and extruded during an explosive but prolonged increase in global biodiversity. Silurian magmatism includes both alkalic and calc-alkaline varieties and coincided with the eastern North American Coastal Volcanic Province that was a major flood basalt event. However, neither alkalic nor calc-alkaline magmatism has ever been correlated with major extinction events. Early Jurassic magmatism was part of the Central Atlantic Magmatic Province (CAMP) that suddenly extruded out of major rift systems associated with the break-up of Pangea. It is tholeiitic and closely resembles the composition of Siberian flood basalt. Both the Siberian and CAMP events coincided with major mass extinctions that may have been caused in part by toxic volatiles introduced into the magmatic sources during previous subductions. Evidence is presented here that indicates that the source of at least the eastern North American portion of CAMP was the same as the source of the Coastal Volcanic Belt. Marine sulfates and chlorides introduced into the mantle during Silurian subduction were finally released during early Jurassic rifting associated with the breakup of Pangea. These toxic volatiles may have been suddenly released from low viscosity magmas on a global scale and together with the Siberian Traps may be responsible for two of Earth's five most severe global mass extinctions.

## **INTRODUCTION**

The most widespread occurrences of igneous rocks exposed in New Jersey are the Mesoproterozoic granitic rocks of the New Jersey Highlands province. Most of these granites and other Mesoproterozoic igneous rocks were intruded or extruded before or during the Grenville Orogeny and were subjected to granulite facies metamorphism. These Mesoproterozoic rocks include amphibolites that may represent basaltic and diabase protoliths. However, to the extent that these mafic (basaltic and diabasic) rocks have been geochemically modified by intense metamorphism and metasomatism they are difficult to interpret, although future studies may provide an insight into their pre-metamorphic composition, tectonic setting, and environmental impact.

Following the emplacement of the Mesoproterozoic mafic rocks, New Jersey was intruded three additional times by mafic rocks that are relatively unaffected by metamorphic alteration and offer much more potential for determining tectonic setting and environmental impact.

The first was a Neoproterozoic diabase dike swarm that was first described by Volkert and Puffer (1995). This dike swarm was determined by Puffer (2002) to be part of a major superplume event centered near Quebec City, Quebec that was involved in the break-up of the Rodinia supercontinent 550 Ma. Huge quantities of continental flood basalt were extruded across the eastern Laurentian and South American portions of Rodinia. The geographic position of Rodinia at the time is uncertain but most evidence reviewed by Puffer (2002) suggests a southern polar location. Michael Hozik (this volume) is attempting to gather new data pertaining to this question. The superplume volcanism coincided with a major expansion in global biodiversity that continued into the Cambrian.

The second post-mesoproterozoic magmatic event occurred during the Silurian. The igneous rocks include a variety of rare and strange types including cortlandite, carbonatite, and various lamprophyres, syenites, and diorites. The tectonic setting of these rocks is discussed by Eby (this volume) and Gates and Lupulescu (this volume). However, as pointed out in this chapter, the Silurian magmatism of New Jersey and New York may be a southern extension of the huge and chemically diverse Coastal Volcanic Belt (CVB) that was also active during the Silurian. The CVB magmatism, however has not been linked to any mass extinction.

The last of the three post-mesoproterozoic magmatic events to affect New Jersey was the early Jurassic intrusion of several quartz tholeiitic diabase dikes and sills including the Palisades Sill and the co-magmatic Watchung basalts. The petrology and geochemistry of these early Jurassic rocks have been described by Puffer (1992, 2003a., 2003b), Puffer and Student (1992), and Puffer and Husch (1996), and placed in a global tectonic context by Puffer (2001). They are part of a large igneous province known as the Central Atlantic Magmatic Province (CAMP) that covered eastern North America, eastern South America, northeastern Africa, and Southern Europe with basalt during the break-up of Pangea 200 Ma. They correlate with a major mass extinction and are geochemically indistinguishable from the Siberian Traps that were responsible for the largest recorded mass extinction of all time.

In addition to serving as a general background for the following chapters in this volume, this chapter is intended to offer some new ideas pertaining to the reasons for the contrasting effects that major igneous events have on the environment. For example, the Neoproterozoic superplume event and other rare but similar superplumes correlate with biological expansion. However, the Silurian magmatism is alkalic and calc-alkaline and magmatism of such composition never correlates with mass extinctions. The early Jurassic CAMP magmatism is chemically the same as Late Permian Siberian Trap magmatism and correlate with two of Earth's most extreme mass extinctions. I will propose that the principal control on environmental damage is the source of the magma in conjunction with the tectonic setting of the melting regime.

## **CONTRASTING GEOCHEMISTRIES**

The three diabase dike and sill populations exposed in New Jersey are geochemically distinct and with the possible exception of the Silurian population are each part of major regional magmatic events or large igneous provinces.

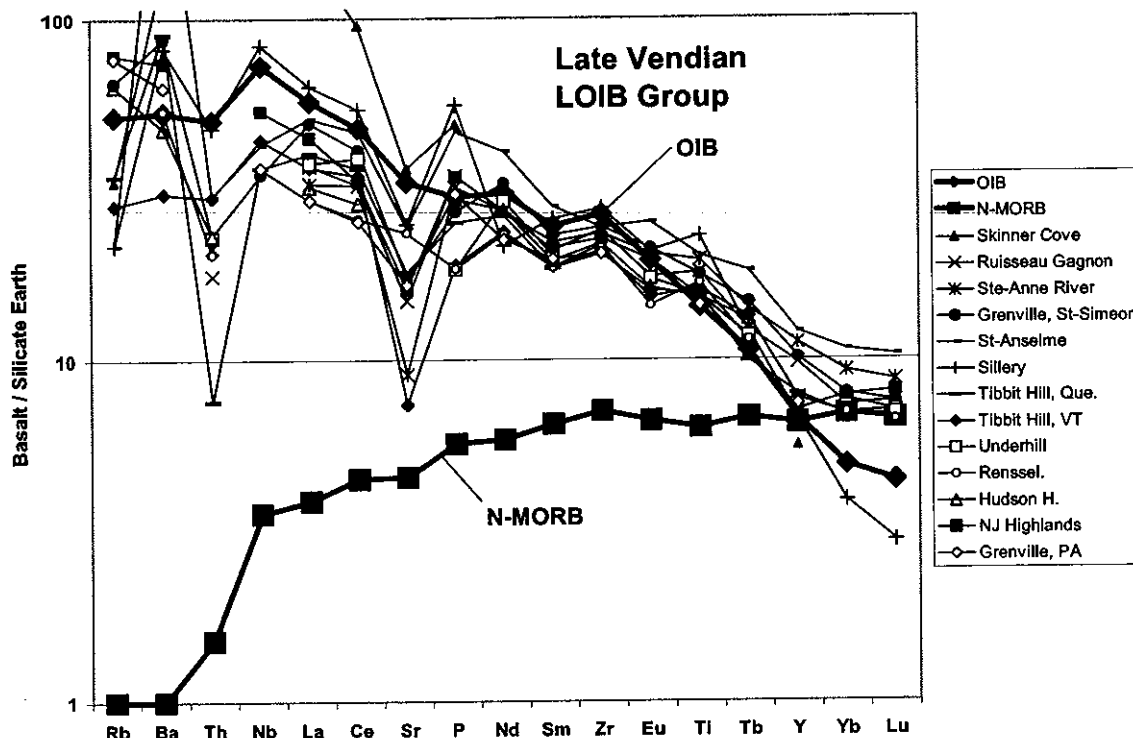
### **Neoproterozoic Magma Geochemistry**

The first of three New Jersey diabase populations is a dike swarm that is part of an important superplume related dike swarm and flood basalt that affected eastern Laurentia about 550 Ma. It is referred to by Puffer (2002) as the LOIB group. The LOIB group includes 13 chemically analyzed occurrences of Vendian eastern North American basaltic rocks. This group of subaerial continental flood basalts and submarine pillow basalts is defined by its restricted range of chemical composition resembling OIB (ocean island basalt, Fig. 1). The interval from Nd through Y (Fig. 1) is a particularly closely spaced cluster and each member of the group contains Ti, Tb, and Y levels that equal or exceed OIB levels. Each of the LOIB rock suites have been either dated radiometrically to lie within a 554 to 550 Ma range or correlate with dated 554 to 550 Ma late Vendian rocks on the basis of compelling stratigraphic evidence. The youngest radiometrically dated extrusions are found at Skinner Cove, Newfoundland dated at 550 Ma (McCausland and others, 1997) while the oldest are the Tibbit Hill basalts dated at 554 Ma (Kumarapeli and others, 1989).

Despite the highly varied degree of secondary alteration that the LOIB suites have been subjected to, the LOIB major element averages (Puffer, 2002) fall within a restricted range, particularly FeO<sub>t</sub> with

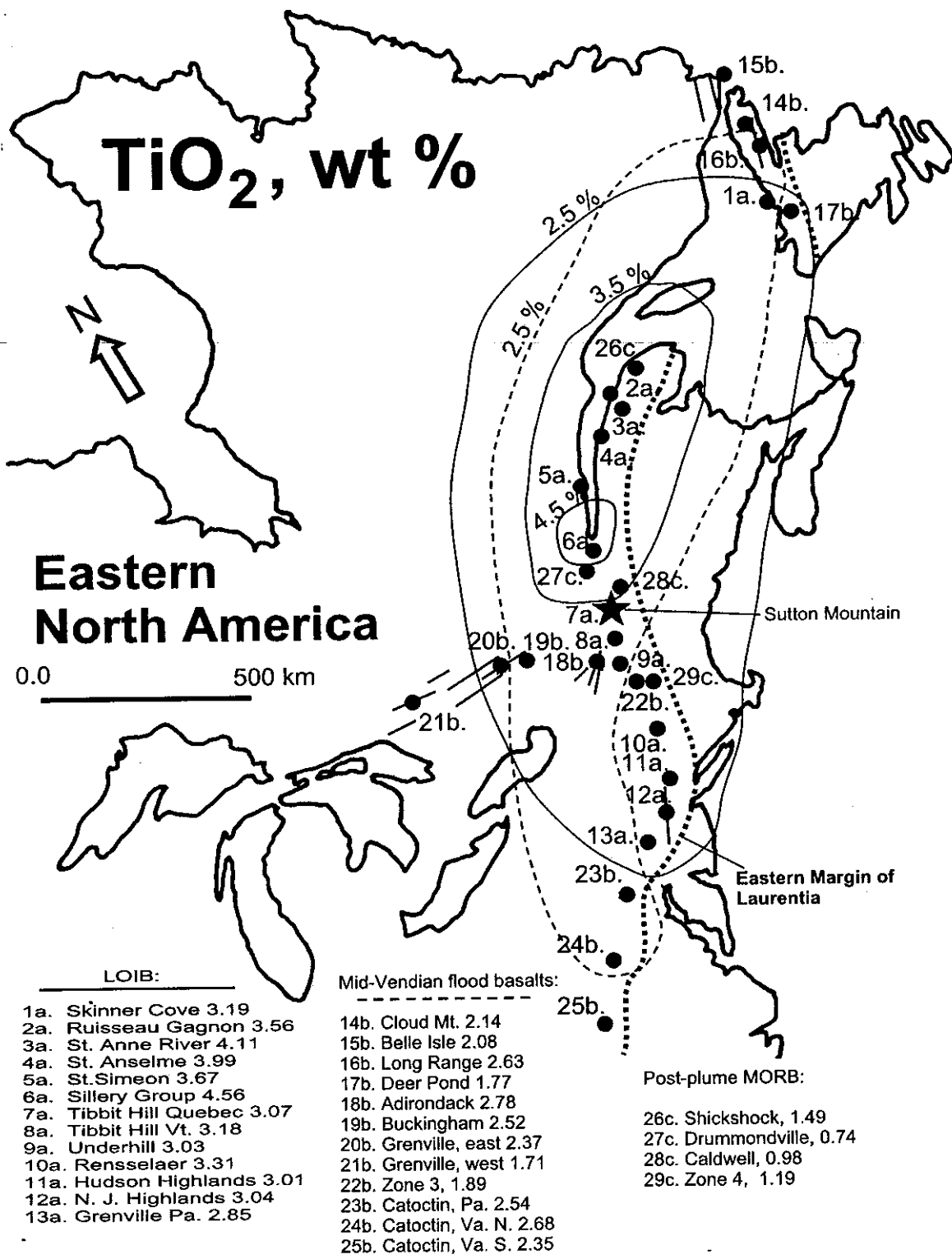


typical values of about 14 weight percent. Most trace element averages also share a restricted range among the LOIB suites (Fig. 1), particularly the HFSEs. Within each suite the standard deviation calculations (Puffer, 2002) are typically much lower for the relatively immobile HFSEs such as  $\text{TiO}_2$  and Zr than for the relatively mobile elements such as  $\text{K}_2\text{O}$ , Ba, and Sr consistent with the varied influence of secondary alteration processes.



**Figure 1.** Chemical composition of LOIB averages normalized to the “Silicate Earth” of McDonough and Sun (1995) compared to the standard OIB and N-MORB of Sun and McDonough (1989) after Puffer (2002).

In general, the chemical composition of the component members of the LOIB resembles OIB, generally regarded as plume derived. The compositional averages generate a cluster of lines that plots close to OIB on a silicate earth normalized (McDonough and Sun, 1995) spider diagram with elements arranged according to compatibility in oceanic basalt (Sun and McDonough, 1989). The degree of clustering is not perfect, with some scatter among the least compatible elements. However, the very tight clustering of Nd, Sm, Zr, Eu, Ti, and Tb (Fig. 1) clearly relates each of the averages to each other and to OIB. In addition, each of the component members of the LOIB shares similar slopes on Fig. 1, with the possible exception of the Skinner Cove basalts which contain extreme and anomalous average Nb (72 ppm), La (101 ppm), and Ce (160 ppm) contents. The LOIB component members are linked by distinct geochemical characteristics. Recognition of these characteristics makes it possible to estimate the approximate shape of the plume related magmatic event. Titanium and zirconium are generally regarded as among the least mobile and least soluble of the HFSEs. When  $\text{TiO}_2$  data (Fig. 2) and Zr data (Puffer, 2002) are contoured onto a map of eastern North America it becomes apparent that the most incompatible



**Figure 2.** Eastern Laurentian rock suites that chemically resemble OIB basalt (the LOIB group), that resemble continental flood basalts (the Mid-Vendian group), and that resemble N-MORB (the post-plume group) after Puffer (2002).

element enriched center of the LOIB is located near, but north, of Sutton Mountain, Quebec. Kumarapeli and others (1981) have determined that Sutton Mountain coincides with major positive gravity and magnetic anomalies that they interpret as due largely to Tibbit Hill volcanism. Although the current surface expression of the Tibbit Hill volcanism is restricted to a narrow belt, Kumarapeli and others (1981) conclude that their gravity and magnetic data is consistent with an 8 km thick pile of Tibbit Hill basalt that probably covered a considerable portion of eastern Canada and New England.

The shape of the contours (Fig. 2) is interpreted to approximate the shape of the superplume that generated the LOIB. Apparently LOIB magmatism was concentrated along the eastern margin of Laurentia generating a thin lensoid shape. Some component members of the LOIB may have been displaced westward during Appalachian thrusting by as much as 100 km from their original locations (Ernst and Buchan, 1997). However, most thrusting estimates, particularly southern Appalachian thrusting, are highly uncertain and probably involve less than 60 km (Alec Gates, personal communication). Since Fig. 2 has not been adjusted for Appalachian thrusting most contours would need to be repositioned eastward to plot on their pre-thrusting locations. On the scale of Fig. 2 this undetermined repositioning would not materially reconfigure the contours.

### **Silurian Dike Magma Geochemistry**

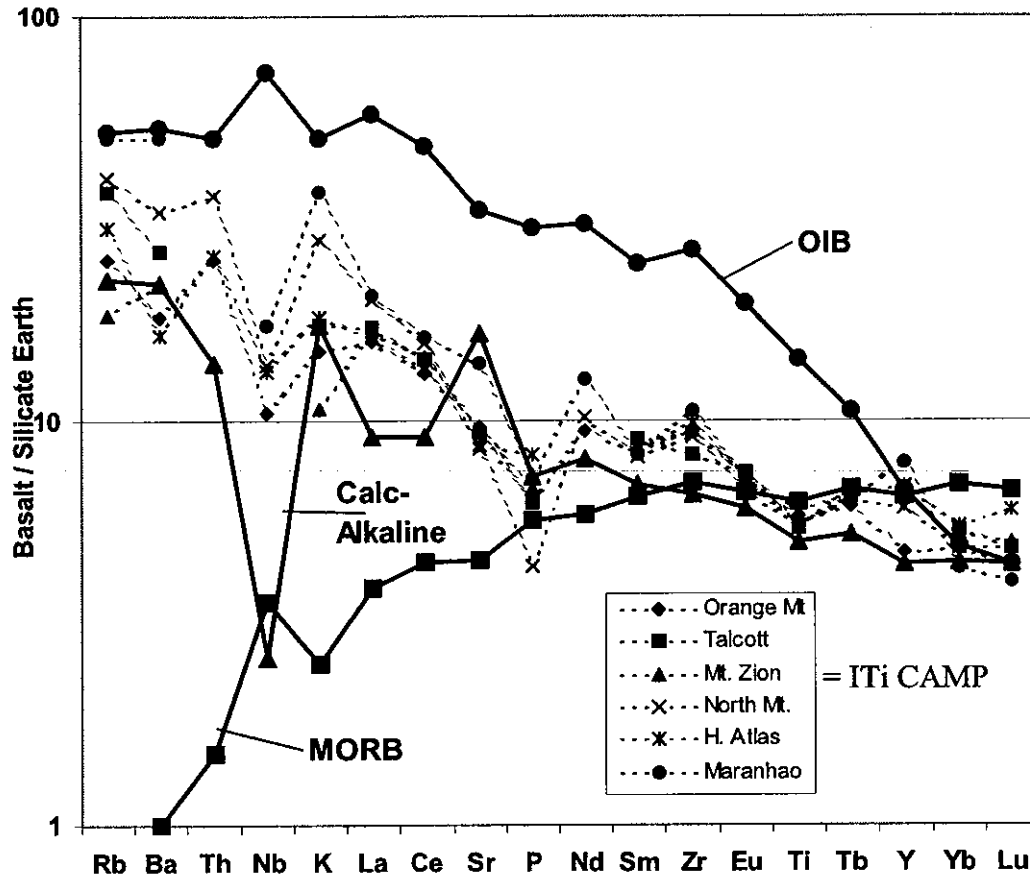
The second dike population is a geochemically diverse population of unusual rock types including norites, pyroxenites, lamprophyres, granodiorites, diorites, syenites, and extremely unusual carbonatites, and cortlandites. These rocks are very alkalic and occur along an east-west belt between Cortland, New York and Beemerville, New Jersey. The Cortland end of the complex has been dated using K-Ar by Long and Kulp (1962) as 435 Ma (minimum age), using Rb-Sr by Ratcliffe and others (1982) as 423 +/- 20 Ma, and using titanite fission-track by Eby and others (1992) as 420 +/- 6 Ma. This places the intrusions in the mid-Silurian, at the close of the Taconic orogeny which according to Hatcher (1989) ended about 423 Ma. Ratcliffe (1968) describes the Cortland complex as clearly discordant and concludes that metamorphism associated with the Taconic orogeny preceded intrusion. The geochemistry of the Beemerville end of the belt is described in detail by Eby (this volume) and the geochemistry of the Cortland end of belt (at Stony Point) is described by Gates and Lupulescu (this volume).

### **Early Jurassic (CAMP) Magma Geochemistry**

The third diabase population exposed in the northern New Jersey / Southeastern New York area occurs as thick sills and less common dikes and a prominent volcanic neck. These diabase occurrences are each quartz tholeiites and are part of the Central Atlantic Magmatic Province (CAMP) that extends north into Nova Scotia, south across eastern North America into Brazil, and east into North Africa, and southern Europe (Marzoli et al., 1999). The diverse Mesozoic intrusive and extrusive rocks included within the boundaries of the CAMP province generally fit into three chemistries as defined by Salters et al. (2003): 1.) Initial rift-related, intermediate titanium ITi-type CAMP; 2.) Secondary drift-stage, low titanium LTi-type CAMP and 3.) Early and late stage high titanium HTi-type CAMP basalts. The diabase exposed in the northern New Jersey / Southeastern New York area included types one and two.

#### *1. Initial rift-related CAMP.*

Only one basalt chemistry is common to the whole CAMP province. The ITi basalt type, first defined as the High-Ti type (HTQ) by Weigand and Ragland (1970) then redefined as ITi by Salters et al. (2003) is exposed as thick extrusive flows on each of the circum-Atlantic continents. This initial extrusive basalt composition represents the peak of CAMP magmatic activity and is represented in the



**Figure 3.** Basalt incompatible element compositions normalized to the Silicate Earth (McDonough and Sun, 1995) and arranged according to compatibility in oceanic basalt. Standard Ocean Island Basalt (OIB) and normal Mid-Ocean-Ridge Basalt (MORB) from Sun and McDonough, (1989), and standard Calc-alkaline basalt sample 272825 from Puyehue within the southern active volcanic zone of the Andes Mountains after Hickey et al. (1986) are compared to the average CAMP values of Puffer (2001).

northern New Jersey / Southeastern New York area as the Palisades sill and related offshoots including Laurel Hill.

Most of the chemical characteristics of ITi type CAMP basalt are displayed on spider diagrams (Fig. 3) that have been normalized to the silicate earth (McDonough and Sun, 1995) with elements arranged according to incompatibility in oceanic basalt. Probably the most obvious characteristic displayed by Fig. 3 is the extreme degree of chemical uniformity among the initial CAMP basalts. The degree of uniformity is particularly close within the range of high field strength elements Nd through Ti (Fig. 3). Major element uniformity may be related to depth controlled eutectic melting in response to extensional Pangean break-up decompression and similar province-wide rapid movement to the surface through rifts. Trace element uniformity may be related to central Pangean mantle homogeneity and a similar degree of subduction related interaction.

Another important observation from Fig. 3 is the lack of any similarity of rift related ITi CAMP with the standard ocean island basalt (OIB) or normal mid-ocean-ridge basalt (MORB) of Sun and McDonough (1989). Compared to N-MORB, CAMP basalts contain similar HFSE concentrations but much higher LIL values (Fig. 3). The standard E-MORB of Sun and McDonough (1989), a hybrid magma type, intermediate between N-MORB and OIB, is a good match for CAMP except for its positive Nb anomaly with respect to Th and K. A positive Nb anomaly is a consistent characteristic of N-MORB, E-MORB and OIB magmas; however, ITi CAMP basalts consistently display negative Nb anomalies.

The best fit for CAMP basalt among the three genetic choices of Fig. 3 is Calc-alkaline, although it is recognized that ITi CAMP magma is tholeiitic. Almost any typical calc-alkaline basalt including typical back-arc-basin-basalt (Puffer, 2003) derived from any of the worlds continental marginal arc systems would plot close to CAMP. The Calc-alkaline basalt plotted onto Fig. 3 is commonly used as a calc-alkaline standard (for example, Wilson, 1989). It is sample 272825 from Puyehue within the southern active volcanic zone of the Andes Mountains after Hickey et al. (1986).

## *2. Secondary LTi drift stage CAMP.*

The ITi flows of eastern North America are locally overlain by a thick sequence of basalt flows defined by as LTi by Salters et al. (2003). These secondary basalts extruded within 580 thousand years of the underlying ITi flows (Olsen et al., 1996). The LTi flows and related dikes occur throughout the central and southern portions of the eastern North American sub-province. However most olivine normative LTi dikes are confined to the southeastern US.

Secondary drift stage CAMP is rarely exposed within the northern New Jersey / Southeastern New York area as intrusive diabase but is exposed as extrusive Preakness Basalt.

## *3. REE enriched dikes (and flows)*

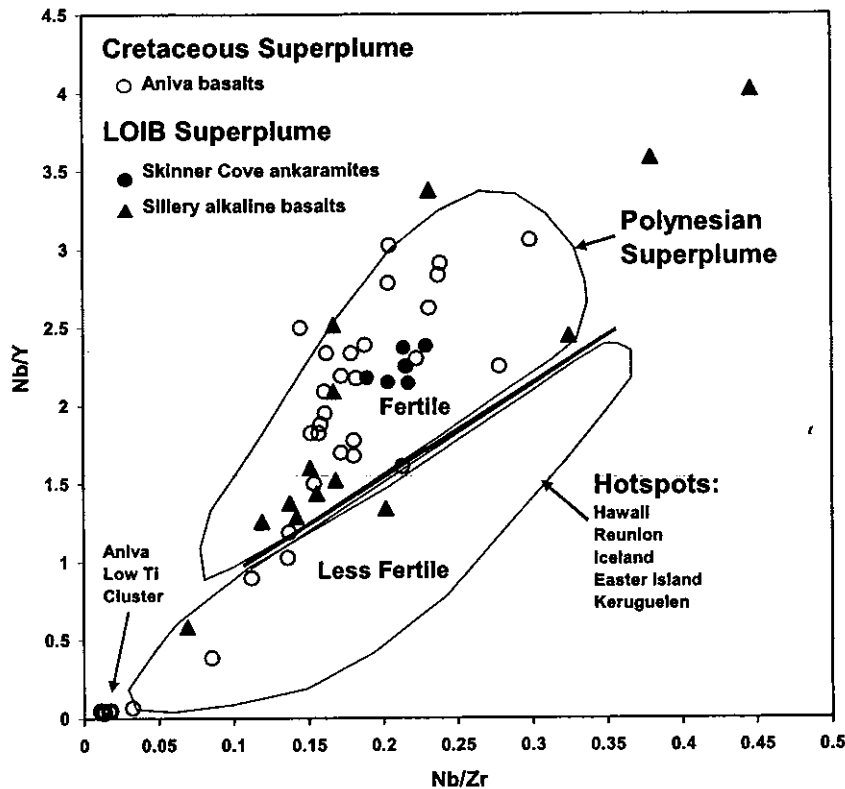
Several Mesozoic and undated intrusions and extrusions characterized by  $\text{TiO}_2$  contents exceeding 2 %, (HTi) and REE contents resembling plume related ocean island basalt, occur within the boundaries of the CAMP province. However, it is unlikely that any of these rocks are genetically related to initial CAMP magmatism. Examples include portions of the Liberian dike swarm (Dupuy et al., 1988), the dikes of the eastern Maranhão province and the Cassipore dikes of Brazil (De Min et al, 2003), the Mesozoic dikes of the coastal New England (CNE) igneous province (McHone, 1992), and some thin upper or "group 3" flows in Morocco including those described by Manspeizer and Puffer (2000).

## **CONTRASTING TECTONIC SETTINGS AND MAGMA SOURCES**

### **Neoproterozoic Tectonic Setting and Magma Source**

The identification of the LOIB group as a superplume is based on extensive geochemical evidence. The center of the LOIB group where volcanic thicknesses reached 8 km thick is particularly diagnostic.

Tatsumi and others (1998) have compared the chemical composition of Cenozoic basalts from the Polynesian superplume with mid-Cretaceous superplume basalts from the western Pacific. Although Polynesian superplume basalts have been characterized as HIMU basalts (high  $\mu$ ;  $\mu = ^{238}\text{U}/^{204}\text{Pb}$ ) by Kogiso and others (1997), Tatsumi and others (1998) point out that Pb isotope geochemical signatures might not be applied to old, altered, and/or metamorphosed samples because of compositional changes during secondary processes. Tatsumi and others (1998) suggest, instead, that HFSE elements are a more appropriate means of comparison because they are less mobile and less subject to fractionation processes. They suggest that plots of Nb/Zr vs. Nb/Y ratios are particularly useful and show that many of the mid-Cretaceous basalts from western Pacific sites plot in the same compositional field as the Polynesian basalts.



**Figure 4.** The Nb, Zr, and Y data from LOIB basalts plotted onto a modified Tatsumi and others (1998) diagram for hotspots and plateau lavas in Pacific, Atlantic and Indian Oceans after Puffer (2002).

Similarly, the composition of basalts from the geographic center of the LOIB plot within the same field described by Tatsumi and others (1998) as the superplume field (Fig. 4) although LOIB basaltic compositions, in general, become progressively less fertile at distal geographic positions. The Nb/Y and Nb/Zr ratios of 14 samples of Sillery basalt located at the LOIB plume center, as analyzed by Olive and others (1997) and Vermette and others (1993) overlap the entire superplume field of Fig. 4. In addition, the Nb/Y and Nb/Zr ratios of 7 samples of ankaramite from the Skinner Cove suite located near the northern end of the LOIB lensoid shape (Fig. 2) plot within the center of the superplume field. The Skinner Cove ankaramite flows are much less fractionated than the other Skinner Cove flows and constitute a distinct sub-suite of relatively uniform composition (low standard deviations).

The Nb/Y and Nb/Zr ratios of superplume basalts are unusually high among terrestrial basalts. Although similar levels of Nb enrichment are reached whenever alkalic magma undergoes high degrees of fractionation, another important characteristic of superplume basalts are their elevated compatible element contents (MgO, Cr, Ni) resulting in a rare combination of elevated compatible and incompatible elements. Each of the superplume basalts is rich in MgO and Ni and represents primary superplume melt. There is a remarkable chemical similarity shared by the four superplume basalt suites.

The chemical composition of basalt from each of the three superplumes plotted onto Fig. 4 is unlike basalt from any other source. Basalt from Earth's largest continental flood basalt provinces, including each of the 401 samples from the 10 flood basalt provinces described by Puffer (2001), are less enriched in Nb at comparable MgO levels than the superplume samples plotted onto Fig. 4. In addition, each of several hotspot-generated ocean island basalts including Hawaii, Reunion, Iceland, Easter Island and Keruguelen plot in a field described by Tatsumi and others (1998) as "less fertile" than the "fertile" superplume field (Fig. 4).

Kosigo and others (1997) suggest that the geochemical signature of Polynesian basalt is consistent with a source reservoir near the core-mantle boundary. They also suggest on the basis of isotopic data that the lower mantle Polynesian basalt reservoir incorporated subducted oceanic crust before ascending back to the Earth's surface as mantle plumes. The incorporation of subducted crust would explain the enrichment of Nb into the lower mantle reservoir. Nb is retained by subducted crust as partial melting and mantle metasomatism deplete subducted crust in other incompatible elements. This retention results in the negative Nb anomaly that is a characteristic of andesites and arc basalts.

### **Silurian Dike Tectonic Setting and Magma Source**

The extremely alkalic and REE enriched chemistry of the Cortland – Beemerville belt is geochemically consistent with anorogenic magmatism generally associated with hot spot tracts that have intersected continental plates or island arcs. Most of the world's hot spot tracts are identified on the basis of the occurrence of similar, although, typically less extreme magmatism. In addition to the alkalic chemistry support for a hot spot tract interpretation includes the linear nature of the belt that cross-cuts the prevailing NE strike of the Appalachian plate sutures and metamorphic foliation. Ratcliffe (1981) recognized the possibility that the Cortland Complex was intruded over a mantle plume but concluded that the Cortland Complex intruded along the junction between two plates in motion during the Taconic. He proposed that the magma was generated in response to transform stresses that ruptured the crust and mantle.

A mantle plume magma source would imply extensional tectonism. However, Gates and Lupulescu (this volume) present geochemical evidence for compressional tectonism and plot new geochemistry on discrimination diagrams that indicate volcanic arc and syncollisional magmatism. Compressional tectonism was also active in New England and eastern Canada during the mid Silurian. It may, therefore, be significant that the Cortland – Beemerville belt was intruded during major igneous activity throughout New England and eastern Canada along the western edge of the Avalon Superterrane (Fig. 5). Therefore, there is a good possibility that Cortland- Beemerville magmatism may be a southern extension of The Coastal Volcanic Belt (CVB).

The Coastal Volcanic Belt (CVB) of New England and New Brunswick is a Siluro-Devonian bimodal (mafic-felsic) province (Fig. 5). It is one of the largest bimodal volcanic provinces in the world (Dadd et al., 2000), and is over 4000 m thick.

Some of the most carefully mapped exposures are located in the Passamaquoddy Bay area of southwestern New Brunswick, the Machias-Eastport and Penobscot Bay areas of coastal Maine, and the Newbury area of northeastern Massachusetts (Fig. 5). In the Passamaquoddy Bay area Dadd et al. (2000) describe four cycles of mafic-felsic volcanism that Van Wagoner and Dadd (2003) determined extruded from 419 to 436 Ma, suggesting that volcanism there was ongoing for a minimum of about 10 Ma.

Gates and Moench, (1981) describe three Late Silurian volcanic formations in the Machias-Eastport area (the Dennys, Edmunds, and Leighton Fms.) an Early Devonian volcanic formation (the Eastport Fm.) and a younger-Devonian post-Acadian flow (the Perry Formation). Gates and Moench (1981) conclude that volcanism lasted 20 M.y. in the Silurian, and a few more million years in the Devonian.

Further south, the 750 m thick Siluro-Devonian Thorofare Andesite is well exposed along the south shore of North Haven Island in Penobscot Bay (Fig. 5).

The CVB flows of the Newbury Formation have been mapped by Shride, (1976) and chemically analyzed by Hon and Thirlwall, (1985) and McKenna et al. (1993). Robinson and Hall (1980) interpret the Newbury volcanism of the Nashoba and Boston tectonic terranes of Massachusetts (Fig. 5) as the product of Silurian subduction related accretion. Paleontological evidence indicates that over 3000 m of Newbury rhyolite, basalt, and andesite volcanism continued post accretion until late Silurian to earliest Devonian time, (Shride, 1976). The basalt and andesite portions of the Newbury complex are

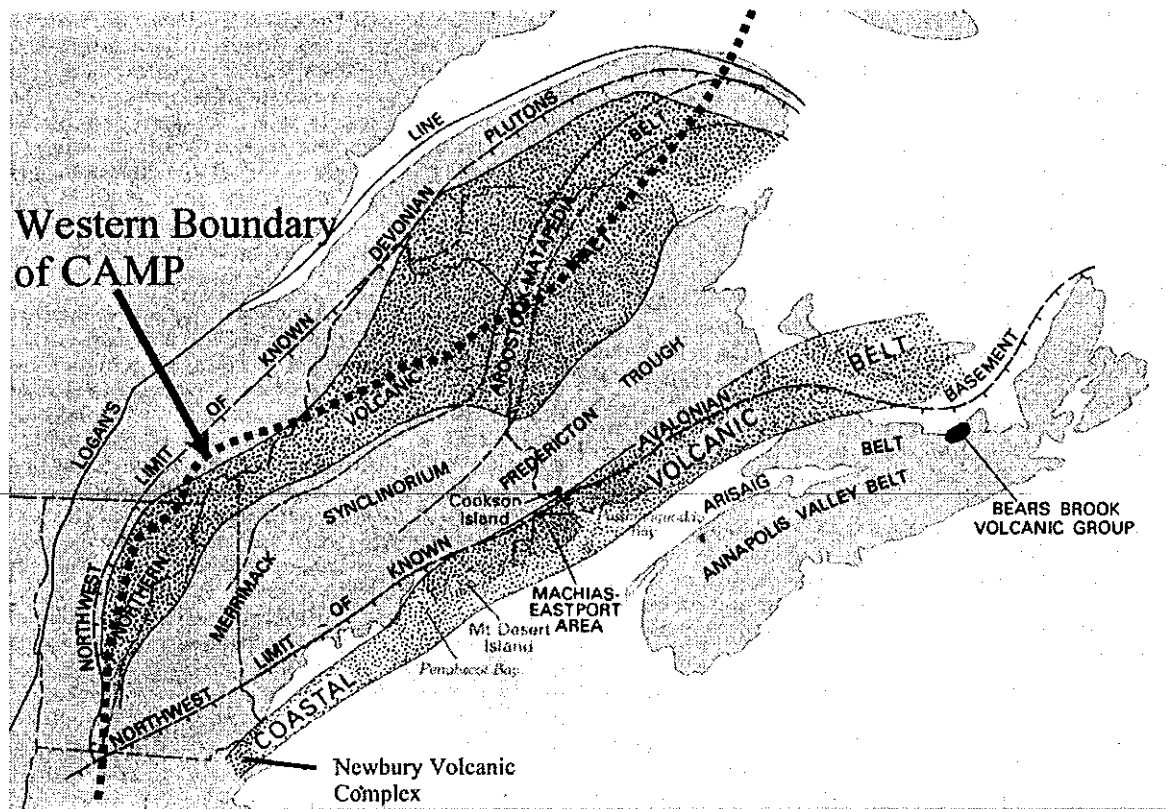


Figure 5. Map of New England and Maritime Provinces modified after Gates and Moench (1981) showing the Silurian and Lower Devonian CVB and the western boundary of the CAMP after McHone (<http://jmchone.web.wesleyan.edu>). The Silurian Cortland complex of New York and New Jersey may be a southern extension of the CVB.

calc-alkaline and display distinct negative Nb anomalies on spider diagrams (McKenna et al., 1993).

At each of these locations the CVB flows are highly diverse and consist of three general lithologies:

1) rhyolitic flows that have a clear arc affinity on granitic discrimination diagrams (Pinan-Llomas, and Hepburn, 2002).

2) basaltic andesites that are calc-alkaline with strong negative Nb anomalies on spider diagrams, rich in  $\text{Al}_2\text{O}_3$  (16 to 18 weight percent), and low in Zr and  $\text{TiO}_2$  (about 100 ppm and 1.2 wt %) interpreted as generated in a subduction related environment (Pinan-Llomas and Hepburn, 2002; Hon and Thirlwall, 1985) consistent with a convergent arc or complex extensional back-arc tectonic setting.

3) highly fractionated basalt characterized by low MgO content (4-6 wt %), high Zr and  $\text{TiO}_2$  (about 200 ppm and 2 wt %) interpreted on the basis of several discrimination diagrams as “within-plate” continental basalt (Gates and Moench, 1981; Pinette, 1983; Pinan-Llomas and Hepburn, 2002; Dadd et al., 2000; Van Wagoner and Dadd, 2003). Although most of the “within-plate” tectonic interpretations are based largely on discrimination diagrams, including Pearce and Cann (1973), an alternative explanation might be that early CVB arc and back-arc volcanism was followed by transtensional magmatism as proposed by Barr et al (2002). These highly fractionated basalts are younger than the calc-alkaline flows that preceded them.

The early CVB flows display clear calc-alkaline characteristics, such as the Leighton Thorofare, and Newbury andesites. During the end stages of CVB volcanism, after Avalonia was docked and subduction processes were replaced by transtensional processes, the character of the magmatism changed. The titanium and zirconium enriched late CVB “within plate” basalts may, therefore, have been



influenced by cooling temperatures and lower degrees of partial melting. To the extent that the late (Devonian) CVB magmatism was generated by shallow or transtensional processes the source was clearly independent from the early (mid-Silurian) CVB magmatism.

Although Cortlandt magmatism is alkalic, Gates and Lupulescu (this volume) have shown that it is nevertheless consistent with a compressional environment like a magmatic arc. This is also consistent with early mid-Silurian subduction related CVB (type 2, above) arc magmatism that was related to subduction under Avalonia (Figs. 5 and 6). The geochemical data of Gates and Lupulescu (this volume), therefore, supports a tectonic relationship between Cortlandt and early CVB magmatism during the mid-Silurian.

### **Early Jurassic (CAMP) Tectonic Setting and Magma Source**

The lithologies and aerial distribution of CAMP rocks have been defined and described by Marzoli et al. (1999), McHone (2000), and Puffer (2001). The tectonic setting was clearly extensional and the magmatism was very short-lived (Olsen et al., 1996; Hames et al., 2000; Et-Touhami et al., 2001; Olsen et al., 2003). These are characteristics of the type of continental flood basalts described by Puffer (2001) that were derived from mantle sources enriched by previous subduction.

Most CAMP magmatism occurred during the initial stages of Pangean break-up along the plate boundaries that separated to form the Atlantic Ocean. The distribution of most CAMP magmatism, however, also approximates the boundaries of plate sutures that closed during the assembly of Pangea, perhaps even more closely than the break-up boundaries. For example, the Paleozoic suture along the boundary of the Piedmont and Avalonian terranes is closer to most CAMP dikes and sills than is the current continental margin of North American. Therefore, the rifts that were the source of CAMP magmatism are commonly described as “failed rifts” and CAMP may be just as accurately described as the Circum-Iapetus-Magmatic-Province. The key to the understanding of the source of CAMP magma may have more to do with what was happening along the Iapitan sutures than what was happening under the initial Mid-Atlantic-Ridge.

There are at least two competing theories pertaining to the source of the CAMP. Some petrologists favor a source that involves an upwelling deep mantle plume or superplume modified to varying degrees by interaction with lithospheric rock while others favor a source that was independent of a plume, generated instead by decompression melting, perhaps edge-driven melting, along previously developed plate sutures.

#### *A Plume Source*

Most published models of continental flood basalt petrogenesis depend heavily on a plume or hot spot to initiate volcanism that is modified by highly varying amounts of contamination or mixing with sub-continental lithospheric material. For example, each of the continental flood basalts (CFBs) described in the AGU Monograph “Large Igneous Provinces: Continental, Oceanic, and Planetary Flood Volcanism” edited by Mahoney and Coffin (1997), with one possible exception, are interpreted as the product of deep mantle plume induced melting. The exception is Sharma (1997) who considers the non-plume or “perisphere” source proposed by Anderson (1994) as a source for the Siberian Traps although he indicates that the existence of the perisphere “...as a viable mantle source is not accepted by many geochemists.”

Several geologists have also proposed that CAMP or at least large portions of CAMP were derived from deep mantle plumes. For example, Wilson (1997) and Ernst et al. (2003) point out the large radiating dike pattern of CAMP dike swarms as evidence of a large plume or super-plume. And Greenough and Hodych (1990) interpret the pattern of lateral dike flow as evidence of a plume centered opening of Pangea. CAMP magma was interpreted by Pe-Piper et al. (1992) as related to a hotspot on geochemical grounds based on the progression from alkalic early (or pre?) CAMP to tholeiitic mid-CAMP magmatism. In addition, Oliveira et al., 1990 has shown that the chemistry of HTi Amapa and Jari

dike swarms of northern Brazil geochemically resemble plume related ocean island basalt magmatism, as do the HTi Cassipore dikes of northern Brazil (De Min et al., 2003). A mantle superplume (the Cape Verde plume) has also been proposed as an important influence on CAMP magmatism (Wilson, 1997). The difficulty with a plume source application to CAMP is fivefold:

1. Continental flood basalt extruded from a plume head or superplume might be expected to be compositionally symmetrical around the center of the plume with plume characteristics inversely proportional to distance. Instead rift related ITi CAMP basalt, as indicated above is extremely uniform. Recent data from Brazil (De Min., 2003) and Africa (Olsen et al., 2003), confirm this uniformity. In addition, as pointed out by McHone (2000) a major plume responsible for all or most of CAMP is inconsistent with the absence of a hotspot tract that would account for the northward drift of Pangea during the Mesozoic.

2. Several interpretations of CAMP as plume related are based on analyses of OIB type flows that are atypical of the province, particularly late and early flows. McHone et al. (1987) present compelling evidence that the early alkalic suite of Triassic (CAMP?) dikes exposed along the coast of New England, as well as subsequent alkalic early Cretaceous dikes and seamounts, are the product of parallel leaky transform fault zones oriented perpendicular to the eastern US coast line.

3, 4. Hames et al. (2000), McHone, (2000), and Puffer, (2001) argue that CAMP is inconsistent with plume magmatism because of a lack of evidence for associated regional Mesozoic uplift and the short duration of volcanism. Plume related flood basalts are instead characterized by prolonged volcanism over several million years.

5. Puffer (2001) has shown that the chemistry of rift-related ITi CAMP is unlike any of the Earth's major continental flood basalts that are clearly plume related, such as the North Atlantic Magmatic province, the Deccan, or the Ethiopian flood basalts. CAMP, as indicated above, is chemically more closely akin to calc-alkaline basalt.

#### *A reactivated calc-alkaline (enriched lithospheric) source*

The alternative to a plume source for ITi CAMP is derivation from a mantle wedge that was chemically modified by interaction during a previous subduction cycle. Arguments made by Puffer (2001, 2003) are based in part on the good fit that CAMP has with the non-plume or edge-driven model of King and Anderson (1995) as also proposed by Hames et al. (2000), and De Min et al. (2003). King and Anderson (1995) propose an upper mantle flood basalt source located under thickened continental lithosphere and partial melting triggered by "edge-driven" convection near plate sutures characterized by different lithospheric thickness. The relative thickness of the Paleozoic terranes that underlie eastern North American CAMP (Fig. 6) is uncertain but countless Paleozoic arching, rapid erosion, back-arc thinning and arc thickening processes have effected some of these terranes more than others.

A CAMP mantle source previously influenced by subduction was first proposed by Pegrarn (1990) on the basis of isotopic evidence and supported by additional geochemical and isotopic data presented by Puffer (1992). Both sets of data indicate a close resemblance of several eastern North American CAMP basalts to calc-alkaline volcanism.

More recently Puffer (2003) supports a similar subduction modified source but one that more closely resembles back-arc basin basalt (BABB) than arc basalt. A back-arc basin basalt (BABB) source interpretation for most CAMP basalt, is based largely on trace element data, particularly high field strength elements (HFSEs). Woodhead et al. (1993) have shown that arc basalts have lower average concentrations of incompatible HFSEs and Y, and consistently higher Ti/Zr ratios than BABB.

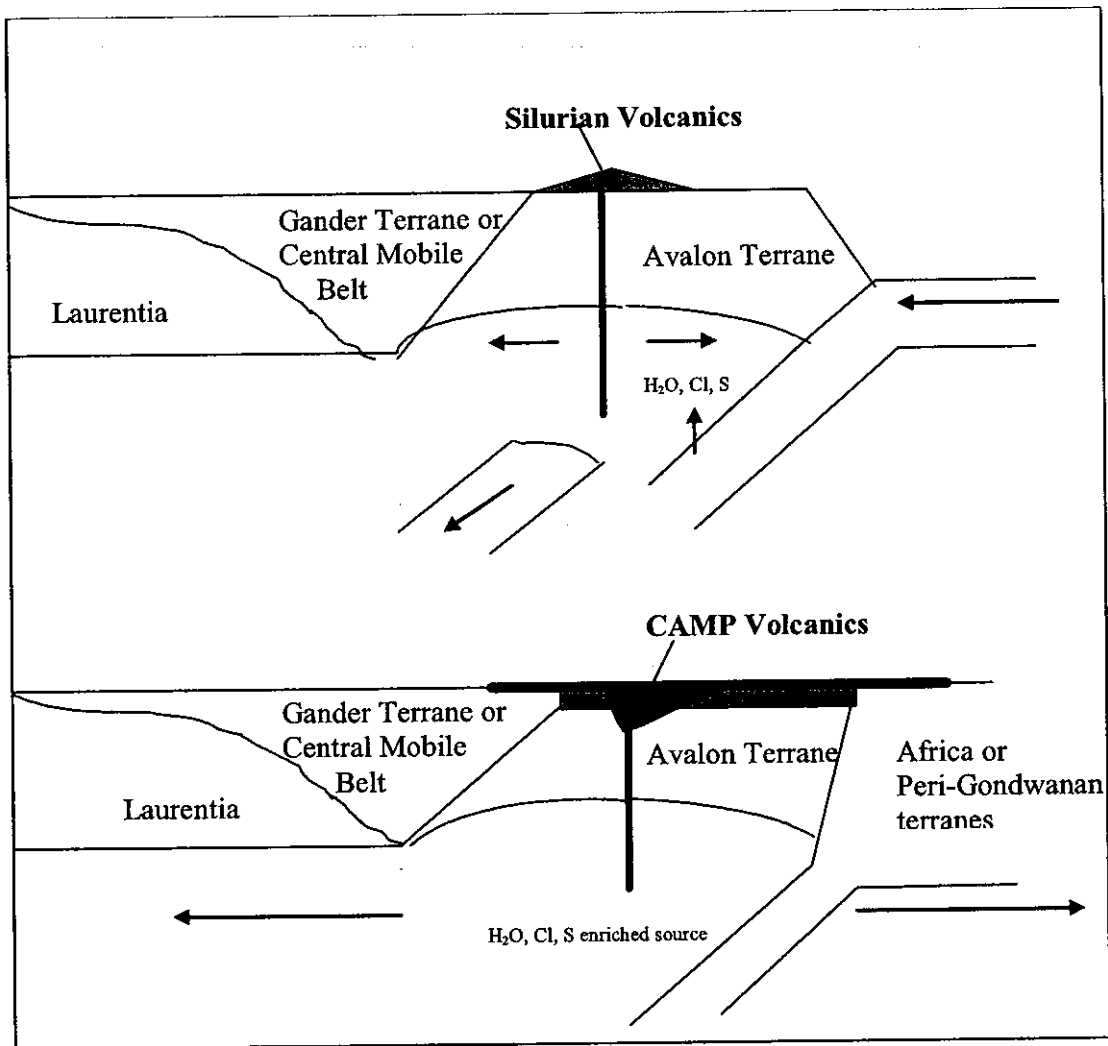
The source of both arc and BABB is mantle that has been modified to varying degrees by subduction processes. It will be interesting to determine if both CVB and CAMP volcanism resemble each other to the degree of detail permitting distinction between arc and BABB.

Ewart et al. (1994) show that BABB magmatism can develop in an extensional tectonic setting. They describe the basin between the Lau Ridge and the Tonga Ridge as it occurred 3 million years ago as a "basin and range" style basin. The Mesozoic basins of Eastern North America are commonly described

in similar terms. In addition, some recent studies of the northern portion of the CVB including Barr et al (1998 and 2002) suggest similar extensional tectonism.

Since the early work of Pegram (1990) and Puffer (1992) more recent data presented by De Min et al. (2003) and Salters et al (2003) also propose a ITi CAMP source that may have been influenced by subduction. In particular, Salters et al. (2003) suggested that both the ITi and LTi magma types have relatively high concentrations of subduction related source components compared with most continental flood basalt. De Min et al. (2003) suggest that Brazilian CAMP magmatism is concentrated along Proterozoic mobile belts at the boundaries of the Amazonia craton and may have been derived from sources enriched during Proterozoic subduction.

However, not enough is currently known about the tectonics of Pangean assembly to rule out Paleozoic subduction activity along each of the CAMP sutures subsequently reopened during Pangean break-up. It has also been proposed (Pegram, 1990) that Proterozoic subduction may have modified the source of CAMP. However, it is likely that most of the mantle wedges active



**Figure 6.** Proposed tectonic model illustrating a similar back-arc source for both the Silurian calc-alkaline flows (such as the Leighton) and for CAMP ITi basalts 200 Ma later. Duration of CAMP volcanism was constrained by the limited flux content of the source.

during the Proterozoic were overridden or depleted during considerable early and mid-Paleozoic continental plate motion and calc-alkaline magmatism during the assembly of Pangea

Instead, it is here proposed that the source of CAMP was mantle enriched during Paleozoic subduction. Evidence is presented that at least the eastern North American portion of CAMP was mantle enriched during the Silurian subduction described above that coincided with the extrusion of the CVB.

McKerrow and Ziegler (1971) and Dewey and Kidd (1974) were among the first to describe the CVB as a volcanic arc above a subduction zone. This interpretation is consistent with most tectonic models that describe subduction related to the accretion of Avalonia during the early stages of the Acadian orogeny. Although there are several competing models, most describe Avalonia as an exotic terrane that can be traced from Nova Scotia to the Southern Appalachians that was accreted in a convergent environment that included subduction as it approached (Hatcher, 1989; Cook et al. (1979); Robinson and Hall (1980). Fig. 6 is a plausible cross section through Avalonia that is consistent with most of these published models. However some tectonic models, including Pinan-Llamas and Hepburn (2003) suggest that the CVB was formed on the Avalonian margin during plate conversion above a SE-dipping slab. Pinan-Llamas and Hepburn (2003) conclude that declination of a downgoing lithospheric slab and related decompression led to the extensional tectonic regime described above by the time Devonian magmatism began. The evidence in support of eastward vs. westward dipping slabs is inconclusive. However in either case the mantle beneath Avalonia was chemically modified by subduction.

The subduction zone depicted in Fig. 6 includes a mantle wedge that was probably the source of at least the Silurian portion of CVB volcanism. This same shallow subcontinental source may have also been the source of subsequent CAMP volcanism (Fig. 6). CVB volcanism was the last major extrusive event along the margin of Avalonia until resumption of volcanism during the CAMP event. Scotese (1997) has shown that by early Devonian time, Gondwanaland had almost merged with Laurentia, only a few kilometers away.

In contrast to subduction related early CVB magmatism, the final closure of the Iapetus ocean during the Allaganian Orogeny was largely transpressional. Some felsic plutonism occurred during the Allaganian orogeny but was probably derived from shallow sources unrelated to CAMP sources. Although only one Avalonian subduction slab is modeled on Figure 6 for the sake of simplicity and in agreement with Scotese (1997) and Torsvik et al. (1996), the actual tectonic scenario may have been much more complex. Hatcher (1989) describes two Avalonian arcs, a western and eastern separated by a "Theic-Rheic Ocean" only one of which was accreted during the Acadian Orogeny. The Ammonoosuc – Piedmont arc was accreted during the Penobscottian Orogeny. The western Avalonian terranes accreted during the Taconic Orogeny and the eastern Avalonian terrane during the Acadian Orogeny according to (Hatcher, 1989). The final complex along the eastern margin of North America may have included at least six large terranes and several additional smaller terranes accreted during the Paleozoic (Williams and Hatcher, 1982).

At the northern end of the CVB tectonic complexities increase still more. Barr et al, (2002) describe a complex collage of originally peri-Gondwanan continental fragments in the outboard parts of the northern Appalachian orogen that include seven terranes separated by zones of transtension, subducted slabs, or undefined sutures. Some of these terranes are presumably underlain by depleted mantle, others by enriched mantle wedges. The tectonic synthesis of the northern Appalachians and plate margins that may have influenced the source of CAMP magma is, therefore, controversial and is the subject of ongoing revision. However, evidence of Silurian subduction related volcanism stratigraphically beneath the eastern North American CAMP is generally agreed upon.

## CONTRASTING ENVIRONMENTAL IMPACTS OF MAGMATIC EVENTS

### *Neoproterozoic Environmental Impact*

Although major volcanic events are commonly associated with environmental hardship and even mass extinction, the LOIB event correlates with the beginning of a major expansion in the diversity and quantity of marine life during the early Cambrian. Larson's (1991) study of the geological consequences of superplumes does not extend as far as the Neoproterozoic largely because of a lack of good data on the magnetic field reversal process that he uses to identify superplume events. However, the two superplume events that he has highlighted occurred during the mid-Cretaceous (124-83 Ma) and the Pennsylvanian-Permian (323-248 Ma). The mid-Cretaceous superplume event has been particularly well characterized (Caldeira and Rampino, 1991; Larson, 1991a, b; and Tatsumi and others (1998) and correlates with global warming, rising sea levels, and mantle outgassing, particularly carbon dioxide and nutrients. Both of the superplumes that were described by Larson (1991a, b) occurred over an extended period of time, comparable to the 615-550 Ma range measured for the LOIB, and were also associated with a major increase in the growth rate and deposition of marine life, comparable to that of the late Vendian to early Cambrian. In addition, both the Mesozoic and Paleozoic superplumes correlate with rapid increases in convection in the outer core and with a rapid increase in seafloor spreading rates as was probably also the case with the early Cambrian (Gurnis and Torsvik, 1994; Soper (1994); Dalziel (1997); Scotese, C. R., 1997).

### **Silurian Environmental Impact**

Silurian magmatism locally and on a regional scale included both alkalic and calc-alkaline activity. However, calc-alkaline magmatism and alkalic magmatism have never been correlated with mass extinction events. Despite the large volumes of volcanic activity during the mid-Silurian there is no evidence of rapid change in global biodiversity (Sepkowski, 1996). Most of earth's large igneous provinces that have been correlated with mass extinctions are instead tholeiitic continental flood basalts.

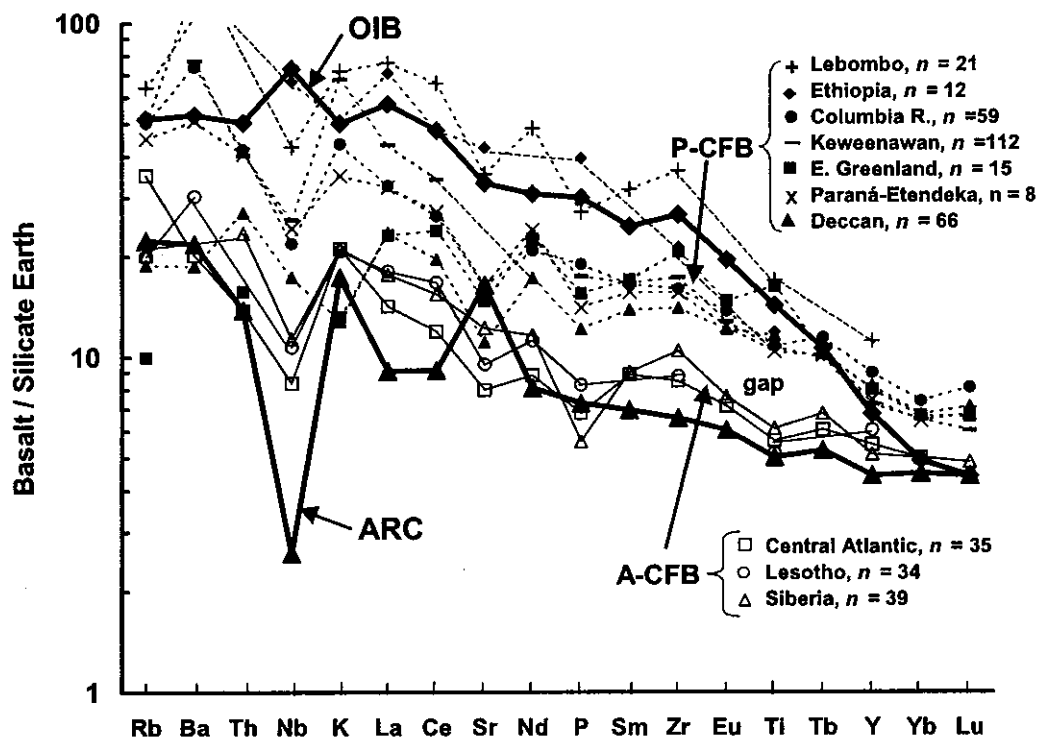
Most alkalic basalt provinces are confined to isolated hot spot tracts that are constrained by the size of a point source (a plume / hot spot). Calc-alkaline (arc) provinces such as the Andies Mountain system are instead generated along subduction zones that may extend several thousand kilometers. However, subduction zones are generally compressional with intrusive systems that rarely permit rapid delivery of magma through the kind of rifts that open up in extensional settings. Massive outpourings of calc-alkaline basalt or andesite on the scale of continental flood basalts are therefore rare if non-existent. Calc-alkaline magmatism is also constrained by the rate of subduction which rarely undergoes sudden change.

Finally, alkalic and calc-alkaline magmas are both characteristically viscous and prevent efficient release of potentially toxic volatiles. The relatively high  $\text{SiO}_2$  and  $\text{Al}_2\text{O}_3$  content of alkalic and calc-alkaline basalts result in lower viscosities than tholeiitic basalt such as Siberian Traps and CAMP basalt (Shaw, 1972).

### **Early Jurassic (CAMP) Environmental Impact**

Evidence of mass extinction at the end of the Permian, and Triassic periods is well documented (Sepkowski, 1996). In both cases major episodes of volcanic activity, the Permian (~250 Ma) Siberian

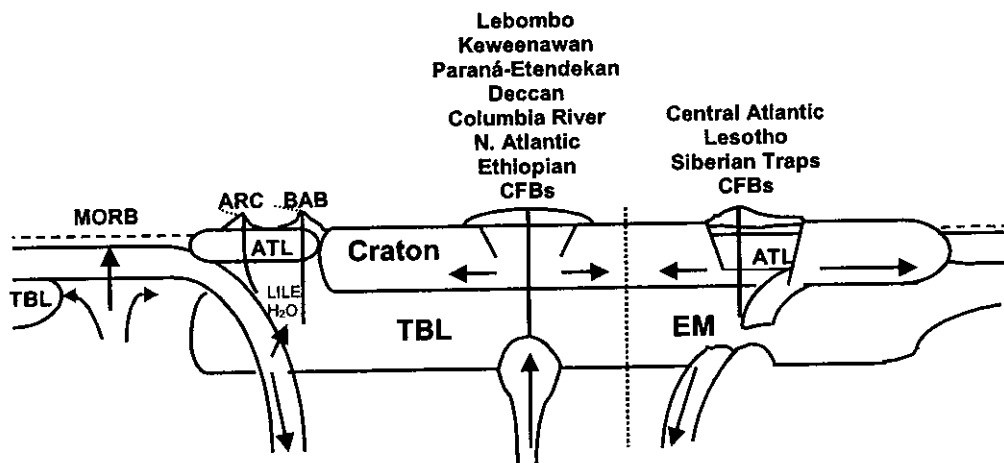
Traps and the early Jurassic (~200 Ma) CAMP, are the most probable cause of these extinctions. The chemical composition of the most widespread portion of each of these volcanic provinces is very similar (Fig. 7) and probably share some toxic characteristic. If both of these volcanic types were enriched in and fluxed by water, sulfur, and chlorine during subduction related processes (Fig. 8) these extinctions may have been caused by extreme degassing and sudden climatic change. McHone and Puffer (2003) and McHone (2003) have shown that CAMP volcanism at the Triassic-Jurassic boundary was accompanied by enormous emissions of sulfur, fluorine, and particularly high chlorine levels that were probably sufficient for major global extinction. Similar ideas have been widely suggested as the cause of the Permian extinction (Courillot et al., 1996; Rampino et al., 1988; Sigurdsson 1990; Stothers 1993, Wignall 2001, and Palfy 2003).



**Figure 7.** Spider diagram after Puffer (2001) illustrating the virtually identical chemical composition of the Central Atlantic Magmatic Province (CAMP), the Lesotho Basalt, and the Siberian Traps; products of similar edge driven sources (Fig. 8). Most other continental flood basalts in contrast are the product of plume activity modified to varying degrees by crustal interaction.

The third major continental flood basalt that is geochemically very similar to the Siberian Trap and CAMP is the Lesotho Basalt (Fig. 7 and 8) also known as the Karroo basalt of Lesotho within South Africa. The Lesotho basalt was also extruded through extensional rifts but is slightly younger (~185Ma) than the largest extrusive CAMP activity (~200 Ma) although late CAMP magmatism continued until ~185 Ma (Salters et al., 2003) and therefore overlaps and perhaps exacerbated the environmental effects of CAMP magmatism.

Of earth's five largest mass extinctions (Late Ordovician, Late Devonian, Late Permian, Late Triassic, and Late Cretaceous) the two mass extinctions that are most likely to have been caused by volcanic activity are the Late Permian (~250 Ma) and Late Triassic (~200 Ma). It is probably not a coincidence that the chemical composition of the two volcanic rocks involved is identical (Fig. 7). Late Ordovician extinctions were probably caused by global cooling and the draining of shallow marginal shelf environments on the basis of glacial sediments found on northern Africa interpreted as the "Gondwana ice caps" (Stanley, 1987). Late Devonian extinctions occurred slowly over a span of about 20 Ma and were probably caused by an ecological crisis, particularly eutrophication, or possibly another global cooling event (Levin, 1993). There is almost general agreement among geologists that the Late Cretaceous mass extinction was caused by the Chicxulub bolide. However, there is no evidence that Late Permian and Late Triassic extinctions were caused by global cooling or eutrophication and although a bolide has been proposed as the cause of the Late Triassic extinction (Olsen, 2004) the extreme level of compositional similarity of CAMP basalt and Siberian Trap (Fig. 7) strengthens the interpretation of McHone (2003) and McHone and Puffer (2003) that the Late Triassic extinction was also caused by the effects of volcanism.



**Figure 8.** Of all the various mafic magma types only continental flood basalts similar to the Central Atlantic Magmatic Province, the Lesotho Province, and the Siberian Traps are derived from a source that has been enriched in H<sub>2</sub>O and other volatiles (enriched mantle, EM) during a previous subduction event and then rapidly extruded out of new rifts in an extensional tectonic setting. Partial melting and magma mixing within the Thermal Boundary Layer (TBL) is the source of Arc basalt, Back Arc basalt (BAB), and most continental flood basalts. The source of Mid-Ocean Ridge Basalt (MORB) is depleted mantle.

## CONCLUSIONS

1. The Neoproterozoic dikes of New Jersey are part of a superplume event that correlates with the beginning of a major early Cambrian expansion of biodiversity.
2. The Silurian intrusions of New Jersey may be a southern extension of the Coastal Volcanic Belt, a major compressional event, but without any significant impact on global biodiversity.
3. The Early Jurassic intrusions of New Jersey are part of the Central Atlantic Magmatic Province (CAMP), a huge continental flood basalt that correlates with a global mass extinction.
4. The source of the eastern north American portion of CAMP was mantle rock enriched in volatiles during Silurian subduction.
5. Of all the genetic types of mafic rock only continental flood basalts similar to CAMP (particularly the Siberian traps) can result in major global mass extinctions for three reasons.
  - a. their mantle sources were enriched in water, sulfur, and chlorine during previous subductions.
  - b. massive volumes of magma were extruded quickly through rifts opened during extension.
  - c. the viscosity of the magma was low enough to release toxic volatiles into the air.
6. The short duration of CAMP and Siberian magmatism was constrained by the quantity of volatiles that acted as a flux.

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# Geochemistry and Tectonic Setting of Late Neoproterozoic Diabase Dikes, New Jersey Highlands and Trenton Prong

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## ABSTRACT

Abundant, widely distributed diabase dikes of late Neoproterozoic age intrude Mesoproterozoic rocks in the New Jersey Highlands and in the Trenton Prong in the west-central part of the state. Although the dikes are undated, a Neoproterozoic age is indicated by the field relations in that dikes postdate the regional granulite-facies metamorphism that ended at ca. 1030 Ma and are absent in cover sequence rocks of Cambrian or younger age. Neoproterozoic dikes strike predominantly toward the northeast and have a mean trend of N50°E in the southwest Highlands and N35°E in the northern Highlands. Dikes range in width from a few inches (cm) to 60 feet (18 m) and have lengths of as much as several miles (km). They are characterized by sharp, largely discordant contacts against enclosing Mesoproterozoic rocks and by aphanitic chilled margins that grade into coarser-grained interiors. Xenoliths of Mesoproterozoic rocks are present in some dikes. Locally developed columnar joints suggest that the dikes were emplaced at a shallow crustal level, perhaps locally breaking the surface, and may have been feeders for mafic volcanics that were removed by erosion. If so, then the mafic volcanics, along with felsic volcanics of the Neoproterozoic Chestnut Hill Formation, would have formed a suite of rift-related bimodal volcanic rocks in the New Jersey Highlands.

Neoproterozoic dikes range from alkalic through transitional alkalic basalt to tholeiitic basalt and are quartz or olivine normative. All dikes are enriched in TiO<sub>2</sub>, P<sub>2</sub>O<sub>5</sub>, Zr, and light rare-earth elements (LREE) compared to Early Jurassic basalt and diabase in the Newark basin. Neoproterozoic dikes form two broad groups based on P<sub>2</sub>O<sub>5</sub> contents, one low (0.29-0.81 percent) and one high (1.19-1.60 percent). Both groups fall into three types that form a geochemical continuum ranging from type 1 (least evolved) to type 3 (most evolved). All three types formed through fractionation of predominantly olivine, clinopyroxene, and plagioclase, but originated from separate magmas that tapped different enriched, Ocean Island Basalt-like mantle sources. Type 1 dikes were derived from a shallower (~30-35 mi, 50-55 km) source that had undergone a greater amount of melting, possibly by rising, hotter type 2 and 3 dike magma from a deeper (35-45 mi, 55-70 km) mantle source. Geochemical similarities of all three types imply that dikes in the Highlands and Trenton Prong were emplaced as part of a single magmatic event as cogenetic intrusions. All of the dikes have geochemical compositions that are consistent with magma formation in a rift-related, within-plate tectonic setting. Regional dike geometries form right-stepping, rhomb-shaped patterns due to emplacement into dilational fractures that were generated through a combination of southeast-directed extension and strike-slip shear stresses. Neoproterozoic dikes in New Jersey were emplaced between 602 and 554 Ma, during rifting of eastern Laurentia (proto North America) from western Gondwana and breakup of the supercontinent Rodinia.

## INTRODUCTION

The occurrence of diabase dikes that intrude Mesoproterozoic rocks in the New Jersey Highlands has long been recognized (e.g., Westgate, 1895; Darton et al., 1908; Bayley et al., 1914). Because of the dark gray color of the dikes, their mineralogical composition, and geographic distribution, workers (e.g., Milton, 1947; Sims, 1958; Baker and Buddington, 1970; Maxey, 1973) commonly assigned them a Mesozoic age and correlated the dikes to basalt and diabase in the Newark basin to the east. A recent exception was the study of Hull et al. (1988) in which the dikes were interpreted to be Paleozoic in age based on structural and geochemical considerations. Ratcliffe (1987) argued for a Neoproterozoic age for

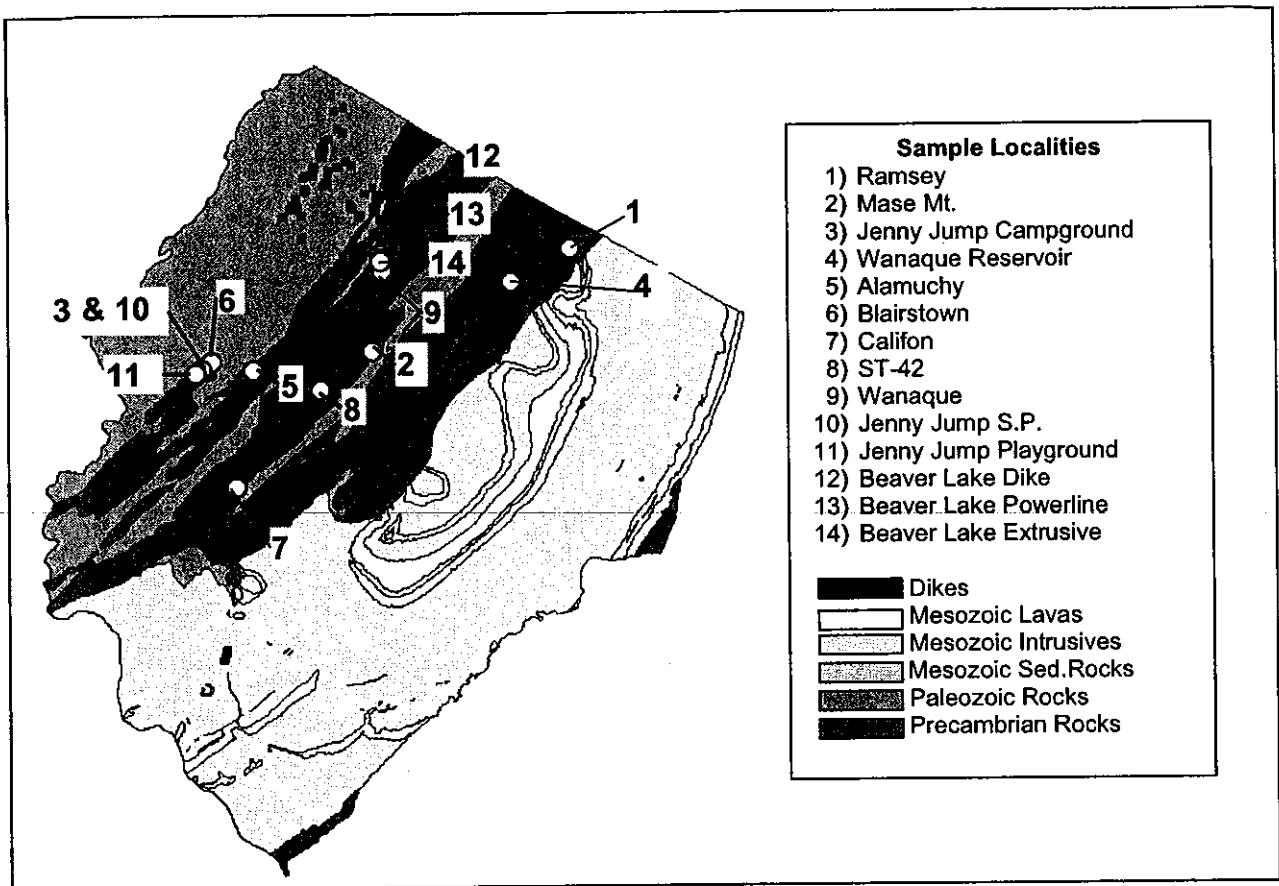


Figure 1. Map showing the location of sampling localities in northern New Jersey.

## METHODOLOGY

- 1) Samples were collected from fifteen mafic dikes and at two sites in one lava flow for paleomagnetic analysis. Cores were drilled in the field and oriented with a Brunton compass (and sun compass if there was sufficient sun light). One or two samples were cut from each core.
- 2) At least one sample from each core was subjected to step-wise, alternating field demagnetization at 2.5, 5, 10, 15, 20, 25, 30, 40, 50, 60, 70, 80, and 90 mT.
- 3) Selected samples were subjected to thermal demagnetization at 100, 200, 300, 400, 450, 500, 530, 560, 600, 625, 650, 660, 670, and 680 degrees C.
- 4) Vector endpoint diagrams were plotted (after Zijdeveld, 1967) to help identify components.
- 5) Principal component analysis (after Kirschvink, 1980) was carried out to isolate components.

## RESULTS

Except for samples from the Cortlandt Complex, which yielded inconsistent results, all of the sites indicated a magnetization that trended to the south and plunged gently. The results are shown in Table 1. Data from each of the sites show more scatter than is usually desirable. This could be due to errors of measurement, inconsistencies in remagnetization, or errors in magnetic compass readings. All of the data are consistent with remagnetization in the Permian, as has been demonstrated in sedimentary rocks for decades (Elmore and Others, 2001).



## DIKE POPULATIONS IN THE NEW JERSEY HIGHLANDS

### Mesozoic Rocks

Early Jurassic basalt is abundantly exposed as flows, and diabase as sills and dikes in the Newark basin in the New Jersey Piedmont. Neoproterozoic diabase dikes in the Highlands are dissimilar to these Mesozoic occurrences in the following respects. The Early Jurassic basalt and diabase are physically different in lacking the ubiquitous alteration present in the Neoproterozoic dikes. They are also geochemically distinct from Neoproterozoic dikes in having lower  $\text{TiO}_2$  (Fig. 2),  $\text{P}_2\text{O}_5$ , Zr, and LREE. Intrusive activity of Early Jurassic age may have occurred in the Highlands but to date the evidence is meager. A single mafic dike mapped by the author in a gas pipeline trench in the Mendham quadrangle, a few miles (km) west of the Ramapo fault, is considered a possible candidate. This dike (Volkert and Puffer, 1995, Table 1, sample 364) is geochemically unlike the Neoproterozoic dikes and resembles Orange Mountain Basalt (Fig. 2). Costa (1991) also recognized other possible Mesozoic dike candidates from within the Ramapo fault zone but the dikes analyzed by him were too highly deformed and altered for meaningful study and correlation. To date, no other dikes of unequivocal Early Jurassic age are recognized in the Highlands.

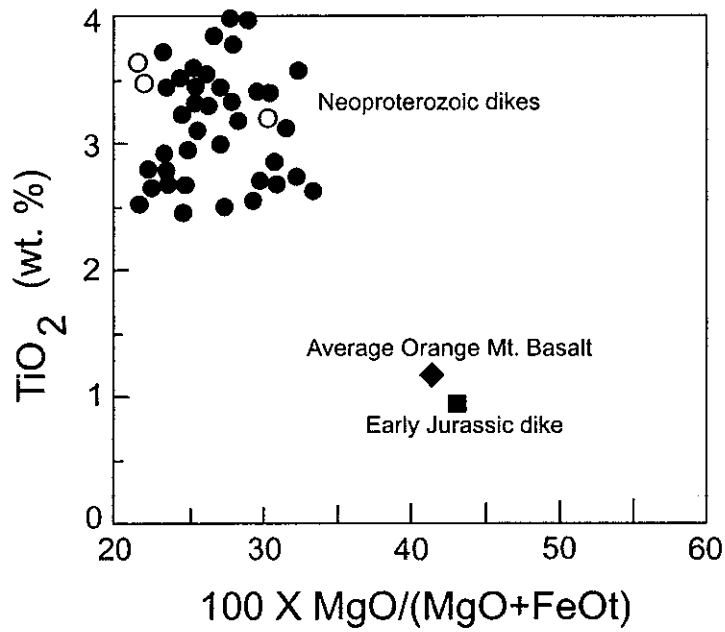
### Paleozoic Dikes

Widely distributed and abundant Paleozoic-age dikes intrude Mesoproterozoic rocks near the lower Paleozoic unconformity predominantly in the northwestern Highlands, although a few of these dikes have also been recognized and mapped by the author in the south-central and eastern Highlands. Paleozoic-age dikes commonly intrude unmetamorphosed rocks of Cambrian and Ordovician age in the Valley and Ridge in New Jersey and become increasingly abundant toward the Beemerville area in Sussex County (Fig. 1). These dikes, and related intrusive rocks, are part of the Beemerville alkaline complex that was renamed the Beemerville Intrusive Suite (Drake and Monteverde, 1992). Paleozoic dikes are dimensionally comparable to the Neoproterozoic dikes and they overlap the geographic occurrence of the Neoproterozoic dikes. However, there are dissimilarities between the two populations that help to distinguish them. Paleozoic dikes of the Beemerville Suite typically trend toward the northwest and occur in a fan-shaped swarm, in contrast to the linear northeast trend and more solitary occurrence of the Neoproterozoic dikes. Paleozoic dikes differ mineralogically in that they contain various assemblages of biotite, clinopyroxene, calcite, plagioclase, K-feldspar, and nepheline. They are also geochemically distinct in having higher concentrations of LREE,  $\text{K}_2\text{O}$ , Ba, Rb, Sr, and Zr, and compositions that are more typically nepheline normative compared to the hypersthene normative Neoproterozoic diabase dikes.

Several successful age dates have been obtained from the alkaline rocks of the Beemerville Intrusive Suite. A radiometric age of  $435 \pm 20$  Ma was obtained by Rb-Sr and K-Ar dating techniques on biotite from nepheline syenite at Beemerville (Zartman et al., 1967). Milton (written communication to F.J. Markewicz, 1972) obtained a K-Ar whole-rock age of  $422 \pm 14$  Ma from a minette dike in Sussex County, and Eby et al. (1992) obtained an age of  $422 \pm 14$  Ma using fission-track dating of titanite from nepheline syenite at Beemerville. All of these ages are in relatively good agreement and suggest an Early Silurian age for dikes and intrusive rocks of the Beemerville Intrusive Suite.

### Neoproterozoic Dikes

Widely distributed diabase dikes of late Neoproterozoic age intrude Mesoproterozoic basement rocks in the New Jersey Highlands and Trenton Prong but are conspicuously absent in Cambrian or younger cover sequence rocks. The dikes occur in about equal abundance along most of the 62 mile (100



**Figure 2.**  $\text{TiO}_2$  versus  $100 \times \text{MgO}/(\text{MgO} + \text{FeO})$  plot of Neoproterozoic diabase dikes from the New Jersey Highlands (solid circles) and Trenton Prong (open circles) and a single dike of Early Jurassic age from the Highlands (square). Data are from Volkert and Drake (1993) and Volkert and Puffer (1995). Shown for comparison is the average composition of Early Jurassic Orange Mountain Basalt (Puffer, 1992) from the Newark basin (diamond).

km) regional strike length of the Highlands, as well as east and west of the Green Pond Mountain Region over a width of 19 miles (30 km) (Fig. 1). Interestingly, dikes are somewhat less abundant, or even absent entirely, in the south-central and extreme northern Highlands. The reason for this is not well understood but may be related to the thickness of widespread and abundant granite bodies that underlie these areas. Quite possibly they were less susceptible to extensional stresses during rifting and dike emplacement and behaved as a barrier to the rising magma. Continuation of Neoproterozoic dikes in New Jersey east of the Highlands has been documented (Volkert and Drake, 1993; Volkert et al., 1996) based on the recognition of tholeiitic to alkalic mafic dikes geochemically similar to the Highlands dikes that intrude Mesoproterozoic rocks in the Trenton Prong (Fig. 1).

Two discrete episodes of Neoproterozoic rifting and igneous activity are recognized in eastern Laurentia, an older 760 to 680 Ma pulse in the southern and central Appalachians (Aleinikoff et al., 1995) and a younger 615 to 554 Ma pulse in the central and northern Appalachians (Rankin et al., 1997; Fail, 1997 and references therein; Cawood et al., 2001). Magmatism associated with the younger pulse includes (from north to south) the  $615 \pm 2$  Ma (Kamo et al., 1989), Long Range dikes in Labrador and Newfoundland,  $602 \pm 10$  Ma (Williams et al., 1985) Round Pond granite in Newfoundland,  $590 \pm 2$  Ma (Kamo et al., 1995) Grenville dikes of the Ottawa Graben in Ontario and Quebec, poorly dated but postulated to be  $\sim 588$  Ma (Isachsen et al., 1988) northern New York Adirondack dikes,  $554 \pm 4/-2$  Ma (Kumarapeli et al., 1989) Tibbit Hill Volcanic Member of the Pinnacle Formation in southern Quebec,  $571 \pm 5$  Ma (Walsh and Aleinikoff, 1999) metafelsite in the Pinney Hollow Formation in Vermont,  $563 \pm 3$  Ma Yonkers Gneiss and  $562 \pm 6$  Ma Pound Ridge Granite Gneiss (Rankin et al., 1997; Tollo et al., 2004) in the southern New York Manhattan Prong, and  $602 \pm 2$  Ma (Smith, 2003) felsite dikes in southeastern Pennsylvania. These ages overlap mafic dikes and flows in the central Appalachians that include the  $570 \pm 36$  Ma (Badger and Sinha, 1988) metabasalt in the Catoclin Formation in northern Virginia, and  $564 \pm 9$  Ma (Aleinikoff et al., 1995) metarhyolite of the Catoclin Formation in northern Virginia. Although precise geochronologic data are lacking for the diabase dikes in the New Jersey



Highlands and Trenton Prong, it logically follows that their age falls within this 615 to 554 Ma pulse of magmatism. Neoproterozoic dikes are discussed in greater detail in the sections that follow.

## NEOPROTEROZOIC DIKES

### Field Relations

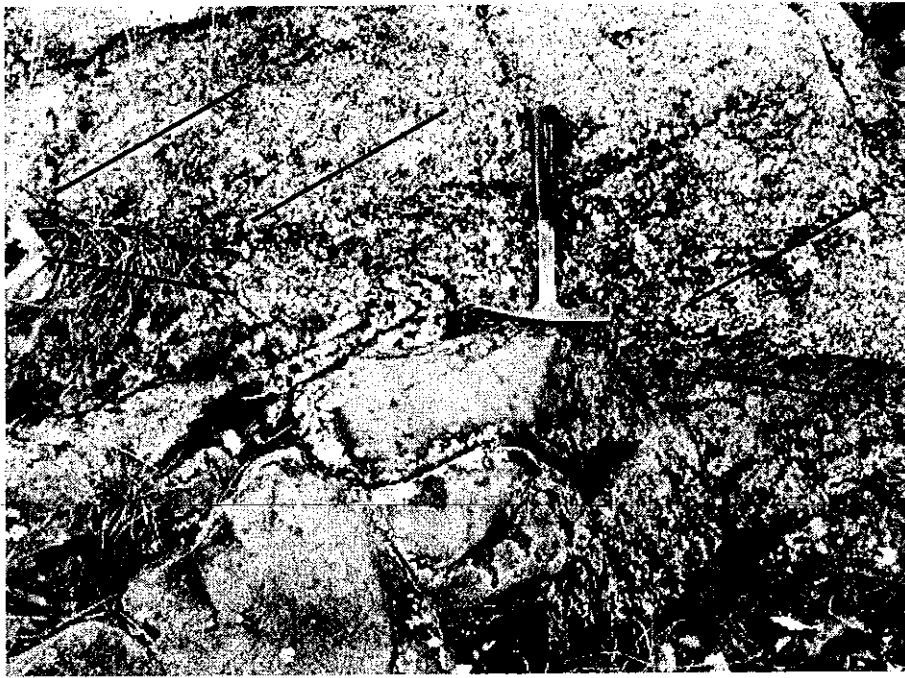
Diabase dikes of Neoproterozoic age in the New Jersey Highlands range in width from a few inches (cm) to about 60 feet (18 m) and have strike lengths of as much as several miles (km). Because the dikes generally weather more readily than the enclosing Mesoproterozoic rocks, they tend to crop out discontinuously and often are reduced to float in areas between outcrops. The majority of the dikes are solitary, but where multiple dikes intrude the same general area they commonly occur en echelon. Diabase dikes characteristically are hard, massive-textured, unmetamorphosed and unfoliated, and are dark-greenish-gray on fresh surfaces. Most dikes are fine-grained to aphanitic, although many of the thicker dikes grade texturally from fine-grained margins into coarser-grained interiors. Diabase dikes typically have chilled margins and sharp contacts against enclosing Mesoproterozoic rocks (Fig. 3). Aphanitic chilled margins along the dike margins imply emplacement into relatively cold Mesoproterozoic rocks. Scattered, typically small, undigested xenoliths of country rock occur along the margins of some dikes (Fig. 4). Some of the xenoliths were likely detached through magmatic stoping, while others are fractured and may have been dislodged through forceful injection during dike emplacement. Thin apophyses of fine-grained diabase commonly intrude the country rock adjacent to the main body of thicker dikes. Closely spaced columnar joints (Fig. 5) that are oriented normal to the dike walls were observed in several dikes from widely separated areas, suggesting that the dikes were emplaced at a fairly shallow crustal level, perhaps locally breaking the surface and feeding flows that have since been eroded.

On a regional scale in the Highlands, dike contacts range from concordant to highly discordant in relation to foliation in the Mesoproterozoic rocks (e.g., Fig. 3). Most dikes are only moderately discordant and, as a result, the predominant orientation of dikes is toward the northeast. Neoproterozoic dike trends average about N44°E (Fig. 6), very close to the overall regional trend of the Highlands. Most of the dikes dip steeply toward the southeast or northwest at between 70° and 90°, and so the majority of dikes are discordant in both strike and dip.

Field relations of Neoproterozoic dikes in the Trenton Prong are more difficult to interpret because of the overall poor exposure. Consequently, these dikes are known mainly through recovery of drill core or rotary cuttings. Nevertheless, they display the same unmetamorphosed, relatively fine-grained, greenish-gray appearance, mineralogy, and ubiquitous sulfide blebs as dikes in the Highlands, and there is little reason to believe that the field relations are any different. Dikes in the Trenton Prong are confined to belts of Mesoproterozoic rocks (Fig. 1) and have not been encountered in the Cambrian Chickies Quartzite or in Wissahickon schist or gneiss on the hanging wall of the Huntingdon Valley fault.

### Petrography

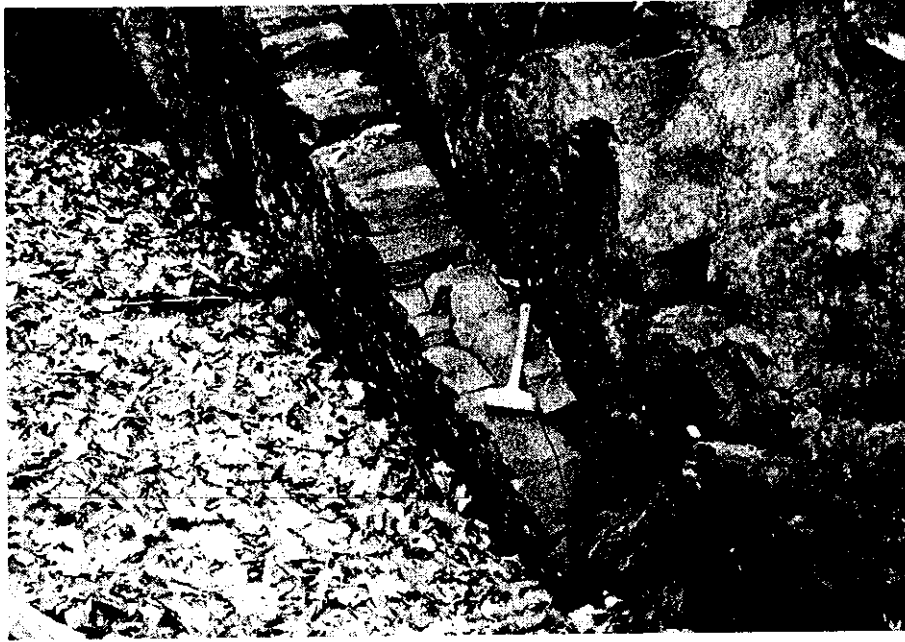
Neoproterozoic diabase dikes in the Highlands and Trenton Prong are composed principally of plagioclase (labradorite to andesine), augite, titanomagnetite, ilmenite, and altered olivine (Volkert and Puffer, 1995). Although apatite was not seen in any of the dikes studied in thin section, the high P<sub>2</sub>O<sub>5</sub> contents of some dikes would seem to demand its presence as a modal phase. The groundmass of all of the dikes is typically felty and contains small laths of plagioclase and small anhedral masses of pyrite. Despite the relatively fresh and unaltered appearance of most dikes in the field, petrographic examination



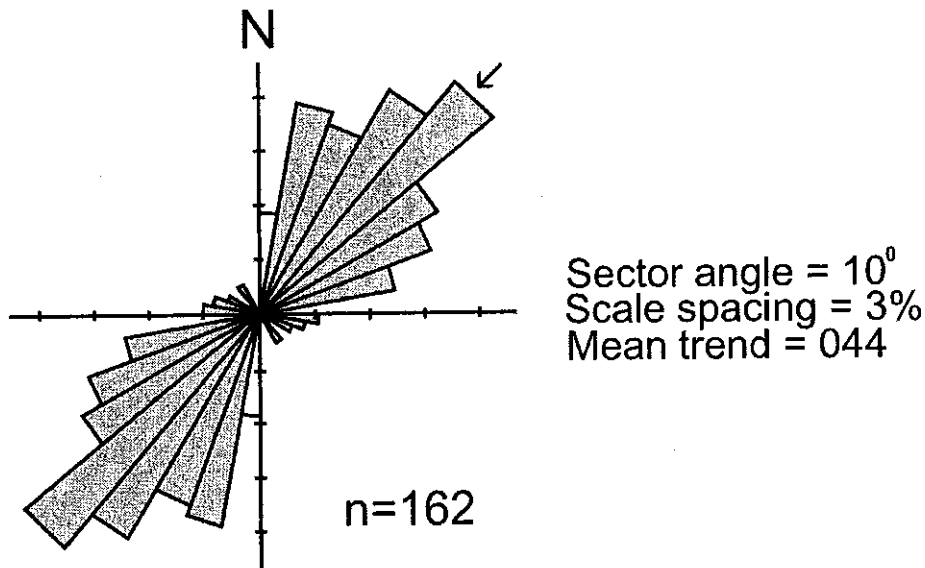
**Figure 3.** Moderately discordant ( $\sim 40^\circ$ ) and sharp contact between Neoproterozoic dike (below head of hammer) and Mesoproterozoic gneiss (above hammer) from the Wanaque quadrangle. Upper lines show foliation in gneiss and lower lines the trend of the dike. Hammer is 11 in (28 cm) long.



**Figure 4.** Xenolith of Mesoproterozoic country rock immediately left of hammer enclosed in Neoproterozoic dike from the Tranquility quadrangle. Xenolith is the same lithology as rock at right of photograph, suggesting that it was dislodged during dike emplacement. Hammer is 11 in (28 cm) long.



**Figure 5.** Well-developed columnar jointing in chilled margin of Neoproterozoic dike from the Tranquility quadrangle. Contact with Mesoproterozoic rock is to the immediate right of hammer. Columns are oriented at a right angle to the contact.



**Figure 6.** Rose diagram of strikes of 162 Neoproterozoic dike segments from the New Jersey Highlands. Note the predominant orientation toward the northeast. Arrow delineates the N44°E statistical mean of the dike segments.

reveals varying amounts of alteration. Main alteration effects include secondary albite rimming plagioclase laths, sericitic replacement of plagioclase, secondary amphibole rimming augite, chloritic replacement of augite, magnetite altered to hematite, and disseminated grains or clusters of pyrite (Volkert and Puffer, 1995). Other alteration phases present in minor amounts include quartz, calcite, chalcopryrite, epidote, and serpentine. Alteration of the original dike composition is interpreted to be the

result of retrograde reactions involving hydrothermal fluids generated primarily during Taconian and Alleghanian orogenesis, but possibly also during Mesozoic rifting as well.

## Geochemistry and Petrogenesis

Neoproterozoic diabase dikes in the New Jersey Highlands and Trenton Prong have major-oxide and trace-element compositions that are consistent with basalt. They range from alkalic, through transitional alkalic basalt, to tholeiitic basalt as determined by ratios of high field-strength elements (HFSE). Of the more than 40 dikes geochemically analyzed to date, 35 are quartz normative and the rest are olivine normative. Values of  $P_2O_5$  broadly divide the dikes in the Highlands and Trenton Prong into two groups, one with low  $P_2O_5$  (LPG dikes) and one with high  $P_2O_5$  (HPG dikes) (Volkert and Puffer, 1995). Concentrations of  $P_2O_5$  range from 0.28 to 0.81 wt. percent in LPG dikes and from 1.19 to 1.6 percent in HPG dikes (Fig. 7). The higher concentrations of  $P_2O_5$  in some dikes may be due to the presence of a phosphate phase such as apatite that was present in the mantle source of the magma. Two dikes analyzed, one from the western Highlands and one from the eastern Highlands, have intermediate  $P_2O_5$  contents of 0.74 to 0.81 and they may reflect a transition between the two groups. Dikes from both groups typically contain 2.34 to 4.02 wt. percent  $TiO_2$ , <14 percent  $Al_2O_3$ , 3.85 to 7.07 percent  $MgO$ , 4.24 to 10.1 percent  $CaO$ , and <10-109 ppm Cr, 112-450 ppm Zr, and 5-30 ppm Ni (Table 1). However, Y is lower in LPG dikes (27 ppm) compared to HPG dikes (55 ppm). Values of Mg number (Mg#), calculated as  $[100(Mg/Mg+Fe^{2+}) \text{ atomic}]$ , range from 42-59 in LPG dikes and from 49-59 in HPG dikes. The relatively low Mg#, combined with the low Cr and Ni of both groups, implies that none of the dikes has a primitive mantle composition, or was formed directly through partial melting of mantle sources, and that all had undergone considerable fractionation prior to emplacement.

The magma that generated the Neoproterozoic dikes in the Highlands and Trenton Prong likely experienced olivine fractionation as suggested by the generally low  $MgO$  and Ni contents of the dikes (Table 1). The effects of clinopyroxene fractionation are likewise evident in the generally low Cr values, positive correlation of  $Al_2O_3$ ,  $CaO$ , and Cr with  $MgO$ , and increasing La/Yb with decreasing Mg# (Volkert and Puffer, 1995). Plagioclase was a much less important fractionating phase because Sr rises slightly with decreasing  $MgO$ , and REE patterns lack a negative Eu anomaly, suggesting the presence of excess plagioclase in the melt.

Using Mg#,  $TiO_2$ , Zr, Cr, and Ni as one measure of an index of fractionation, the dikes fall into three distinct types. Five of the dikes, two from the Franklin quadrangle and one each from the Tranquility, High Bridge, and Wanaque quadrangles (Fig. 1), have compositions that are the most primitive of those sampled and they fall into type 1. Type 1 dikes are alkalic and have high Mg# (56.4), Cr (101 ppm), Ni (38 ppm), low  $TiO_2$  (2.83), Zr (138 ppm) and low total REE (Ce/Yb = 12.91) (Table 2). Two other dikes have somewhat higher  $MgO$  of 7.34 (Mg# = 66.2) and 8.87 (Mg# = 76.1) but low Cr and Ni and their  $MgO$  values are probably inflated by high chlorite contents. Seven of the dikes, one each from the Blairstown, High Bridge, Tranquility, Newton East, Franklin, Pompton Plains, and Wanaque quadrangles, have intermediate values and fall into type 2. Type 2 dikes are alkalic to transitional alkalic and have intermediate values of Mg# (52.1), Cr (60 ppm), Ni (16 ppm), and slightly higher  $TiO_2$  (2.91), and Zr (201 ppm) (Table 2). The rest of the Highlands dikes, and all of the Trenton Prong dikes, are the most geochemically evolved and they fall into type 3. Type 3 dikes are mainly tholeiitic or transitional alkalic. They have the lowest Mg# (49.7), Cr (23 ppm), Ni (8 ppm), and highest  $TiO_2$  (3.18), Zr (267 ppm) and total REE (Ce/Yb = 17.06) (Table 2). In correlating these three types to  $P_2O_5$  content, one HPG dike falls into type 1, two into type 2, and eight into type 3. The rest are LPG dikes. Of considerable significance is the absence of geographic control on the distribution of the three types of dikes; all overlap and occur in both the western and eastern Highlands. Likewise, there is a complete absence of structural control or correlation of dike geochemistry in relation to dike trend.

TABLE 1. AVERAGE GEOCHEMICAL COMPOSITION OF NEOPROTEROZOIC DIABASE DIKES FROM THE NEW JERSEY HIGHLANDS AND TRENTON PRONG

number	New Jersey Highlands dikes		Trenton Prong dikes			
	High P <sub>2</sub> O <sub>5</sub> group	Low P <sub>2</sub> O <sub>5</sub> group	MCP-228	MCP-158	TE-11	PR-14
	10	25	1	1	1	1
wt. percent						
SiO <sub>2</sub> (range)	49.27 (47.73-51.15)	49.01 (45.85-51.74)	49.00	47.80	49.60	49.90
TiO <sub>2</sub>	3.45 (2.68-4.02)	2.91 (2.34-3.72)	3.67	3.25	3.47	3.25
Al <sub>2</sub> O <sub>3</sub>	13.33 (12.19-14.32)	13.19 (10.98-15.98)	12.60	15.50	12.60	13.50
Fe <sub>2</sub> O <sub>3</sub>	5.82 (4.87-6.40)	5.48 (3.50-7.00)	5.30	4.20	5.40	3.30
FeO	7.23 (4.95-8.60)	8.76 (6.68-10.80)	10.60	6.60	10.50	10.90
MnO	0.22 (0.12-0.28)	0.20 (0.16-0.24)	0.26	0.14	0.26	0.22
MgO	5.11 (4.08-8.87)	5.10 (3.99-7.34)	4.30	4.64	4.47	4.86
CaO	6.58 (4.24-7.70)	8.21 (5.48-10.10)	8.15	8.46	7.09	7.43
Na <sub>2</sub> O	3.20 (2.14-5.34)	2.91 (2.19-3.96)	2.94	3.26	3.03	1.71
K <sub>2</sub> O	1.68 (0.13-2.88)	1.40 (0.36-2.36)	0.86	2.35	1.59	0.66
P <sub>2</sub> O <sub>5</sub>	1.37 (1.19-1.60)	0.50 (0.28-0.81)	0.50	1.58	0.56	0.42
ppm						
Cr (range)	38 (10-102)	48 (10-109)	12	50	10	50
Ni	17 (5-46)	25 (1-47)	24	--	22	--
Nb	35 (30-40)	35 (10-70)	50	40	50	20
Rb	50 (40-60)	47 (30-80)	10	60	30	20
Sr	287 (93-473)	346 (240-592)	250	1310	280	100
Y	55 (50-60)	27 (18-40)	20	20	30	40
Zr	258 (184-300)	234 (112-450)	280	370	360	210
Mg#*	55.7 (48.6-76.1)	50.7 (42.1-66.2)	41.9	55.6	43.1	44.3

\* Mg# = [100(Mg/Mg+Fe<sup>2+</sup>) atomic]

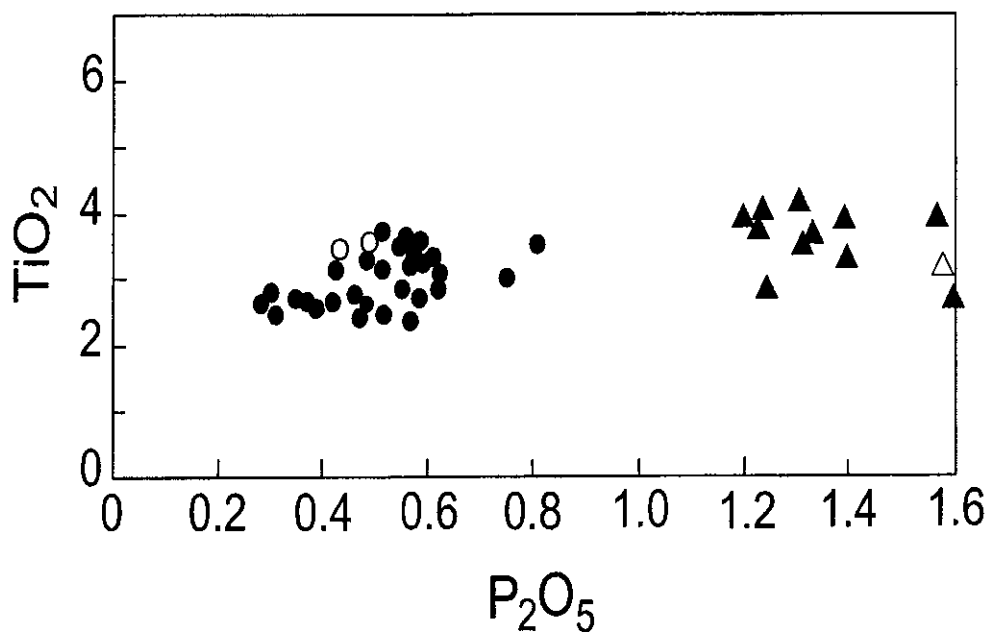
Data sources: New Jersey Highlands dikes, Volkert and Puffer (1995); Trenton Prong dikes, Volkert and Drake (1993).

TABLE 2. COMPARISON OF THE GEOCHEMISTRY OF NEW JERSEY NEOPROTEROZOIC DIKES WITH OCEAN ISLAND BASALT, ENRICHED AND NORMAL MID-OCEAN RIDGE BASALT

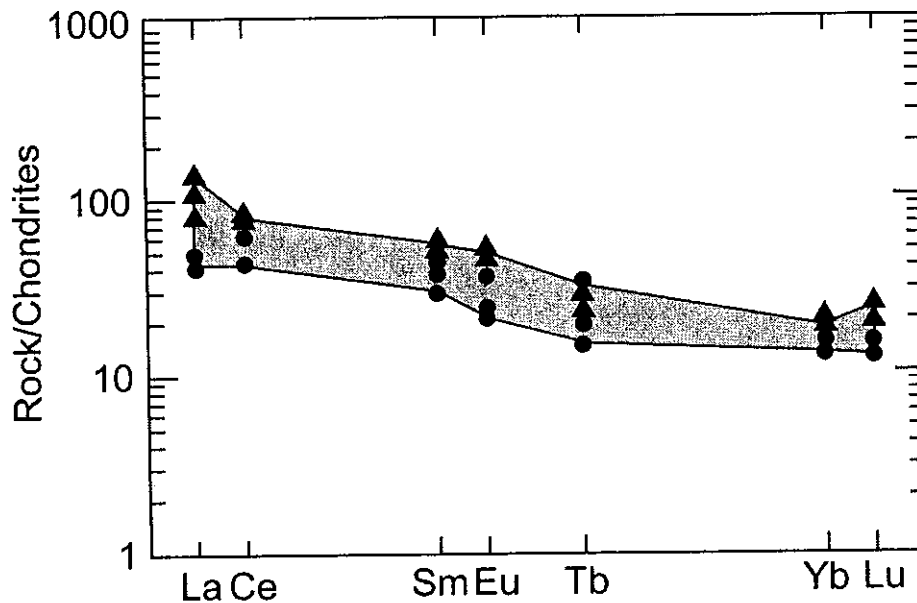
	Type 1	Type 2	Type 3	OIB	E-MORB	N-MORB
TiO <sub>2</sub>	2.83 (2.34-4.02)	2.91 (2.52-3.97)	3.18 (2.47-3.89)	3.62	1.69	1.36
MgO	5.82 (4.74-7.34)	5.22 (3.99-7.07)	4.72 (3.85-6.60)	5.52	6.90	8.96
Mg#	56.4 (52.3-66.2)	52.1 (43.8-59)	49.7 (42.1-59.2)	--	--	--
Cr	101 (92-109)	60 (42-85)	23 (10-40)	81	225	346
Ni	38 (23-47)	16 (2-27)	8 (1-15)	78	132	177
Zr	138 (112-154)	201 (170-230)	267 (220-450)	280	73	74
Ti/Y	--	655 (504-797)	657 (287-1186)	593	273	271
Ti/V	66 (58-83)	142	127 (119-136)	--	--	--
Ti/Zr	112 (95-142)	84 (69-94)	71 (33-97)	61	82	103
Zr/Nb	--	10.6 (4.2-22)	8.1 (3.7-14.7)	5.83	8.79	31.76
Zr/Y	--	7.8 (7.3-8.5)	9.2 (5-13)	9.65	3.32	2.64
Nb/Y	--	1.22 (0.33-2.0)	1.28 (0.60-3.5)	1.65	0.38	0.08
La/Ce	0.38 (0.37-0.39)	--	0.55 (0.37-0.65)	0.46	0.42	0.33
La/Sm	2.41 (2.33-2.50)	--	3.64 (2.34-4.56)	3.70	2.42	0.95
Ce/Yb	12.91 (12.26-13.57)	--	17.06 (14.21-19.17)	37.04	6.33	2.46
Sm/Yb	2.04 (1.94-2.14)	--	2.63 (2.32-3.11)	4.63	1.10	0.86

Note: Major elements in wt. percent; trace elements in ppm. Ranges shown in parentheses. Average OIB, E-MORB and N-MORB values from Sun and McDonough (1989), except TiO<sub>2</sub>, MgO, Cr, and Ni from Wilson (1989).

Total concentrations of REE in all of the dikes overlap and display a somewhat enriched pattern relative to chondrites (Fig. 8). The LPG dikes are enriched in LREE from 43 to 112 times chondrite and have  $\text{La/Yb}_N$  (chondrite-normalized) ratios of 3.5 to 8.7, whereas HPG dikes are enriched 77 to 140 times chondrite and have  $\text{La/Yb}_N$  of 4.5 to 9.0. Contents of Ce also help to distinguish the somewhat more LREE enriched HPG dikes ( $\text{Ce} = 68$  ppm) from the LPG dikes ( $\text{Ce} = 46$  ppm). Partitioning of heavy REE (HREE) is different in basalts from a deeper mantle source in which garnet is stable ( $>47$  mi,  $>75$  km) than from a shallower mantle source where spinel is stable ( $<47$  mi,  $<75$  km) (Ellam, 1992). Also, ratios of Sm/Yb are less affected by fractionation than by the amount of partial melting or the mantle source characteristics. Therefore, abundances of Sm and Yb are useful for determining the approximate depth of magma genesis because they are higher in melts from a deeper source, and so lower Sm/Yb and Ce/Yb ratios indicate a shallower source (e.g., Ellam, 1992). Ratios of Sm/Yb are slightly higher in HPG dikes (2.81) compared to LPG dikes (2.29). A similar relationship is seen in type 1 dikes (least evolved) that have Sm/Yb = 2.04 and Ce/Yb = 12.91 compared to type 3 dikes (most evolved) that have Sm/Yb = 2.63 and Ce/Yb = 17.06 (Table 2). Evaluation of all dike compositions using primitive mantle-normalized ratios of Yb/Sm and Ce/Sm (Fig. 9) indicates that they formed through the mixing of different mantle sources produced by  $\sim 4$ -6 % melting of garnet peridotite and  $\sim 2$ -4 % melting of spinel peridotite. The Ce/Yb, Sm/Yb, and Ce/Y (1.8-2.75) ratios of the dikes support this and imply magma formation at a depth of  $\sim 30$ -35 mi (50-55 km) for type 1 dikes and 35-45 mi (55-70 km) for type 3 dikes applying the model of Ellam (1992).



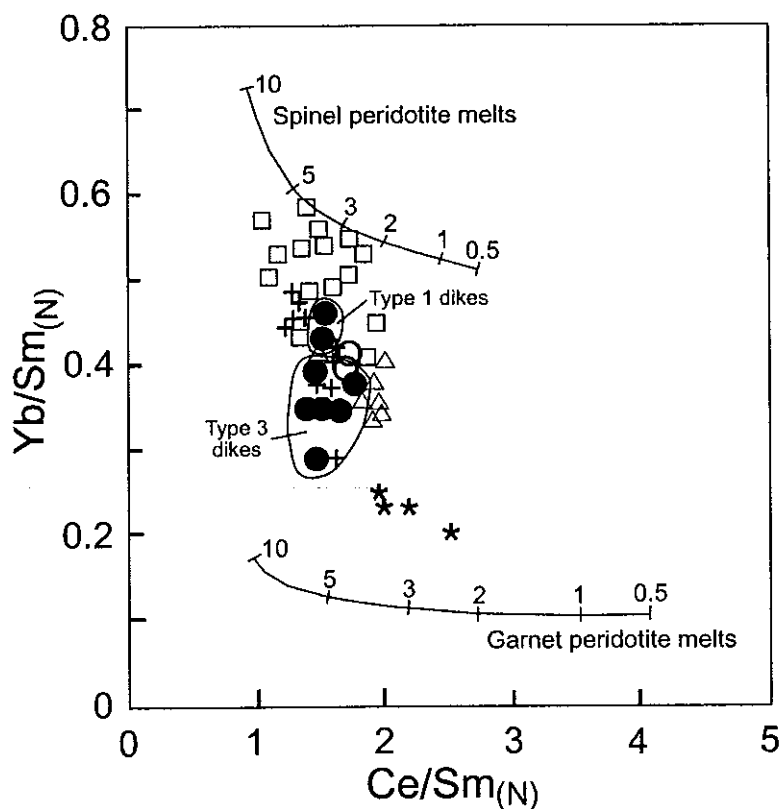
**Figure 7.**  $\text{TiO}_2$  versus  $\text{P}_2\text{O}_5$  plot of Neoproterozoic dikes from the Highlands (solid symbols) and Trenton Prong (open symbols). Note the separation of the dikes into two groups characterized by low  $\text{P}_2\text{O}_5$  (circles) and high  $\text{P}_2\text{O}_5$  (triangles) that occur over a relatively restricted range in  $\text{TiO}_2$ .



**Figure 8.** Chondrite-normalized REE patterns of Neoproterozoic dikes from the Highlands and Trenton Prong showing their compositional range. Circles are LPG dikes and triangles are HPG dikes. Normalizing values are from Nakamura (1974).

Comparison of the geochemistry of the New Jersey dikes to average Ocean Island Basalt (OIB), enriched Mid-Ocean Ridge Basalt (E-MORB) and normal Mid-Ocean Ridge Basalt (N-MORB) (Table 2) reveals the close similarity of the dikes to an OIB-like mantle source, as observed also by Puffer (2003), and to some extent E-MORB, and the dissimilarity of the dikes to an N-MORB-like source. Based on the geochemistry of the dikes, a reasonable interpretation for the generation of their magmas may be summarized as follows. Earliest rifting in the Highlands tapped an enriched OIB-like deeper mantle source that had undergone a smaller amount of partial melting, giving rise to magma that fed the type 3 dikes (most evolved). As rifting progressed and the crust became more extended, somewhat more depleted OIB-like mantle from a shallower source that had undergone a greater amount of partial melting, possibly by the rising hotter, more enriched type 3 and type 2 dike mantle source with which it may have mixed, produced the magma that fed the type 1 dikes (least evolved). This is supported by the systematic decrease in  $\text{TiO}_2$ ,  $\text{Zr}/\text{Y}$ ,  $\text{Ce}/\text{Yb}$ , and  $\text{Sm}/\text{Yb}$ , and increase in  $\text{Zr}/\text{Nb}$  and  $\text{CaO}/\text{Al}_2\text{O}_3$  from type 3 dikes, through type 2, to type 1 dikes (Table 2). Geochemical similarities of all dikes, particularly the  $\text{TiO}_2$ ,  $\text{FeO}$ ,  $\text{MgO}$ ,  $\text{Al}_2\text{O}_3$ , and  $\text{CaO}$  contents, as well as the overall similarity of the REE patterns, hint at a magmatic lineage between the three dike types. However, the contrasting  $\text{P}_2\text{O}_5$  contents and variations in HFSE and REE preclude the fractionation of all of the dikes from a single magma. A far more likely explanation to account for geochemical differences in the dikes is the mixing of at least two different mantle sources (Fig. 9). Geochemistry of the dikes is consistent with the derivation of type 1 dikes from a shallower, more depleted mantle source and types 2 and 3 dikes from a deeper, more enriched mantle source. Nevertheless, the widespread and overlapping distribution of type 1, 2, and 3 dikes in the Highlands implies that they represent a continuum of magmatic compositions that were emplaced as part of a single magmatic event. Hence, all three of the dike types are cogenetic but not comagmatic.

An interesting and noteworthy relationship exists between geographic location of the dikes and their  $\text{P}_2\text{O}_5$  content. Despite the fact that HPG and LPG dikes overlap spatially, HPG dikes are restricted



**Figure 9.** Plot of New Jersey dikes and selected Neoproterozoic mafic dikes and flows from the central and northern Appalachians on a diagram of  $Yb/Sm_{(N)}$  versus  $Ce/Sm_{(N)}$  normalized to primitive mantle values of Sun and McDonough (1989). Curves show compositions produced by melting of spinel or garnet peridotite at increments of 0.5 to 10 percent (from Storey et al., 1997 and references therein). Symbols are: solid circles, Highlands dikes (Volkert and Puffer, 1995); open circles, Trenton Prong dikes (Volkert and Drake, 1993); asterisks, Adirondack dikes (Coish and Sinton, 1992); crosses, Hudson Highlands dikes (Ratcliffe, 1987); triangles, Tibbit Hill volcanics (Coish et al., 1985); and squares, Catoclin volcanics, Virginia (Badger and Sinha, 2004). Note the vertical trend of the New Jersey dikes intermediate between garnet and spinel mantle sources.

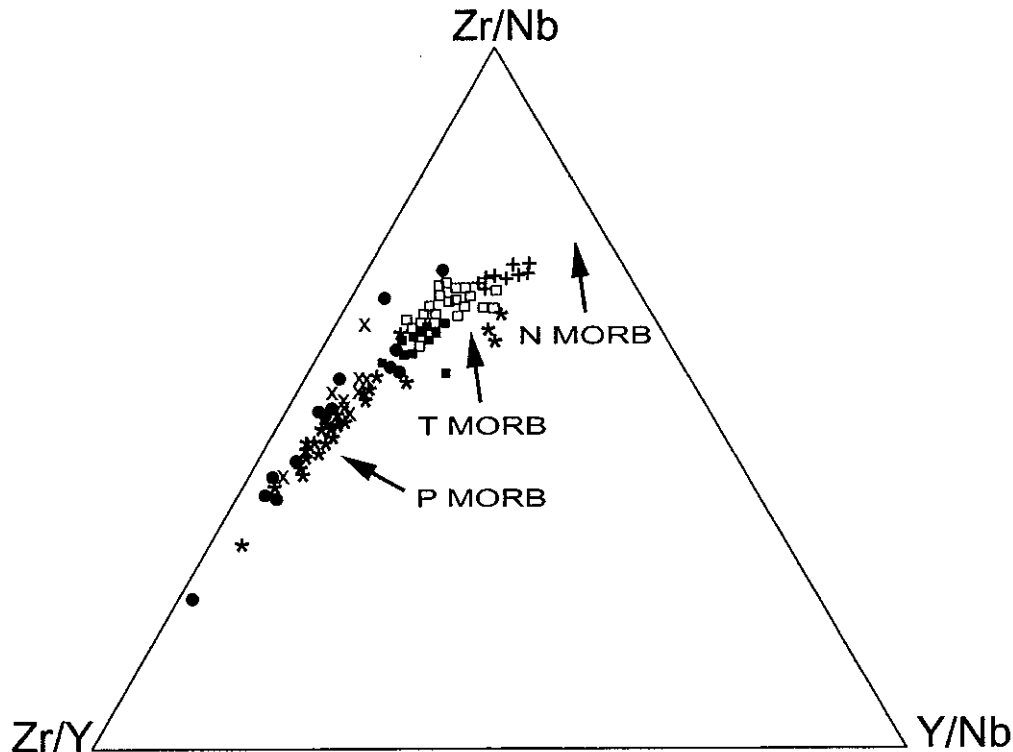
exclusively in their occurrence to the extreme western Highlands (Fig. 1). Equally noteworthy is another array of HPG dikes along heavily diked Jenny Jump Mountain to the west (Fig. 1). Jenny Jump Mountain has long been recognized as a northwest-transported segment of the Highlands (Bayley et al., 1914; Stose and Kummel, 1929). Restoring the HPG dike array and Mesoproterozoic rocks of Jenny Jump Mountain southeast to the current position of the HPG dikes in the western Highlands provides an estimate of about 4 mi (~6.5 km) of transport for Jenny Jump Mountain. Alternatively, the more heavily diked Jenny Jump crustal block may have been tectonically transported over the more sparsely diked eastern Highlands from a location to the southeast during Paleozoic orogenesis, as has been proposed for the Cranberry terrane in northwestern North Carolina (Rankin et al., 1989). In this interpretation the Jenny Jump crustal block would represent the more extensively thinned and rifted, and hence, diked crust closer to the edge of the Laurentian margin.

## COMPARISON WITH NEOPROTEROZOIC MAFIC DIKES AND FLOWS

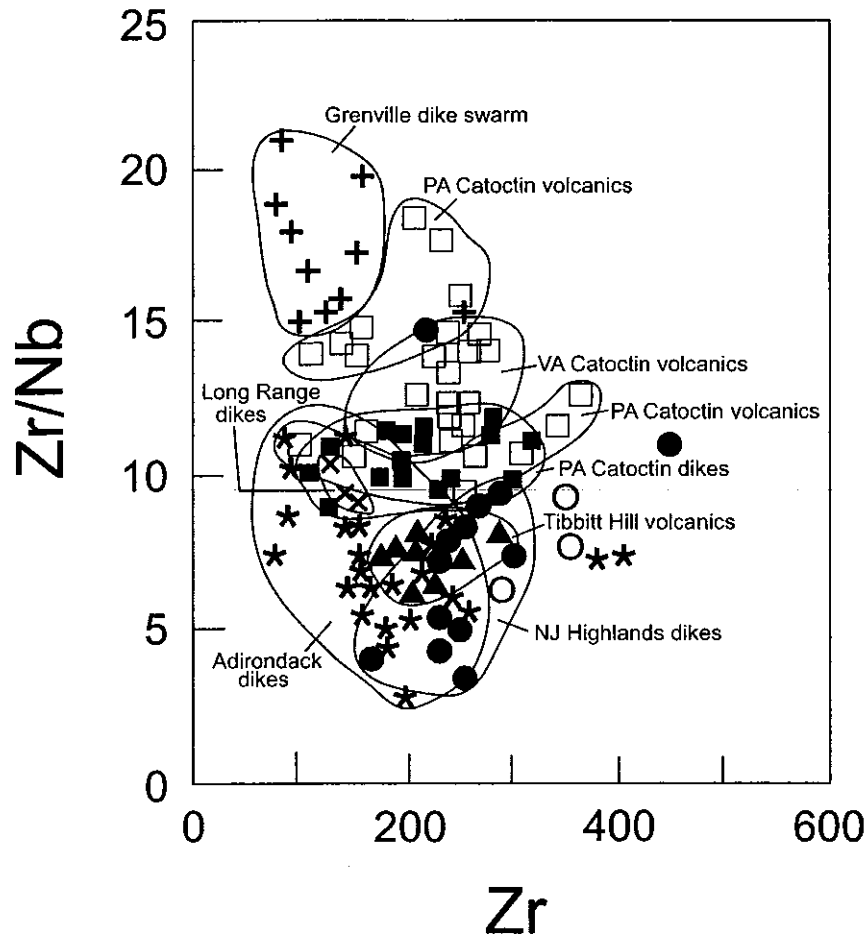
New Jersey diabase dikes have high concentrations of  $TiO_2$ , Zr, Y, Nb, and LREE that are similar to, and overlap, some late Neoproterozoic mafic dikes and flows in the north-central and northern Appalachians. Evaluation of HFSE abundances of other Appalachian occurrences, particularly Zr, Nb,



and Y, is quite useful in comparing them to the New Jersey dikes. On diagrams of Zr/Nb-Zr/Y-Y/Nb (Fig 10) and Zr-Zr/Nb (Fig. 11), dikes from the Highlands consistently display a geochemical affinity to dikes from the Adirondack Mountains and the Tibbit Hill Volcanic Member of the Pinnacle Formation in Vermont and Quebec. HFSE data are incomplete for dikes from the southern New York Hudson Highlands, preventing a comparison on these figures. Nevertheless, they are quite similar to the New Jersey dikes in most other elements. Thus it is not surprising that most of the closest regional occurrences are similar geochemically to the New Jersey dikes and they probably represent the continuation of a contemporaneous swarm. New Jersey dikes share some similarities to, but are less like, Catoctin



**Figure 10.** Plot of Neoproterozoic mafic dikes and volcanic rocks from the central and northern Appalachians on a diagram of Zr/Nb-Zr/Y-Y/Nb. Arrows point to fields for normal (N) MORB (Mid-Ocean Ridge Basalt), transitional (T) MORB, and plume (P) MORB. Symbols are: solid circles, New Jersey dikes; asterisks, dikes from the Adirondack Mountains, New York; x's, Long Range dikes, Labrador, Canada (Murthy et al., 1992); solid squares, Catoctin dikes, Pennsylvania (Smith and Barnes, 1994); open squares, Catoctin volcanics in Pennsylvania (Smith et al., 1991) and northern Virginia (Spencer et al., 1989); and crosses, Grenville dike swarm from the Ottawa graben, Canada (St. Seymour and Kumarapeli, 1995). Note the geochemical similarity of the New Jersey dikes to plume-derived basalt.



**Figure 11.** Plot of Neoproterozoic mafic dikes and volcanic rocks from the central and northern Appalachians on a diagram of Zr/Nb versus Zr. Symbols as in figures 9 and 10 except solid triangles are Tibbit Hill volcanics, Vermont. Data sources as in figures 9 and 10.

dikes and flows, and even less like the Grenville dike swarm from the Ottawa Graben (Figs. 10, 11).

The similarity of the New Jersey dikes to the Adirondack dikes and Tibbit Hill volcanics is interesting given their distance from the Highlands. The Tibbit Hill is the most proximal to the location of the triple junction and formed from enriched plume-source mantle near the plume center (St. Seymour and Kumarapeli, 1995; Coish and Sinton, 1992). Tibbit Hill volcanics differ from Grenville dikes in the Ottawa Graben that were emplaced into the failed rift arm to the west of the triple junction. Grenville dikes display generally less enriched compositions from a shallower, more depleted mantle source (St. Seymour and Kumarapeli, 1995). This is seen on Fig. 9 as well as Fig. 11, in which Grenville dikes plot at high Zr/Nb (15-20) compared to dikes from New Jersey, the Adirondacks, and Tibbit Hill volcanics, all of which have lower Zr/Nb (5-10). The New Jersey dikes clearly were formed from the same, or a similar mantle source(s) as the Tibbit Hill volcanics and Hudson Highlands dikes (Figs. 9-11). However, the 554 Ma age and lithologic associations of the Tibbit Hill volcanics may be better explained by late rift-related volcanism that immediately preceded opening of the Iapetus Ocean. These rocks are likely not part of the earlier plume-related magmatism (St. Seymour and Kumarapeli, 1995), although they may have formed from the same or a similar source. The Adirondack dikes seemingly tapped a much deeper asthenospheric mantle source with garnet in the residue (Coish and Sinton, 1992) (Fig. 9), and they differ from the New Jersey dikes mainly in this respect.

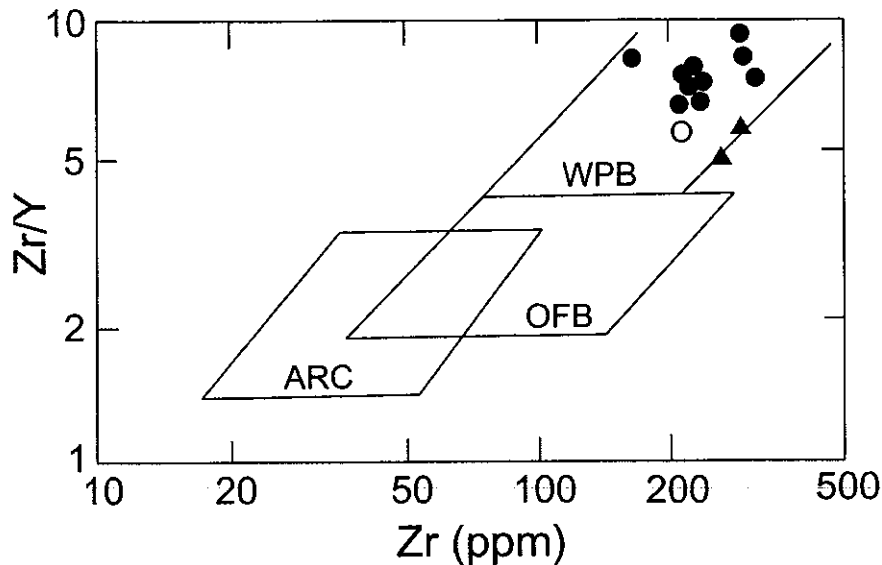
Based on the geochemical data, regional compositional differences in Neoproterozoic dikes and flows from the central to the northern Appalachians appear to correlate to variations in the mantle sources of each of the occurrences related to crustal thickness. Thicker crust during rifting would have prevented the rapid ascent of magma through the lithosphere, resulting in smaller amounts of partial melting and more enriched compositions from a deeper asthenospheric mantle source. Conversely, thinner crust would have resulted in a greater amount of melting from a shallower, somewhat more depleted source. Fig. 9 demonstrates a clear progression of greater crustal thickening northward, from thin crust and a shallower mantle source in the Blue Ridge (Catoctin dikes and volcanics), through intermediate crustal thickness and mantle sources in New Jersey and the Hudson Highlands, to thicker crust and a deeper mantle source for dikes in the Adirondacks.

## TECTONIC SETTING

Various discrimination diagrams using immobile elements were used to define a specific tectonic setting that is consistent with the geochemistry of the New Jersey dikes. All of the dikes analyzed plot in the field of within-plate basalts on diagrams of Zr-Zr/Y (Fig. 12) and Nb-Zr-Y (Fig. 13), as well as on a number of other diagrams (not shown). Ratios of HFSE in the dikes, particularly Zr/Y (9.0) and Ti/Y (657), are consistent with basalts formed in a within-plate setting. Both the LPG and HPG dikes display similar elemental abundances on an incompatible trace element diagram (see Volkert and Puffer, 1995, Fig. 10) and “humped” shaped patterns that are characteristic of basalts from within-plate settings (Pearce, 1993). Emplacement of the dikes into a continental setting is supported further by their occurrence within Mesoproterozoic rocks of the Highlands and Trenton Prong that represent the eastern Laurentian continent and their absence in overlying Paleozoic-age continental margin shelf clastic and carbonate rocks. The Lower Cambrian Hardyston Quartzite records the fluvial-to-marine transition in the north-central Appalachians (Simpson et al., 2002) related to the onset of drift-facies sedimentation and the Cambrian marine transgression that occurred at ca. 540 Ma (Bowring et al., 1993; Cawood et al., 2001).

In general, earliest Neoproterozoic rift-related rocks in the northern Appalachians display high values of TiO<sub>2</sub>, Zr, Zr/Y, and LREE and are alkalic to transitional alkalic with OIB-like compositions in contrast to later rift-related rocks that are tholeiitic and have MORB-like compositions (Coish et al., 1991). Although the New Jersey dikes are a considerable distance from the triple junction and plume axis, their compositions are more consistent with magmatism that formed during the early stages of rifting rather than the later stages that produced more depleted N-MORB-like compositions. Therefore, emplacement of the diabase dikes in the Highlands must have occurred during rifting in a continental setting prior to opening of the Iapetus Ocean. This is supported by the geochemical affinity of the dikes to early rift-related magmatism and an enriched OIB-like source, rather than a later rift-related N-MORB-like source.

As noted earlier, rifting along the eastern margin of Laurentia between ca. 615 and 554 Ma has been attributed to the effects of a mantle plume northeast of the Adirondack Mountains and east of Montreal that resulted in the formation of a triple junction (Kumarapeli, 1993). The southern arm of this triple junction extends through the eastern Adirondack Mountains and projects southward, through the Hudson Highlands and then into New Jersey. The geochemical affinity of the Tibbit Hill volcanics and mafic dikes in the Adirondacks, Hudson Highlands and New Jersey suggests that these occurrences were likely emplaced into the same extensional rift arm, although not from the same mantle source(s) and not synchronously. Magmatism along this southern rift arm may have been continuous or it may have occurred in separate pulses. Evidence supporting the latter is proposed by Ernst and Buchan (2001), who point out that continuous rift-related magmatism seldom spans intervals of more than 10-20 Ma.

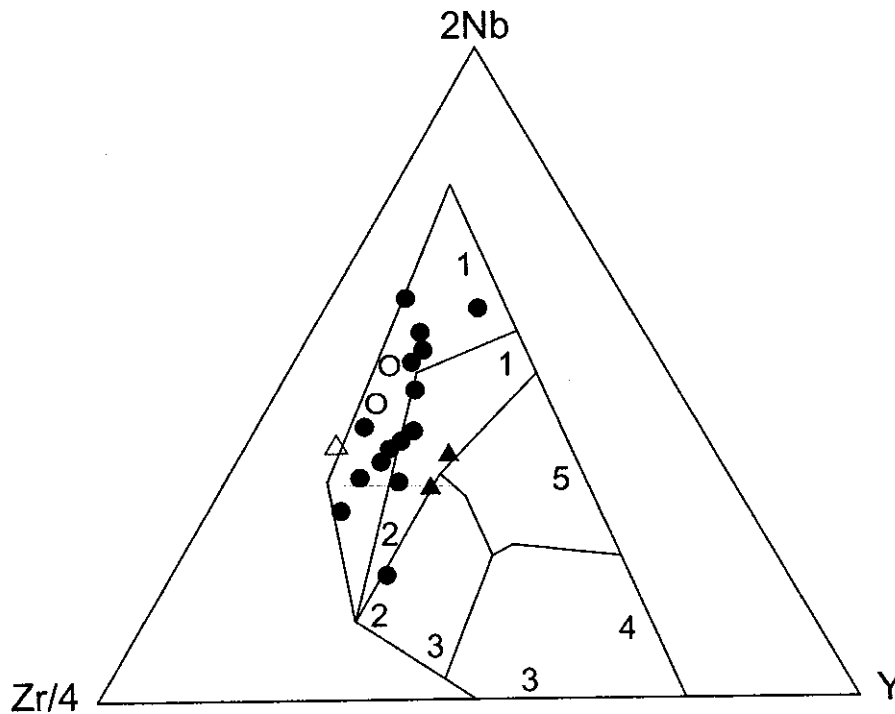


**Figure 12.** Plot of Neoproterozoic dikes from the Highlands and Trenton Prong on Zr/Y versus Zr diagram of Pearce and Norry (1979). WPB, within-plate basalt; OFB, ocean-floor basalt; and ARC, island-arc basalt. Symbols as in figure 7.

Therefore, a continuous rifting event of 65 Ma (615-550 Ma) is perceived by them as unreasonably long and was broken up into an early Central-Iapetus event (617-613 Ma), middle Central-Iapetus event (592-589 Ma), and late Central-Iapetus event (575-550 Ma). Age determinations of 602 Ma from the northern Appalachians (Williams et al., 1985) and the north-central Appalachians (Smith, 2003) suggest that the age span of the middle Central-Iapetus event may need to be revised. Regardless, the time of emplacement of the New Jersey dikes very likely falls within the middle or late events.

The New Jersey Highlands dikes may have been feeders for mafic volcanic rocks that have since been removed by erosion. Contact relationships between the diabase dikes and the Neoproterozoic Chestnut Hill Formation are never seen. Thus, it is difficult to fix the timing of magmatism and dike emplacement relative to clastic sedimentation and felsic volcanism during the Neoproterozoic in this rift basin in the Highlands. Elsewhere regionally, bimodal mafic and felsic volcanism and rift-facies clastic sedimentation were formed after emplacement of mafic dikes. Such is the case in the Catoctin Formation (Dilliard et al., 1999), as well as in the Ottawa Graben, where crustal extension and dike emplacement preceded volcanism and sedimentation and occurred ~35 Ma prior to formation of the Iapetus Ocean (Kumarapeli, 1993; Kamo et al., 1995). If these successions are an appropriate analog for the Chestnut Hill Formation, then felsic volcanism and clastic sedimentation in the Highlands would have similarly followed emplacement of the voluminous dikes.

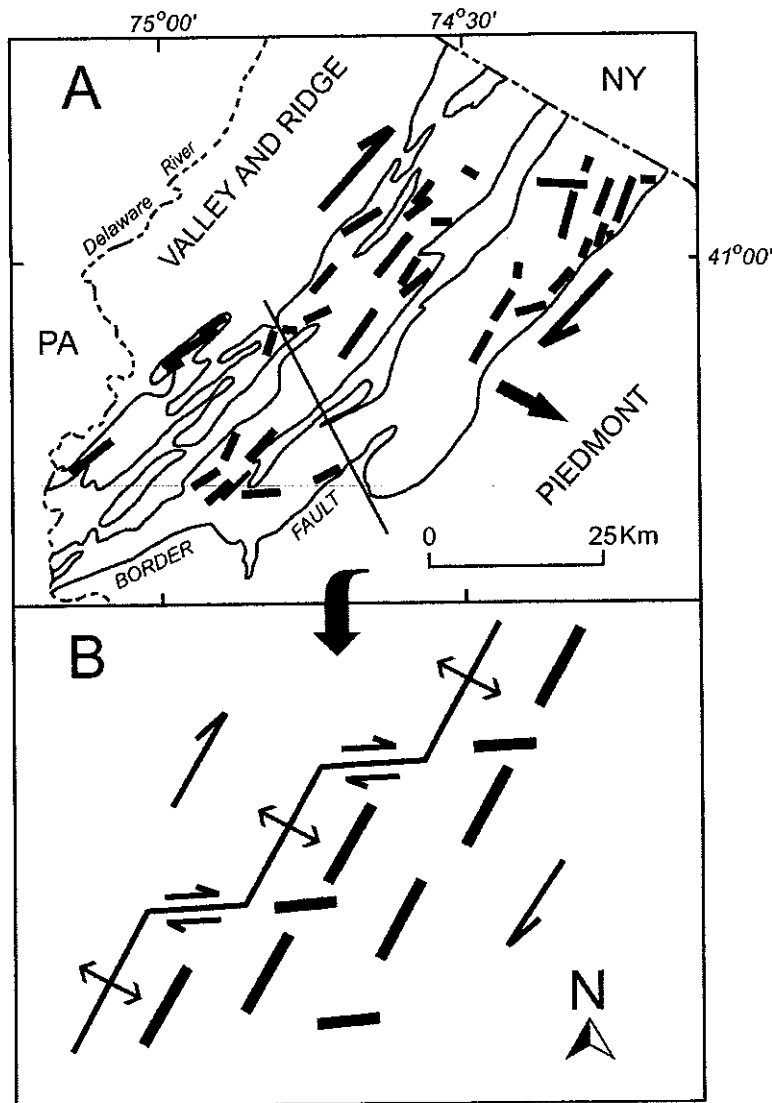
Orientations of Neoproterozoic dikes in the Highlands may be used, albeit with caution, to estimate conditions of regional stresses associated with crustal extension during the late Neoproterozoic. Dike orientations do not match the dominant N50°W to N60°W joint trend that formed in Mesoproterozoic rocks during uplift following the Grenville orogeny, precluding the localization of dike emplacement into these pre-existing fractures. Similarly, the slight to moderate discordance of many dikes in relation to foliation in Mesoproterozoic rocks rules out the emplacement of dikes along lithologic contacts or zones of lithologic heterogeneity. Significant features of the dikes are their linear, predominantly northeast trend and common en echelon occurrence. However, the interspersed east-west-trending dikes between northeast-trending dikes provides an important piece of information.



**Figure 13.** Plot of Neoproterozoic dikes from the Highlands on Nb-Zr-Y diagram of Meschede (1986). Fields are: 1, within-plate alkalic; 2, within-plate tholeiitic; 3, volcanic-arc basalt; 4, N-type MORB; and 5, P-type MORB. Symbols as in figure 7.

East-trending dikes are geochemically indistinguishable from the northeast-trending dikes and the regional trend of the two populations defines a rhomb shape with long northeast segments and short east-west segments (Fig. 14). Southeast-directed extensional stresses alone cannot account for this dike pattern. However, right-lateral strike-slip shear stresses, combined with extension to the southeast, are consistent with the pattern. If correct, then rifting in the Highlands and Trenton Prong likely resulted in the development of rhomb-shaped pull-apart basins in response to the combination of extensional and shear stresses during rifting. "Z"-shaped rhomboidal basins are common along active continental margins, resulting from right-lateral strike-slip and extensional stresses, accompanied by crustal thinning, high heat flow, and volcanism (e.g., Mann et al., 1983). Neoproterozoic dikes in New Jersey were emplaced into crustal-scale, dilational, en echelon fractures that developed oblique to lithologic contacts and that began to intersect along east-west right-stepping segments as rifting progressed and the crust became more attenuated. A similar pattern of right-stepping segments in Neoproterozoic dikes has also been observed in the Hudson Highlands (Ratcliffe, 1987).

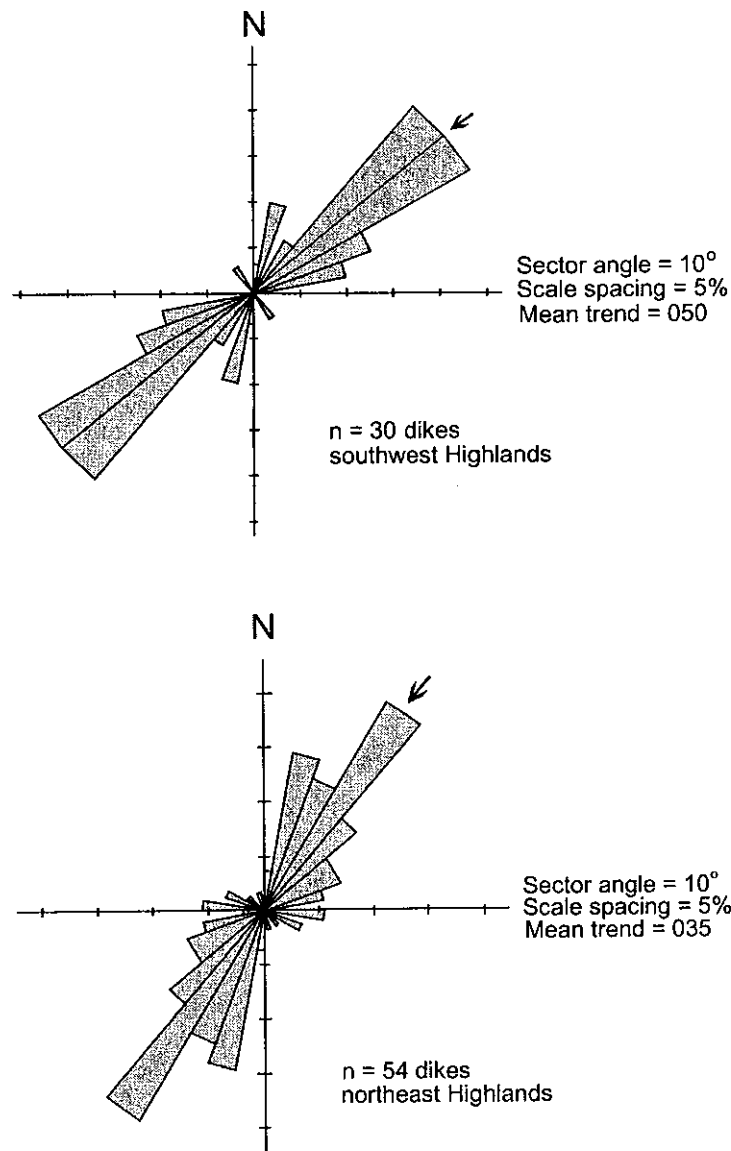
Although most of the New Jersey dikes have relatively uniform northeast trends, reorientation of at least some of the dikes may have occurred during Paleozoic orogenesis. The overall N44°E trend of the dikes conforms closely to the regional fabric in Mesoproterozoic rocks in the Highlands and Paleozoic rocks in the Valley and Ridge. This trend may reflect a clockwise shift of 10° to 20° in orientation toward the east from dominant trends of N25°E to N35°E for Catoclin dikes in northern Virginia (Bartholomew, 1988) and N20°E to N30°E for dikes in the Hudson Highlands (Ratcliffe, 1987). Interestingly, 24 percent of the New Jersey dikes match this regional Neoproterozoic trend of N20°E to N35°E compared to 27 percent of the Highlands dikes that trend N40°E to N50°E. Comparison of dike trends in the southwest Highlands, where Mesoproterozoic rocks display clear evidence of tectonic transport, to dike trends in the northern Highlands, where tectonism was mainly vertical, reveals an interesting pattern.



**Figure 14.** A) Map showing dike trends in relation to inferred crustal movement during the late Neoproterozoic in the New Jersey Highlands. Dikes are not to scale and not all can be shown. Nevertheless, the trends depicted are regionally representative. Diagonal line divides the Highlands into southwest and northeast domains for structural plots in figure 15. B) Diagram illustrating crustal stresses during rifting and dike emplacement in the Highlands. En echelon dike patterns with characteristic northeast long segments and east-west short segments are best explained by southeast-directed extension in combination with right-lateral shear couples. The resulting configuration fits that of a rhomb-shaped pull-apart basin.

Dikes in the southwest have a mean trend of N50°E compared to the mean trend of N35°E for dikes in the north (Fig. 15). The latter is more consistent with other Appalachian Neoproterozoic dike trends. Consequently, it is possible that Paleozoic reorientation of the dikes was mainly localized to the southwest Highlands and had only a minor effect elsewhere in the Highlands.

The diachronous age of Neoproterozoic siliciclastic successions and volcanic and plutonic rocks between the southern Appalachians (760 to 680 Ma) and the northern



**Figure 15.** Rose diagrams showing trends of 30 dikes from the southwest Highlands (top) and 54 dikes from the northeast Highlands (bottom). Arrows delineate the statistical mean of the populations. Note the 15° variance in dike trends from the southwest to the northeast.

Appalachians (615 to 554 Ma) has been suggested to reflect a two-stage history for Iapetan rifting of Eastern Laurentia (e.g. Aleinikoff et al., 1995). Earliest rifting occurred in the southern and central Appalachians and preceded the initiation of rifting in the north-central and northern Appalachians by >100 Ma. Rocks of this older rift event are not recognized north of Virginia. Although this earliest rifting was a significant and widespread event, it did not lead to separation of the continental margin and opening of the Iapetus ocean basin. However, Cawood et al. (2001) proposed that this earlier rifting event extended as far north as the Adirondack Mountains and may have created crustal weaknesses in the northern Appalachians that were later reactivated during the younger 615 to 554 Ma rifting event. The younger stage of rifting is well represented in the north-central and northern Appalachians. Exclusion of radiometric ages of 615 Ma from the Canadian Appalachians in Newfoundland, Labrador and Quebec results in a tighter range of ages of 602 to 554 Ma for emplacement of mafic dikes and flows. In all likelihood it was during this interval of time that the Neoproterozoic dikes in New Jersey were emplaced.

In addition to the predominant mafic magmatism in the north-central Appalachians, small volumes of sheet-like metaluminous, A-type granite (Pound Ridge Granite Gneiss and Yonkers Gneiss) were emplaced during the Neoproterozoic at 562 to 563 Ma at fairly shallow crustal levels into Mesoproterozoic rocks of the Manhattan Prong north of New York City (Tollo et al., 2004). Similar metaluminous to peralkaline magmatism is common in the older 760 to 680 Ma rifted successions in the southern and central Appalachians but is unknown in the north-central Appalachians outside of these two bodies. This may be due to the smaller volumes of mantle that impinged on the lower crust in this part of the Appalachians during rifting and crustal extension, resulting in relatively little lower crustal melting and the paucity of felsic magma production. An interesting question arises as to whether these A-type granites may have fed the felsic volcanics in the Chestnut Hill Formation, or whether the 602 Ma felsite dike (Smith, 2003) in the Pennsylvania Reading Prong may have been a feeder for the Chestnut Hill volcanics. More work is clearly needed to answer this and other questions pertaining to the geologic evolution of Neoproterozoic rocks in New Jersey.

## CONCLUSIONS

Abundant, widespread, unmetamorphosed diabase dikes of late Neoproterozoic age intrude Mesoproterozoic rocks of the New Jersey Highlands and Trenton Prong but are conspicuously absent in rocks of Cambrian or younger age. Dikes display a mean trend of N35°E in the northern Highlands and N50°E in the southwest Highlands, with the latter reflecting possible reorientation during Paleozoic orogenesis. East-west-trending dike segments are less common but are ubiquitous throughout the Highlands. All dikes dip vertically or steeply toward the southeast or northwest.

Neoproterozoic dikes have geochemical compositions that range from alkalic to transitional alkalic basalt to tholeiitic basalt. Dikes have typically high TiO<sub>2</sub>, Zr, Y, and LREE and fall into a low P<sub>2</sub>O<sub>5</sub> group (0.29-0.81 wt. %) and high P<sub>2</sub>O<sub>5</sub> group (1.19-1.60 wt. %). Both groups formed through fractionation of olivine, clinopyroxene, and plagioclase, but from separate enriched OIB-like mantle sources. All dikes form a geochemical continuum ranging from type 1 (least evolved) to type 3 (most evolved). Type 1 dikes were derived from a shallower (~30-35 mi/50-55 km) source that had undergone a greater amount of melting, possibly by rising, hotter type 2 and 3 dike magma from a deeper (35-45 mi/55-70 km) source. Geochemical similarities of all three types imply that dikes in the Highlands and Trenton Prong were emplaced as part of a single magmatic event as cogenetic intrusions. Dike geochemistries are consistent with magma genesis in a rift-related, within-plate tectonic setting. Emplacement of the dikes into deep, crustal-scale extensional fractures that developed during rifting in the Highlands enabled magma from a mixed garnet-spinel peridotite source to access the upper crust during breakup of the Rodinia supercontinent. Dike geometries form rhomb-shaped patterns, suggesting that dilational fractures were generated through a combination of extension to the southeast and strike-slip shear stresses.

New Jersey dikes have a geochemical affinity to mafic dikes in the New York Hudson Highlands, Tibbit Hill volcanics in Vermont and Quebec, and mafic dikes in the Adirondacks, but are much less like Catoctin mafic dikes and volcanics in the central Appalachians. The New Jersey dikes were emplaced into the southern arm of a triple junction and possibly form a cogenetic swarm with dikes from the Adirondacks, Hudson Highlands, and Reading Prong. Although the New Jersey dikes have not been dated radiometrically, timing of their emplacement most likely overlapped that of other regional dikes in the north-central and northern Appalachians that were emplaced between 602 and 554 Ma.

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# Petrology, Geochronology, Mineralogy, and Geochemistry of the Beemerville Alkaline Complex, Northern New Jersey

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## ABSTRACT

The Beemerville alkaline complex, New Jersey, occurs at the western end of the Cortlandt-Beemerville magmatic belt. The eastern end of the belt is defined by the Cortlandt complex, New York, which largely consists of ultramafic and mafic rocks. Between these two intrusive centers is a linear almost east-west trending zone of lamprophyre and felsic dikes. The Beemerville complex consists of two large nepheline syenite bodies, several diatremes, and associated lamprophyre and phonolite dikes, all of which intrude the shales and graywackes of the Ordovician Martinsburg Formation. The emplacement of these igneous bodies is post-tectonic and titanite fission-track ages indicate an emplacement age of  $420 \pm 6$  Ma. The mean apatite fission-track age is  $156 \pm 4$  Ma, which represents the time the rocks of the complex were last at temperatures of  $60^{\circ}\text{C}$  to  $100^{\circ}\text{C}$ .

The major rock forming minerals are orthoclase, nepheline, and clinopyroxene. Minor minerals are biotite and titanite. Accessory minerals are sodalite, fluorite, calcite, and opaque oxides. Cancrinite alteration of nepheline is common. Based on electron microprobe data the pyroxenes range in composition from diopside through aegirine-augite to aegirine. The feldspars in the nepheline syenites and phonolites are orthoclase ( $>Or_{80}$ ), with albite noted in some specimens. Two biotite populations can be identified: one encompasses the nepheline syenites and phonolites and the other the lamprophyre dikes and matrix material from the diatremes, an observation which suggests a genetic linkage between the lamprophyres and the diatremes.

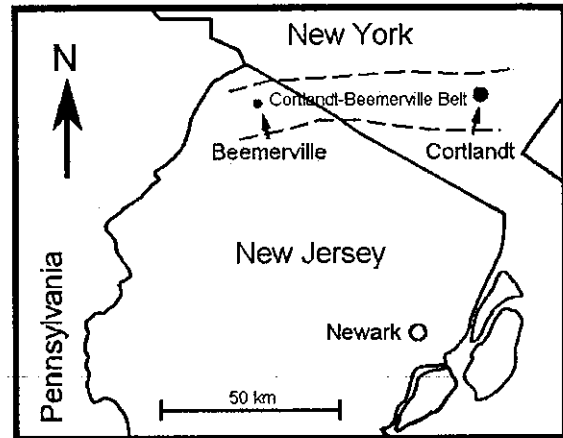
Based on textural, phase equilibria, and chemical criteria, the phases fractionating from the syenitic magmas were aegirine, nepheline, and titanite. The nepheline syenites are essentially crystal cumulates with some trapped interstitial melt. Few, if any, of the phonolites represent liquid compositions, and this is partly due to post-magmatic alteration. Nb/Ta and Zr/Hf ratios are very variable, in part due to the fractionation of titanite and in part due to hydrothermal alteration. Three distinct REE patterns are observed for the phonolites: uniform slope downwards from LREE to HREE, flattening at the HREE end, and U-shaped. These patterns are partly due to the fractionation of titanite, but a more detailed explanation most likely involves the role of fluids in altering these patterns.

## INTRODUCTION

The Beemerville complex is a member of the Cortlandt-Beemerville magmatic belt (Ratcliffe, 1981). This belt extends approximately 100 km from the Peach Lake-Croton Falls area of Westchester County, New York,  $S80^{\circ}W$  to Beemerville, New Jersey, (Fig. 1) and transects the structural grain of the Appalachians. The calc-alkalic to alkalic Cortlandt and Rosetown complexes mark the eastern end of the belt and the strongly alkalic Beemerville complex marks the western end, although the belt may extend farther to the west beneath Silurian strata. A 30 km wide belt of abundant mafic and intermediate dikes extends from Cortlandt to Beemerville. At the eastern end of the belt these dikes are weakly alkalic (spessartite, kersantite, andesite, and rhyodacite) while at the western end they are strongly alkalic (phonolite, and camptonite). Individual dikes typically have NE or NW strikes. Ratcliffe (1968, 1981) concludes that the Cortlandt complex was emplaced syntectonically relative to Taconic dynamo-thermal metamorphism and the Beemerville complex and associated dikes were emplaced post tectonically relative to folding and slaty cleavage in the Martinsburg. Nowhere (with a possible exception at Beemerville) have the rocks been observed to intrude strata younger than Ordovician age, thus fixing a minimum age for this period of igneous activity. Ratcliffe (1981) suggests that the juncture of two distinct

Taconic salients controlled the location of the belt and that the magmatic activity occurred in response to differential stresses that produced local rupturing of the continental crust and mantle. These stresses arose because of different directions or rates of plate consumption and collision during the Ordovician.

The Cortlandt complex, the largest of the alkalic – calc-alkalic complexes, intrudes metamorphosed Cambrian-Ordovician sedimentary rocks of the Manhattan Prong. Ratcliffe et al. (1982) subdivided the complex into geographic components, east of the Hudson River and Stony Point. East of the Hudson River the Cortlandt complex consists of 6 individual plutons: (1) hornblendite and kaersutite gabbro, (2) diorite, (3) clinopyroxenite, (4) cortlandtite and amphibole pyroxenite, (5) norite and biotite-amphibole norite, and (6) clinopyroxenite (Ratcliffe, 1968, 1982). The Stony Point portion of the Cortlandt complex consists of diorite, cortlandtite, and amphibole pyroxenite. The Rosetown complex consists of hornblendite, amphibole pyroxenite, diorite, and granodiorite. Models for the origin of these plutons have largely involved fractional crystallization plus crustal contamination of mantle-derived magmas (Domenick and Basu, 1982; Bender et al., 1984). Bender et al. (1982) suggested that large-scale liquid immiscibility may have played a role in the evolution of the granodiorite and diorite plutons of the Rosetown complex. The Beemerville complex consists of several nepheline syenite plutons, diatremes, and associated phonolite and lamprophyre dikes (Iddings, 1898; Arousseau and Washington, 1922; Wilkerson, 1946; Davidson, 1948; Maxey, 1976; Eby et al., 1994). In this paper new mineralogical, geochemical, and geochronological data are reported for the Beemerville complex. The emphasis will be on the geochemistry and petrogenesis of the nepheline syenites and phonolites.



**Figure 1. Location map for Beemerville and Cortlandt complexes and the Cortland-Beemerville magmatic belt. Modified from Maxey (1976).**

## **GEOLOGY OF THE BEEMERVILLE COMPLEX**

The following geologic description of the Beemerville complex (Fig. 2) is summarized from Maxey (1976). The nepheline syenite occurs in two large bodies and as several small bodies, one located in the Shawangunk conglomerate and the other associated with the diatreme of Rutan Hill. The large bodies intrude the Ordovician Martinsburg shale and to the west are overlain by the Silurian Shawangunk conglomerate. The nepheline syenite that crops out in the Shawangunk is stratigraphically higher than the basal Shawangunk and may represent an intrusion of nepheline syenite from the southern large nepheline syenite body into the Shawangunk. The Martinsburg shale has been metamorphosed to a hornfels near the boundary with the nepheline syenite plutons, but actual contacts between the nepheline syenite and the shale are not observed. The nepheline syenite is medium- to coarse-grained and consists of nepheline, orthoclase, clinopyroxene, biotite, titanite,  $\pm$ melanite, magnetite, apatite, and trace amounts of pyrite and zircon.

Several diatremes are found in the Martinsburg shale, and the largest comprises Rutan Hill. The diatremes consist of a variety of angular to subangular xenoliths and autoliths in a dark, dense matrix. Xenoliths noted by Maxey (1976) include shale and graywacke (Martinsburg Formation), fine-grained pale-blue dolomite (Kittatinny Formation?), cream-colored fine-grained limestone (Jacksonburg Formation), and gneiss (similar to that observed in the Reading Prong). Autolithic inclusions are nepheline syenite, micromelteigite, and carbonatite. The matrix consists of an extremely fine-grained groundmass with fine- to coarse-grained megacrysts of biotite, diopside, aegerine-augite, orthoclase, magnetite, apatite, and nepheline. A nepheline syenite plug, approximately 30 m in diameter, intrudes the Rutan Hill diatreme. Fenitization of the greywacke in the Martinsburg is noted in a fracture zone

associated with the small diatremes. The fenitized greywacke consists of albite, aegirine, sodic amphibole, and trace amounts of biotite and calcite.

Phonolite, tinguaita and lamprophyre dikes occur in the Beemerville area. Phonolite dikes occur in the southern nepheline syenite body, crosscut diatremes, and intrude the Martinsburg Formation. Dikes vary from a few cm to up to 50 m in width and both fine-grained and porphyritic varieties are noted. The phonolites associated with the Rutan Hill diatreme are extensively altered. Tinguaita dikes are found in both nepheline syenite bodies and the Martinsburg formation. The dikes vary in width from 0.25 to 50 m. Most are porphyritic and the common phenocrysts are nepheline, orthoclase, clinopyroxene, and titanite. The lamprophyre dikes occur throughout the complex and range in width from 10 cm to 7 m and most trend northwest. Many of the dikes are extensively altered and the primary minerals are largely replaced. Where observable, the primary minerals are diopside, aegirine-augite, biotite, nepheline, orthoclase, titanite, melanite, magnetite, and apatite. Ocelli are common in some of the dikes.

Based on gravity data, plus supporting aeromagnetic data, Ghatge et al. (1992) modeled the Beemerville complex as a thin body near the surface that broadened and elongated with depth. The Beemerville body extends to the southwest and plunges at high angle to the southeast. From their data they also inferred the existence of a second, subsurface, intrusive body approximately 9 km southeast of the Beemerville complex.

## PETROGRAPHY

### Phonolite and tinguaita

Mineralogically phonolite and tinguaita are the same rock. Historically the distinction was based on the presence of phenocrysts (tinguaita). The term tinguaita is now considered archaic. As used on the geologic map (Fig. 2), phonolite indicates a fine-grained dike rock while tinguaita indicates the presence of phenocrysts in the dike rock. The major minerals in the phonolites are orthoclase, nepheline, and clinopyroxene. Minor minerals are biotite, titanite, and opaque minerals. A fine-grained phonolite, with a texture indicative of rapid cooling, is illustrated in Figures 3a and 3b. In the porphyritic varieties, orthoclase, nepheline, clinopyroxene, and titanite are the common phenocryst phases (Figs. 4a, 4b).

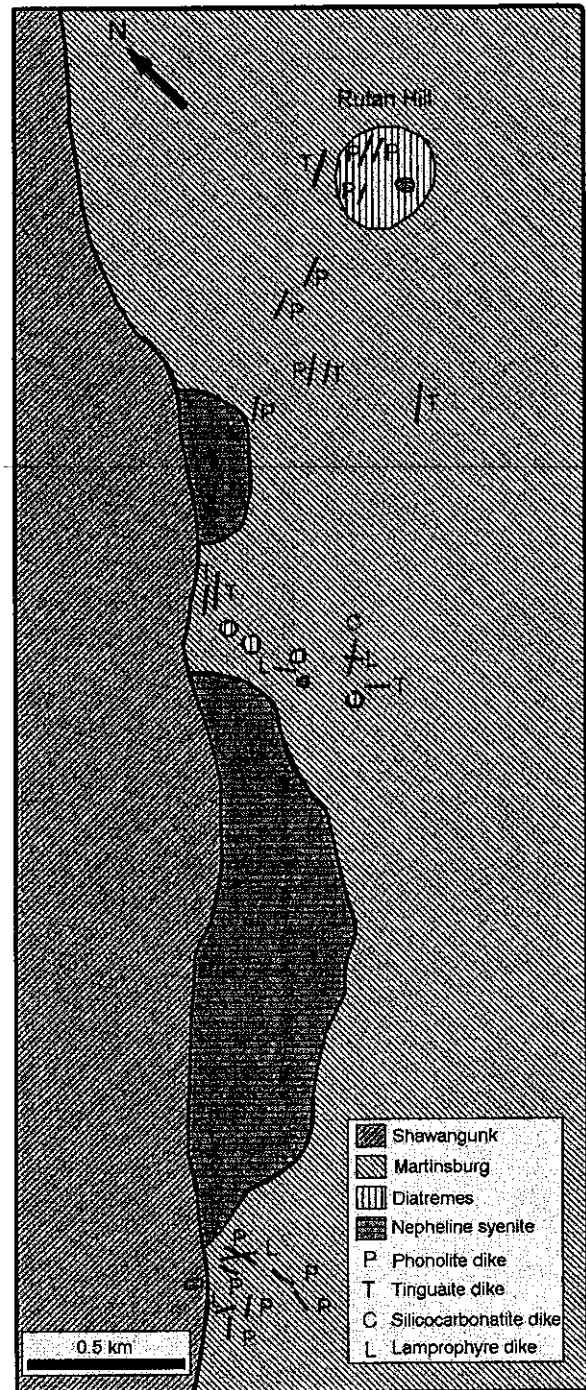
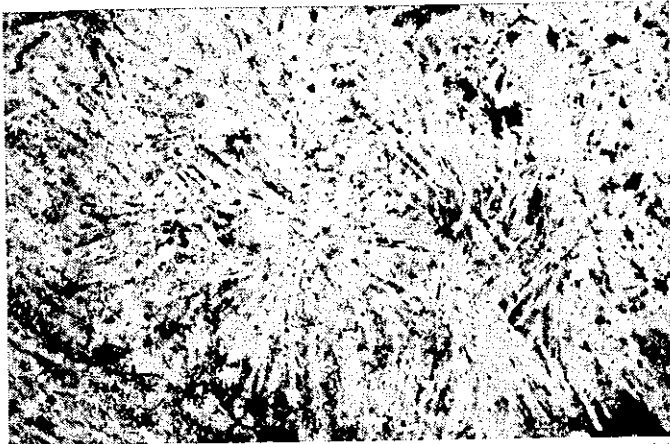


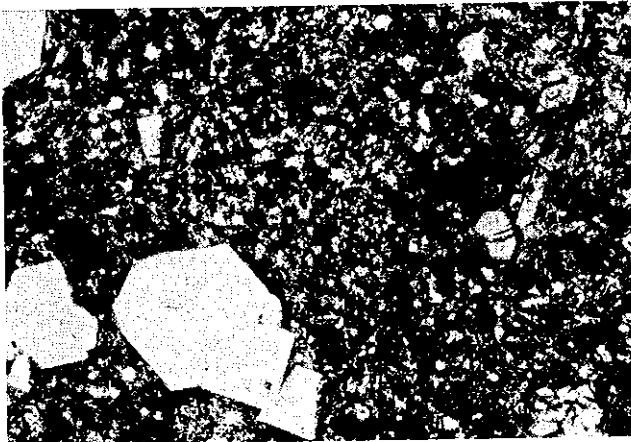
Figure 2. Geological map of the Beemerville complex. Modified from Maxey (1976).



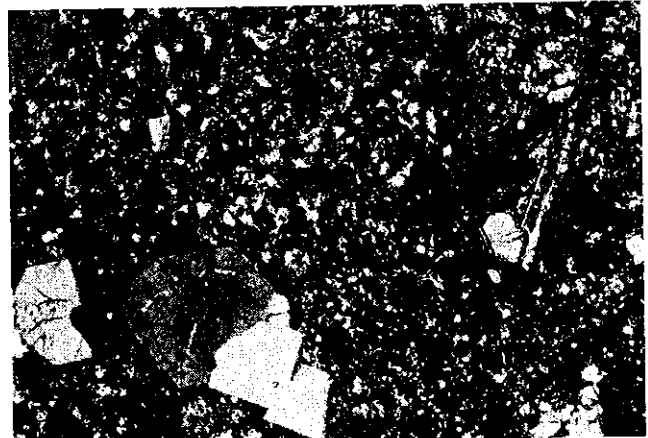
**Figure 3a.** Phonolite dike at base of Rutan Hill (BEM16). Major minerals are nepheline and orthoclase with minor aegirine-augite. Plumose texture indicative of rapid cooling. Width of field, 2 mm. Plane light.



**Figure 3b.** Crossed polars.



**Figure 4a.** Phonolite (*tinguaite*) dike (BEM19). Nepheline, aegirine, and titanite phenocrysts in a fine-grained matrix of nepheline, orthoclase, and aegirine. Width of field, 5 mm. Plane light.

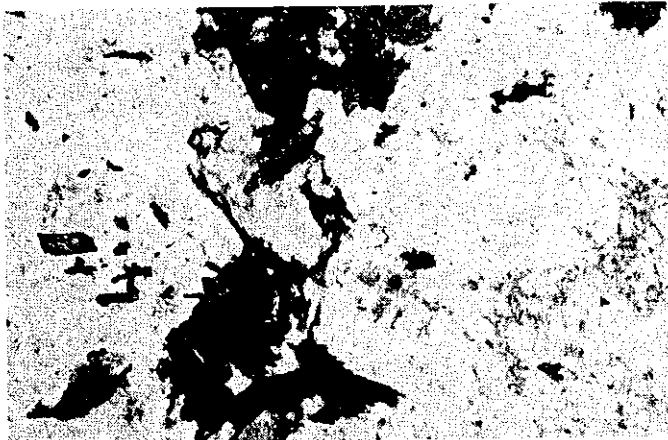


**Figure 4b.** Crossed polars.

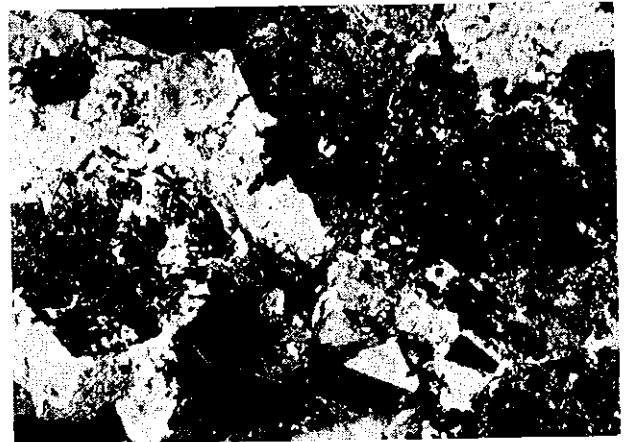
### Nepheline syenite

The nepheline syenite is medium- to coarse-grained, equigranular to subporphyritic. Major minerals are orthoclase, nepheline, and clinopyroxene. Minor minerals are biotite, titanite,  $\pm$ melanite, fluorite, cancrinite, sodalite, calcite, magnetite, and apatite. Based on probe data, the feldspars are orthoclase ( $\geq$  Or<sub>80</sub>). Optically, the feldspars do not show unmixing textures. The pyroxenes vary in composition from diopside to aegirine. Zoned clinopyroxenes have diopside cores and aegirine-augite rims or aegirine-augite cores and aegirine rims. Biotite occurs as a minor mineral either replacing clinopyroxene or as discrete grains. Cancrinite and sodalite replace nepheline. Fluorite is a common trace mineral. Figures 5 – 8 illustrate the various types and textures of nepheline syenite.

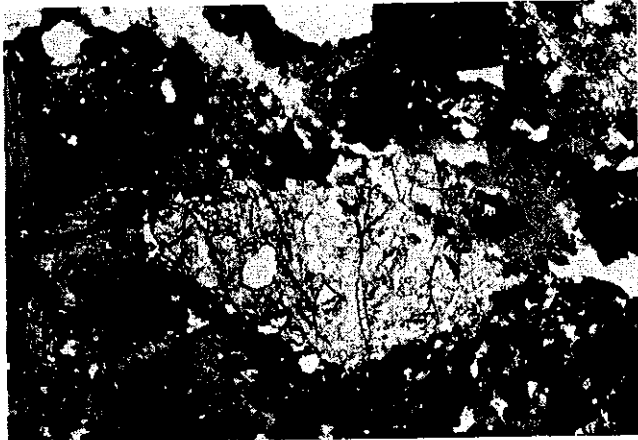




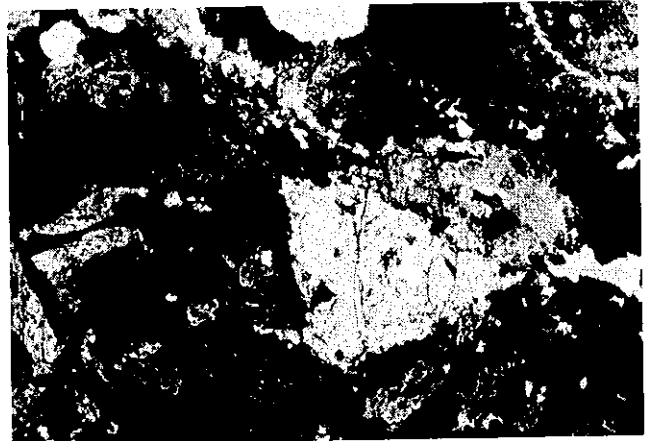
**Figure 5a.** Coarse-grained nepheline syenite (BEM4). Nepheline, orthoclase, aegirine-augite, and minor biotite. Width of field, 5 mm. Plane light.



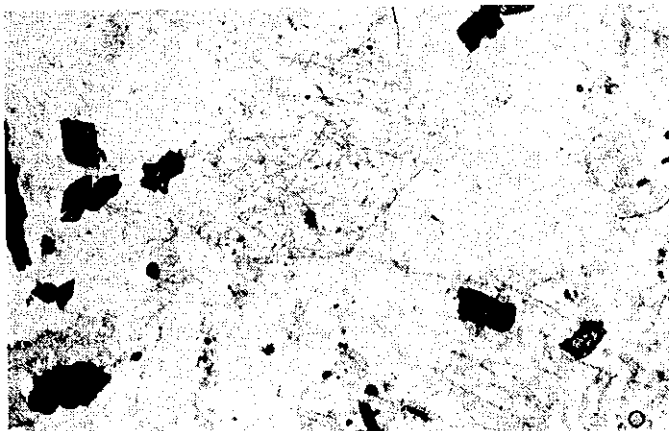
**Figure 5b.** Crossed-polars. Sodalite replaces nepheline (isotropic mineral).



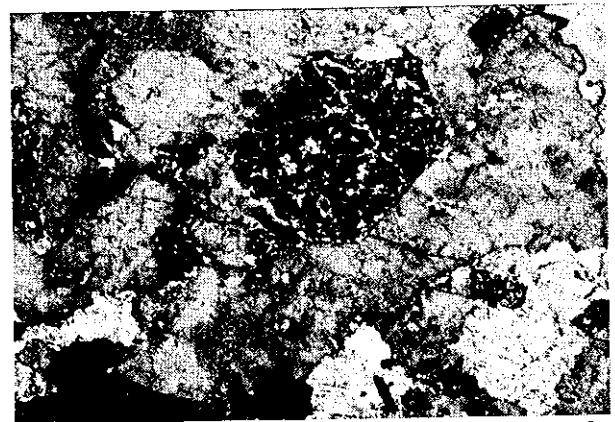
**Figure 6a.** Nepheline syenite (BEM29). Intergrown melanite and aegirine-augite, minor biotite, feldspar, titanite. Width of field, 5 mm. Plane light.



**Figure 6b.** Crossed polars. The high birefringence grain in the center of the field of view is calcite.



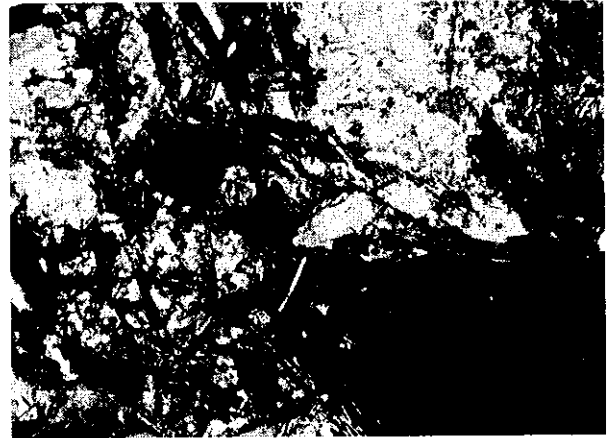
**Figure 7a.** Nepheline syenite (BEM33). Euhedral to subhedral nepheline, aegirine-augite, and titanite in a large orthoclase. Width of field, 5 mm. Plane light.



**Figure 7b.** Crossed polars. Note that the mineral inclusions in the orthoclase are the same as the phenocryst minerals in the phonolite (Fig. 4a,4b).



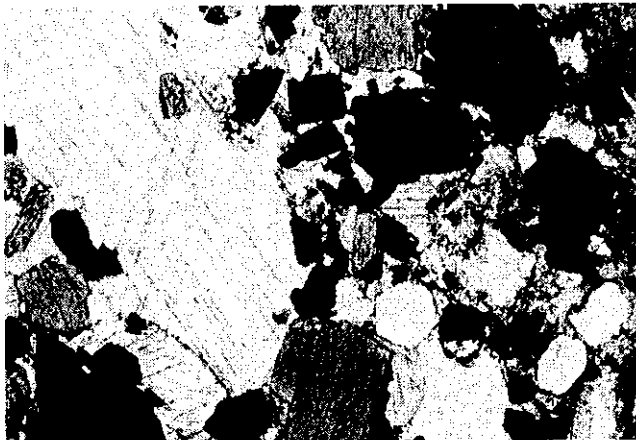
**Figure 8a.** Border phase of northern nepheline syenite body (BEM2). Phenocrysts of nepheline and orthoclase in a fine-grained matrix of feldspar and aegirine. The nepheline has been extensively altered by cancrinite. Width of field, 2 mm. Plane light.



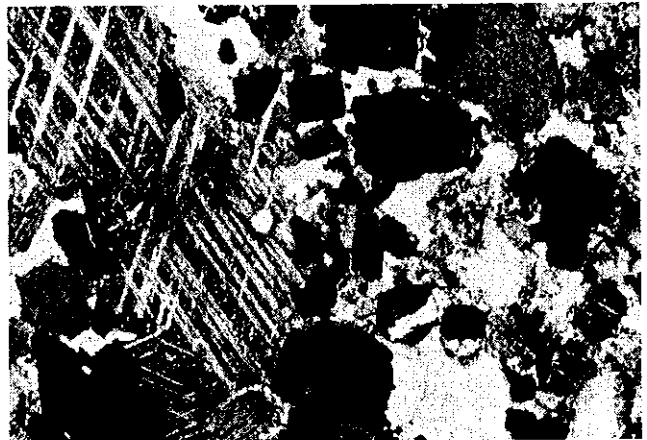
**Figure 8b.** Crossed polars. The high birefringence mineral is cancrinite.

### **Diatreme breccia**

The diatreme breccias are heterogeneous at all scales and show a wide range of xenoliths and autoliths. Locally calcite is an important component. An example of this carbonate breccia is shown in Figures 9a and 9b.



**Figure 9a.** Carbonate breccia from Rutan Hill (BEM11). Biotite, apatite, and opaque minerals in a coarse-grained matrix of calcite. Width of field, 5 mm. Plane light.



**Figure 9b.** Crossed polars.

### **Lamprophyre dikes**

Many of the lamprophyre dikes are intensely altered obscuring the primary mineralogy. For relatively unaltered specimens the major minerals are diopside or aegirine-augite, and/or biotite phenocrysts in a fine-grained matrix of clinopyroxene, biotite, nepheline, orthoclase, titanite, magnetite, calcite, and apatite. Some of the dikes contain ocelli (Fig. 10a, 10b). A more complete description of the lamprophyre dikes will be given elsewhere.

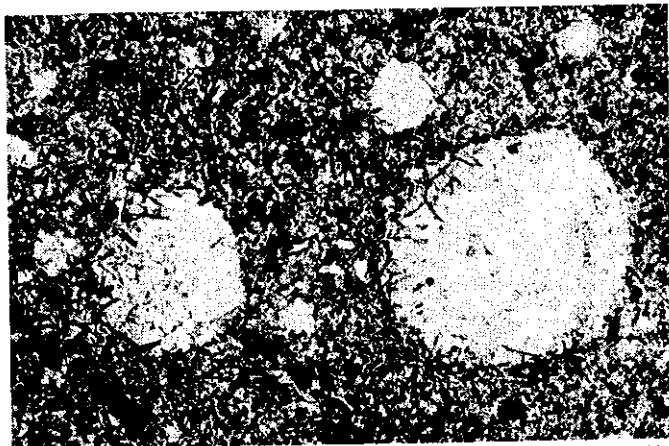


Figure 10a. Ocelli in lamprophyre dike (BEM20).  
Width of field, 5 mm. Plane light

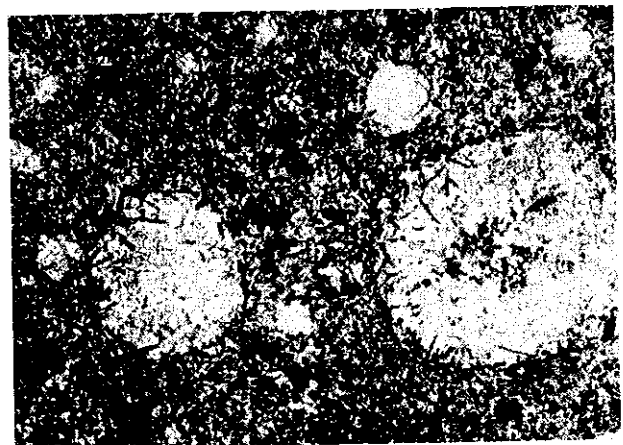


Figure 10b. Crossed polars.

## GEOCHRONOLOGY

Fission-track ages were determined for various lithologic units from the Beemerville complex. Apatite and titanite were separated using standard heavy liquid and electromagnetic techniques. Apatite fission-track ages were determined for nepheline syenites and the carbonate breccia from Rutan Hill using the population method. Titanite fission-track ages were determined for nepheline syenites using the external detector method. The zeta calibration method described by Hurford and Green (1983) and Hurford (1990) was used to calculate the fission-track ages. A detailed description of our procedures can be found in Eby et al. (1995). The analytical results are reported in Table 1.

TABLE 1. DATA FOR APATITE AND TITANITE FISSION-TRACK AGES

Sample	Number crystals	Spontaneous		Induced		Dosimeter		Age (Ma) ( $\pm 1 \sigma$ )
		$\rho_s$ ( $10^5$ t/cm $^2$ )	( $N_s$ )	$\rho_i$ ( $10^5$ t/cm $^2$ )	( $N_i$ )	$\rho_d$ ( $10^5$ t/cm $^2$ )	( $N_d$ )	
Apatite								
FC	100/50	1.37	(329)	4.78	(573)	3.01	(1877)	27.0 $\pm$ 2.2
DUR	100/50	1.12	(266)	3.46	(415)	3.16	(1877)	31.9 $\pm$ 2.7
BEM11	50/50	13.1	(1574)	7.92	(950)	3.02	(1877)	155.0 $\pm$ 10.6
BEM13	50/50	12.8	(1534)	7.48	(898)	3.03	(1877)	160.4 $\pm$ 11.4
BEM18	50/50	9.23	(1108)	5.76	(691)	3.04	(1877)	151.2 $\pm$ 13.8
BEM20	50/50	1.53	(184)	0.93	(112)	3.06	(1877)	155.5 $\pm$ 15.7
Titanite								
DROM	8/8	52.3	(1005)	49.8	(957)	6.00	(4227)	99.0 $\pm$ 2.5
FC	9/9	4.01	(346)	14.0	(1210)	6.39	(4227)	28.8 $\pm$ 1.8
ESM1	10/10	5.46	(419)	1.21	(93)	6.03	(4227)	417 $\pm$ 48
ESM2	8/8	5.44	(418)	1.25	(96)	6.42	(4227)	426 $\pm$ 48
DROM	10/10	42.7	(1025)	2.17	(520)	3.16	(1877)	97.8 $\pm$ 1.6
BEM4	10/10	11.8	(1136)	1.42	(136)	3.20	(1877)	410 $\pm$ 63
BEM5	10/10	13.7	(1317)	1.61	(155)	3.24	(1877)	421 $\pm$ 115
BEM18	12/12	8.44	(972)	1.02	(117)	3.28	(1877)	417 $\pm$ 48
BEM29	10/10	10.6	(1019)	1.26	(121)	3.31	(1877)	428 $\pm$ 75
BEM33	10/10	18.6	(1782)	2.25	(216)	3.35	(1877)	424 $\pm$ 62

Note: FC (Fish Canyon), DUR (Durango), and DROM (Mount Dromedary) are fission-track age standards. Accepted ages are FC = 27.8 Ma (Green, 1985), DUR = 31.4 Ma (Hurford and Green, 1983), and DROM = 98.8 Ma (Williams et al., 1982). All dated samples are nepheline syenites except BEM11 and BEM13, which are carbonate

breccias from Rutan Hill. ESM1 and ESM2 are the drill core samples reported in Eby et al. (1992). Zeta calibrations: apatite =  $314 \pm 10$  and titanite =  $317 \pm 6$ .

The mean apatite fission-track age for all samples is  $156 \pm 4$  Ma. There is no statistical difference between the individual ages. Apatite shows a complex history of track retention as it cools from  $120^\circ\text{C}$  to  $60^\circ\text{C}$  through the partial annealing zone (PAZ). Above  $120^\circ\text{C}$  no tracks are retained and below  $60^\circ\text{C}$  all tracks are retained (Wagner, 1990). Thus, the apatite apparent age represents the time when the sample cooled through the PAZ. The exact significance of the age depends on the cooling history (long versus short) and can be addressed using a track length study (not done here).

The mean titanite age is  $420 \pm 6$  Ma. The annealing temperature for titanite fission-tracks is  $\sim 275 \pm 25^\circ\text{C}$  (Fitzgerald and Gleadow, 1988), similar to the closure temperature for the K-Ar biotite system. The Beemerville complex was emplaced in a post tectonic setting, and given that some of the phonolite dikes have quench textures, presumably at relatively shallow depths. Hence, it is reasonable to infer that 420 Ma is the emplacement age for the Beemerville suite. A Rb-Sr biotite age of  $424 \pm 20$  Ma and a K-Ar biotite age of  $443 \pm 22$  Ma were obtained by Zartman et al. (1967) for a syenite from the Beemerville complex. Within error these ages agree with the titanite fission-track age of  $420 \pm 6$  Ma.

A large number of age determinations have been done for the various intrusions comprising the eastern end of the Cortlandt-Beemerville belt. Dallmeyer (1975) obtained  $^{40}\text{Ar}/^{39}\text{Ar}$  biotite and hornblende ages for the Cortlandt and Rosetown complexes ranging between 420 and 450 Ma. A K-Ar age of  $448 \pm 14$  Ma was reported by Long and Kulp (1962) and a Sm-Nd isochron age of  $430 \pm 34$  Ma was reported by Domenick and Basu (1982) for the Cortlandt complex. The most extensive set of geochronological data is found in Ratcliffe et al. (1982). These authors report Rb-Sr biotite ages of  $417 \pm 15$  and  $423 \pm 20$  Ma for the Cortlandt complex and a Rb-Sr isochron age of  $422 \pm 18$  Ma for the Rosetown complex. K-Ar ages reported by Ratcliffe et al. (1982) are listed in Table 2. If we eliminate the two significantly older K-Ar ages, the remaining K-Ar ages reported by Ratcliffe et al. (1982) give a mean age of  $413 \pm 12$  Ma, in reasonable agreement with the Rb-Sr ages and the Sm-Nd isochron age. Given the data currently available it is reasonable to conclude that the Cortlandt-Beemerville belt intrusions were emplaced at about 420 Ma. This is a Silurian age, and the observation by Maxey (1976) that nepheline syenite appeared to intrude the Silurian Shawangunk conglomerate is in agreement with this age. However, a Silurian age is not in agreement with the syntectonic setting suggested for the Cortlandt complex and Ratcliffe et al. (1982) interpreted the younger ages as cooling ages. Resolution of these geochronological – geological discrepancies will most likely require the acquisition of high precision U-Pb zircon ages.

TABLE 2. K-Ar AGES FOR EASTERN INTRUSIONS

Sample	Mineral Dated	Age (Ma)
Stony Point		
OC4	Hornblende	$483 \pm 10$
OC5	Biotite	$422 \pm 10$
OC6	Biotite	$396 \pm 8$
East of Hudson River		
OC9	Hornblende	$496 \pm 7$
C1	Hornblende	$419 \pm 8$
C6	Hornblende	$426 \pm 8$
C6b	Biotite	$412 \pm 8$
Rosetown biotite rhyodacite dikes		
RD-1	Biotite	$393 \pm 14$
RD-2	Biotite	$423 \pm 15$

TABLE 2. K-Ar AGES FOR EASTERN INTRUSIONS

Sample	Mineral Dated	Age (Ma)
RD-3	Biotite	413 ± 14

*Note:* Ages are from Ratcliffe et al. (1982).

## MINERAL CHEMISTRY

The chemical composition of nepheline, feldspar, biotite, pyroxene, garnet, and titanite from the various lithologic units was determined by electron microprobe. The complete data listing (as Excel spreadsheets) is available on the author's website ([http://faculty.uml.edu/Nelson\\_Eby](http://faculty.uml.edu/Nelson_Eby)). The feldspar, biotite, and pyroxene data will be discussed here and the nepheline data will be used in a later section.

### Pyroxenes

Pyroxenes show a complete range of compositions from diopside through aegirine-augite to aegirine (Fig. 11). The compositional range is similar to that of other alkaline provinces (see Eby et al., 1998), with the notable exception of extreme iron enrichment in the more Na-rich compositions. Based on the existing data, there seems to be no distinction between pyroxenes from the nepheline syenites and pyroxenes from the phonolites, with the possible exception that Na-rich pyroxenes are more common in the phonolite. Also, pyroxenes from the southern nepheline syenite body show a more restricted range in composition (diopside to aegirine-augite) than those from the northern nepheline syenite body.

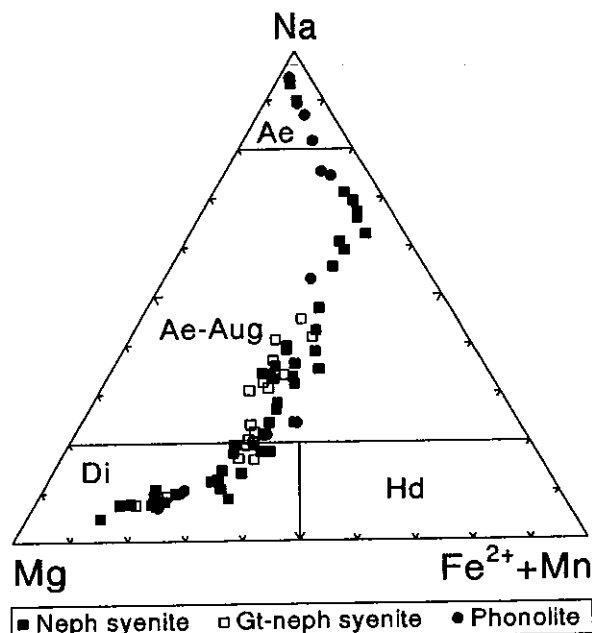


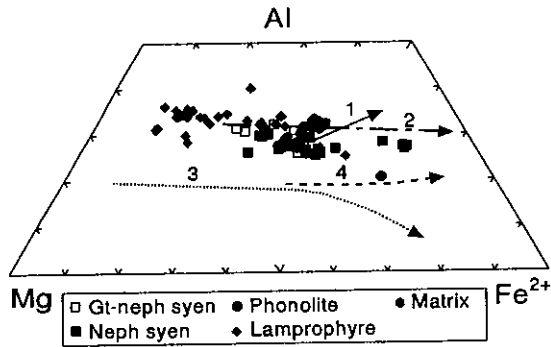
Figure 11. Compositions of Beemerville pyroxenes projected into the Mg-Na-(Fe<sup>2+</sup>+Mn) classification diagram.

### Feldspars

Feldspar compositions are plotted in the Ab-An-Or ternary feldspar diagram in Figure 12. Of note is that the feldspars from the nepheline syenites and phonolites have virtually no calcium and are K-rich (Or>80). At the microscopic level the feldspars are homogeneous, and if any unmixing has occurred it is submicroscopic (i.e., cryptoperthite). Feldspar solvi contoured on the diagram indicate that these are relatively low temperature feldspars. Only a few plagioclase grains (An<90) were analyzed and these were from a lamprophyre dike. Given the composition of the dike, these are not primary plagioclases, but are the product of hydrothermal alteration (albitization), which removed Ca from more Ca-rich plagioclases.

## Biotite

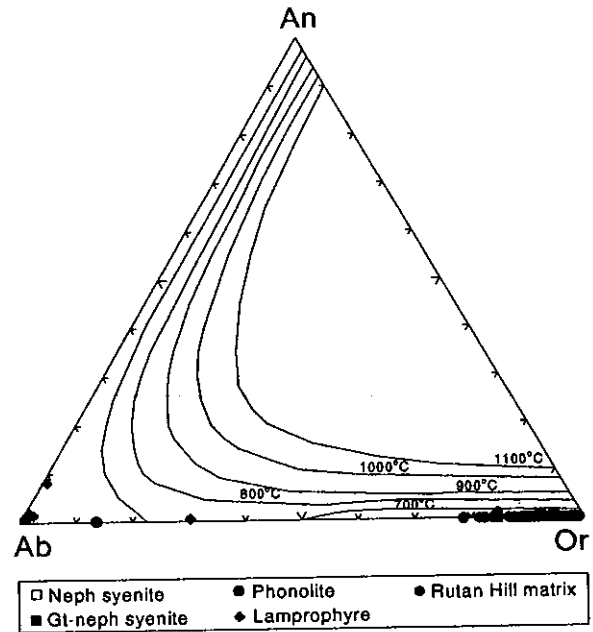
The biotite compositions are projected into the Mg-Al-Fe<sup>2+</sup> diagram in Figure 13. The Beemerville biotites tend to be more aluminous than those found in a number of other alkaline provinces. They are most similar to biotites from Igdlersfigsalik, South Greenland. There is a slight trend towards decreasing Al with increase in Fe<sup>2+</sup>. Mn and Al (atomic) are plotted versus the Mg/Mg+Fe+Mn ratio in Figure 14. In terms of both of these elements the biotites form two groups: (1) lamprophyres and matrix material from the diatremes and (2) nepheline syenites and phonolites. The phonolite biotite analyses that plot in the mafic field are from the phonolite intruded into the Rutan Hill diatreme and the inference is that these are biotite xenoliths. A second inference is that the matrix material in the diatremes and the lamprophyres are related, while the nepheline syenites and phonolites represent a separate magmatic group.



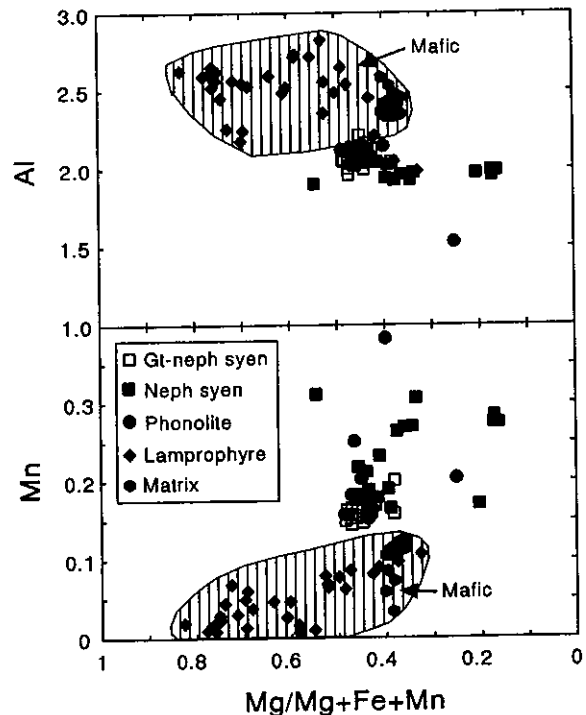
**Figure 13.** Biotite compositions plotted in the Mg-Al-Fe<sup>2+</sup> diagram. Composition trends from other alkaline provinces are included for comparison. (1) Ilomba-Ulindi, North Nyasa Alkaline Province, Malawi (Eby et al., 1998); (2) Igdlersfigsalik, South Greenland (Finch, 1995); (3) Magnet Cove, Arkansas Alkaline Province, Arkansas (Flohr and Ross, 1990); (4) Junguni, Chilwa Alkaline Province, Malawi (Woolley and Platt, 1988).

## WHOLE-ROCK GEOCHEMISTRY

Representative rock samples were reduced to fine powder using a jaw crusher (with hardened steel plates), a rotary pulverizer with ceramic plates, and final reduction by hand grinding in an agate mortar. Major elements were determined by X-ray fluorescence (XRF) analysis of fused beads using standard



**Figure 12.** Feldspar composition plotted in the Ab-An-Or ternary. Feldspar solvi (°C) are from Nekvasil (1992).



**Figure 14.** Al and Mn atomic for biotites plotted versus the Mg/Mg+Fe+Mn ratio.

(XRF) analysis of fused beads using standard

techniques. Trace elements were determined by XRF on pressed powders and by instrumental neutron activation analysis (INAA). Replicate analysis of international rock standards returned expected precision and accuracy for these analytical methods (XRF major elements,  $\pm 2\%$ ; XRF trace elements,  $\pm 5-10\%$ ; INAA trace elements,  $\pm 5-10\%$ ). LOI was determined by drying the samples at  $1000^\circ\text{C}$  for 1 hour.  $\text{CO}_2$  was determined using a CNH analyzer. A listing of the chemical analyses can be found on the author's website.

In agreement with the observed mineralogy, chemically all the nepheline syenite and phonolite samples are silica undersaturated (Fig. 15). There is an inverse relationship between the silica abundance and the percent foidal minerals. This is not an unexpected result, but the important point is that given the silica content, many of these samples, even in the case of many of the phonolite dikes, cannot represent liquid compositions.

Harker diagrams are plotted for the nepheline syenites and phonolites in Figures 16a and 16b. There is considerable scatter in the data, but  $\text{TiO}_2$  and  $\text{CaO}$  decrease with increasing  $\text{SiO}_2$ . A phonolite trend (dashed line) can be defined for  $\text{CaO}$ , but not for  $\text{TiO}_2$ . Ca principally resides in the titanites and

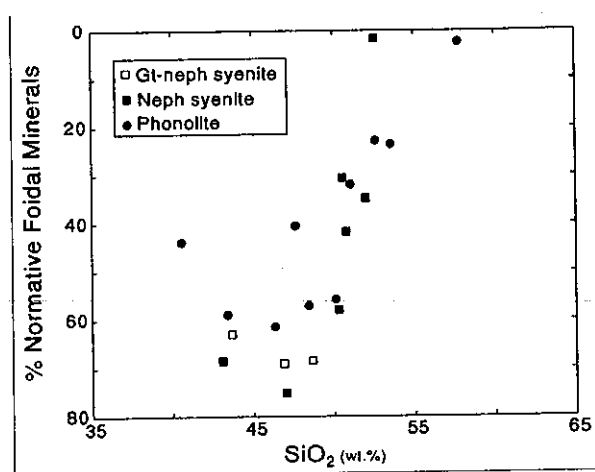


Figure 15.  $\text{SiO}_2$  versus % normative foidal (nepheline and leucite) minerals.

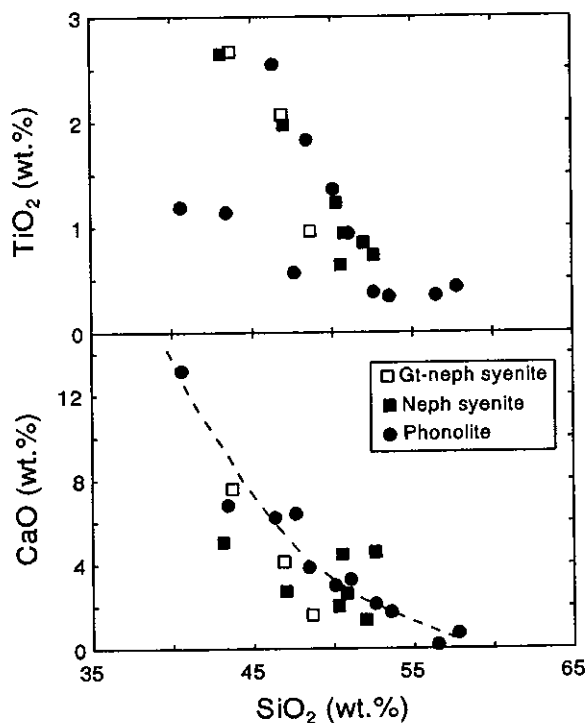


Figure 16a.  $\text{SiO}_2$  versus  $\text{TiO}_2$  and  $\text{CaO}$

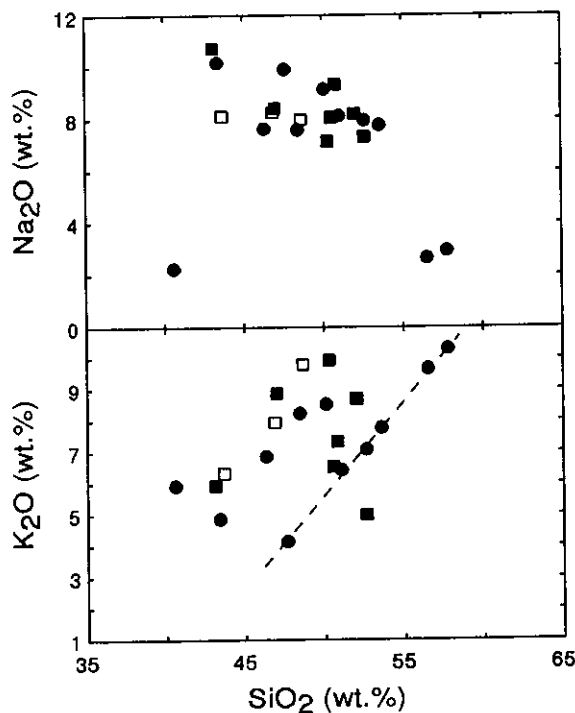


Figure 16b.  $\text{SiO}_2$  versus  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$ .

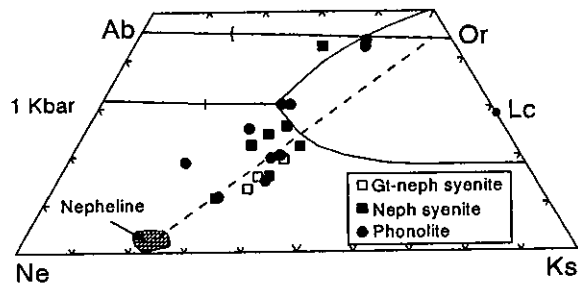
Ca-clinopyroxenes while the major repository for Ti is titanite. A plot of  $\text{CaO}$  versus  $\text{TiO}_2$  (not shown) shows that for most samples these two oxides covary. Hence, one should not interpret these variations as representing a liquid line of descent. For the alkalis,  $\text{Na}_2\text{O}$  decreases with increasing  $\text{SiO}_2$  and  $\text{K}_2\text{O}$  increases with increasing  $\text{SiO}_2$ . As before, in the case of the nepheline syenites these variations are largely

due to changes in the modal abundance of clinopyroxene and alkali feldspar. For the phonolites, a  $K_2O$  trend (dashed line) can be defined for some of the samples, but as will be discussed subsequently there has been a significant amount of element mobility.

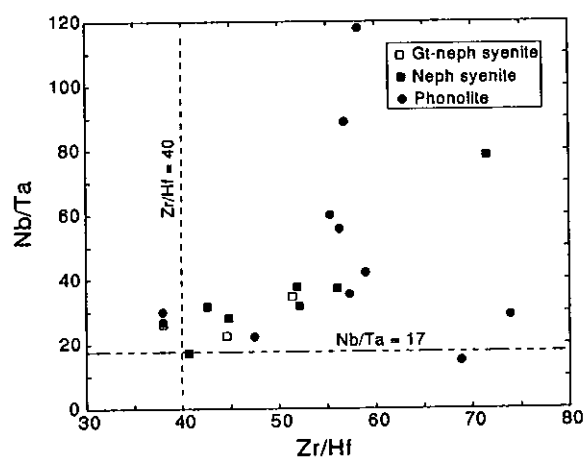
Nepheline syenite and phonolite compositions are projected into the 1 Kbar Neph-Or-Ab-Kal phase diagram in Figure 17. Also shown on the diagram is the composition field for the nephelines from the nepheline syenites and phonolites. The dashed line connects the average K-feldspar composition with the nepheline field. Of note is that many of the samples, particularly the nepheline syenite samples, lie along or near this line suggesting that the rock chemistry can largely be explained as a result of crystal accumulation of orthoclase and nepheline. A few of the phonolite dike samples plot at what might be considered liquid compositions, at pressures  $>1$  Kbar, but for most of the samples the inference is that the rocks do not represent liquid compositions. Maxey (1976) notes that many of the phonolite dikes have been albitized. In addition, cancrinite alteration of nepheline is common. Hence, there is ample evidence that the primary chemistry of the dikes has been modified by hydrothermal alteration.

Nb, Ta, Zr, and Hf are usually immobile elements in igneous rocks and the distribution of these elements has proven to be a powerful tool in unraveling petrogenetic histories. A standard ratio plot for these elements is shown in Figure 18. Typical mantle ratios are shown with dashed lines on the plot. Samples that plot in the upper right quadrant of the diagram are said to have "superchondritic" ratios, which is the case for virtually all the Beemerville nepheline syenite and phonolite samples.

Chondrite normalized rare earth element (REE) plots for nepheline and garnet-nepheline syenite (Fig. 19a), phonolites (Figs. 19b and 19c), and the rocks from Rutan Hill (Fig. 20) are shown below. None of the samples show either positive or negative Eu anomalies, which indicates that feldspar/plagioclase fractionation did not play an important role in the evolution of these magmas. Three types of phonolites (labeled I, II, and III) can be distinguished on the basis of REE patterns. Type I phonolites have a linear REE pattern, type II phonolites have patterns that flatten at the heavy REE end of the spectrum, and type III patterns are concave down (show a relative depletion in the middle REEs). The matrix material from Rutan Hill (Fig. 20) shows a linear REE pattern while the phonolite dike has a type II phonolite pattern.



**Figure 17. Nepheline syenite and phonolite compositions projected into the 1 Kbar Neph-Or-Ab-Kal phase diagram (from Henderson, 1984).**



**Figure 18. Zr/Hf versus Nb/Ta plot for nepheline syenites and phonolites.**



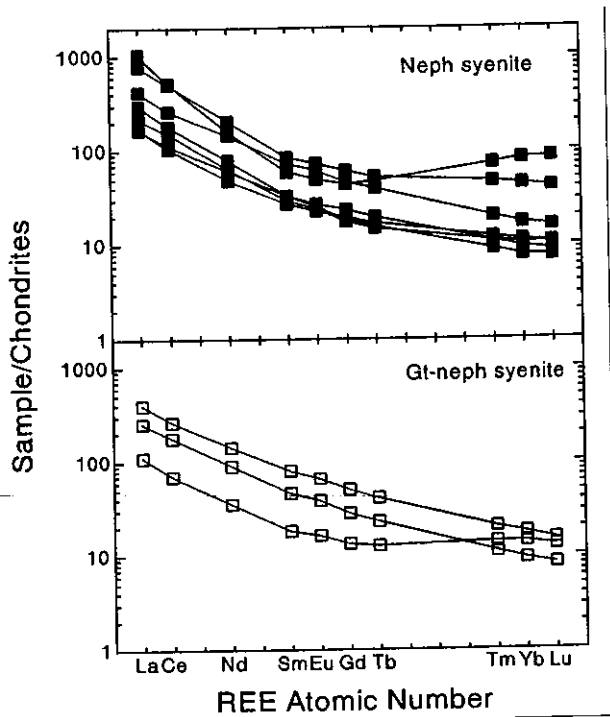


Figure 19a. Chondrite (Nakamura, 1974) normalized REE patterns for nepheline syenite and garnet-nepheline syenite.

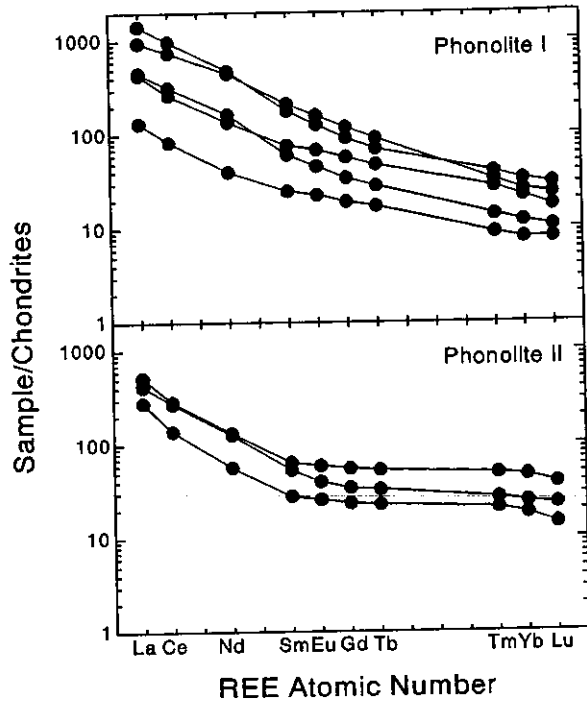


Figure 19b. Chondrite normalized REE patterns for type I and II phonolites.

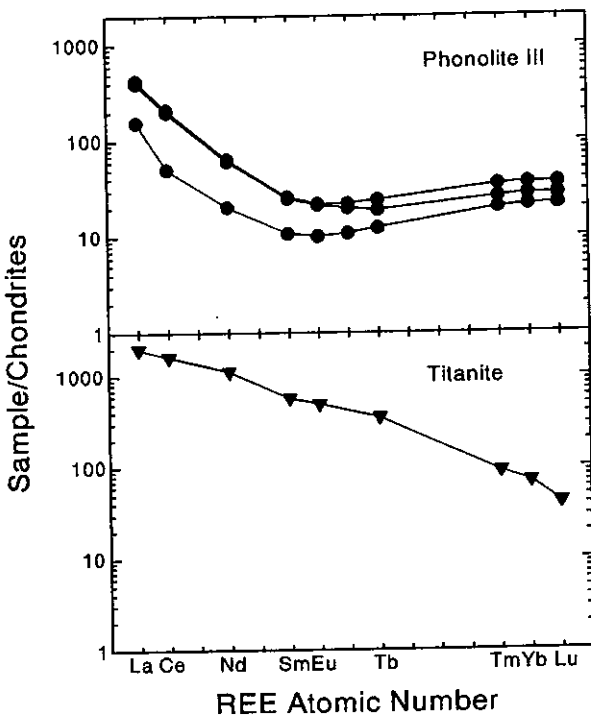


Figure 19c. Chondrite normalized REE patterns for type III phonolite and titanite from Beemerville nepheline syenite. Titanite data from Eby and Sclar (1993).

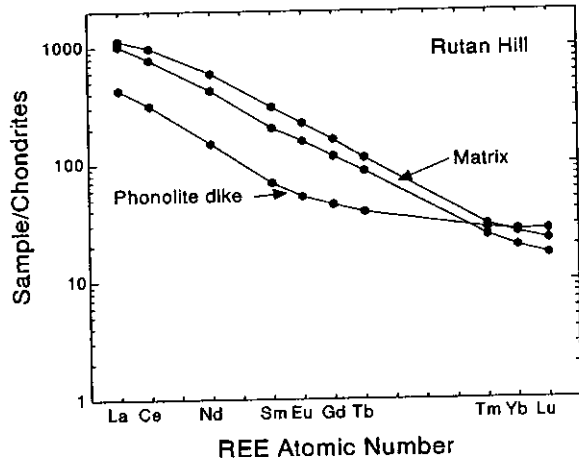


Figure 20. Chondrite normalized REE patterns for matrix material and phonolite dike, Rutan Hill.

## DISCUSSION

The major emphasis of this paper is the geochemistry and petrogenesis of the felsic alkaline rocks in the Beemerville complex, and this topic will be considered further here. The common phenocrysts in the phonolite dikes are nepheline, clinopyroxene, and titanite (Fig. 4). These minerals often occur as inclusions in orthoclase (Fig. 7) in the coarse-grained nepheline syenites. Hence, it seems reasonable to conclude that these are the major minerals fractionating from the syenitic melt. In this interpretation, orthoclase is not a fractionating phase, which agrees with the observation that none of the syenite or phonolite samples have negative (or positive) Eu anomalies. The presence of these anomalies would indicate that feldspar and/or plagioclase were fractionating (negative Eu anomaly) or accumulating (positive Eu anomaly) phases. Given this interpretation, the nepheline syenites are partially crystal cumulates and do not represent liquid compositions. As illustrated on the phase diagram (Fig. 17), many of the nepheline syenites can be explained as simple mixtures of orthoclase and nepheline.

So-called "chondritic" ratios are,  $Nb/Ta = 17$  and  $Zr/Hf = 40$ . While the data for the lamprophyre dikes is not presented in this paper, it is worth noting that for the dikes the majority of the samples plot near the chondritic ratios with scatter towards higher ratios, but not as high as those obtained for the nepheline syenites and phonolites. If we consider the mineralogy of the nepheline syenites and phonolites, the only mineral that contains significant amounts of these elements is titanite (and garnet for the garnet-bearing nepheline syenites). Zircon is the other potential reservoir of Zr and Hf, but it is rare in the Beemerville suite compared to the almost ubiquitous occurrence of titanite. For a titanite from nepheline syenite,  $Nb = 3300$  ppm,  $Ta = 164$  ppm,  $Nb/Ta = 20$ ,  $Zr = 4430$ ,  $Hf = 14.4$ , and  $Zr/Hf = 308$  (Eby and Sclar, 1993). Given the high  $Zr/Hf$  ratio for titanite, the presence of cumulate titanite could contribute to a high rock  $Zr/Hf$  ratio. The same is not the case for the  $Nb/Ta$  ratio. Green (1995) noted that in alkaline suites  $Nb/Ta$  ratios tended to fall into two groups: one characterized by consistent  $Nb/Ta$  ratios close to chondritic or mantle values and the other characterized by highly variable ratios that significantly exceed mantle values. The Beemerville nepheline syenites and phonolites belong to the latter group. Green and Pearson (1987) experimentally investigated the effect of titanite fractionation on the  $Nb/Ta$  ratio and concluded that fractionation of this mineral could increase the ratio by about 20%. Their experiments did not include silica-undersaturated liquids. Data obtained for glassy volcanic rocks (e.g., Weaver, 1990) suggest that titanite fractionation can lead to  $Nb/Ta$  ratios of 50 to 60. Given the titanite/liquid partition coefficients for Nb and Ta, fractionation of this mineral leads to a decrease in the absolute Ta abundance and an increase in the  $Nb/Ta$  ratio. For the Beemerville nepheline syenites and phonolites there is a crude inverse relationship between the absolute Ta abundance and the  $Nb/Ta$  ratio (plot not shown) and, thus, it is possible that some of the variation in  $Nb/Ta$  ratios can be explained by titanite fractionation. However, the very high  $Nb/Ta$  ratios are not readily explained by this mechanism. Vard and Williams-Jones (1993) investigated inclusions in vug minerals found in phonolite sills. They concluded that the vug filling was the result of mineral precipitation from a primary orthomagmatic fluid and that this fluid was enriched in HFSE ions, particularly Zr, Hf, Nb, and Ti. Thus, it is possible that the high  $Nb/Ta$  ratios (and  $Zr/Hf$  ratios) may be due to late stage hydrothermal processes. This would be in agreement with the observed alteration in these rocks. Hence, the high  $Nb/Ta$  and  $Zr/Hf$  ratios may represent a combination of titanite fractionation (or accumulation) plus hydrothermal alteration.

The variety of REE patterns shown by the nepheline syenites and phonolites of the Beemerville complex have been observed in a number of other alkaline provinces, e.g., Monteregian Hills (Eby, 1984, 1985), Magnet Cove, Arkansas Alkaline Province (Flohr and Ross, 1990), and the North Nyasa Alkaline Province (Eby et al., 1998). Eby et al. (1998) discussed the possible origin of these patterns in some detail. This discussion will not be repeated here, other than to note that three possibilities were considered: titanite fractionation, zircon accumulation, and fluid transport. Fluid transport by F- and  $CO_3$ -rich solutions could modify primary linear REE patterns to HREE flattening and U-shaped patterns, and the authors concluded that this was the most likely process. It is, therefore, relevant to note the occurrence of both fluorite and calcite in a number of the Beemerville nepheline syenites and phonolites, which provides mineralogical evidence for this process.

Given the degree of alteration of many of the samples, plus the variability of trace element distributions, it is difficult to draw any definitive conclusions about the petrogenetic history of the nepheline syenites and phonolites. However, based on textural evidence, we can conclude that nepheline, clinopyroxene, and titanite were the major fractionating phases. Both textural and phase equilibria arguments support the conclusion that the nepheline syenites are largely cumulates. We can also conclude that these rocks underwent significant alteration by late stage magmatic and F- and CO<sub>3</sub>-rich hydrothermal fluids that significantly altered both the major and trace element chemistry. Whether or not the phonolites are the end result of fractionation of the magmas that gave rise to the lamprophyre dikes is still an open question, but mineralogical evidence suggests that there is a relationship between the lamprophyres and the magmatic phase of the diatremes.

## ACKNOWLEDGEMENTS

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# Geochemistry and Structural Setting of the Cortlandt Igneous Complex, Stony Point, NY.

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## INTRODUCTION

The Cortlandt Complex of southeastern New York is one of the classic areas of the Appalachians. It was first seriously studied by Dana (1881) and has been periodically revisited by researchers and teams of researchers numerous times since then. The Cortlandt igneous Complex lies primarily on the east side of the Hudson River in Westchester County with only a small extension on the west side of the Hudson in Stony Point, New York (Fig. 1). The complex is a composite body with six distinct plutons ranging from granodiorite to ultramafic rocks, many with gradational contacts (Williams, 1886, Rogers, 1911, Ratcliffe et al., 1982). In addition, there is a dike swarm associated with the complex that extends westward across the Hudson Highlands and into the Valley and Ridge Province to another igneous complex at Beemerville, New Jersey (Ratcliffe, 1981; Maxey, 1976). These dikes are primarily lamprophyres but exhibit a large range in compositions.

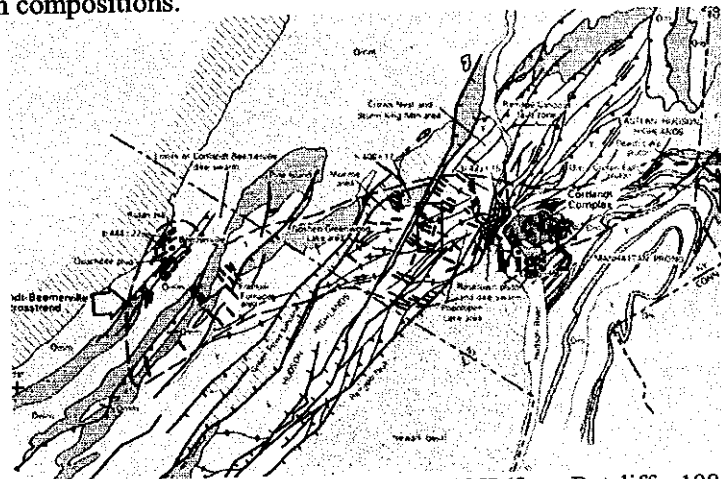


Figure 1. Geologic map of the Cortlandt-Beemerville trend in NY-NJ (from Ratcliffe, 1981) showing location of Fig. 2.

The igneous rocks intrude rocks of the northern Manhattan Prong subsequent to deformation and metamorphism that resulted from the middle Ordovician Taconic Orogeny (Ratcliffe et al., 1982). Metamorphic zonation from this event increases from greenschist through upper amphibolite facies eastward across the area of the Cortlandt Complex. There is extensive contact metamorphism of the country rock as a result of the intrusions. The most famous of these rocks are the emery deposits from ultrametamorphism of xenoliths that are completely engulfed by mafic igneous rocks (Friedman, 1956; Baker, 1964). This field trip will visit the Cortlandt Complex in the Stony Point Battlefield, NY (Ratcliffe, 1968).

## REGIONAL GEOLOGY

The country rock to the Cortlandt Complex at Stony Point consists of two units, the Wappinger Group and Annsville Formation, both of Cambrian-Ordovician age (Fig. 2) (Ratcliffe, 1968). The rocks

were metamorphosed to greenschist facies and deformed during the middle Ordovician Taconian Orogeny. The intrusion of the Cortlandt igneous complex occurred either in the late stages of the orogeny or after cessation at ca. 430±34 Ma (Domenick and Basu, 1982). There are three major intrusive units at Stony Point, plutons of diorite and hornblende (Cortlandite) and dikes of variable composition. Mesozoic faulting along the Ramapo Fault juxtaposes sedimentary rocks of the Newark basin against the older crystalline in the western part of the field area.

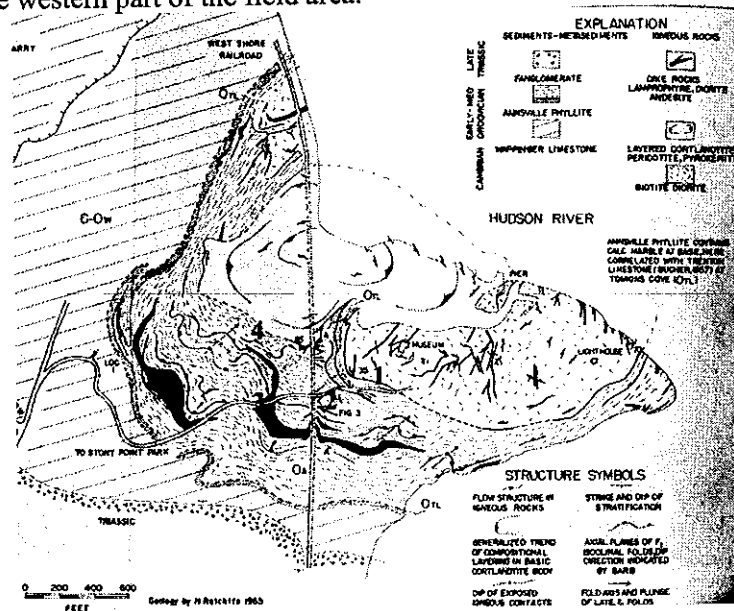


Figure 2. Geologic map of the Stony Point Battlefield area (from Ratcliffe, 1968) showing field trip stop locations.

### *Wappinger Group*

The Wappinger Group rocks consist primarily of tan dolomite and gray limestone. These rocks range from thinly bedded to massive with white veins of calcite that are locally dense. The Wappinger Group rocks are primarily Cambrian in age but range to lower Ordovician. They were deposited in a shallow water carbonate bank setting mostly as algal mats. We will not observe Wappinger Group rocks during this field trip.

### *Annsville Formation*

The Annsville Formation ranges from phyllite to migmatitic gneiss depending upon position relative to the igneous complex. In the least deformed and metamorphosed state farthest from the igneous complex, the Annsville Formation is a thinly interbedded psammite-pelite with rhythmic layering. The pelite layers are phyllite to fine muscovite schist whereas the psammites range from quartzite to quartz-muscovite schist. Even at this grade the rock is intensely deformed with isoclinal intrafolial folds almost ubiquitous to every outcrop.

In the vicinity of the igneous complex, the Annsville Formation is a biotite schist to gneiss. It contains biotite, muscovite, quartz and plagioclase with minor accessory phases. The foliation is transposed and multiply crenulated. Small quartz veins commonly cross foliation either along or crossing crenulation cleavages.

In contact with the diorite, the Annsville Formation contains garnet and sparse granite veins. The garnet is fine to medium grained and increases towards the contact. The granite veins are commonly thin (2-3 mm) though thicker veins (1-2 cm) can also be found. The thin veins are composed of quartz, K-feldspar and minor plagioclase. Thicker veins can contain tourmaline and muscovite as well.

**Table 1. Electron microprobe data for biotite**

	Diorite	Kersantite
SiO <sub>2</sub>	35.5	36.33
TiO <sub>2</sub>	2.97	2.96
Al <sub>2</sub> O <sub>3</sub>	15.4	15.99
FeO	26.08	22.48
MgO	5.82	8.26
MnO	0.26	0.35
CaO	0.08	0
Na <sub>2</sub> O	0.03	0.05
K <sub>2</sub> O	8.86	9.55
Cl	0.17	0.12
O=Cl	-0.04	-0.03
<b>Total</b>	<b>94.49</b>	<b>95.41</b>

**Normalized composition**

Si <sup>4+</sup>	5.61	5.59
Ti <sup>4+</sup>	0.35	0.34
Al <sup>3+</sup>	2.86	2.9
Fe <sup>2+</sup>	3.44	2.89
Mg <sup>2+</sup>	1.37	1.89
Mn <sup>2+</sup>	0.04	0.05
Ca <sup>2+</sup>	0.01	0
Na <sup>+</sup>	0.01	0.02
K <sup>+</sup>	1.79	1.87
Cl <sup>-</sup>	0.05	0.03
<b>Total</b>	<b>15.48</b>	<b>15.55</b>
Mg/(Mg+Fe)	0.24	0.4

**Table 2. Electron microprobe data for plagioclase**

	Diorite	Kersantite
SiO <sub>2</sub>	64.86	59.84
Al <sub>2</sub> O <sub>3</sub>	22.5	26.06
FeO	0.05	0.27
CaO	4.17	8.16
Na <sub>2</sub> O	9.01	7.36
K <sub>2</sub> O	0.13	0.13
<b>Total</b>	<b>100.72</b>	<b>101.81</b>
An	20.22	37.73
Ab	79.05	61.58
Or	0.73	0.70
<i>Name</i>	<i>Oligoclase</i>	<i>Andesine</i>



The veins are commonly ptymatically folded indicating that there may have been significant deformation of the country rock at the time of intrusion.

Near the ultramafic intrusion, within highly deformed gneiss, muscovite begins to breakdown to sillimanite+K-spar. The rock contains more quartz and granite veins than in other areas though they are commonly still very thin and locally ptymatic.

### ***Diorite***

The diorite is generally homogenous in texture and composition. It is medium gray with medium to coarse grains and primarily hypidiomorphic granular in texture with only local flow foliation. The main mineral components are plagioclase with oligoclase ( $An_{20.22}$ ) composition (Table 1) and biotite (annite) (Table 2) with  $Mg/(Mg+Fe)$  ratio of 0.28. It also contains quartz, hornblende, garnet, and accessory phases. The diorite is progressively garnet-rich toward the contact with Annsville Formation where it contains abundant xenoliths. Composition in this area approaches granodiorite. Mafic schlieren can be seen in the main exposure of the diorite body as a result of the igneous differentiation. Dikes of 2 cm to 1 m thickness intrude the diorite. They are primarily kersantite and aplite. The kersantite dikes are the extension of apophyses of an intrusion with the same composition.

### ***Hornblendite "cortlandite"***

The ultramafic rock that is exposed at Stony Point is a variety of pyroxene olivine hornblendite. It is black and massive and composed of large randomly oriented crystals of hornblende that poikilitically enclose olivine, enstatite and augite. Shand (1942) recognized the secondary (metasomatic) origin of the amphibole. In places, the amphibole displays rutile exsolutions (Fig. 2) and changes in composition from pargasite to edenite and actinolite (Table 3). We interpret this feature as a decomposition of an initial kaersutite composition at cooling. This interpretation is in good agreement with the Ratcliff (1968) observation that the amphibole in hornblendite on the border of the pluton exposed at the railroad cut is "allotriomorphic green hornblende" (actinolite). The main hornblendite body displays variations from pyroxene-rich to hornblende-rich varieties.

### ***Dikes.***

Fine to medium-grained dikes of 5 cm to 15 m thickness occur in both diorite and country rocks (Ratcliff, 1968). Compositionally, the dikes are lamprophyres (spessartite and kersantite) and less commonly aplites. The lamprophyres have a high mafic index, a small sized groundmass and mafic phenocrysts (annite, amphibole, and pyroxene). In the kersantite dikes cutting the diorite body and cropping out at the railroad cut, the plagioclase (Table 2) in the groundmass is andesine ( $An_{37.73}$ ) and annite (Table 1) has more phlogopite in solid solution with  $Mg/(Mg+Fe)$  ratio of 0.40. The leucocratic dikes are tourmaline-rich aplites.

## **GEOCHEMISTRY**

Samples of the main igneous rock types at Stony Point were analyzed for trace elements: hornblendite, diorite (in contact with Annsville Formation and the core of the diorite body), kersantite and aplite dikes. The samples were analyzed by ICP-MS at Union College, Schenectady, NY, by low-pressure HF digestion using a PerkinElmer/Sciex Elan 6100 DRC. We used the NIST-278 (obsidian) and NIST-688 (basalt) as standards. The results obtained for trace elements are presented in Table 4.

**Table 3.** Electron probe data on amphiboles from cortlandite and kersantite

	1	2	3	4	5
SiO <sub>2</sub>	41.63	45.32	50.59	56.35	41.4
TiO <sub>2</sub>	1.66	2.45	0.63	0.27	0.87
Al <sub>2</sub> O <sub>3</sub>	15.47	10.98	7.49	2.91	12.69
V <sub>2</sub> O <sub>3</sub>	0.12	0.08	0.04	0.04	0.05
Cr <sub>2</sub> O <sub>3</sub>	0.01	0.01	0.02	0.02	0
Fe <sub>2</sub> O <sub>3</sub> *	1.31	0	0	0	4.35
FeO	10.7	11.16	9.49	7.32	18.57
MgO	11.81	13.31	16.07	16.07	6.01
MnO	0.13	0.14	0.12	0.14	0.57
CaO	12.2	12.03	12.45	13.04	11.51
Na <sub>2</sub> O	2.04	1.96	1.24	0.51	1.3
K <sub>2</sub> O	1.21	1.21	0.64	0.12	1.3
Cl	0.02	0.03	0.03	0.01	0.15
H <sub>2</sub> O*	2.04	2.06	1.91	2.17	1.93
O=Cl	0	-0.01	-0.01	0	-0.03
<b>Total</b>	<b>100.36</b>	<b>100.73</b>	<b>100.71</b>	<b>101.5</b>	<b>100.66</b>
<b><u>Normalized composition</u></b>					
Si <sup>4+</sup>	6.09	6.57	7.17	7.77	6.31
<sup>IV</sup> Al <sup>3+</sup>	1.91	1.43	0.83	0.23	1.69
Σ(T)	8	8	8	8	8
<sup>VI</sup> Al <sup>3+</sup>	0.76	0.45	0.42	0.24	0.59
Fe <sup>3+</sup>	0.14	0	0	0	0.5
Fe <sup>2+</sup>	1.31	1.35	1.12	0.84	2.37
Mg <sup>2+</sup>	2.58	2.88	3.39	3.82	1.37
Mn <sup>2+</sup>	0.02	0.02	0.01	0.02	0.07
Ti <sup>4+</sup>	0.18	0.27	0.07	0.03	0.1
V <sup>3+</sup>	0.01	0.01	0.01	0	0.01
Σ(M1,M2,M3)	5	4.98	5.02	4.95	5.01
Ca <sup>2+</sup>	1.91	1.87	1.89	1.93	1.88
Na <sup>+</sup>	0.09	0.13	0.1	0.07	0.12
Σ(M4)	2	2	2	2	2
Na <sup>+</sup>	0.49	0.42	0.23	0.07	0.26
K <sup>+</sup>	0.23	0.22	0.12	0.02	0.25
Σ(A)	0.72	0.64	0.35	0.09	0.51
OH	2	1.99	1.99	2	1.96
Cl <sup>-</sup>	0	0.01	0.01	0	0.04
Mg/(Mg+Fe)	0.64	0.68	0.75	0.82	0.32
<b>Name</b>	<b>Pargasite Edenite Edenite Actinolite Ferro-pargasite</b>				

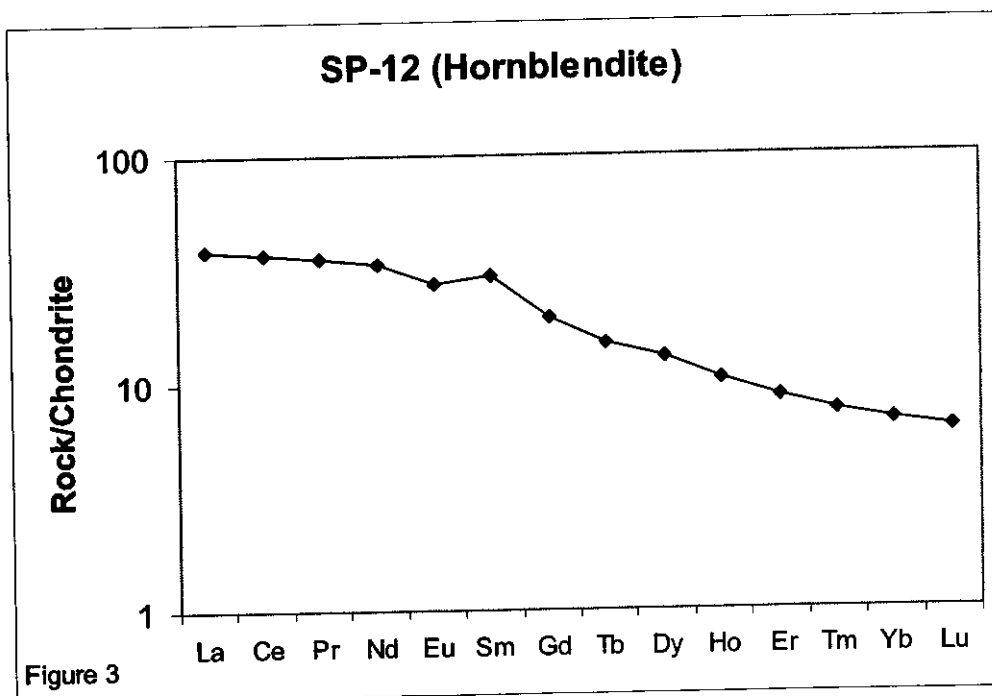
\* Fe<sub>2</sub>O<sub>3</sub> and H<sub>2</sub>O have been calculated by stoichiometry from the EMPA data.

Table 4. Trace element composition of rocks from Stony Point. K=Kersantite, D=Diorite, C=Cortlandtite, A=Aplite

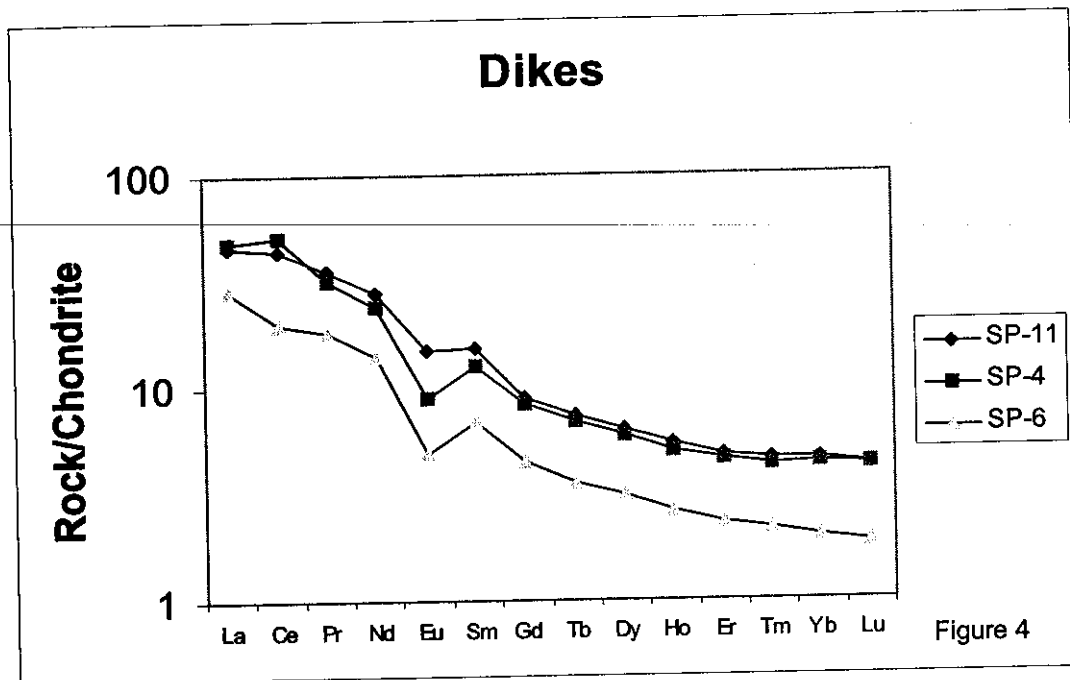
Sple	Rock	Li	V	Cr	Ni	Cu	Zn	Rb	Sr	Y	Zr	Hf	Nb	Mo	Sn	Cs	Ba
SP-11	K	22.71	162.32	3.32	25.16	23.78	97.91	55.86	374.09	6.98	128.09	3.13	27.27	0.96	1.06	1.11	657.22
SP-03	D	15.71	9.82	0.93	4.61	6.95	80.32	51.47	1047.41	4.16	18.42	0.52	19.46	0.12	0.08	0.44	1580.20
SP12	C	3.45	399.40	66.46	115.42	78.74	79.25	3.85	212.65	15.22	85.38	3.00	17.36	0.36	1.01	0.05	167.71
SP-01	D	12.22	41.12	3.42	6.64	11.47	85.17	48.89	2187.41	10.75	10.31	0.28	9.27	0.41	0.17	0.48	2940.49
SP-05	D	14.49	27.32	1.14	3.77	10.20	114.18	48.87	1346.76	5.30	23.58	0.85	16.62	0.17	0.10	0.47	2264.63
SP-04	K	12.58	61.22	5.99	5.58	10.30	72.00	34.86	489.46	6.45	165.77	4.69	42.09	0.40	0.94	0.30	1066.95
SP-06	A	0.72	67.46	0.69	2.30	15.52	24.53	34.75	253.73	1.76	79.68	2.41	21.32	0.80	0.13	0.10	690.41
SP-08	D	18.23	50.44	14.64	8.77	35.48	80.16	53.81	250.25	71.71	146.40	2.64	23.68	0.22	0.86	1.14	331.94

Sple	Rock	La	Ce	Pr	Nd	Eu	Sm	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Pb	Th	U
SP-11	K	14.43	36.03	4.32	16.91	1.11	2.97	2.34	0.35	2.05	0.39	1.01	0.15	0.96	0.14	9.45	3.15	1.90
SP-03	D	9.35	25.33	3.28	14.30	1.16	2.17	2.15	0.27	1.36	0.21	0.45	0.06	0.30	0.04	8.40	0.55	0.79
SP12	C	11.90	30.19	4.32	20.20	2.00	5.74	5.05	0.71	4.11	0.74	1.82	0.24	1.41	0.20	2.44	0.96	0.32
SP-01	D	21.82	64.68	7.36	33.27	2.67	4.76	4.42	0.57	3.03	0.53	1.27	0.17	1.04	0.16	11.04	3.69	0.66
SP-05	D	12.40	33.28	4.52	19.61	1.47	3.00	3.04	0.38	1.93	0.30	0.64	0.08	0.43	0.06	9.23	1.38	0.80
SP-04	K	14.93	41.39	3.91	14.60	0.66	2.46	2.17	0.33	1.92	0.36	0.96	0.14	0.93	0.14	10.28	4.48	1.62
SP-06	A	8.85	16.03	2.23	8.53	0.36	1.33	1.18	0.17	1.00	0.19	0.49	0.07	0.42	0.06	13.03	1.47	0.90
SP-08	D	18.24	157.47	9.35	44.87	2.82	14.34	15.07	2.99	20.15	4.00	10.45	1.49	12.98	1.42	5.94	1.68	1.87

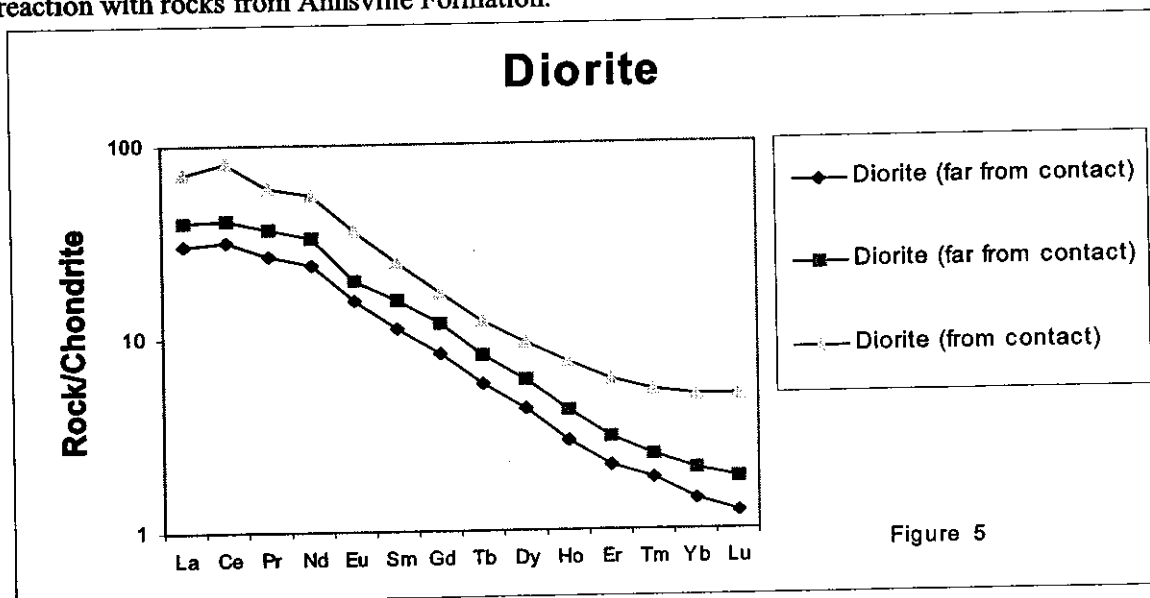
Rock/Chondrite diagram for (Fig. 3) hornblendite has a smooth slope from LREE to HREE. A very small negative Eu\* anomaly and a small positive anomaly for Sm are other features of the curve. If the Eu\* anomaly can be attributed to a low amount of plagioclase in the source region, the positive Sm anomaly could be the result of the oxidation state in the source region.



The mafic intrusion in diorite (SP-4), the kersantite from the railroad cut (SP-11) and the aplite (SP-6) are plotted on a Rock/Chondrite diagram (Fig. 4). The curves for the mafic intrusion and for the kersantite dike are very close indicating that they represent the same composition and are likely from the same magma. The differences are only in the texture and morphology. The kersantite has a deeper Eu\* negative anomaly than the mafic rock because it has less plagioclase. The aplite dike shows the same pattern as the kersantite, but at lower values, suggesting that these dikes may come from the same parental magma and simply differ by igneous differentiation.



Diorite samples plotted in Fig. 5 display a similar pattern especially for the rocks sampled far from the contact. They have an almost flat pattern for La, Ce, Pr, Nd, and an accentuated slope from Eu to Lu. The garnet that was probably left behind in the source region can explain the depletion in the HREE in the diorite. Diorite from the contact is richer in HREE because it contains garnet as a result of the reaction with rocks from Annsville Formation.



Rock/mantle diagram (Fig. 6) shows very close patterns for all samples. Some lithophile moderate volatile elements (Li, Cs) show no depletion or strong enrichment. Some chalcophile highly volatile elements (Pb) have the same behavior or show a light depletion (Sn). The refractory lithophile elements, Rb, Ba, Nb, Sr, Hf, U, Th show a strong enrichment as a result of crustal contamination. The transitional elements, Cr and Ni display a strong depletion showing that they were removed from the magma or were left behind in the source region.

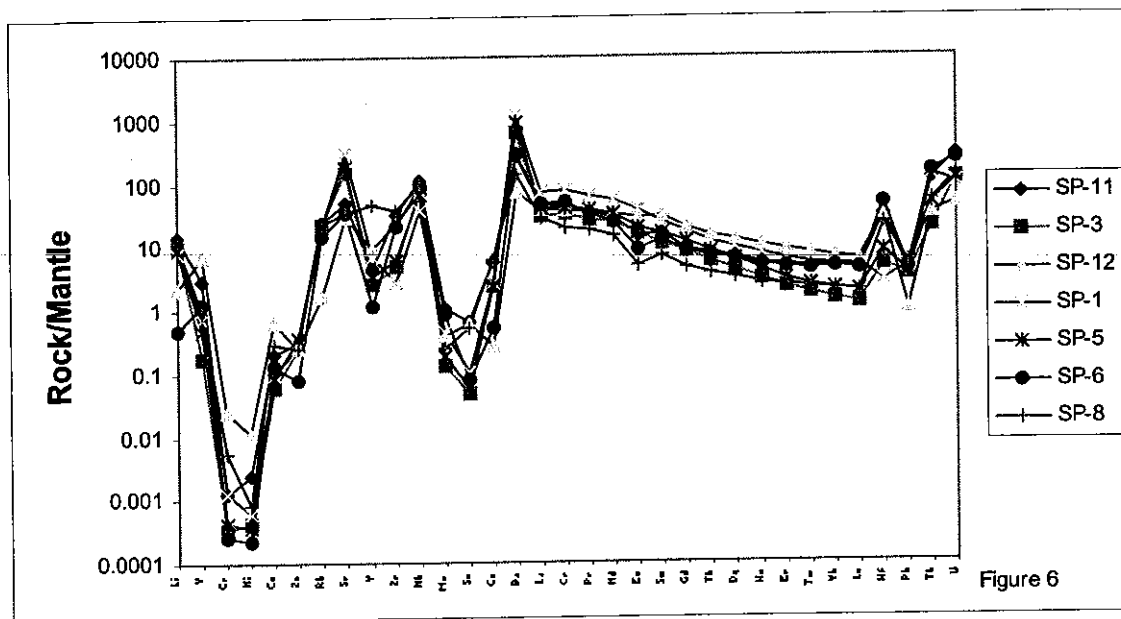


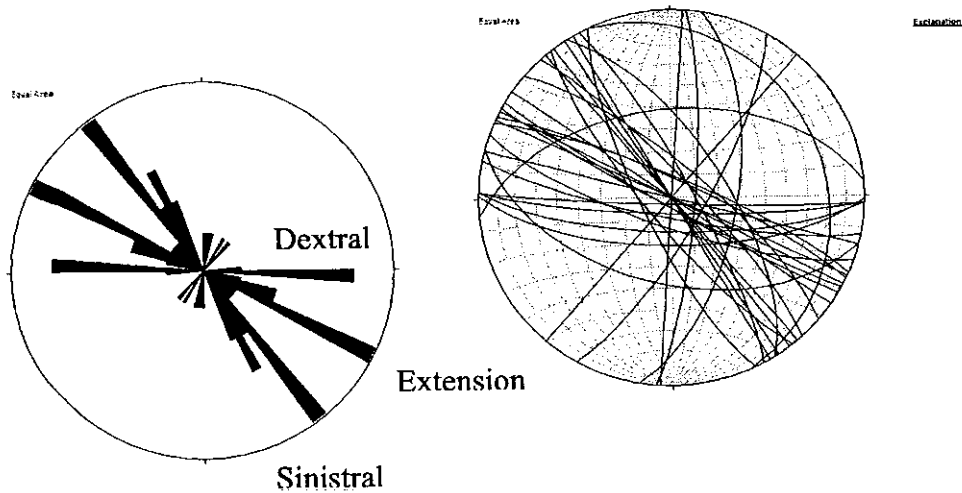
Figure 6

The very close patterns, displayed by the rocks that were analyzed, in the rock/chondrite and rock/mantle diagrams, suggest that they originated from the same parental magma and diversified by igneous differentiation.

## DIKE EMPLACEMENT

The lamprophyre and related dikes that occur at Stony Point continue in a West to WNW trend across the Hudson Highlands and into the Valley and Ridge Province to the Beemerville alkaline igneous complex in New Jersey (Fig. 1) (Maxey, 1976; Ratcliffe, 1981). The orientation and kinematics of dikes were analyzed in the Hudson Highlands from just west of Stony Point, westward to the New York State Thruway near Sloatsburg, NY. In many cases, the walls of the dikes are in pavement outcrops or are not exposed. The kinematics of these dikes could not be determined. In some dikes, however, stepped surfaces on the dike walls, rare slickensides and offset pegmatite dikes show the direction of movement. In all but a few cases the sense of movement is strike-slip or oblique with a prevailing strike-slip component.

Plotted on a rose diagram, three dominant orientations of dikes emerge: 325, 300 and 270 (Fig. 7a). The 300 group of dikes show no offset along their walls. Those dikes from the 325 group that show offset are consistently sinistral or sinistral oblique. The few dikes that show offset from the 270 group are dextral strike-slip or oblique. Plotted on a stereonet (Fig. 7b), the three groups are also distinct by their dip. The 300 group is essentially vertical whereas some of the 270 group dips steeply North and some of the 325 group dip steeply Southwest.



**Figure 7.** a) Rose diagram for Cortlandt dikes in Hudson Highlands with wall shear senses shown. b) Stereonet of dike orientations.

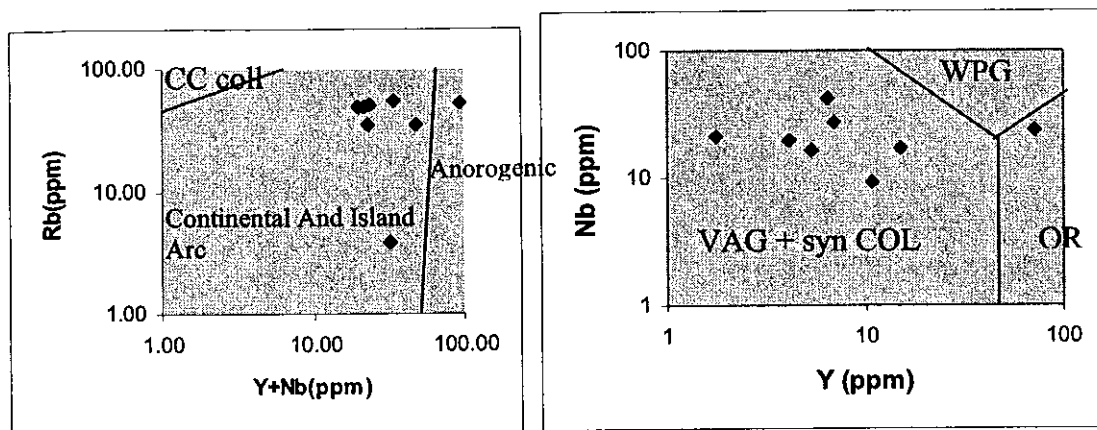
The strikes and shear senses on the walls of the three groups of dikes supports an interpretation that the dikes were emplaced into an active stress field. The dikes with the 300 strike are interpreted as extensional because they are not sheared. They would then lie parallel to the maximum compression direction. The other two sets, 270 and 325 opened as conjugate shear joints as indicated both by their orientation and the shear sense of the wall rocks. The converging dip of these two sets of dikes indicates that the plunge of the maximum compression direction was approximately  $10^\circ$  to the Northwest (300). Otherwise, all dikes would be vertical like the 300 set (parallel to  $\sigma_1$ ).

## TECTONIC IMPLICATIONS

There are more questions about the plate tectonic setting of the Cortlandt Complex than there are answers. Few researchers even attempt such an interpretation. Ratcliffe (1981) and Domenick and Basu (1982) concluded that the Cortlandt-Beemerville trend is anorogenic based upon its geometry and alkaline chemistry. This chemistry (Bender et al., 1984) and the geometry of spaced circular composite plutons in a linear trend is also suggestive of magmatism over a mantle plume (Ratcliffe, 1981). Either of these interpretations would seem reasonable but other data complicate the issues.

Cortlandt plutonism post-dates the middle Ordovician Taconian orogeny, overprinting both metamorphism and deformation. Geobarometric studies by Waldron and Tracy (1986) however, found that the pressures in the contact aureole significantly exceed (approx. 2 kbar) those of the regional Taconian metamorphism. If this was an extensional environment, pressures would be expected to have significantly decreased after the Taconian tectonism rather than increase. Considering the fine grain size of the dikes and their mode of emplacement into crust undergoing brittle deformation, it is clear that they must have been emplaced at significantly shallower levels than the earlier main plutons. Therefore, there must have been uplift and erosion at that time possibly as the result of later extension. However, the stress field at the time of emplacement was inconsistent with rifting where  $\sigma_1$  would be vertical. Instead, it was near horizontal and consistent with the regional Taconian stress field.

Tectonic discrimination diagrams of Pearce et al. (1984) and based on immobile trace elements in granitoid rocks, are inconsistent for the Cortlandt complex. They primarily reflect a compressional environment like a magmatic arc with few analyses reflecting an anorogenic environment (Fig. 8). These results are also inconsistent with rifting.



**Figure 8.** Tectonic discrimination diagrams based upon immobile trace elements of Pearce et al. (1984). Diamonds show Cortland analyses. CC coll = continent-continent collision, VAG = volcanic arc granite, syn COL = syncollisional granite, WPG = within plate granite.

The orientation of the Cortlandt-Beemerville trend is basically perpendicular to the Taconian orogen rather than parallel to it as would be expected in a successor rift system like in the Basin and Range Province. Clearly, compression continued to thicken the crust subsequent to Taconian metamorphism as shown in the increase in pressure in the contact aureole. The stress field into which the dikes were emplaced was clearly not reflective of rifting though significant uplift and unroofing must have been occurring at the time. Based upon these data, we favor an impactogen (Sengor et al., 1978) origin for Cortlandt-Beemerville. An impactogen is an early stage failed rift like an aulocogen except that it is produced by the impact of one continent into another and lies perpendicular to the trend of the orogen. Extensional systems perpendicular to the orogen are well documented in the Himalayas (Tapponnier et al., 1982) among others where such rifts can be large as in the case of Lake Baikal, Siberia. The Cortlandt-Beemerville system was in the early stages. Whether the impactogen was generated by the same collision as that which produced Taconian tectonism in the area or the impact of a second island arc like that described in Massachusetts (Karabinos et al., 1998) cannot be determined without further research.

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## FIELD GUIDE

### Stop 1. Diorite

Medium-coarse grained diorite from the interior of the pluton contains plagioclase, biotite, quartz, amphibole and minor garnet and other accessories. The texture is hypidiomorphic granular with no flow foliation. Thin veins and dikes of lamprophyre intrude the diorite.

### Stop 2. Diorite – Annsville Contact

At the contact of diorite and gneiss of the Annsville Formation each unit is affected by the other. Annsville Formation is a biotite and biotite-garnet schist/gneiss. It contains pygmatically folded granite veins of 0.5 to 2 cm thickness and locally containing tourmaline. The schist/gneiss is highly deformed with multiple crenulation cleavages. It contains muscovite, biotite, quartz, feldspar and accessories. The diorite contains numerous xenoliths of Annsville Formation in various stages of assimilation. The density of garnets is much higher at the contact where it forms clots with grains up to 1.5 cm. The composition of the diorite approaches granodiorite. Contamination of the magma with country rock has clearly changed the composition.

### Stop 3. Dike in Annsville Formation

Fine-grained lamprophyre dike intruding Annsville Formation schist/gneiss. The schist/gneiss is highly deformed with multiple crenulation cleavages. It contains muscovite, biotite, quartz, feldspar and accessory minerals. Locally, chill margins may be observed. The dikes have been drilled for paleomagnetic studies.

### Stop 4. Cortlandite

Massive black pluton of Cortlandite is in fault contact with Annsville Formation. It is composed of large randomly oriented crystals of hornblende that poikilitically enclose olivine, enstatite and augite.



# Migmatites Formed by Fusion of Metasediments Intruded by Cortlandt Complex, Stony Point New York

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## ABSTRACT

Our petrographic examinations of several country rock fusion products generated along the margins of Jurassic diabase intrusions indicates a restricted range of melt types despite the wide range of pelitic sedimentary country rocks that were intruded. Partial fusion of these pelitic rocks has resulted in three types of melt: 1) trondhjemite that plots near the eutectic point of the Quartz-Albite eutectic, 2) granite that plots near the ternary minimum of the Quartz-Albite-K-Spar phase diagram, and 3) syenite that plots near the ternary minimum of the Quartz-Nepheline Kalsilite phase diagram. Partial fusion of Ordovician Annsville metapelite intruded by diorite at Stony Point, New York extends our data-base to a new set of conditions. The leucosome of Annsville migmatite sampled along the margins of Cortlandt Complex diorite is granite unlike the granite leucosome in Triassic Lockatong migmatites generated at the base of the Palisades Sill. In both cases partial fusion has resulted in retention of CaO, Na<sub>2</sub>O, FeO and Sr in the refractory residue or melansome portion of the migmatites and in both cases the leucosome portions display most of the characteristics of S-type granite. However, in the case of the Annsville granite leucosome K<sub>2</sub>O, Rb, Ba and HREEs are highly partitioned into the melt phase with strong retention of Zr in the refractory residual phase. Just the reverse is true of the Lockatong leucosome. These differing apparent partitioning behaviors are probably due to hydrothermal introduction of potassic phases into the Annsville leucosome. The granite leucosome generated at Stony Point does not plot near the ternary minimum of the Quartz-Albite-Orthoclase phase diagram but instead plots in a potassium enriched area (8.55 % K<sub>2</sub>O) at temperatures that are probably not realistic.

## INTRODUCTION

This study is focused on the Annsville metasediments in contact with the Cortlandt Complex on the west side of the Hudson River at Stony Point NY (see Fig. 1) Rockland County. For other investigations of the Complete Cortlandt complex the reader is referred to Williams (1888) Shand (1942), Streeland and Woollard 1952, Barker(1964), Friedman (1956) Dolgoff (1958) Bender et al. (1982,1984) and Tracy et al. (1987) and Gates and Lupulescu (this volume). The Cortlandt Complex consists (Fisher et al 1970) of the following complex mafic intrusives: biotite-augite-norite, hornblende and or biotite-diorite, poikilitic hornblende-norite, hornblendite, olivine-pyroxenite, pyroxenite, and complex gabbroic to noritic rocks.

Ratcliffe (1968) determined the geometrical and temporal relationships among the igneous rocks of the Stony Point Cortlandt Complex. Here the igneous rocks intruded the Annsville Phyllite and crosscut F<sub>1</sub> and F<sub>2</sub> isoclinal folds. His spatial relationships among these rocks and the host are shown both in the plan view (Fig. 1) and the cross sections Fig. 2.

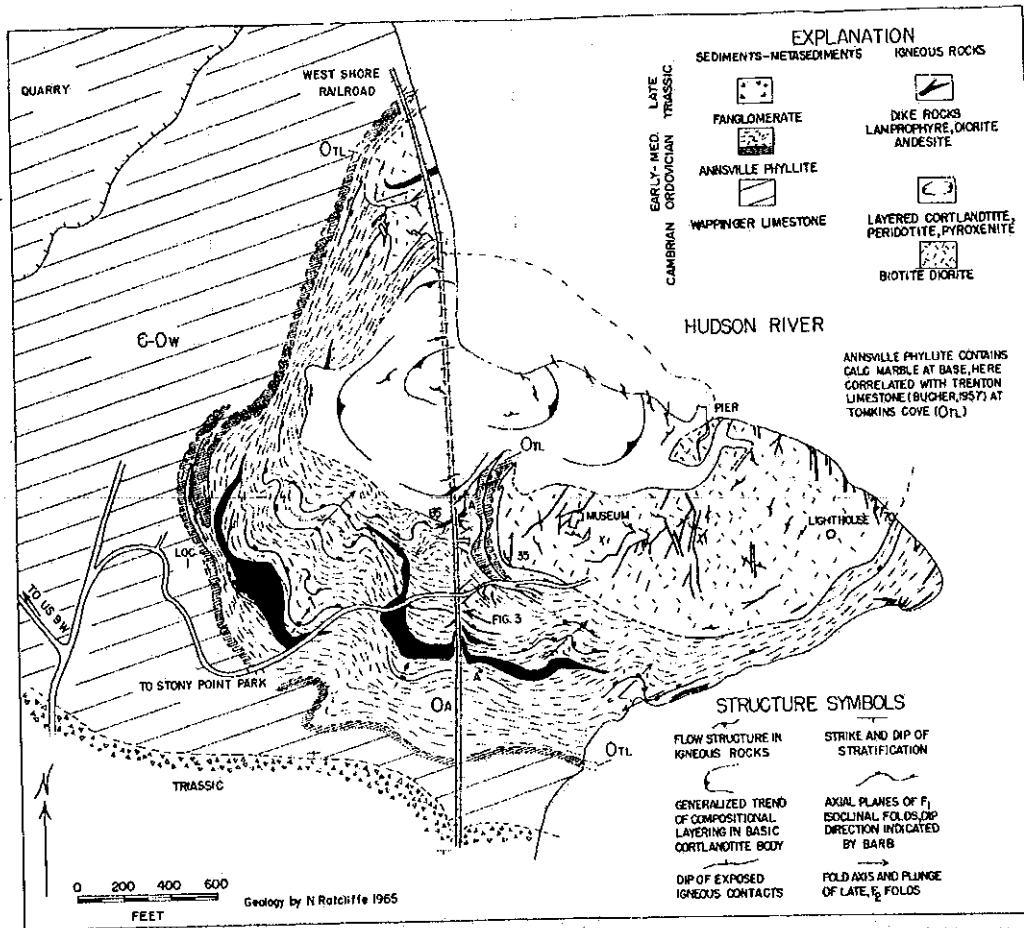


Figure 1: Map of Cortland Complex at Stony Point from Ratcliffe (1968a).

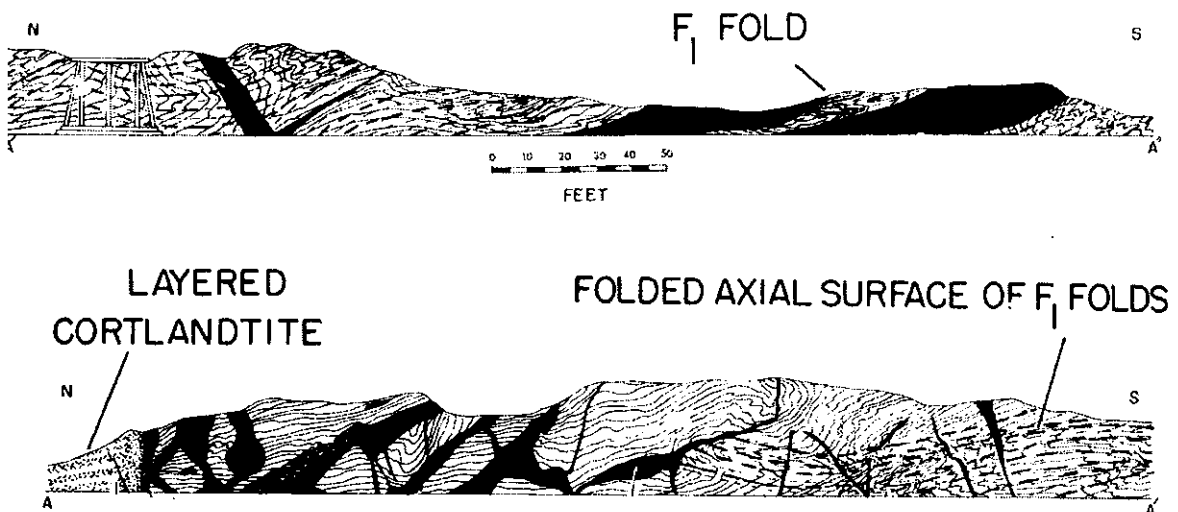


Figure 2: Cross Section along railroad cut, from Ratcliffe, 1968a,b)

## HOST ROCKS AT STONY POINT

The host rocks of the Stony Point mafic intrusions are the Ordovician Annsville Formation (metapellites) and the Wappinger Limestone (Ratcliffe 1968, Fisher et al. 1970).

### *DIORITE*

Diorite from the interior of the pluton at Stony Point (Fig. 1) is medium grained and contains plagioclase, biotite, quartz, amphibole, garnet and accessory minerals. The texture is hypidiomorphic and does not exhibit any flow foliation.

### *LAMPROPHYRE*

Fine-grained lamprophyre, occurs as dikes cross cutting the diorite. The most common lamprophyres are Kersantites composed largely of plagioclase, biotite, quartz, K-spar, muscovite, and accessory minerals.

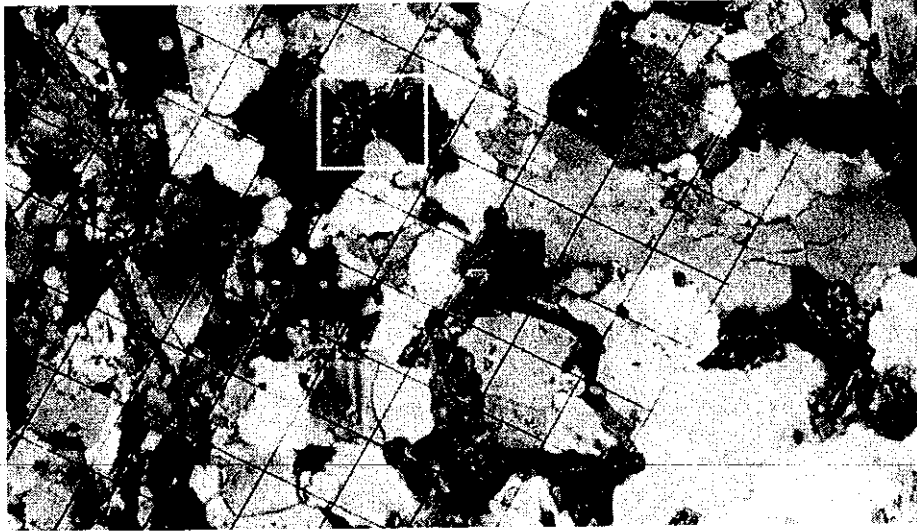
### *CORTLANDITE*

A massive dark pluton of Cortandite is in fault contact with the Annsville phyllite/manhattan schist. It is composed of large randomly oriented crystals of hornblende that poikolitically enclose olivine, enstatite, augite and plagioclase.

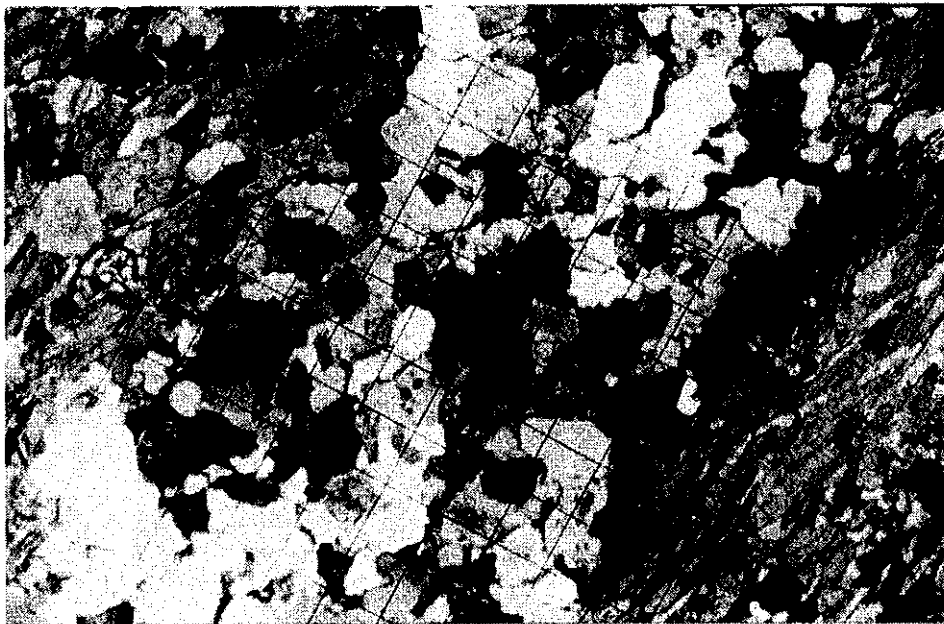
## ANATEXIS AT STONY POINT

Migmatites are nicely exposed at several locations along a railroad cut and at several rock outcrops at Stony Point. The migmatites sampled and chemically analyzed for this study (Table 1) were collected adjacent to kersantite dikes along the railroad cut, however migmatites were also observed along a hill-slope near the Museum at Stony Point adjacent to a large diorite body. In both cases these migmatites exhibit leucosomes and melansomes and are found in the host Annsville Phyllite adjacent to Silurian igneous intrusions that are part of the Cortland Complex. Ratcliffe (1968) notes that the occurrence of 0.5 to 1.0 cm poikiloblastic muscovite and biotite within 100 ft of the igneous contacts. Gardien et al. (1995) showed experimentally that the melting of biotite-plagioclase-quartz-muscovite is initiated at 750° C and biotite-quartz-plagioclase assemblages is initiated at 800°C. Their model one-mica biotite bearing protolith yielded melt fractions of <15 volume% melt and a two mica protolith yielded melt fractions approximately 50 volume % melt. They concluded that the starting mineralogy is an important factor in controlling the amount of melt generated. The two mica Annsville protolith at Stony Point, therefore, greatly increases the probability of considerable fusion at temperatures that may have been reached close to the contacts of Cortland intrusions.

Melt segregation in migmatites was studied by Brown et al. (1995). They note that although the mechanisms and rates of melt segregation and transfer during crustal anatexis remain poorly understood many granitoid rocks show mineralogical, chemical and isotopic evidence of an anatectic origin.



**Figure 3.** Mymekite (in white square) in granite leucosome portion of Annsville migmatite sampled at Stony Point indicating fusion. Grid lines are spaced at 1.0 mm.



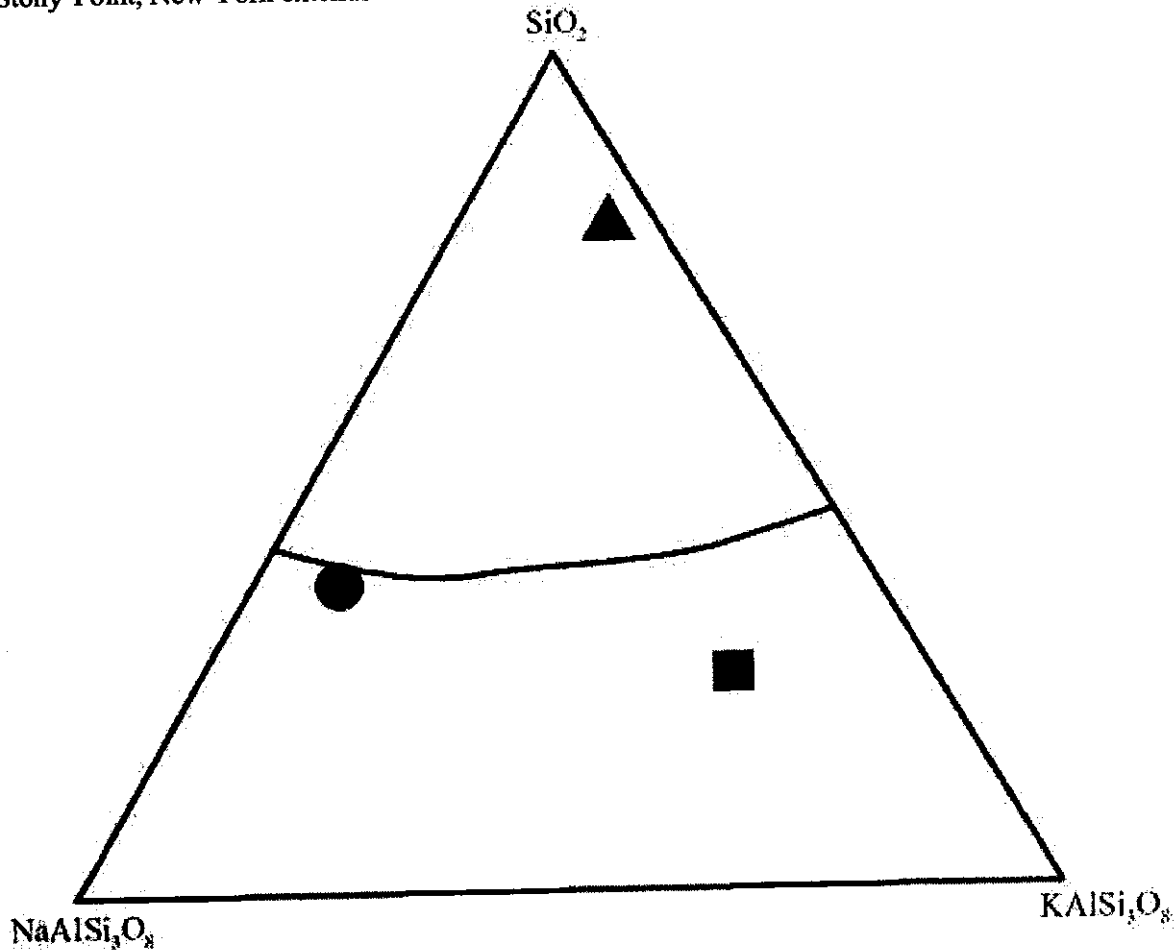
**Figure 4.** Granite leucosome composed largely of quartz and potassium feldspar surrounded by melanosome enriched in biotite and muscovite grains aligned in a north-east orientation. Grain size is about 1 mm on the basis of 1.0 mm grid lines.

Our petrographic examination of these migmatites reveals myrmekites. The presence of these myrmekites in the leucosomes of this migmatite is conclusive evidence that fusion took place. Myrmekite is an intergrowth of vermicular quartz in plagioclase, which forms a few volume percent of many igneous and metamorphic rocks (Barker, 1970).

## GEOCHEMICAL INTERPRETATION

Petrographic examinations of several country rock fusion products generated along the margins of Jurassic diabase intrusions indicates a restricted range of melt types despite the wide range of pelitic sedimentary country rocks that were intruded (Puffer and Benimoff, 1997; Benimoff and Puffer, 2000; 2001; and in press). Partial fusion of these pelitic rocks has resulted in three types of melt: 1) trondhjemite that plots near the eutectic point of the Quartz-Albite eutectic; 2) granite that plots near the ternary minimum of the Quartz-Albite-K-Spar phase diagram, and 3) syenite that plots near the ternary minimum of the Quartz-Nepheline Kalsilite phase diagram.

Partial fusion of Ordovician Annsville metapelite intruded by Silurian diorites and kersantite at Stony Point, New York extends our data-base to a new set of conditions. The leucosome of Annsville



**Figure 5:** Stony point migmatite and quartzite plotted on the 500 bar system  $\text{NaAlSi}_3\text{O}_8 - \text{KAlSi}_3\text{O}_8 - \text{SiO}_2$  (Luth et al. 1964); triangle is SP02 (quartzite); circle is SP06 (melanosome); closed square is SP05 (leucosome) of Annsville migmatite.

migmatite sampled along the margins of Cortlandt Complex diorite is granite unlike the granite leucosome in Triassic Locketong migmatites generated at the base of the Palisades Sill. In both cases partial fusion has resulted in retention of CaO, Na<sub>2</sub>O, FeO and Sr in the refractory residue or melansome portion of the migmatites and in both cases the leucosome portions display most of the characteristics of S-type granite. However, in the case of the Annsville granite leucosome K<sub>2</sub>O, Rb, Ba and HREEs are highly partitioned into the melt phase with strong retention of Zr in the refractory residual phase (Table 1). Just the reverse is true of the Locketong leucosome (Benimoff and Puffer, in press). These differing apparent partitioning behaviors are probably due to hydrothermal introduction of potassic phases into the Annsville leucosome.

The granite leucosome generated at Stony Point does not plot near the ternary minimum of the Quartz-Albite-Orthoclase phase diagram (Fig. 5) but instead plots in a potassium enriched area (8.55 % K<sub>2</sub>O) at temperatures that are probably not realistic. The Annsville leucosome, therefore, is probably not the direct result of in-situ partial fusion but is instead interpreted as the result of partial fusion combined with subsequent hydrothermal introduction of K<sub>2</sub>O, Rb, Ba, and HREEs.

The data presented in table 1 includes new analyses of kersantite lamprophyre dike rock which was the heat source of partial fusion. Some of this data is plotted onto Spider diagrams to illustrate the contrasting slopes of the various lithologies (Figs. 6 – 11).

**TABLE 1 CHEMICAL ANALYSES OF STONY POINT INTRUSIVES (this study)**

	SP01	SP02	SP03	SP04	SP05	SP06
<i>Major Oxides, weight % oxides</i>						
SiO <sub>2</sub>	48.82	87.78	51.38	48.02	72.17	70.31
Al <sub>2</sub> O <sub>3</sub>	18.8	4.07	20.43	17.72	14.77	15.93
CaO	7.01	0.16	4.24	7.49	0.89	3.18
MgO	3.88	0.21	1.76	5.63	0.35	1.04
Na <sub>2</sub> O	4.64	0.48	5.9	3.97	2.16	5.02
K <sub>2</sub> O	2.11	2.31	2.72	1.84	8.55	1.17
Fe <sub>2</sub> O <sub>3</sub>	10.81	0.66	9.38	10.8	0.64	2.37
MnO	0.13	0.01	0.07	0.16	0.01	0.04
TiO <sub>2</sub>	1.9	0.11	1.2	1.83	0.08	0.35
P <sub>2</sub> O <sub>5</sub>	0.71	0.06	0.78	0.52	0.05	0.22
Cr <sub>2</sub> O <sub>3</sub>	0.01	0.01	0.01	0.01	0.01	0.01
LOI	0.35	0.45	0.55	0.69	0.3	0.4
Sum	99.42	96.29	98.96	98.93	100.1	100.1
<i>Minor Elements, ppm</i>						
Rb	44	...	56	38	104	38
Sr	880	...	1730	825	257	389
Y	27	...	11	25	40	13
Zr	271	...	739	190	30	53
Nb	33	...	31	27	5	11
Ba	870	...	1970	1060	707	220
<i>REE, ppm</i>						
Ce	106	8	83.3	94.2	7.7	11.7
Dy	5.61	0.88	2.14	5.06	3.57	2.32
Er	3.1	0.35	0.75	2.99	3.97	1.54
Eu	2.72	0.99	3.71	2.84	0.81	0.79
Gd	8.3	1.02	5.18	7.46	1.55	1.85
Ho	1.2	0.16	0.33	1.1	1.14	0.57
La	51.8	4	39.1	44.1	4.8	5.6
Lu	0.44	0.06	0.1	0.39	0.99	0.28
Nd	48.7	4	43	45	3.6	6.1
Pr	14.2	1.16	11.9	13.4	1.09	1.7
Sm	9.5	1.1	7.9	9.5	1	2
Tb	1.26	0.19	0.68	1.17	0.44	0.41
Th	2.4	0.9	0.7	2.6	1.7	1.3
Tm	0.43	0.05	0.09	0.43	0.78	0.26
U	1.05	1.15	0.51	0.92	1.24	0.74
Y	30.4	4.4	9	28.8	34.8	14.6
Yb	2.8	0.3	0.6	2.7	5.7	1.8

SP01 – kersantite

SP02 – metasedimentary quartzite

SP03 – kersantite

SP04 – kersantite

SP05 – migmatite – granite leucosome

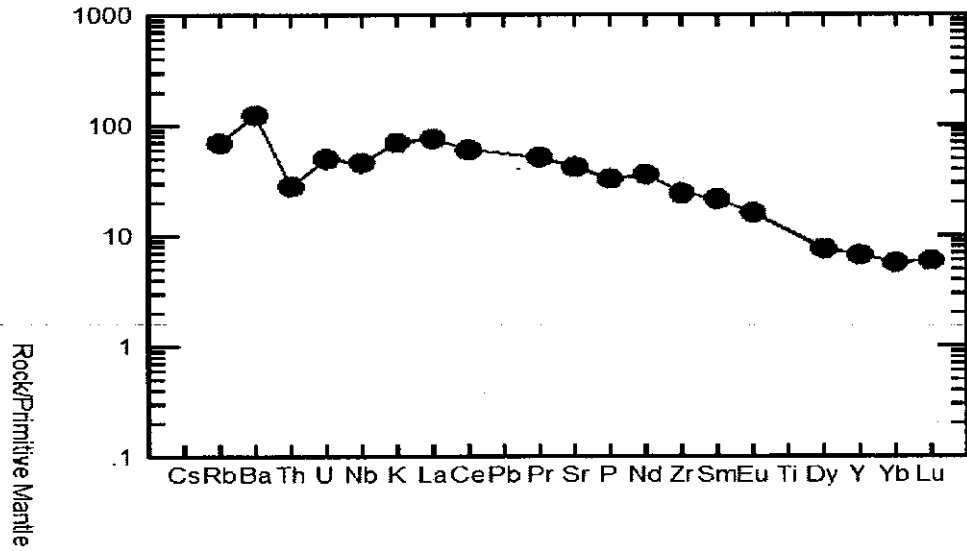
SP06 – migmatite – melanosome

<b>TABLE 2 CIPW norms Of Stony Point Intrusives (this study)</b>						
	SP01	SP02	SP03	SP04	SP05	SP06
Q	0.00	75.66	0.00	0.00	24.59	28.89
or	12.47	13.65	16.07	10.87	50.53	6.91
ab	32.70	4.06	43.09	30.27	18.28	42.48
an	24.24	0.40	15.94	25.10	4.09	14.34
lc	0.00	0.00	0.00	0.00	0.00	0.00
ne	3.56	0.00	3.70	1.80	0.00	0.00
kal	0.00	0.00	0.00	0.00	0.00	0.00
C	0.00	0.63	1.94	0.00	0.46	1.15
di	4.82	0.00	0.00	7.01	0.00	0.00
hy	0.00	0.52	0.00	0.00	0.87	2.59
wo	0.00	0.00	0.00	0.00	0.00	0.00
ol	10.10	0.00	8.44	12.67	0.00	0.00
ac	0.00	0.00	0.00	0.00	0.00	0.00
mt	4.93	0.00	3.91	4.83	0.00	0.62
il	3.61	0.02	2.28	3.48	0.02	0.66
hem	0.00	0.66	0.00	0.00	0.64	1.42
ti	0.00	0.00	0.00	0.00	0.00	0.00
ap	1.64	0.14	1.81	1.20	0.12	0.51
cc	0.00	0.00	0.00	0.00	0.00	0.00
pero	0.00	0.00	0.00	0.00	0.00	0.00
wus	0.00	0.00	0.00	0.00	0.00	0.00
ru	0.00	0.10	0.00	0.00	0.07	0.00

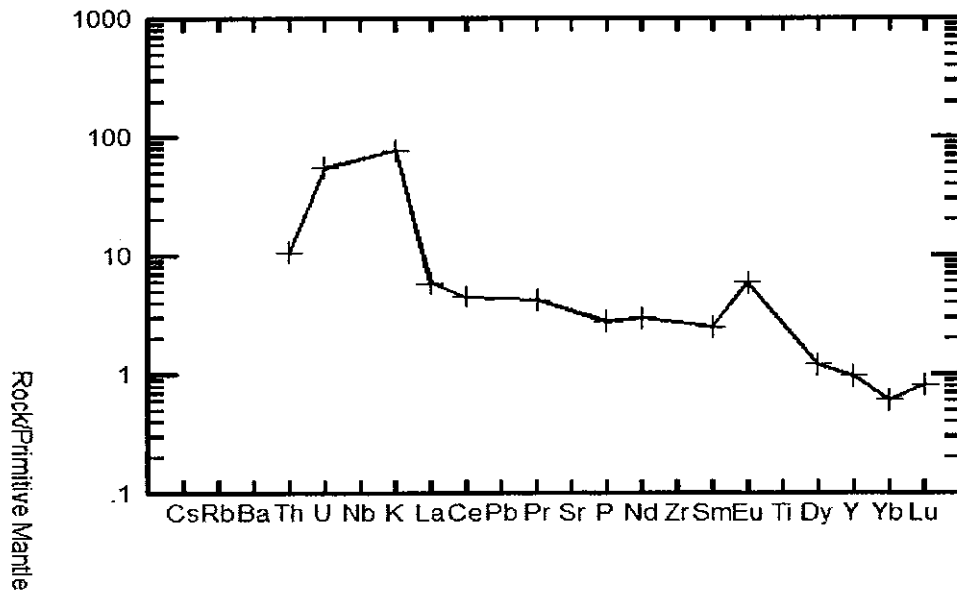
SP01 – kersantite  
 SP02 – metasedimentary quartzite  
 SP03 – kersantite  
 SP04 – kersantite  
 SP05 – migmatite – granite leucosome  
 SP06 – migmatite – melanosome



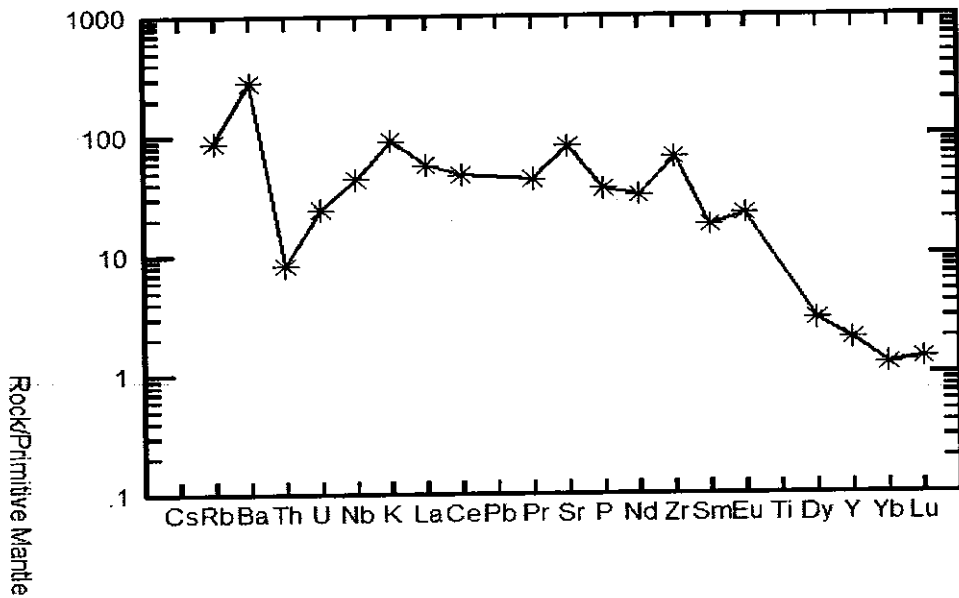
## SPIDER DIAGRAMS OF STONY POINT INTRUSIVE ROCKS



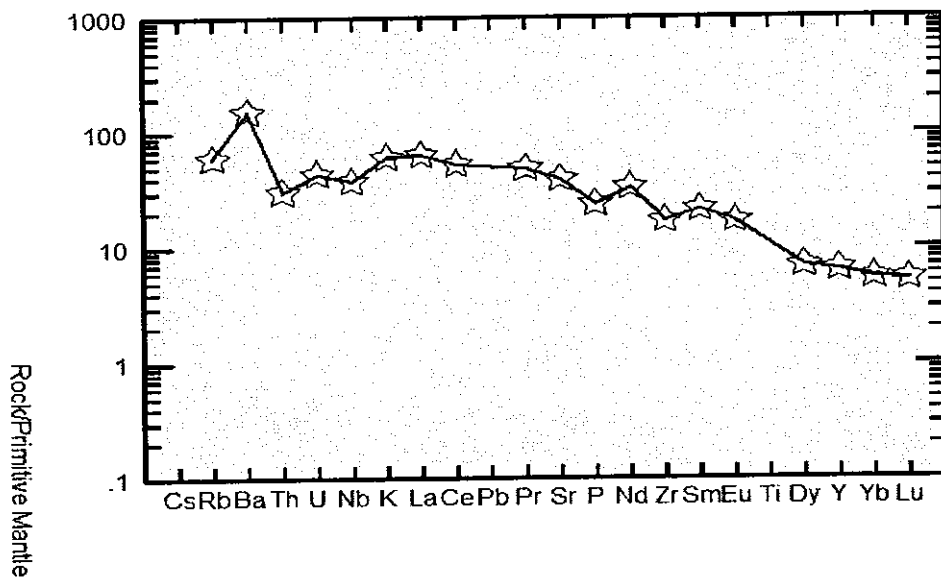
**Figure 6:** Kersantite sample SP01 plotted on a mantle normalized Spider Diagram (Sun and McDonough, 1989)



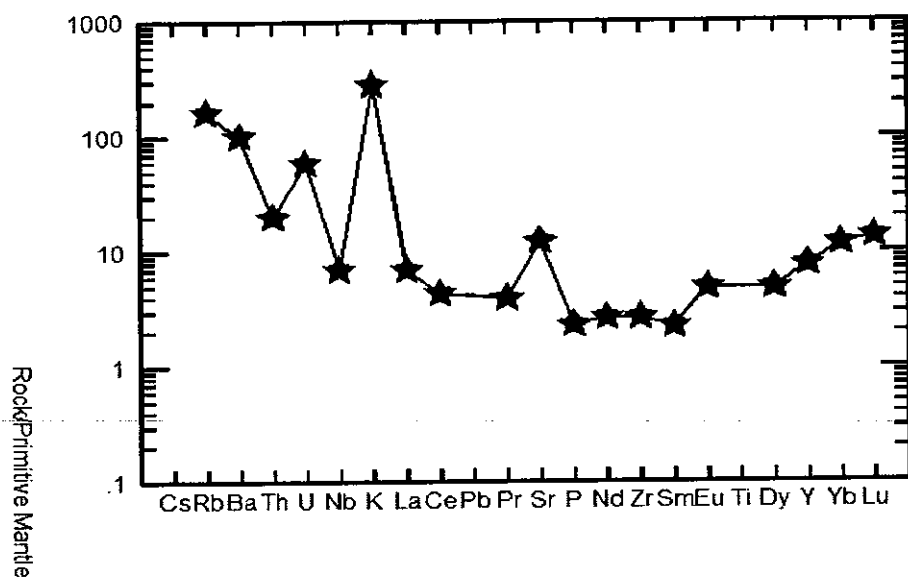
**Figure 7:** Metasedimentary quartzite sample SP02 plotted on a mantle normalized Spider Diagram (Sun and McDonough, 1989).



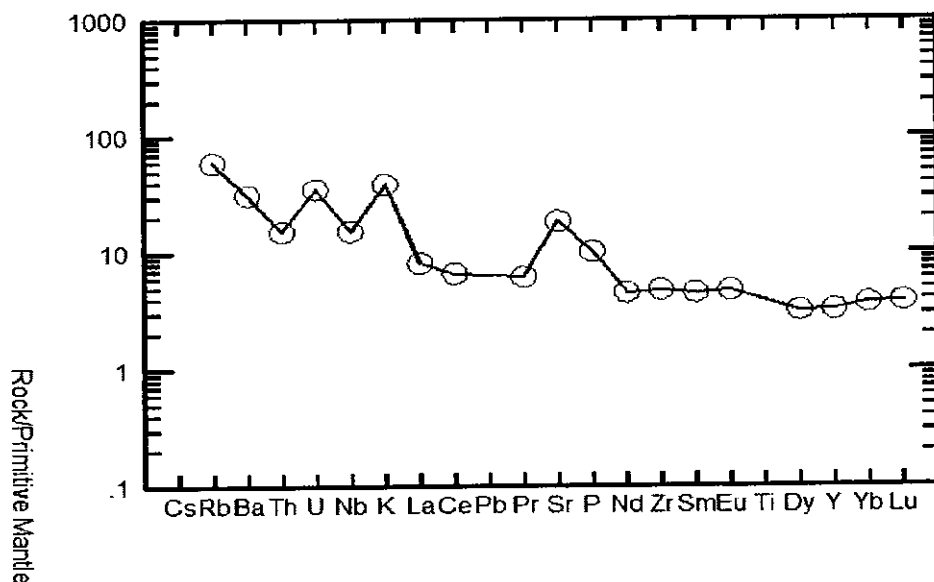
**Figure 8.** Kersantite sample SP03 plotted on a mantle normalized Spider Diagram(Sun and McDonough, 1989)



**Figure 9:** Kersantite sample SP04 plotted on a mantle normalized Spider Diagram (Sun and McDonough, 1989)



**Figure 10.** Granite leucosome portion of migmatite (sample SP05) plotted on a mantle normalized Spider Diagram (Sun and McDonough, 1989)



**Figure 11:** Refractory melanosome portion of migmatite (sample SP06) plotted on a mantle normalized Spider Diagram (Sun and McDonough, 1989)

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# Toward a Comprehensive Model for the Palisades: Tracing New Internal Contacts

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## INTRODUCTION

The Palisades is emplaced as an intrusive sheet and therefore the evolution of thermal boundary layers can be vastly different from those described for surface flows. Zones of rhythmic layering, for example, have recently been noted for the Holyoke basalt, and cited as evidence for the presence of foundered blocks in the Palisades by Dickson and Philpotts (2001). However, the exact location of these rhythmically layered horizons relative to other well-established horizons is not well documented. The present field trip examines rhythmically banded layers prominently displayed in the Alpine section (stop one), suggesting that features within this horizon are consistent with post-intrusion convective cooling, flow banding or similar structural events. This interpretation is supported by over 200 bulk geochemical analyses and complementary electron microprobe analyses (Block, PhD thesis, in preparation).

The famous olivine layer of the Palisades has been the center of attention for almost a century. In the present study we consider the implication of finding quench features in a mafic layer exposed in Piermont, New York. The main olivine layers reside near the base of the Sheet, ascribed variously to (i) differentiation of a single pulse of magma (Walker, 1940, Latypov, 2003), (ii) multiple pulse model (Walker, 1969, Husch, 1990, and Shirley, 1987), and (iii) flow differentiation or cumulus-transport-deposition (Battacharji, 1969, Steiner et al., 1992 and Gorrington and Naslund, 1995). These models generally suppose that the olivine concentrations are situated at nearly the same horizon(s) along a stratigraphic level that roughly parallels the base of the Sill. The present study collates the existing data and suggests that the olivine horizons may actually plunge to the north, and are associated north of Englewood Cliffs, New Jersey, with a new mafic zone, the hypersthene-phyric layer (Steiner et al., 1992).

The second stop (at Piermont) marks the locality where the hypersthene-phyric layer can be observed in quench contact with the base of the Sheet. This contact is close to a second, less magnesium-rich, holocrystalline quench lithology. Piermont facies are shown to be compositionally and petrographically dissimilar to other contact lithologies.

Time permitting, a third stop will take place along Route 9W to document a second type of quench feature, the glomeroporphyritic quench horizon of Steiner et al. (1992) that is now identified with the "boa" texture proposed by Philpotts and Dickson (2002).

Key points for this field trip:

- *The rhythmically layered section may rest on a platform provided by the transport-emplacement of the mafic layer, and therefore may represent a flow feature perhaps related to convection in the stratigraphic middle of the intrusion (Stop 1).*
- *A mafic layer defined as Cr- and Mg-enriched ( $Mg\# > 0.6$ ,  $Cr > 500$  ppm) maps at various levels in studied sections and tends to pinch out to the north (Stop 2).*
- *Boa texture appears to be consistent with an alternative differential fusion (decompression melting) hypothesis. (Stop 3)*

## RHYTHMIC BANDING: AN OVERVIEW

Rhythmic banding at the cm-scale is a prominent feature of many basic intrusive rocks (e.g. Bushveld Complex (Rice, 1995); Stillwater laccolith of Montana, (see Morse, 1994 for a review) though the formative mechanisms are not well understood. Theories vary: Jackson's (1970) Stillwater model

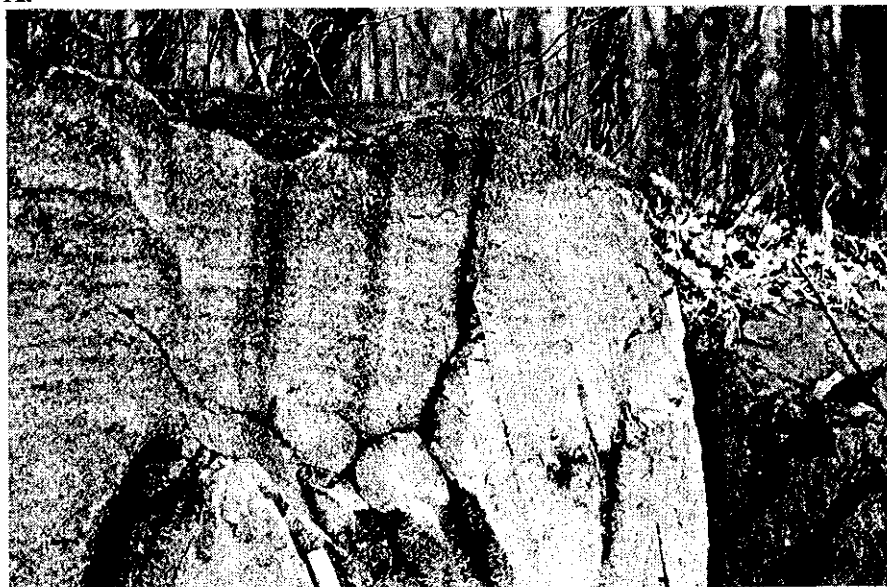
requires systematic episodes of magma replenishment and theorizes that a system of convective overturn operates to reverse the crystallization order. Here, the main magma mixes with each batch and then fractionates in such a way that the primary phase resumes crystallization after the secondary phase has deposited in a cumulus layer above the original horizon. Irvine (1970) presents an argument against a purely kinetic model for rhythmic layering by suggesting a thermodynamic explanation for the Stillwater cyclic units. This model proposes a cycling deposition controlled by the cotectic-peritectic relationship. Similar mechanisms are proposed by Dickson and Philpotts (2001) and Philpotts and Dickson (2002) for the formation of cm-scale rhythmic banding in the Palisades and mm-scale banding in the Holyoke basalts. Rice (1981) proposed rhythmic banding to be similar to solute banding or microsegregation seen in the solidification of industrial melts. In this case, oscillating temperatures due to turbulent convection caused cotectic shifting at the solidification front.

Recent studies of the chromite bands and facies changes in the Bushveld are consistent with models for flow differentiation and internal convection where a chemically and mineralogically distinct, syn-emplaced mafic layer evolves during initial stages of stratification (Rice, 1999; and Rice et al. 1999). Middle and upper layers are significantly more evolved and contain considerable amounts of assimilated country rock. Raia and Spera's (1997) numerical model for intrusion and subsequent anatexis of the lower crust reaches similar conclusions.

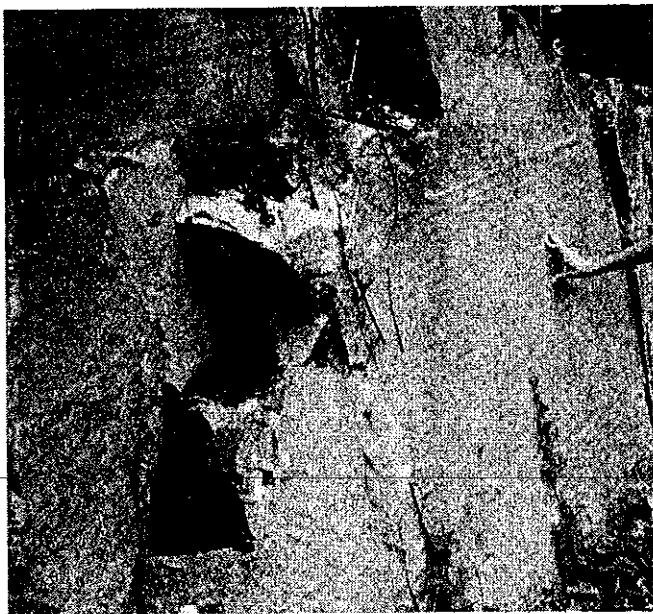
### Stop 1: Centimeter-Scale Banding at Alpine

The stratigraphic central portion of the Palisades at Alpine contains an extensive section extending from about 60 meters above the basal contact to approximately 150 meters that comprises a chemically and mineralogically distinctive unit (Block and Steiner, in preparation). Portions of this unit are characterized by cm-scale rhythmic banding (Fig. 1). Thin sections reveal that associations of plagioclase and clinopyroxene tend to be separated by finer-scale plagioclase chains (Fig. 2). The banding may thus represent a type of 'foundered block' as proposed by Dickson and Philpotts (2001) for presumably similar layered horizons, but ongoing consideration of mineral chemistry and texture is also consistent with an evolution via a convective flow model (Block, Rice and Steiner, in preparation).

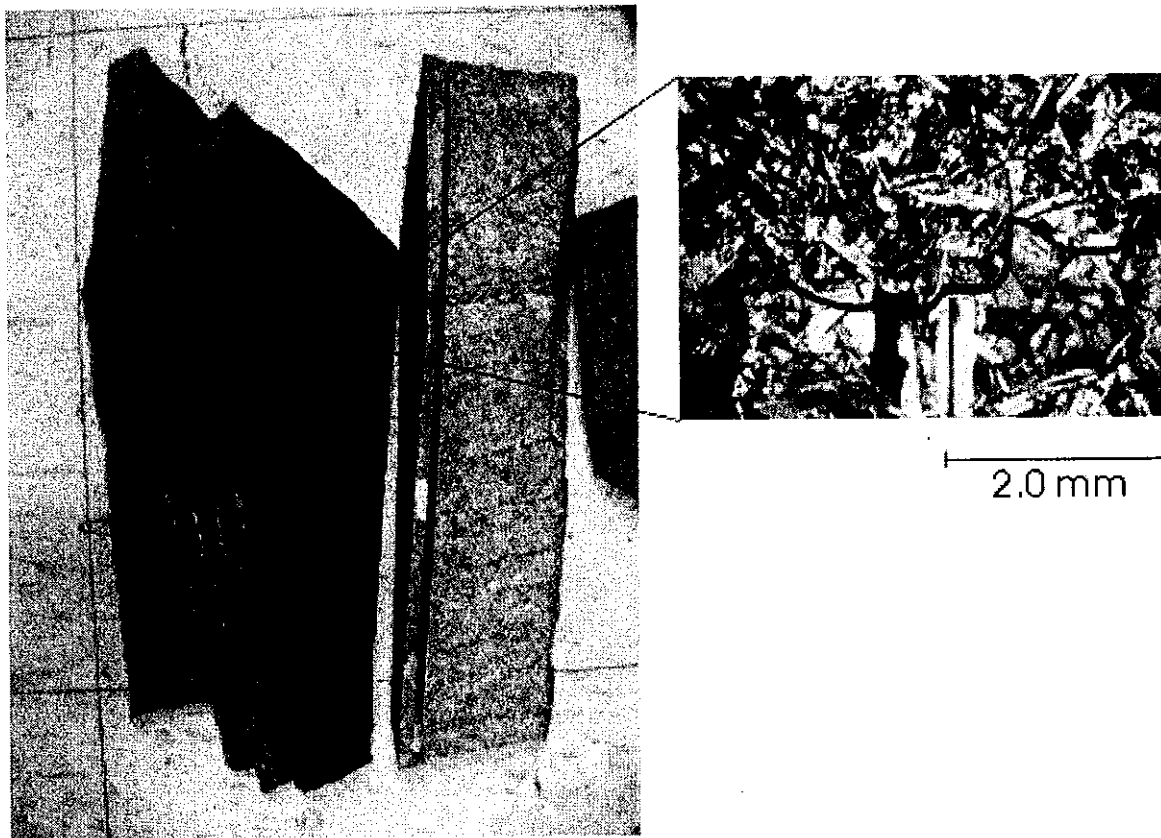
A.



**B.**



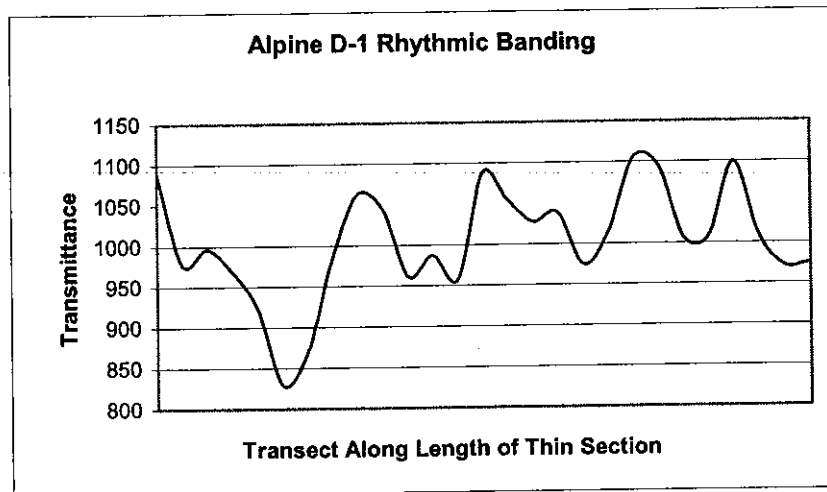
**Figure 1.** Centimeter-Scale Rhythmic Banding in Alpine Section of the Palisades Sill. Figure 1A is approximately 150m from lower contact; Figure 1B is 100 m from lower contact.



**Figure 2** Alpine D rhythmically banded specimen. Cut face has been moistened to clearly show alternating layers of light and dark minerals. Right: Polarized light micrograph of showing a finer-scale plagioclase horizon adjoining a pyroxene-plagioclase matrix; lighter plagioclase horizons can be traced from slab to thin section.



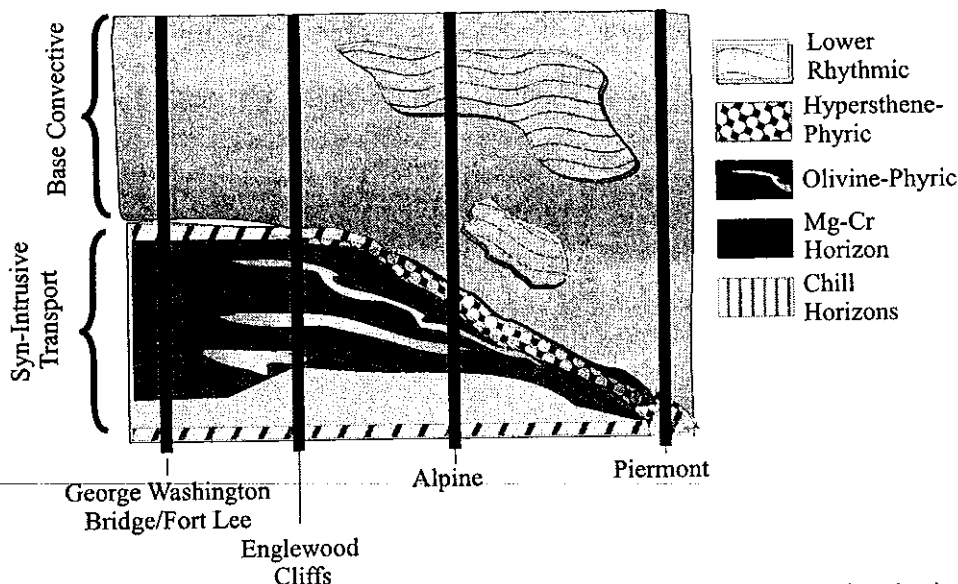
Thin section study also reveals that the Alpine section's rhythmic banding is complicated by a cryptic mm-scale banding, possibly a more-evolved version of Philpotts and Dickson's "boa" texture. Thin sections were optically analyzed by microscopic densitometry, channeling an incident light beam through the ocular of an epi-plan fluorescence microscope. The transmitted light is recorded using an Ocean Optics spectrometer. By averaging scans taken in raster fashion back and forth along the east-west direction (width), the traverse along the length of the thin section can be shown to possess millimeter-scale banding (Fig. 3).



**Figure 3.** Stepwise (0.02 mm) optical transmission measurements through a polished section. Each point represents an average of more than 300 data points taken along a set of east-west traverses (width); peaks represent mafic mineral enrichment. Grain size in plagioclase bands is approximately 0.6mm and consistent from band to band.

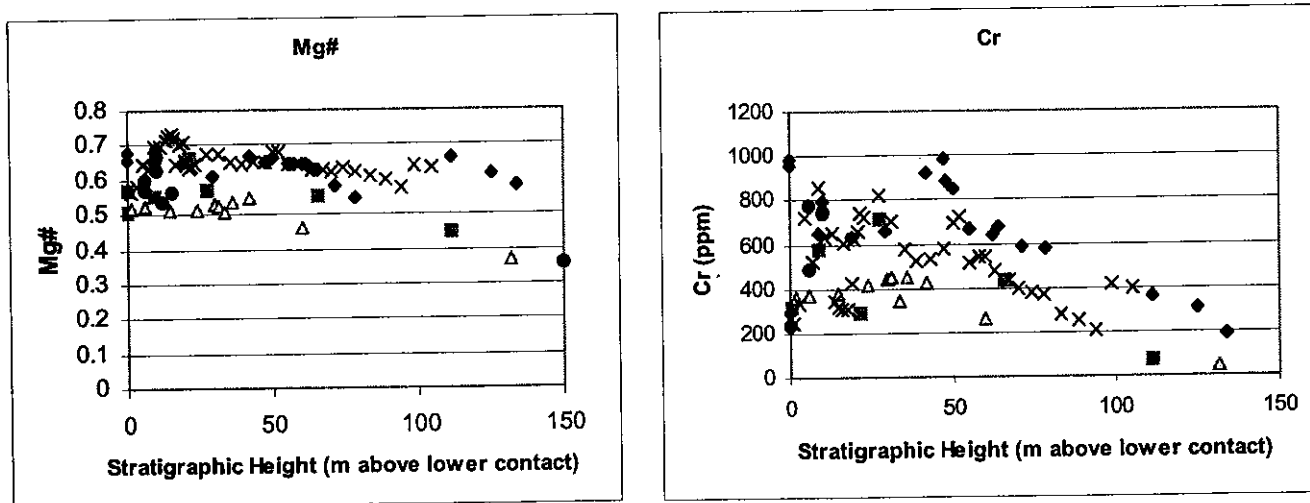
## Stop 2: Mafic Layer at Piermont

The mafic layer in the lower 100 meters of the Palisades is tentatively defined as having a  $Mg\# > 0.60$ ,  $Cr > 500$  ppm, and exhibiting a reversal in grain size along its upper margin at several localities. Facies within the mafic layer include the olivine pulses of the George Washington Bridge Section (Shirley, 1987), the hyalosiderite facies of the Englewood Cliffs Section of Walker (1969), and the geochemical reversals of the Alpine section (Gorring and Naslund, 1995) (Fig. 4). Steiner et al. (1992) note that the olivine-layer in the northern sections is replaced by a more hypersthene-enriched facies with subordinate altered olivine (iddingsite) with similar bulk chemistry as defined by Cr-Mg ratios (Fig. 5). Ongoing mapping carried out as part of the present study seeks to establish whether this Cr-Mg horizon represents an extension of the hyalosiderite layer, or a new facies in structural contact with the Olivine Horizon at some localities.



**Figure 4.** Mafic horizon (Mg-Cr Horizon) characterized by  $MgO\# > 0.6$ ,  $Cr < 500$  ppm and grain size reversal at several localities; also shows hypsthene phyric unit is quenched at Piermont, NY (Stop 2).

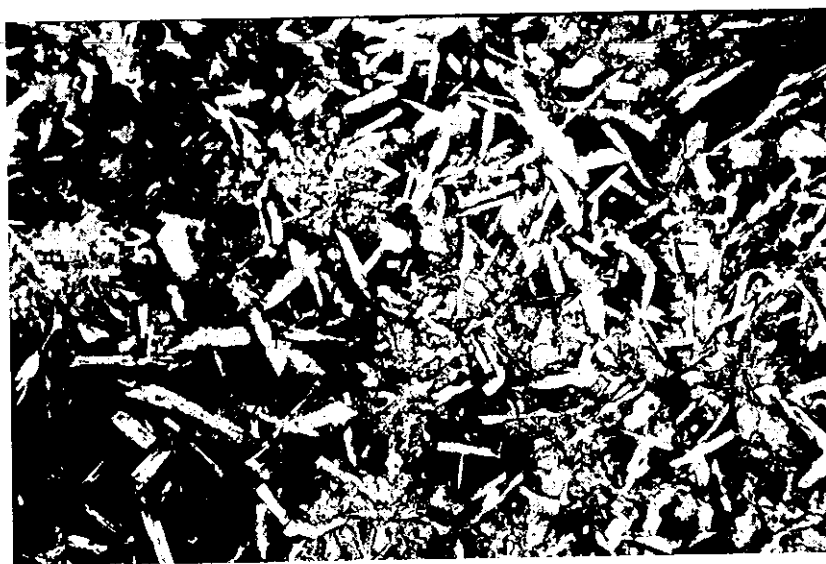
The various pulses, reversals and occurrences of olivine are taken together and mapped as one mafic zone, based not on mineralogy but on mafic components, in this case Mg and Cr. A narrowing or tapering of the layer occurs northward with its terminus at the Piermont lower contact (Figure 5). Extensive electron microprobe (200+, Block, PhD Thesis, in preparation) and whole rock analysis (500+, Block and Steiner, to be published elsewhere) of samples obtained from the lower 150 m of Palisades rock from the Piermont and Upper Nyack sections contain high levels of Mg and Cr ( $Mg\# = 0.67$ ,  $Cr = 979$  ppm) at the lower contact, adjacent to contact rocks possessing a quench texture and much lower levels of Mg and Cr. As little as 5-10 m above the contact there is an abrupt decrease in Cr concentration and  $Mg\#$ , indicating the absence of the mafic layer.



**Figure 5.**  $Mg\#$  and Chromium plots along stratigraphic height. Walker's hyalosiderite facies, Shirley's olivine pulses, Goring and Naslund's geochemical reversals, as defined by Chromium and Mg concentrations are taken together along strike to map out the pinching out of the mafic layer toward the northern portion of the intrusion. Symbols as follows: George Washington Bridge section, blue diamonds (Shirley, 1987); Englewood Cliffs, squares (Walker, 1969); Haverstraw, triangles (Walker, 1969); Fort Lee, crosses (Goring and Naslund, 1995); Upper Nyack, circles (Steiner, et al., 1992); Piermont, red diamonds, (This study).

### Stop 3: Millimeter-scale banding

A recent review of the petrographic differences across thick basalt flows finds that the colonnade comprises intergranular clusters of pigeonite and augite separated by plagioclase chains whereas the entablature develops more ophitic associations involving subcalcic augite (Philpotts and Dickson, 2000). The lower section of the Palisades (up to 150 m above the lower contact) contains linear quench textures similar to those found in the Holyoke flows (Philpotts and Dickson, 2002 after Steiner, et al., 1992). Partial melting experiments conducted by Dickson and Philpotts (2001) yield *boa* textures similar to those found in the Palisades roof zone (Philpotts and Dickson, 2002) and are in line with Steiner's (1992) hypothesis on decompression melting's role in recrystallization in portions of the Palisades as illustrated in his Cumulus-Transport-Deposition (C-T-D) model (Fig. 6).



**Figure 6.** Quench Texture in the Palisades Upper Nyack Section. Chains of ophitic clusters surrounded by dark mesostasis are similar to Dickson and Philpotts' (2002) "*boa*" texture. Partial melting due to decompression is a possible additional mechanism for producing this texture and is associated with entablature structures in flood basalts and large mafic intrusions.

In the Holyoke flows, mm-scale rhythmically banded horizons occur at the upper quench zone of the flows and contain ophitic clusters of decreasing size surrounded by dark mesostasis documenting the advance of the thermal boundary layer (Dickson and Philpotts, 2001). Dickson and Philpotts (2001) have found similar banding in the Palisades roof zone, suggesting that rhythmically banded sections at 100m from the lower contact (exact location unknown) are actually foundered blocks from the roof of the intrusion (Dickson and Philpotts, 2001).

### CONCLUSION

The Palisades Sheet continues to reveal unexpected complexity in the structure of the internal layers as research continues to focus on the lower horizons of the Palisades. Unlike many references that refer to the Palisades and other intrusive basalts as the result of simple internal bi-directional cooling and compression, the growing body of information suggests that the major internal layers are a complex expression that involves a combination of processes that may include cumulus and transport, decompression melting, convective processes and filter pressing.

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# Permian Remagnetization of Mafic Dikes in Northern New Jersey and Southeastern New York

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## ABSTRACT

My students and I sampled 13 sites in mafic dikes and two in mafic extrusives in the Jersey Highlands. Most of these igneous bodies had been mapped as Triassic or Jurassic intrusives. All of them yielded a Permian pole position, consistent with a remagnetization widely recognized in Paleozoic sedimentary rocks in the Appalachians. This work documents the existence of such a remagnetization in igneous rocks, and raises questions about possible causes of the remagnetization. Most of the proposed remagnetization mechanisms involve the movement of fluids through the rocks, something that is difficult to conceive of in igneous rocks. It is suggested that more probable remagnetization mechanisms are either a thermoviscous remagnetization or stress- or strain-induced one.

## INTRODUCTION

It has been well established that many Paleozoic sedimentary rock units in the Appalachian orogen were remagnetized in the Permian. Particular attention has been paid to studies of this remagnetization in carbonates. In many cases, the carrier of the secondary magnetization is magnetite, and numerous theories have been advanced to explain remagnetization of magnetite at temperatures low enough to be consistent with geothermal data in the carbonates such as vitronite reflectance. There have been no reports of remagnetization in mafic igneous rocks.

In the late 1980s I embarked on a study of some mafic dikes in the Jersey Highlands in which John Puffer (personal communication, 1989) had observed that the geochemical composition was inconsistent with their mapped Mesozoic ages. He suggested that they might be early Paleozoic or Precambrian. In the hope that paleomagnetism might enable us to estimate the ages of the dikes, I sampled seven localities. In all cases where the results were consistent from sample to sample, the dikes yielded a paleomagnetic declination and inclination consistent with a Permian magnetic pole. This clearly demonstrated that the dikes were pre-Mesozoic, but did not allow any finer resolution of the age.

Recently, my students and I sampled four additional localities. The locations of from which all samples were collected are shown in Figure 1. With the exception of two mafic dikes in the Cortlandt complex which yielded inconsistent results, all of the additional localities yielded results consistent with a Permian (Kiaman) magnetic overprint.

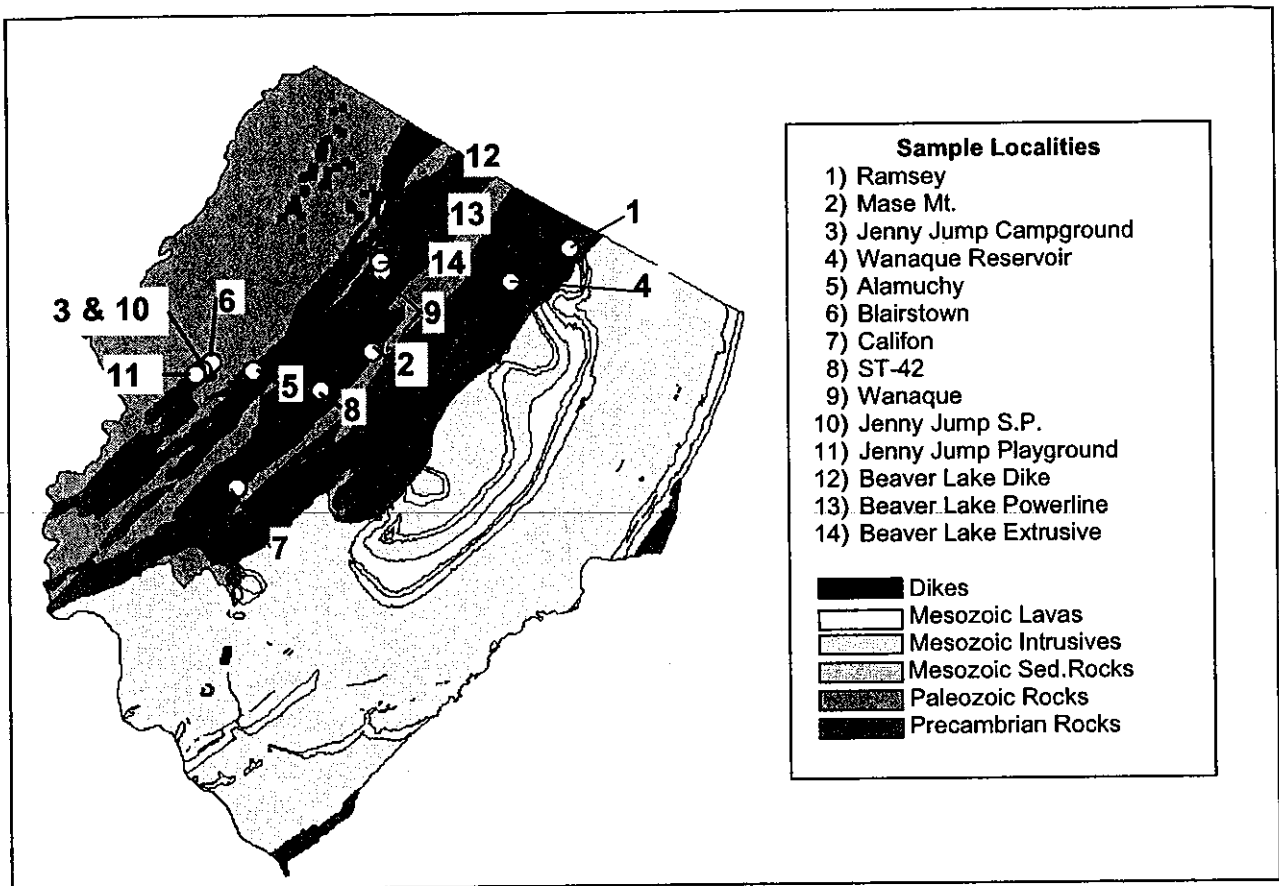


Figure 1. Map showing the location of sampling localities in northern New Jersey.

## METHODOLOGY

- 1) Samples were collected from fifteen mafic dikes and at two sites in one lava flow for paleomagnetic analysis. Cores were drilled in the field and oriented with a Brunton compass (and sun compass if there was sufficient sun light). One or two samples were cut from each core.
- 2) At least one sample from each core was subjected to step-wise, alternating field demagnetization at 2.5, 5, 10, 15, 20, 25, 30, 40, 50, 60, 70, 80, and 90 mT.
- 3) Selected samples were subjected to thermal demagnetization at 100, 200, 300, 400, 450, 500, 530, 560, 600, 625, 650, 660, 670, and 680 degrees C.
- 4) Vector endpoint diagrams were plotted (after Zijdeveld, 1967) to help identify components.
- 5) Principal component analysis (after Kirschvink, 1980) was carried out to isolate components.

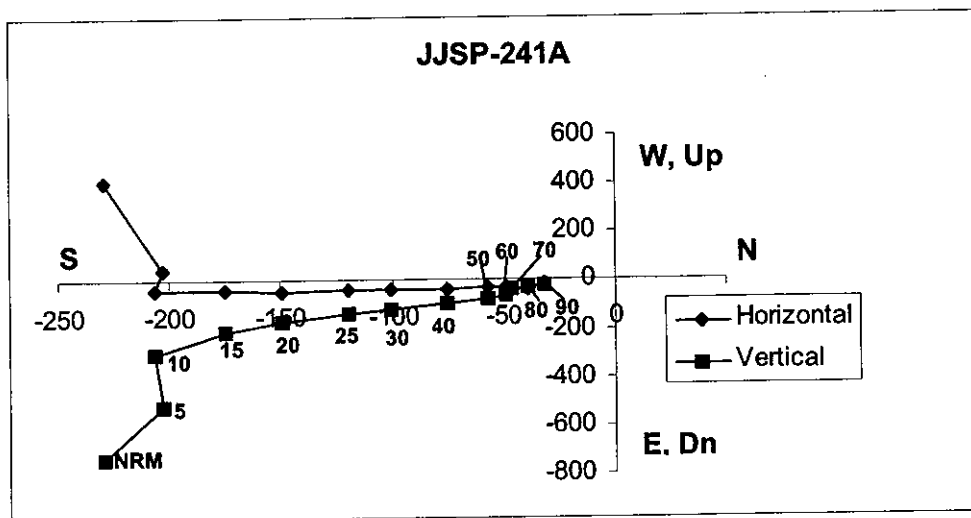
## RESULTS

Except for samples from the Cortlandt Complex, which yielded inconsistent results, all of the sites indicated a magnetization that trended to the south and plunged gently. The results are shown in Table 1. Data from each of the sites show more scatter than is usually desirable. This could be due to errors of measurement, inconsistencies in remagnetization, or errors in magnetic compass readings. All of the data are consistent with remagnetization in the Permian, as has been demonstrated in sedimentary rocks for decades (Elmore and Others, 2001).

**Table 1.** Paleomagnetic data from mafic dikes sampled in northern New Jersey. The number of cores is the number of independently oriented cores collected. Number of samples refers to the number of individual specimens analyzed. Number used refers to the number of specimens used in the analysis.

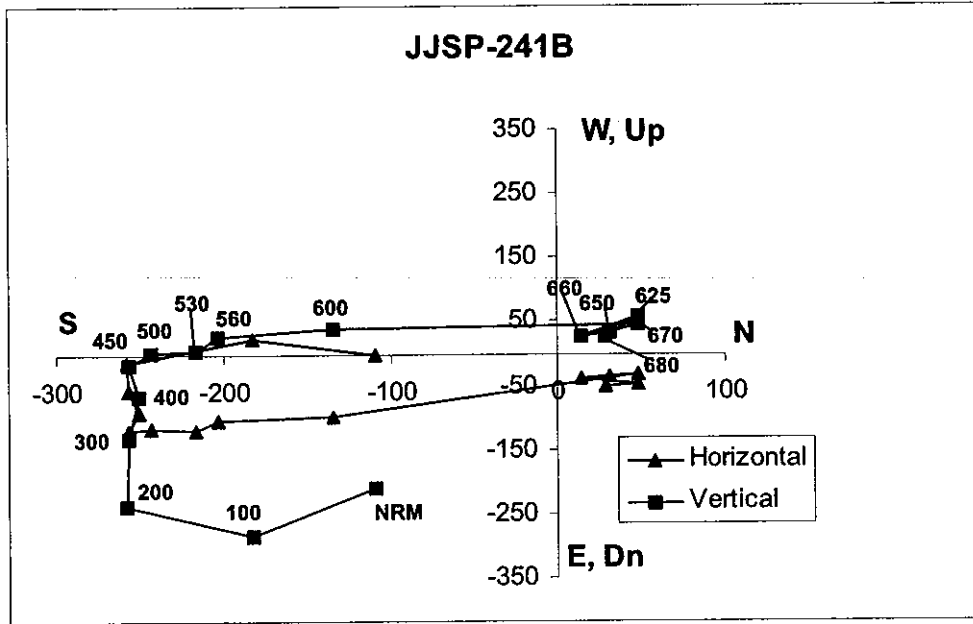
	Site Lat	Site Long	# of Cores	# of Samples	# Used	Mean Dec	Mean Inc	Alpha-95	Pole Lat	Pole Long
Ramsay	41.0894	74.2189	7	7	6	181	3.6	6.32	47.0992	104.3130
Mase Mt.	40.9450	74.5913	6	6	5	175	-6.3	52.04	51.9471	113.5252
Cortlandt Complex Lamprophyre										
Cortlandt Complex Mafic Dike										
Jenny Jump	40.9200	74.9100	12	12	12	164.6	0.3	6.98	46.6169	127.8342
Wanaque Reservoir	41.0433	74.3281	8	8	8	183.5	-11.7	18.76	54.7294	99.6357
Alamuchy	40.9184	74.8151	5	6	6	175.74	-15.24	7.55	56.6228	112.8737
Blairstown	40.9297	74.8934	4	6	6	171.83	5.93	6.64	45.4683	116.7821
Califon	40.7497	74.8456	5	6	6	164.47	9.55	19.32	42.2700	126.2885
ST-42	40.8902	74.6881	3	4	4	190.45	-23.67	10.24	60.0357	84.5355
Wanaque	41.0602	74.5723	8	10	10	167.3	0.1	12.9	47.3072	124.3465
Jenny Jump State Park	40.9203	74.9110	5	7	7	166.8	32.98	23.31	29.8414	119.5907
Jenny Jump Playground	40.9141	74.9238	3	5	5	151.05	-5.99	22.83	44.0007	147.2971
Beaver Lake Dike	41.0836	74.5754	4	5	5	178.78	-14.91	5.42	56.4823	107.6153
Beaver Lake Power Line	41.0792	74.5723	3	3	3	23.87	16.9	65.19	51.2787	145.1875
Beaver Lake Extrusive I	41.0748	74.5745	2	2	2	175.92	20.38	14.32	38.2653	110.5372
Beaver Lake Extrusive II	41.0748	74.5745	2	2	2	175.47	-18.78	14.48	58.3202	113.9521

Vector end point diagrams are shown in Figure 2 for stepwise alternating field demagnetization of samples 241A and 33A, and stepwise thermal demagnetization for 241B. Samples 241A and 241B are cut from the same core. One was subjected to alternating field demagnetization, and the other was subjected to thermal demagnetization. These are typical of the results from all of the sites.

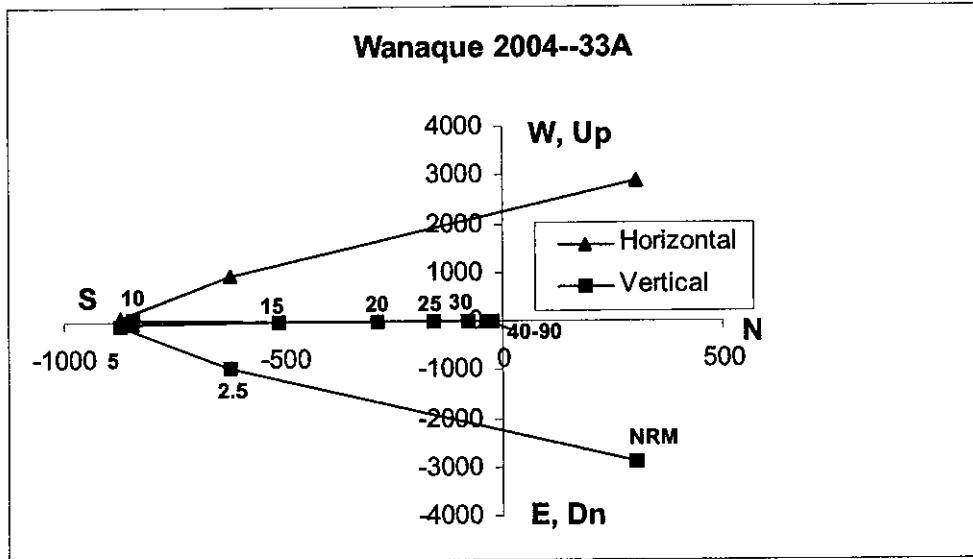


**Figure 2A.** Vector endpoint diagram for Jenny Jump State Park sample 241A. Specimen was taken from the same core as 241B. The sample was subjected to alternating field demagnetization. Boldface numbers indicate demagnetization level in mT.

First, an easily removed component is identified, and is interpreted to be a modern viscous overprint. Next, a very persistent reversed component, consistent with a Permian pole position appears as the characteristic magnetization. In some thermal demagnetizations (see Figure 2B, for example), a weak component that could be a primary thermal remanent magnetization is suggested in the plots, but we have not been able to completely resolve it as yet.



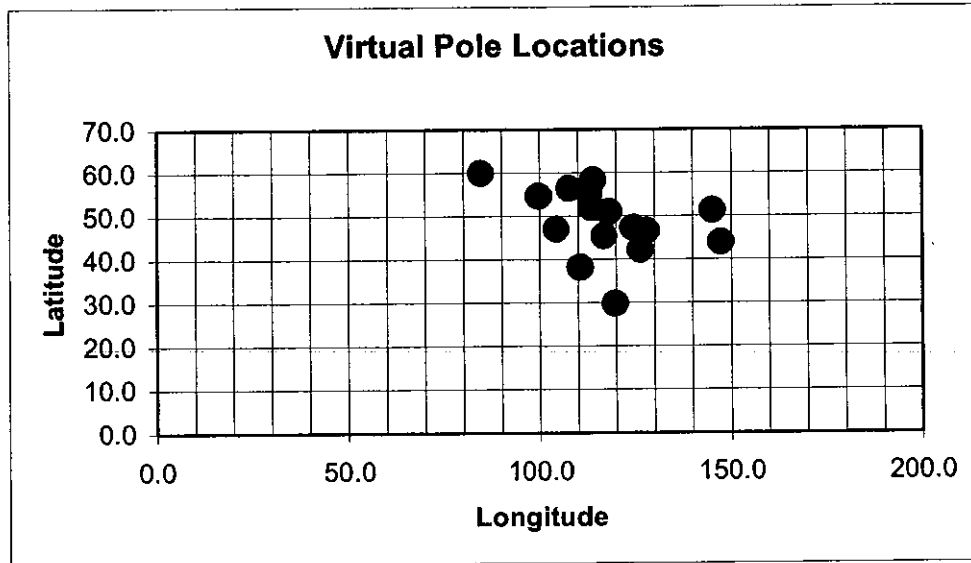
**Figure 2B.** Vector endpoint diagram for Jenny Jump State Park sample 241B. Specimen was taken from the same core as 241A. The sample was subjected thermal demagnetization. Boldface numbers indicate demagnetization temperature in degrees C.



**Figure 2C.** Vector endpoint diagram for Wanaque dike 33. The sample was subjected to alternating field demagnetization. Boldface numbers indicate demagnetization level in mT.



Figure 3 shows the virtual pole positions calculated from the dike data. Although there is considerable scatter, they are generally consistent with a Permian pole position of 50°N, 120°E, the general location suggested by most studies.



**Figure 3.** Latitude and longitude of magnetic poles calculated for the dike localities used in this study. The mean Permian position of the pole is 50°N, 120°E.

## DISCUSSION OF RESULTS

The most important result of this work is the clear demonstration that neoproterozoic dikes have been remagnetized during the Permian Kiaman Reversed Superchron. Previous work has repeatedly demonstrated such remagnetization in sedimentary rocks, and the mechanisms proposed were developed for sedimentary rocks. This is the first report of a Kiaman remagnetization in igneous rocks, and its presence puts constraints on the mechanisms available for causing the remagnetization.

According to McCabe and Elmore (1989) remagnetization of redbeds was suggested at least as long ago as 1968, by Creer. He believed that near-surface weathering produced hematite that carried a chemical remanent magnetization in those rocks. Remagnetization has since been recognized in both sandstones and carbonates (McCabe and others, 1983; Kent, 1985; Miller and Kent, 1988; McCabe and Elmore, 1989; McCabe and Others, 1989; Dorobek, 1989; Saffer and McCabe, 1992; Elmore and Others, 2001). In the sandstones, hematite is generally the carrier. In carbonates, the carrier is often magnetite, although examples where hematite is the carrier are not unusual.

The main difficulty with these remagnetizations is that all of the available data suggest that the rocks were not heated to high temperatures. When the carrier was thought to be only hematite, chemical remagnetization was seen as the likely mechanism. At first, weathering was suggested, but it soon became clear that surface processes were not the likely explanation in most cases. Nevertheless, hematite could be produced by diagenetic processes as well as by the migration of fluids through the sandstones after they were lithified, even at low temperatures.

Once magnetite was identified as the carrier of the remanence in most of the carbonates, the picture became even less clear. Available data suggested that these rocks were not heated above 200°C, yet mechanisms to remagnetize magnetite at low temperatures were unknown.

More recent work has produced a number of theories for the mechanism to remagnetize rocks in which magnetite is the carrier of the magnetization. Some workers (e.g. Saffer and McCabe, 1992), argue that the remagnetization is a chemical remanence acquired when heated fluids migrating through the rocks triggered the formation of magnetite grains. In the case of the Kiaman remagnetization, these fluids are usually linked to the Alleghanian orogenic event. In a study of remagnetization in the Helderberg carbonates, Dorobek (1989) suggested that fluid migration occurred over a relatively short period of time, that the fluids migrated along fractures and intracrystalline deformation microstructures, and that the fluids were quite warm (200° to 300°C).

Kent (1985) has suggested that thermoviscous remagnetization could occur at temperatures lower than had previously been suggested (Pulliah, 1975), and that those temperatures were consistent with burial temperatures in at least some portions of the Appalachian orogen. This mechanism involves the thermal activation of pre-existing magnetite grains. To be consistent with data observed in many of the studies, the rocks would have to have remained at elevated temperatures long enough to completely destroy the original depositional remanent magnetization.

Perhaps the most thorough discussion of the late Paleozoic remagnetization was a review by McCabe and Elmore (1989). They noted that remagnetization mechanisms fell into two broad categories: chemical remanence and thermoviscous remanence. Chemical remanence could have been produced by the migration of meteoric waters through the rocks, by basinal brines buried with the sediments and forced through the rocks, or, in some cases, migration of hydrocarbons. Thermoviscous remagnetization could have been due to heating on burial and uplift, or by heat transfer from hot fluids migrating through the rocks. Most workers associate the remagnetization, no matter what the mechanism, with the Alleghanian orogeny.

Hudson and others (1989) raised one more possibility, that of piezoremanent magnetization. The idea here is that stresses, presumably associated with the Alleghanian orogeny, triggered the remagnetization of magnetite grains. They suggest two main mechanisms: stress-induced movement of crystal lattice defects, and stress-induced acquisition of viscous remanent magnetization. They note that piezoremanent magnetization has been produced experimentally, but it has not been identified in rocks. Their work offers circumstantial support for piezoremanent magnetization as the cause of a syn-folding secondary magnetization in the Preus sandstone.

The identification of a Kiaman remagnetization of the neoproterozoic dikes in northern New Jersey and southern New York, puts an added constraint on the mechanism by which the magnetization was acquired. All of the chemical remanent explanations, and even one of the thermoviscous mechanisms require the migration of warm to hot fluids through the rocks. Such fluid migration is easy to understand in sandstones, but is more difficult to understand in limestones, not to mention igneous or metamorphic rocks. Pathways that have been suggested include fracture systems, a suggestion supported by fluid inclusion data, and intracrystalline deformation microstructures (Dorobek, 1989). Fluid migration through the neoproterozoic dikes and the crystalline rocks into which many of them are intruded seems much more difficult to explain.

To recapitulate, the mafic dikes discussed in this study carry a magnetization that is consistent with a late Paleozoic magnetic pole for North America. Most sites show a relatively weak viscous remanence thought to be related to recent pole positions, which is easily removed by thermal demagnetization to 625°C or by alternating field demagnetization to 20-30 mT. In most samples, only a single additional component can be identified, and it is oriented in a southerly direction which a shallow plunge, either positive or negative. In some samples there is the suggestion that there might be an additional component, but in no cases have we been able to clearly resolve it.

Thermal demagnetization spectra and the mineralogy of the rocks suggest that the principle carrier of the magnetization is magnetite. Because the rocks are igneous, with no pore

space, and they intruded (in many cases) crystalline metamorphic rocks which also lack pore space, it is difficult to envision fluids migrating through the rocks and producing a chemical remanent magnetization. It is possible that the remagnetization is an example of thermoviscous remanence, possibly due to burial, and less likely due to the introduction of hot fluids. Piezoremanent magnetization offers an interesting alternative hypothesis, although the present work provides no evidence on which to evaluate this hypothesis.

## CONCLUSIONS

Neoproterozoic dikes in the Jersey Highlands were remagnetized during the Kiaman Reversed Superchron. Even though some of these dikes have previously been mapped as Jurassic, they are clearly older than that. The carrier of the magnetization in these dikes appears to be magnetite, and mechanisms requiring the migration of fluids through the rocks seem unlikely given the lack of permeability of the dikes and most of the rocks they intrude. Chemical remanent mechanisms are probably not the cause of the remagnetization. Other possibilities include thermoviscous remagnetization or piezoremanent magnetization.

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# Road Log for Field Trip of Proterozoic, Paleozoic, and Mesozoic Mafic Intrusions of Northern New Jersey and Southern New York

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## MILEAGE

### Incr. Cum.

0.0	0.0	Exit Ramapo College parking lot and turn right onto Route 202 North.
1.4	1.4	Cross under Rt. 17 and bear right onto Rt. 17 North.
0.8	2.2	Bear left at split following signs for Routes 17 and 87. Avoid Route 287 South.
0.6	2.8	Move to right lane after split and bear right onto Routes 287 East and 87 South, following signs for Tappan Zee Bridge.
0.5	3.3	Roadcuts on left of hornblende granite of Mesoproterozoic age.
2.1	5.4	Outcrop on left of reddish-brown cobble to pebble conglomerate of Upper Triassic and Lower Jurassic Passaic Formation. Beds here dip toward the west.
1.3	6.7	Outcrops on both sides of road of interbedded cobble to pebble conglomerate and coarse-grained sandstone of the Passaic Formation.
1.2	7.9	Outcrops on both sides of road of the same.
1.4	9.3	Intersection with Garden State Parkway.
2.5	11.8	Take exit 13 onto Palisades Interstate Parkway <u>South</u>
12	23.8	Take Exit 2 (Alpine, NJ)
0.5	24.2	Follow signs to Palisades Interstate Park

## STOP 1 PALISADES SILL

### STOP 1A: CENTIMETER-SCALE BANDING IN PALISADES SILL AT ALPINE, NJ

**Leaders: Karin Block, Jeff Steiner, and Alan Rice**

The stratigraphic central portion of the Palisades at Alpine contains an extensive section extending from about 60 meters above the basal contact to approximately 150 meters that comprises a chemically and mineralogically distinctive unit (Block and Steiner, in preparation). Portions of this unit are characterized by cm-scale rhythmic banding (Fig. 1).



**Figure 1.** Centimeter-Scale Rhythmic Banding in Alpine Section of the Palisades Sill.

0.5	24.7	Return to Palisades Interstate Parkway <u>North</u>
3.0	27.7	Take Exit 4 onto 9W <u>North</u>
3.9	31.6	Exit at Piermont N.Y.
0.5	32.1	Proceed to base of Palisades Sill

**STOP 1B: MAFIC LAYER IN PALSIADES SILL AT PIERMONT, NY.**

The second stop (at Piermont) marks the locality where a mafic hypersthene-phyric layer can be observed in quench contact with the base of the Sheet. This contact is close to a second, less magnesium-rich, holocrystalline quench lithology. Piermont facies are shown to be compositionally and petrographically dissimilar to other contact lithologies.

**STOP 1C: GLOMEROPORPHYRITIC QUENCH HORIZON, ROUTE 9W**

Time permitting, a third stop will take place along Route 9W to document a second type of quench feature, the glomeroporphyritic quench horizon of Steiner et al. (1992) that is now identified with the “boa” texture proposed by Philpotts and Dickon (2002).

0.5	32.6	Return to 9W north
14	46.6	Turn right onto Park Road
1.0	47.6	Cross RR tracks and enter Stony Point Battlefield State Park.

## STOP 2. CORTLAND COMPLEX

Leaders: Alec E. Gates, Marian V. Lupulescu, Nelson Eby, Michael J. Hozik, Alan I. Benimoff, and John H. Puffer

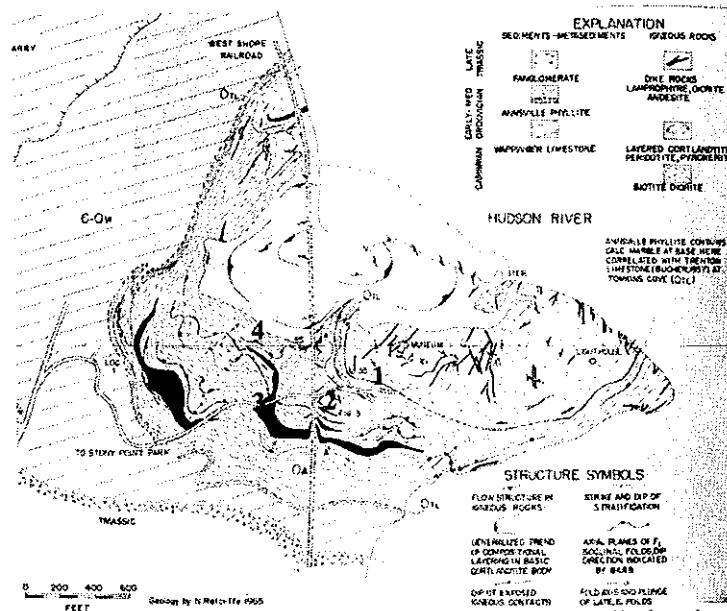


Figure 2. Geologic map of the Stony Point Battlefield area (from Ratcliffe, 1968) showing field trip stop locations.

### Stony Point Stop 1. Diorite

Medium-coarse grained diorite from the interior of the pluton contains plagioclase, biotite, quartz, amphibole and minor garnet and other accessories. The texture is hypidiomorphic granular with no flow foliation. Thin veins and dikes of lamprophyre intrude the diorite.

### Stony Point Stop 2. Diorite – Annsville Contact

At the contact of diorite and gneiss of the Annsville Formation each unit is affected by the other. Annsville Formation is a biotite and biotite-garnet schist/gneiss. It contains ptymatically folded granite veins of 0.5 to 2 cm thickness and locally containing tourmaline. The schist/gneiss is highly deformed with multiple crenulation cleavages. It contains muscovite, biotite, quartz, feldspar and accessories. The diorite contains numerous xenoliths of Annsville Formation in various stages of assimilation. The density of garnets is much higher at the contact where it forms clots with grains up to 1.5 cm. The composition of the diorite approaches granodiorite. Contamination of the magma with country rock has clearly changed the composition.

### Stony Point Stop 3. Dike in Annsville Formation

Fine-grained lamprophyre dike intruding Annsville Formation schist/gneiss. The schist/gneiss is highly deformed with multiple crenulation cleavages. It contains muscovite, biotite, quartz, feldspar and accessory minerals. Locally, chill margins may be observed. The dikes have been drilled for paleomagnetic studies.

#### Stony Point Stop 4. Cortlandite

Massive black pluton of Cortlandite is in fault contact with Annsville Formation. It is composed of large randomly oriented crystals of hornblende that poikilitically enclose olivine, enstatite and augite.

- |     |      |  |
|-----|------|--|
| 0.0 | 46.6 | Exit Park via Park Road and return to Route 9W.  |
| 0.7 | 47.3 | Turn left onto Route 9W South and follow to Haverstraw.  |
| 2.8 | 50.1 | Turn right onto Gurnee Ave., following signs for Route 202 West.   |
| 0.0 | 50.1 | Turn right onto Route 202 West.  |
| 3.4 | 53.5 | Passing beneath Palisades Interstate Parkway   |
| 3.2 | 56.7 | Outcrop on left of dark brown cobble to boulder conglomerate of the Passaic Formation. Cobbles are predominantly light tan to gray cherty limestone and dolomite, and less abundant pink or gray quartzite and purple-red quartz-pebble conglomerate (Savage, 1968). From here south, we are driving subparallel to the Ramapo fault. The uplands of Ramapo Mountain to the right are on the footwall of the fault, whereas the rocks exposed along Route 202 are on the hanging wall. |
| 0.5 | 57.2 | Outcrop on left of Lower Jurassic Ladentown basalt, considered to be the western extension of, and extrusive equivalent to, the Palisades sill (Ratcliffe, 1980; Puffer et al., 1982).   |
| 0.9 | 58.1 | Outcrop on right of more dark brown cobble to boulder conglomerate of the Passaic Formation.   |
| 3.4 | 61.5 | Crossing Ramapo fault. Brittle deformed exposure on right of hornblende granite of Mesoproterozoic age.  |
| 0.3 | 61.8 | Suffern, NY. Turn right at light onto Route 59 West. Outcrops of less deformed hornblende granite along road from here north.  |
| 1.6 | 63.4 | Traffic light. Continue straight on joint Routes 17 North and 59 West.   |
| 0.2 | 63.6 | Passing beneath New York Thruway.  |
| 1.2 | 64.8 | Bear right at exit onto Route 72 West (Sterling Mine Rd). DO NOT stay on Route 17 North.   |
| 1.8 | 66.6 | Entering Orange County, NY.  |
| 1.6 | 68.2 | Entering Ringwood Township, Passaic County, NJ.  |
| 1.0 | 69.2 | Ringwood State Park on right.  |
| 0.7 | 69.9 | Turn right onto Margaret King Ave.   |
| 2.3 | 72.2 | Turn left onto Route 511 South.  |
| 0.6 | 72.8 | Turn right onto Stonetown Road   |
| 0.2 | 73.0 | Crossing dam at Monksville Reservoir (right), created in 1985 by damming of the Wanaque River. Monksville Reservoir has a capacity of 7 billion gallons. Wanaque Reservoir is downstream (to the left).  |
| 3.7 | 76.7 | Bear right at three-way intersection to stop sign. Continue straight ahead onto West Brook Road Wanaque Reservoir on left.   |
| 0.5 | 77.2 | Turn left onto Snake Den Road, opposite sign for NJ Audobon Society Weis Ecology Center.   |
| 0.3 | 77.5 | Turn left, following signs for Weis Ecology Center.  |
| 0.3 | 77.8 | Park in large gravel lot on right at Weis Ecology Center and walk east along Snake Den Road a short distance to combined yellow-red trail on right. Follow trail south past prominent ridge on left and take yellow trail (on left) to top of ridge.   |



### STOP 3. LATE NEOPROTEROZOIC DIABASE DIKE

Leaders: Rich Volkert, Mike Hozik, and John Puffer

At this stop we will examine a diabase dike of late Neoproterozoic age that is quite representative of dikes of this age in the New Jersey Highlands. This dike is exposed continuously along the ridge east of Blue Mine Brook where it intrudes a sequence of Mesoproterozoic gneiss, marble and metaquartzite. This dike maintains an orientation of N10°E to N25°E and has a thickness of as much as 30 ft (9.1 m) along strike for a distance of about 2.5 mi (4 km). To the west, at the base of the ridge, the dike contains a large xenolith of pinkish-white, coarse-grained gneiss along its contact. Other characteristics of the dike (best observed toward the ridge crest) include moderately discordant and sharp contacts against the enclosing Mesoproterozoic rocks, aphanitic chilled margins in the dike that grade into a coarser-grained interior, and possible incipient columnar jointing along the eastern contact. Note the absence of fracturing in the gneiss here along the contact implying passive, rather than forceful, emplacement of the dike into extended country rock. Along the ridge crest the dike has been streamlined, polished and striated by glacial ice. Glacial striations here trend S26°E.

Geochemical analysis (Table 1) of the dike from this stop (sample 503), from outcrop along strike to the southwest (sample 437), and from two other dikes in the Wanaque quadrangle (see Fig. 3 map for sample locations) reveals the following compositions:

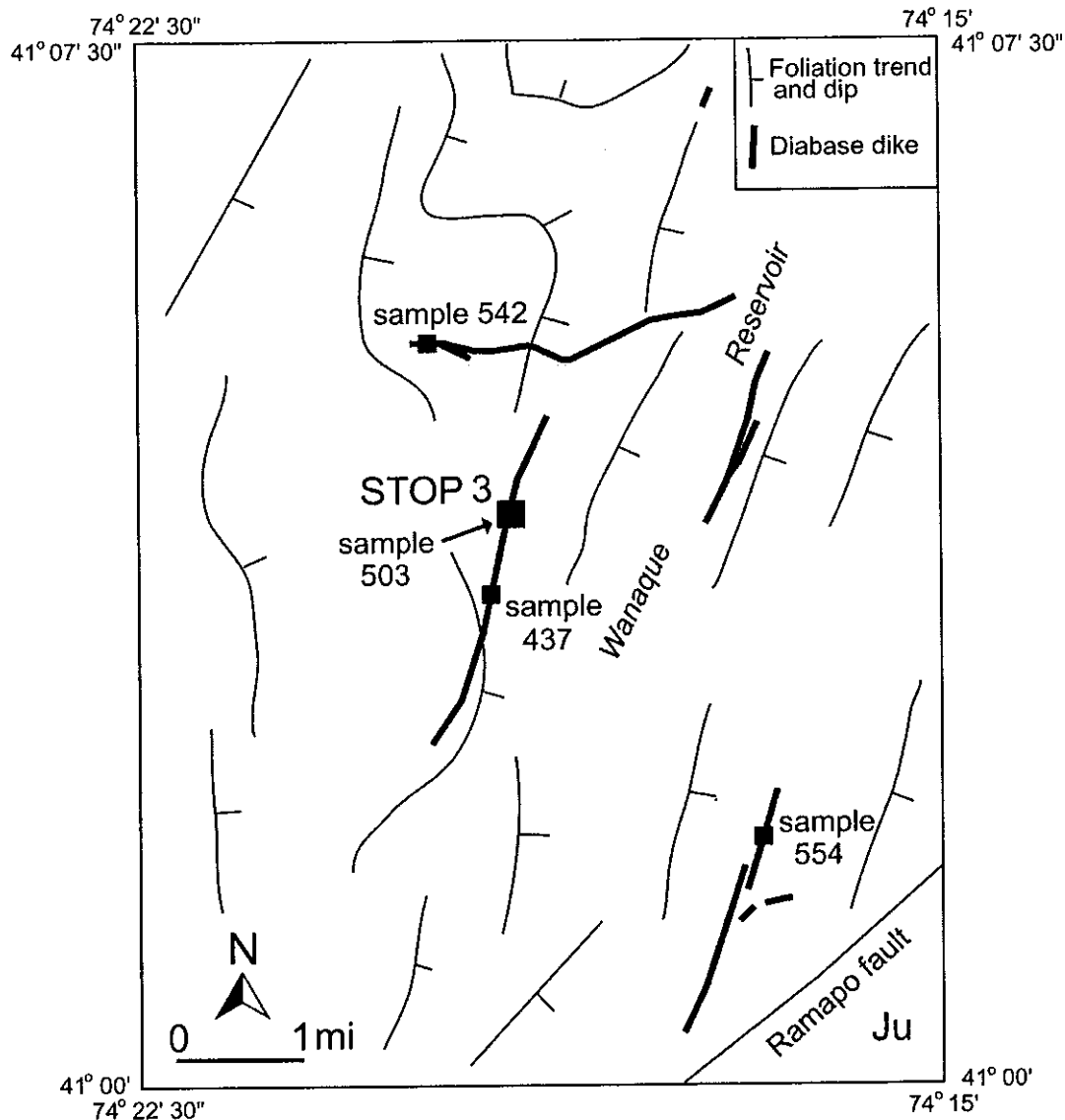
#### GEOCHEMICAL ANALYSES OF NEOPROTEROZOIC DIABASE DIKES FROM THE WANAQUE QUADRANGLE

Sample no.	503	437	542	554
SiO <sub>2</sub> (wt %)	48.50	47.98	49.10	50.01
TiO <sub>2</sub>	2.66	3.72	2.66	3.56
Al <sub>2</sub> O <sub>3</sub>	15.98	12.98	15.00	12.14
Fe <sub>2</sub> O <sub>3</sub>	4.61	6.25	4.20	6.01
FeO	7.44	9.24	10.00	8.12
MnO	--	0.24	0.20	0.23
MgO	5.04	4.57	4.38	6.60
CaO	8.91	7.65	8.92	7.31
Na <sub>2</sub> O	3.04	3.27	2.68	3.33
K <sub>2</sub> O	1.33	1.00	1.40	1.22
P <sub>2</sub> O <sub>5</sub>	0.37	0.51	0.28	0.81
LOI	1.67	2.12	1.23	1.28
Total	99.55	99.53	100.05	100.62
Ba (ppm)	--	--	380	270
Cr	109	--	58	40
Ni	31	--	--	15
Nb	--	--	40	15
Sr	301	--	350	370
Y	--	--	20	18
Zr	112	--	170	220

Table 1. Geochemical data from Volkert and Puffer (1995). LOI, loss on ignition.

Of particular significance is the geochemical similarity of dike samples 503 and 542 despite nearly orthogonal trends, implying the absence of correlation of dike trend and composition. All of the Neoproterozoic dikes in the Highlands are enriched in TiO<sub>2</sub>, P<sub>2</sub>O<sub>5</sub>, Zr, and light rare-earth elements compared to Early Jurassic basalt and diabase in the Newark basin. Contents of P<sub>2</sub>O<sub>5</sub> divide the dikes into

two broad groups, one low and one high (see Volkert this volume for discussion). This particular dike is part of the low  $P_2O_5$  population. All of the dikes have geochemical compositions that are consistent with magma formation in a rift-related, within-plate tectonic setting.



**Figure 3.** Simplified bedrock geologic map of the Wanaque quadrangle (modified from Volkert, 2000) showing the distribution of Neoproterozoic diabase dikes in relation to structural trends in the Mesoproterozoic country rock. Samples are numbered as in the table. Ju, undifferentiated Lower Jurassic rocks of the Newark basin.

Although the Neoproterozoic dikes in the Highlands have not been dated radiometrically, the field relations provide compelling evidence for a Neoproterozoic age. This is supported by the fact the dikes postdate the regional granulite-facies metamorphism that had deformed the Mesoproterozoic rocks by ca. 1030 Ma, as well as by the absence of the dikes in cover sequence rocks of Cambrian or younger age. Neoproterozoic dikes in the Highlands were likely emplaced sometime between 602 and 554 Ma during rifting of eastern Laurentia and breakup of the supercontinent Rodinia, prior to development of the Iapetus Ocean.

- 0.0 77.8 Exit Weis Ecology Center via Snake Den Road  
0.3 78.1 Turn right, staying on Snake Den Road  
0.3 78.4 Turn right onto West Brook Road  
0.5 78.9 Turn left at three-way intersection onto Stonetown Road  
3.7 82.6 Crossing over dam for Monksville Reservoir.  
0.2 82.8 Turn left onto Route 511 North.  
0.6 83.4 Turn right at Gulf station onto Margaret King Ave.  
2.3 85.7 Turn left onto Sloatsburg Road  
1.7 87.4 Entering Orange County, NY. Sloatsburg Road becomes Route 72 East.  
3.2 90.6 Bear right onto Route 17 South.  
1.2 91.8 Bear right onto ramp for Route 287 South.  
1.2 93.0 Bear left at split, following signs for Route 17 South.  
0.6 93.6 Turn right at exit for Route 202.  
0.1 93.7 Turn left at bottom of ramp onto Route 202 South.  
1.3 95.0 Turn left into Ramapo College  
End of field Trip

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# **Interactive Rock Cycle Activity: A Multimedia Learning Environment**

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## **ABSTRACT**

The rock cycle is a basic concept for the understanding of Earth Science. This interactive computer-based program allows students and teachers to simulate the manipulation of rocks through the rock cycle. The program will model how igneous, sedimentary and metamorphic rocks are formed and continuously transformed on Earth. The program presents the basic concepts of the rock cycle in an Introductory Level and then allows the student to apply these criteria to a variety of rocks as they explore the rock cycle in the Exploration Level. Each section provides the students with “just in time” definitions of important earth science terms. They also can answer questions about the rock types and their relationship to the rock cycle in the “Test Your Knowledge” sections that are also available. Finally, students can find where these samples occur on Earth by matching the 3 rock types on generalized geologic formation maps of the world.

## **INTRODUCTION**

Over the past 14 years Martha Schoene has introduced the rock cycle to her students using standard textbooks, supplemental handouts which include additional diagrams and critical thinking questions, and animations of rock formation. Upon reflection of these teaching experiences, she has discovered that students have a very difficult time mastering this fundamental Earth Science concept. Students will often confuse parts of the rock cycle and misinterpret its effects. This low level mastery of the material results in an inability to apply these facts and draw conclusions from presented data. When interpreting static diagrams of the rock cycle provided in textbooks, students are immediately focused on the directional arrows that are used to represent the cycle more than on the actual scientific process. Since this concept is the foundation for more complicated theories in geology, grasping the rock cycle would greatly enhance students’ fundamental understanding of this knowledge domain.

The difficulty that students face in mastering this challenging concept is assisted by misconceptions that are formed by their everyday experience with the world. Since most rocks form over a long period of time and in exotic and often inaccessible locations, most students have no first hand experience of how rocks are formed. Students take for granted the ubiquity of rocks and experience their seemingly static nature in the environment. According to John T Bruer (1993) students learn by watching the world around them which results in the formation of schema. These schemas, cognitive network structures that store general knowledge, are often inconsistent with formal science. “Effective science instruction has to start from students’ informal science conceptions, build on those informal ideas that are consistent with formal science, and correct or eliminate others.” (Bruer, 1993, p.132)

The Interactive Rock Cycle Activity allows student learning to follow a progression, building on existing knowledge to master more difficult concepts. For example, students begin the program at the Introductory Level which exposes them to rock samples that are easily found in nature, i.e. granite, and therefore can be obtained for in-class use. This allows teachers to provide students with the opportunity to actually handle and become familiar with these samples. Beginning with this familiar rock, a learning path is provided to help students experience the variety of circumstances involved in forming a new rock. Compared to reading and memorizing material in a textbook, this activity allows students to spend more

time on task and allows opportunity for more meaningful engagement of the rock cycle. "Compelling problems and activities provide vivid fixed points around which conceptual change can occur." (Bruer, 1993, p.132)

## **INTERACTIVE ROCK CYCLE ACTIVITY**

### **Introductory Level**

The Introductory Level provides background information about the formation of rocks and begins with a sample image of granite. Students will follow this granite sample through a linear, time-line progression that highlights the weathering, melting or exposure to heat and pressure processes. This allows students to gain a simple, but fundamental, understanding of the igneous, sedimentary or metamorphic steps needed to change the original granite sample into a different rock or to reform it into a similar piece of granite.

While exploring the igneous process students are presented with choices that affect the end result of the cycle. By selecting from a set of criteria, students can observe what could happen to the piece of granite as it is weathered, melted and cooled or subjected to heat and pressure. After guiding the sample through one phase change the student can fit this pathway into a circular model of the entire rock cycle. They can also choose to answer follow-up questions in the "Test Your Knowledge" section and/or view the maps provided to find where that rock type might be found in the world. Students are encouraged to complete the linear activities provided in the Introductory Level in order to become familiar with the terms and processes that are a part of the formation of igneous rocks to form metamorphic, sedimentary or other igneous rocks.

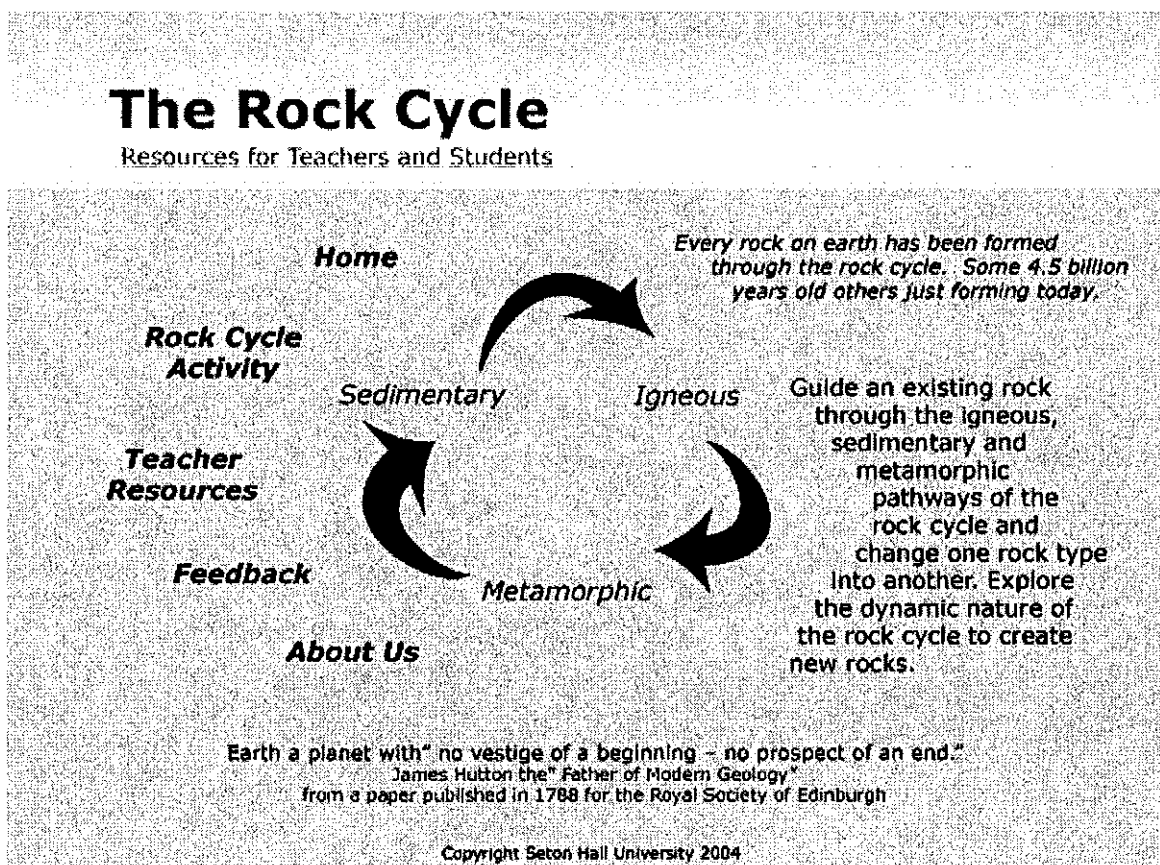
### **Exploration Level**

This level allows students to apply and expand on the information presented in previous level. Students are presented with a choice of rock types such as the igneous rocks Granite, Basalt and Peridotite. Students can then guide the selected rock sample through a variety of steps to form a different rock. Students are given a list of possible choices and have to select an appropriate next step in the process. The list of criteria will contain at least one inappropriate step to increase the level of difficulty of this activity and promote critical thinking. Students can consult the included glossary or ask for clues built into the activity to help them to eliminate the incorrect steps and select a possible pathway. Students are able to explore the igneous, sedimentary and metamorphic processes by also manipulating temperatures and selecting locations for the samples selected. The students are encouraged to fit their selections into a complete rock cycle model, test their knowledge and observe where igneous, sedimentary and metamorphic rocks are most likely found in New Jersey, North America and on each continent.

According to Cary, Smith and Wisner, as cited in Bruer (1993), "Children learn science better through activities than lectures because activities allow them to acquire knowledge while solving problems." (Bruer, 1993, p.141) These activities allow students to become active learners where previously they were only able to passively acquire information from reading and memorizing their textbooks. Students will be able to work in pairs or small groups to solve the problem of converting rocks into new rocks through the rock cycle. The "Test Your Knowledge" section will give students instant feedback on the progress of their learning and they can return to the particular parts of the program that they need to review. In addition, supporting lab activities in the classroom, available in the Teachers Resources section, will give the students a chance to transfer their knowledge to the real world.

## Teacher Resources

The Teacher resources section of this program provides diagrams, demonstrations and suggested lab activities that supplement the information in the Interactive Rock Cycle Activity. Students can make a sedimentary rock, observe weathering, relate cooling rates to crystal size and create edible metamorphic treats. These activities give students a chance to work with the concepts and apply them in the classroom and/or at home. This section also provides a list of appropriate books and Field Guides, as well as suggested video clips, external links to web sites and field trips. Teachers are also encouraged to email any comments or suggestions for future versions of this activity.



**Figure 1.** The Interactive Rock Cycle Activity is part of a larger project dedicated to providing teaching and learning resources. Teachers can obtain additional materials through the Rock Cycle; Resources for Teachers and Students Homepage, pictured above.

## CONCLUSION

This Interactive Rock Cycle Activity is designed to be available via the Internet but is also available as freeware on a CD-Rom to accommodate teachers who need to local copies for their computer labs. At present there are two levels available for grades K-12. The third level for math based geology classes grades 9-16 is not yet completed.

The goal of this program is to allow students to interact with processes that would be otherwise impossible to interact with in the real world or to recreate in a laboratory setting. Based on constructivist learning principles, students are transformed from passive to active learners in the construction of their own knowledge. By providing students with opportunities to correct faulty assumptions that were formed

by their every day experiences with the world, they build correct mental models that enable them to apply this knowledge in other settings.

### **ACKNOWLEDGEMENT**

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