Brooklyn Injection Gneiss Complex: Geochemical and Tectonic Synthesis

By

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Abstract

BROOKLYN INJECTION GNEISS COMPLEX: GEOCHEMICAL AND TECTONIC SYNTHESIS

By

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Middle Proterozoic rocks of the granulite facies Brooklyn Injection Gneiss (BIG) Complex represents a diverse suit of rocks associated with the stepping back of the Grenville subduction zone and subsequent development of an island arc/ back-arc basin situated along the Laurentian margin at ca. 1,035 to 980 Ma. A depleted mantle model of partial melting and subsequent fractionation producing LIL element (K, Rb, Ba) enrichments with HFS element (Ta, Nb, Zr, and Ti) depletions, similar to that of modern island arcs, is responsible for generating the Ravenswood bimodal suite which has an intrusive relationship to the Fordham Gneiss units of the Manhattan Prong. Magmas with calc-alkaline signatures were generated during initial arc/back-arc development where the role of the previous New Jersey Highlands calc-alkaline magmatism associated with the sub-continental lithosphere was still influential. As the basin matured and widened transitional N-type MORB and tholeiitic island arc basalts entered the basin via mantle upwelling. The transformation of the Laurentian margin from a passive continental margin to a back-arc depositional setting is reflected in the variability and character of the Fordham gneiss lithologies as well as their associated REE profiles. The establishment of a volcanic arc/back arc ophiolite complex along this section of the Laurentian margin during the Grenvillian orogeny, suggests that the end stage Grenvillian event did not involve a continent-continent collision but an island arc-Laurentia collision at ca. 960 Ma. It is suggested that the New York Recess is a pre-Grenvillian feature centered on a persistent mantle plume associated with repeated cycles of compressional orogenics and continental breakups. Reactivation of the Ramapo – East River border fault system is responsible for the development and collapse of both the Middle Proterozoic Fordham and Paleozoic Manhattan marginal basins. Taconian overthrusting of the Middle Proterozoic BIG Complex is conducted along a low angle detachment in a similar manner to other Appalachian models. This study suggests that eastern seaboard complexes, with volcanic arc/back-arc ophiolitic affinities that have not been carefully dated, may need to be reevaluated for a possible Middle Proterozoic association.

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I would also like to extend a sincere appreciation to my colleagues at the Department of Environmental Protection who were supportive of my research and who share in the realization of how fortunate we are to have worked at the "heading" on such an enormous undertaking as City Tunnel #3.

Most of all, my heartfelt appreciation is to my lovely wife, Amy, without whose support and understanding I would not be able to function let alone attain my goals. To Linnea, Ellie and Jack, who I think, were quite amused that their Dad still had to go to school and who would greet me with so much love after long days. To my mother and brother who are always there to help. To my father, my hero, whom I'm sure, is looking down with great pride.

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The age and field relationships of the metamorphic rocks of the New York City group have been debated for over a century. Perhaps the most puzzling section exists on the Long Island side of the East River where complicated mixtures of meta-igneous and meta-sedimentary gneiss, the Brooklyn Injection Gneiss, are buried by unconsolidated sediments. An opportunity to clarify the geology of this section of New York City has been provided by the recent excavation of a 17 km (10.5-mile) segment of New York City's Water Tunnel No. 3 (CT#3) through the Boroughs of Brooklyn and Queens (Figure 1). This segment of CT#3, termed Stage 2, was separated into two sections, a 4.9m (16 ft) finished diameter tunnel of 8.9 km (5.5 miles) through Brooklyn and a 6.1m (20 ft) finished diameter of 8.1 km (5 miles) through Queens. Excavated at a depth ranging from 168 m (550 ft) at its terminus in the Red Hook section of Brooklyn, Stage 2 grades to 240 m (785 ft) beneath the surface of the Astoria section of Queens where it connects to the already completed Stage 1. The tunnel has a rock cover ranging from 128m (420ft) to 218m (715ft) with an additional amount of unconsolidated sedimentary overburden ranging from 72m (235ft) to 21m (70ft) respectively. Exposures created by the Stage 2 tunnel are the focus of this petrologic study of the Brooklyn Gneissic rocks.

An integral part of the author's responsibility in mapping the geology of the tunnel and shafts for the City of New York was to conduct a thorough search of all the published reports related to the regional geology of New York City. Not only did the magnificent exposure of this restricted section of New York's geology provide excellent subject matter but the historical debates on the correlation and age of these geologic units did as well. In an effort to clarify the geology of this section, this dissertation employs

Figure 1. Location map of CT#3's route through western Long Island. Dashed lines represent Cameron's Line in the study area (after Baskerville, 1992, 1994). CT#1, CT#2 and the Richmond tunnel are also marked.

geologists, as well as U-Pb isotopic age dating, petrography and electron microprobe studies. Information will be used to provide a petrologic and tectonic synthesis of the evolution of the Manhattan Prong[1], and in particular the Proterozoic Y[2] BIG Complex. The evolution of the BIG complex is used to create a new model for the formation of the Paleozoic Manhattan basin.

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[1] Delineated in Long and Kulp (1962, pg. 971) as encompassing Westchester County and the City of New York, **"bounded on the North by the Highlands, on the west by the Triassic Lowland, on the south essentially by Long Island Sound, and on the east by an adjoining metamorphic province in Connecticut."**

[2] Proterozoic Y corresponds to the Mesoproterozic of ca. 1.6 Ga to 1.0 Ga.

The metamorphic rocks of the Manhattan Prong bedrock have historically consisted of five formations (Fordham Gneiss, Yonkers Gneiss, Lowerre Quartzite, Inwood Marble and Manhattan Schist) after Merrill (1890) and collectively were termed the 'New York City Group' by Prucha (1956). The Precambrian Fordham Gneiss, basement rocks of the New York City Group, is divided by Hall (1966, 1968 and 1976) into 10 mapable units north of New York City. Ratcliffe (1968, as reported in Hall, 1976) subdivided the Fordham into three units south of Peekskill. However, the transition between the Hall and Ratcliffe lithologies has yet to be defined. Similarly, four informal members within New York City have been defined by Baskerville (1992; see Figure 2) and no clear correlations were offered by Baskerville between either the Hall or Ratcliffe series.

Hall (1966, 1968, and 1976) considered the Fordham Gneiss to be a metamorphosed Precambrian eugeosynclinal sequence comprising both para- and ortho-gneiss components. The predominate member of the Fordham Gneiss is a well-banded garnet, biotite, quartz, plagioclase gneiss with granoblastic texture and subordinate inter-layers of amphibolite, impure marbles and mica schist (Berkey, 1910, Prucha, 1956). The Fordham D of Hall (1968) contains sillimanite and biotite with distinctive lavender colored garnets. The other Fordham units contain both hornblende and biotite with an overall general lack of clinopyroxene (Prucha, 1956). The Yonkers Gneiss and the Pound Ridge Gneiss of Westchester County are Avalonian age meta-igneous granitic units with an intrusive relationship against the Fordham Gneiss (Long 1969; Mose and Hayes, 1975).

Figure 2. Geologic map of a portion of lower Manhattan and western Brooklyn from Baskerville (1994) showing interpreted geologic formations and the location of Cameron's line.

The Ravenswood Granodiorite (Kemp 1895; Berkey 1910; Zeigler 1911) of the Manhattan Prong is situated on the Long Island side of the East River, extending to the eastern margin of lower Manhattan. This section has been overlooked in most regional geologic models, in part due to the extensive overburden. The studies of earlier tunnel exposures by Kemp (1895); Berkey (1910, 1931, 1933) and Blank (1930, 1973) studied the crystalline bedrock of the extreme western portion of Long Island and give petrographic descriptions of a meta-igneous suite ranging from a garnet diorite to garnet granodiorite. The Ravenswood Granodiorite is presently redesignated the Ravenswood meta-tonalite[3] (or enderbite[4]) in better accord with its metamorphic texture (see Blank 1973).

During the excavation of CT#2 (Figure 1), Berkey (1931; see appendix I for actual memorandum) introduced the term Brooklyn Injection Gneiss to the system that he described as being represented by the Pre-Cambrian age Ravenswood 'diorite' intrusive, which penetrated the older already metamorphosed Fordham gneiss to such an extent that is was difficult to separate the exposure into mapable units. Therefore, Berkey proposed naming the rock complex in western Long Island as the Brooklyn Injection Gneiss[5] (shown in Figure 3) herein referred to as the BIG Complex.

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[3] The term tonalite is used herein to refer to garnetiferous enderbite representing the metamorphic equivalent of initial tonalite magma.

[4] Pyroxene bearing plagioclase rich member of the charnockite family.

[5] A discussion on the term "injection gneiss" was presented by Barth (1936, pg. 840) in his study of the Precambrian and Paleozoic rocks of the Manhattan Prong in Dutchess County, **"...a few words should be said about the term injection gneiss. It is a good word insofar as it conveys the idea that igneous matter has been injected into the rock."**

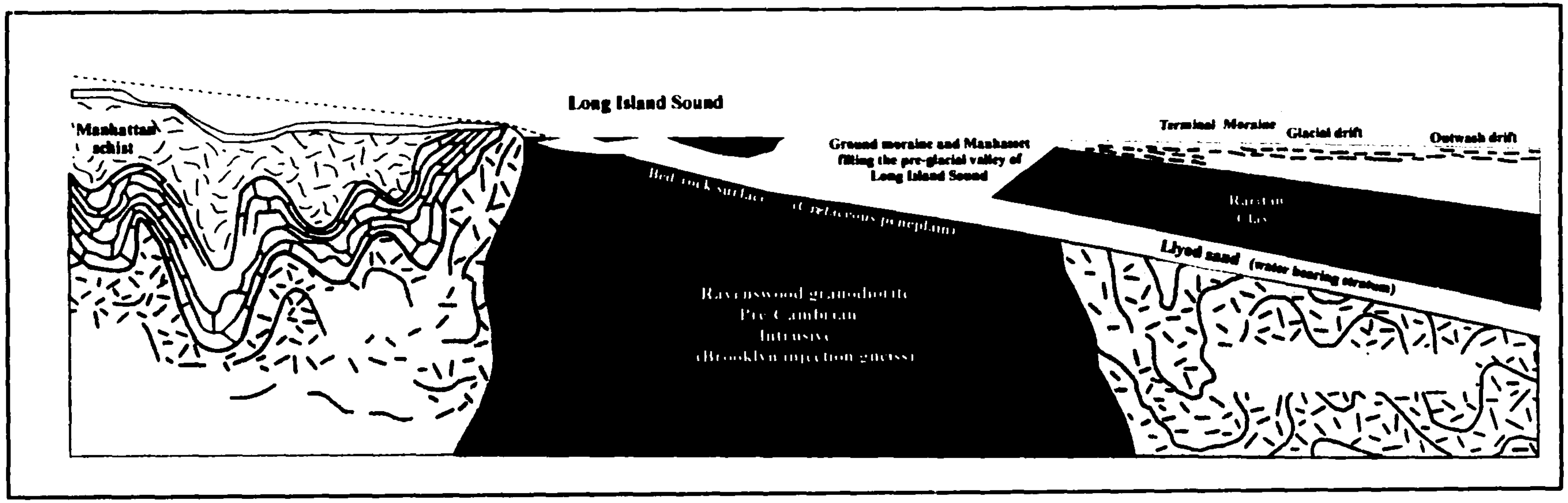


Figure 3. Generalized cross-section of Long Island across the sound into the Bronx by Berkey (ca. 1933) showing the Precambrian designation of the Ravenswood and its intrusive relationship into the Fordham Gneiss to form the Brooklyn Injection Gneiss. Notice the interpretation for Precambrian gneiss units to the east of the Brooklyn Injection Gneiss.

The remainder of the Manhattan Prong sequence comprises lower Paleozoic cover rocks; the Inwood marble is sandwiched in beds of both the East and West branches of the East River around Roosevelt Island (Figure 2). Roosevelt Island (formerly named Welfare Island and Blackwell's Island) is made of the resistant Fordham Gneiss unit B formation (Baskerville, 1992). The majority of Manhattan Island is underlain by the Manhattan schist, which in recent work is shown to be a highly variable formation with abundant gneissic sections grading to quartzite (Fettke, 1914; Prucha, 1956, see appendix for Chesman 1997).

Hall (1968) includes the allochthonous Hartland Formation (Rodgers, et al. 1959) as a third category of rocks within the Manhattan Prong based on his maps of the White Plains area, twenty miles to the north of Brooklyn/Queens. Hall cites the Hartland's similar appearance to the Manhattan Formation, its thrust contact with the Manhattan, Inwood and Fordham formations, and the presence of the Hartland in Westchester as his basis for including this unit within the Manhattan Prong.

Recent efforts have been made to reorganize the geology of New York City by extrapolating the amphibolite grade allochthonous eugeosynclinal schists, gneisses and amphibolites of the Cambrian-Ordovician Hartland formation of western Connecticut (Rodgers, et al. 1959) into Long Island and across to Manhattan (Mose and Merguerian, 1985; Merguerian and Baskerville, 1987; Baskerville and Mose, 1989; Baskerville 1992 and 1994). Emplacement of the Hartland, as interpreted by Hall (1968), occurs via thrusting along Cameron's Line. Cameron's Line, a regionally important Taconic suture, is accepted by many researchers to represent the demarcation between autochthonous parts of the Cambrian-Ordovician Manhattan Formation in the west and allochthonous Cambrian-Ordovician eugeosynclinal Hartland terrane (Hall, 1968) to the east. The Hartland proposals, although depicting different contact locations by researchers, show Hartland rocks as comprising a large portion of Manhattan Island (Merguerian and Baskerville, 1982; Baskerville 1992 and 1994).

The Hartland proposals appear to have been accepted by the U.S.G.S. without much debate or supporting geochemical evidence. This is in stark contrast to the historic debates that generally accompanied Manhattan Prong proposals. These historic debates have helped to form our present understanding of the Manhattan Prong and a brief review of the debates is presented below to assist in laying the outline of this thesis.

1.1 Uncertainty and Historical Debate

At the turn of the century the initial debate centered on the interrelationship of the Hudson Highlands, the Fordham Gneiss and the Manhattan Series (Berkey, 1907). Regional correlations of the Manhattan Prong rocks with similar series along the eastern seaboard were proposed (Schöpf 1787; Knopf and Jonas, 1929; Higgins 1972). These deliberations continued through the first half of the twentieth century with the debates turning structural in nature, concentrating on the possible existence of a depositional unconformity between the Fordham Gneiss and the Lowerre Quartzite (Balk 1936; Scotford 1956; Prucha 1956 and 1959; Norton 1959). It was felt that if a definitive stratigraphic correlation could be established between the un-metamorphosed Paleozoic Wappinger Series (Poughquage Quartzite-Wappinger Limestone-Hudson River pelites) to the north with the more highly metamorphosed Manhattan series (Lowerre Quartzite-

Inwood Marble-Manhattan Schist) to the south then an age relationship between the Fordham Gneiss and the overlying Manhattan series could be established (Balk 1936; Prucha 1956). No consensus was reached in part because of the intermittent nature of the Lowerre Quartzite.

This debate was largely settled during the second half of the century through isotopic age dating of the respective units. The Manhattan Series is established as Cambro-Ordovician in age by both fossil records within the less metamorphosed Manhattan Schist-Inwood Limestone sequence north of New York City (Ratcliffe and Knowles, 1968) and by isotopic age dating (Langer and Bowes, 1969; Mose and Merguerian, 1975). The Fordham Gneiss is established as Precambrian on the basis of various isotopic age dating methods (Long and Kulp, 1962; Mose 1982; Aleinikoff 1985). However, there remains a high degree of uncertainty as to the exact age of the Fordham Gneiss due to the complex stratigraphy, variable source rocks and retrograde metamorphic events (Blank 1973; Aleinikoff 1985).

Further south in New York City, the Ravenswood Granodiorite occupies a significant portion of lower Manhattan Island (Figure 4) and an extensive section of Brooklyn and Queens (Berkey, 1910, 1931, 1933; Baskerville and Mose, 1989; Baskerville 1994). However, the important work of Zeigler (1911) and Berkey (1910) on the Ravenswood Granodiorite, as well as Berkey's (1931 and 1933) and Blank's (1931) work on CT#2 was not taken into account in either Prucha's (1956 and 1959) designation of the New York City Group or Hall's (1966 and 1968) collective definition of the Manhattan Prong. This failure to consider the BIG Complex was made despite the regional expertise of Berkey and other authors, and the unobstructed ~11 mile geologic exposure provided by CT#2 (Figure 1).

In an effort to reestablish the Brooklyn Injection Gneiss in the geologic literature Blank (1973) provided a comprehensive petrographic and structural report on his findings as the primary geologist responsible for mapping CT#2. Of significant regional importance was his identification of the Ravenswood Granodiorite's charnockitic affinity. Unfortunately even Blank's (1973) published article has been largely overlooked and as a result a large segment of regional geology has been excluded from recent structural models for the origin of the Manhattan Prong.

1.2 The "Hartland Problem"

Toward the end of the century a revised scenario involving thrust faulting of Paleozoic meta-sediments (i.e. Hartland Terrane) from the east (Balk 1936) over the New York City Group has taken precedent (Hall 1968; Merguerian and Baskerville, 1982 and 1987). Although there are numerous exposures in Manhattan, as well as the Bronx and Westchester County, rigorous correlations of the various Manhattan units with the correlative Hartland units (Gates and Martin, 1976) have been difficult to establish. The mapped location of the thrust contact between the Manhattan Schist and Hartland Formations, termed Cameron's Line, has shifted numerous times in New York City since Hall's (1968) proposal.

Merguerian (1986), writing on a visit to NYC Water Tunnel #3 under Long Island City, describes recognition of Cameron's Line as the ductile contact between the

Hartland Formation and "coeval allochthonous transitional parts of the Manhattan Formation". Merguerian also describes a location where the Hartland is in mylonitic contact with the Fordham Gneiss. Baskerville (1992) depicts the same section of the tunnel and Cameron's Line as a contact between the Manhattan Formation and the Ravenswood Granodiorite with the Hartland Formation further to the east.

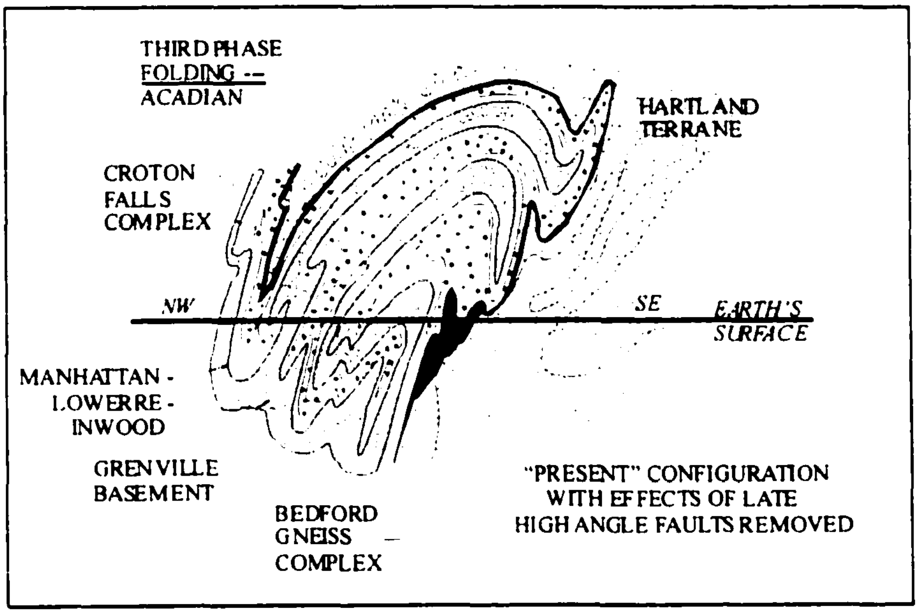


Figure 4. Schematic of Hall's (1979) 'highly interpretive' cross section illustrating the involvement of basement and cover in the Fordham Terrane near the Connecticut – New York state line. Inferences are based on Hall's work south of the Hudson Highlands.

The difficulty in establishing the Hartland correlation was cited by Hall (1968) who termed it the "Hartland problem" and may in fact be due to complex folding as proposed by Hall (1979) in Figure 4. However, the difficulty is probably best summarized by Balk (1936) in his work on the same units in Dutchess County, NY. Balk pointed out that;

> **"within non-fossiliferous formations, of similar lithologic compositions and devoid of any key stratigraphic markers, an infinite number of thrust sheets can be construed. Unfortunately, no positive proof of their existence can be given, nevertheless, it is also impossible to disprove them without reliable criteria".** (Balk 1936 pg. 766)

Even with this uncertainty Baskerville and Mose (1989) suggested that Cameron's Line follows the East River in the New York City area, and represents the easternmost extension of the Middle Proterozoic Fordham Gneiss. The authors argued for abolishing Berkey's (1931, 1933) formational name, the Brooklyn Injection Gneiss, for Roger's et al. (1959) Hartland Formation. This proposal was based on the belief that all of the rocks east of the East River were Ordovician in age, except for the possible occurrence of mafic rocks which Baskerville and Mose (1989) considered to be oceanic crust of a similar age to that of the Fordham Gneiss. Unpublished Rb-Sr isotopic data was used as evidence for an early Paleozoic age, stated as probably Early Ordovician, for the Ravenswood Granodiorite (Baskerville and Mose, 1989). These opinions ultimately lead to a USGS revision of the New York City geology (Baskerville, 1992, 1994) shown in part by Figures 2 and 5.

1.3 Recent Work

The recent excavation of New York City's Third Water Tunnel has allowed a 10.5 mile section of relatively unweathered bedrock to be mapped and systematically sampled and analyzed in order to provide some insight into the tectonic history of this extreme easterly section of the Manhattan Prong.

The author's geologic mapping of the Brooklyn and Queens section of CT#3 documents the exposure of a migmatitic terrane containing mostly orthogneiss with a subordinate amount of paragneiss units, very similar to that documented by Berkey (1910, 1933) and Blank (1931, 1973) during the mapping of CT#1 and CT#2 (see Figure 1). The finding of granulite grade metagabbro and associated enderbite[6] was used by Chesman (1989; Steiner and Chesman, 1996) to bring into question the relationship to granulite terranes that are common in the Appalachian core and Hudson Highlands but not present in the Hartland.

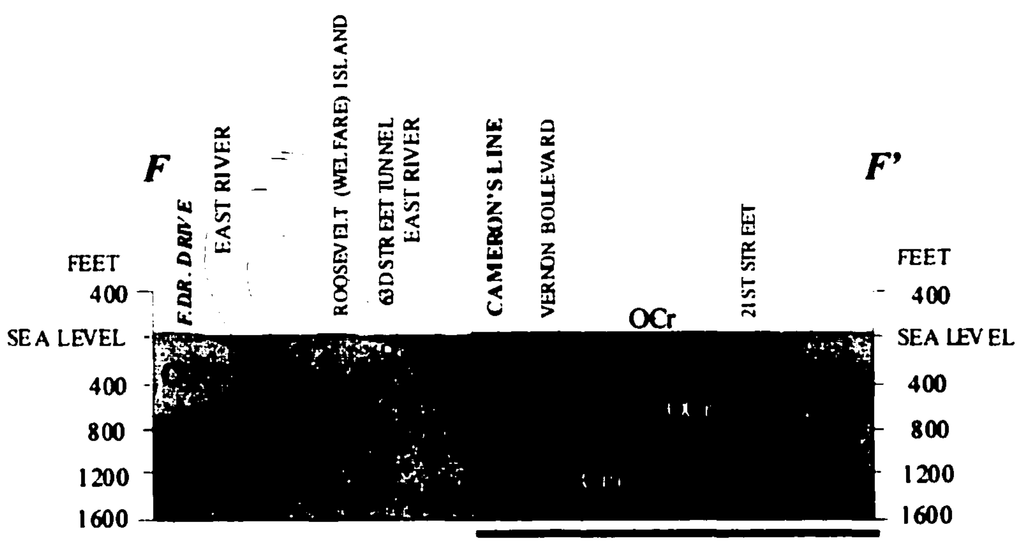


Figure 5. Interpretive cross-section from Baskerville (1992) between Long Island City, Queens and Manhattan Island. Cameron's Line is depicted as an southeast dipping thrust contact between the Manhattan schist (Cm) and Ravenswood Granodiorite (OCr).

[6] Blank (1973) notes the presence of hypersthene and clinopyroxene to indicate a charnockitic (granulite) affinity for the Ravenswood Granodiorite found in Queens and Brooklyn (pg. 647).

Extensive amphibolite grade retrogression, and metasomatism involving the mobile elements, Rb, Sr, and K (Chesman, 1997), erases much of the granulite grade assemblages and possibly alters Rb-Sr ratios. Nevertheless, the use of relatively immobile trace elements indicates that enderbites and diorites belong to the Ravenswood Granodiorite unit (Steiner and Chesman, 1996) of Kemp (1895), Berkey (1910) and Ziegler (1911). Mafic and ultra-mafic rocks with geochemical profiles consistent with a dismembered ophiolite (Steiner and Chesman, 1996) are also identified, as are fairly extensive exposures of coronitic meta-gabbros. The distinctive lavender garnet, sillimanite bearing Fordham Gneiss D unit of Hall (1968) and interbedded marbles, typical of the Fordham Gneiss (Berkey 1933; Prucha 1956; Blank 1973) has also been encountered and mapped.

Multiple injections of cross-cutting mafic dikes (see also Blank 1973) were at first considered to be associated with the Palisades Sill, but granoblastic fabrics establishes a probable pre-Tectonic event. Numerous granite pegmatites of several generations were also recognized and mapped (Chesman 1989; see also Blank 1973).

The youngest exposed rock type is a porphyritic dacite dike swarm intruded along NW striking reactivated faults (Chesman 1998). Dacites, first documented during the excavation of New York City's Water Tunnel No.2 (CT#2) in the Bronx (BWS 1932; Colony and Blank, 1943) are present in both Brooklyn and Queens.

In general, foliation of the bedrock in Brooklyn and Queens strikes N40E, 55SE varying locally due to the presence of isoclinal folds, ductile shear zones and the localized effects of faulting. Late stage deformational events resulted in the imposition of numerous brittle faults on the bedrock.

The author's preliminary studies reveals these and other inconsistencies with Paleozoic horizons in the region, and establishes instead an excellent correlation with East Coast Middle Proterozoic complexes, such as the Housatonic and Berkshire Highlands gneiss (Ratcliff and Zartman, 1976; Gates and Martin, 1976). Furthermore, the presence of granulite grade felsic rocks is, as noted, consistent with other Middle Proterozoic terranes in the region (Dodd 1965; Drake et al., 1991b; Puffer and Volkert, 1991; Ratcliff 1992). The preliminary petrologic, lithologic and isotopic age data thus strengthens Berkey's original interpretation.

In a recent abstract by Brock et al., (2001) on the rock exposed in the Queens section of the CT#3, the authors purport the finding of a new granulite grade Proterozoic orthogneiss complex beneath western Queens, naming it the Queens Tunnel Complex. Although the Queens Tunnel section closely parallels and actually crosses ~70 m below CT#2 no reference is made to either Berkey's earlier work or to Blank's (1973) petrographic study on the Brooklyn Injection Gneiss where he emphasizes the BIG Complex's granulite grade mineral assemblages. With due consideration to Berkey (1931, 1933) and Blank's (1973) establishment of the Precambrian Brooklyn Injection Gneiss, as a complex mixture of Fordham Gneiss and intruded Ravenswood Granodiorite, it is recommended that that the Brooklyn Injection Gneiss is reestablished a formational name for the bedrock of western Long Island (see Baskerville and Mose, 1989).

1.4 COURSE OF ACTION

Stratigraphic correlations and petro-genetic models are vastly improved using geochemical models proposed by Pearce (1982 and 1983) and Pearce et al. (1984a and 1984b). It is particularly important to integrate these new granulite terranes into existing models for passive margin evolution. Passive margins such as the Atlantic Continental margin are typically sites of repeated cycles of compressional orogenics and continental breakups (Bond and Kominz, 1988). In southwestern New England allochthonous Phanerozoic rocks are in thrust contact with the Laurentian margin producing a structural patchwork of various lithologies with contrasting ages, metamorphic histories yet similar geochemistries (Karabinos 2000). Further complicating the stratigraphy of the region are the remnant effects of the Elzevirian, Ottawan, Post-Grenvillian, Avalonian, Taconic, Acadian and Alleghanian orogenies, all within a relatively narrow band along the former Laurentian margin (Clark and Kulp, 1968; Hall 1968; Long 1969; Zartman et al., 1970; Mose and Hayes, 1975; Brock and Mose, 1979; Wintsch and Aleinikoff, 1987; Ratcliffe et al., 1991; Aleinikoff 1985; Karabinos and Aleinikoff, 1990; Ratcliffe et al., 1998; Karabinos et al., 1999).

In the Manhattan Prong of the Central Appalachians, Paleozoic crystalline rocks have been found to be isoclinally folded with the Proterozoic Y Fordham Gneiss (Hall 1968). Similar sequences occur further north with the coeval Berkshire (Ratcliffe and Zartman, 1976) and Green Mountain Massif (Stanley and Ratcliff, 1985) as well as the Baltimore Gneiss-Glenarm Series (Knopf and Jonas, 1929; Higgins 1972) and the Goochland Terrane (Farrar 1984) to the south.

The complicated structure and repetitive orogenics along the eastern seaboard still made it unclear if the meta-igneous rocks of the Ravenswood (Berkey 1931; Steiner and Chesman, 1996) were Proterozoic in age and coeval with the Hudson Highlands (Berkey 1910) or, if the Ravenswood was Ordovician in age (Baskerville and Mose, 1989) and therefore related to the Harrison Gneiss (Zeigler, 1911, Ries, 1898). Age dates, to be definitive, must consider the numerous deformational events and presence of Rb, Sr, and K metasomatism imposed on many of the rock lithologies.

1.5 Approach

The methods employed are: 1) field measurements and mapping, 2) standard petrographic methods, 3) electron microprobe analysis, 4) whole-rock chemical analysis, 5) trace-element analysis and 6) isotopic studies.

1.5a Field Work: As is typical with most tunnel projects, geologic mapping of the entire exposure is performed as a source of information for possible construction-related litigation. The author carried out a detailed mapping and sampling program for the City of New York, due to the anticipated mining difficulties associated with the chaotic migmatitic structure of the rock, the presence of faults and jointing as well as the hardness and stressed condition of the rock. An added benefit of the deep seated tunnel exposures is the access to fresh unweathered samples and clean exposures provided by the smooth bore of the tunnel boring machine (TBM). Mapping followed the excavation closely, with mining of the 10.5 miles taking a total of 75 months. Samples for petrographic study, microprobe and geochemical analysis were obtained from the tunnel and from exploratory drill cores conducted along tunnel line. Rock sampling occurred at 400 ft intervals and in areas of special interest. Of the 10.5 miles excavated, approximately 3.5 miles were excavated across the regional foliation with the remaining 8 miles excavated sub-parallel to foliation. Field maps of 1" = 20' were made for the first 5.5 miles and field maps of 1" = 10' were made for the last 5 miles. The author performed all maps and sampling.

1.5b Petrography: Some 200 thin sections were studied by standard petrographic methods in order to more accurately classify the rock based on mineralogy and texture. These features can have a profound effect on the rocks' hardness and abrasivity, characteristics important in TBM mining.

1.5c Micro-probe analysis: Twenty-five polished thin sections were prepared for electron microprobe analysis. Analyses were performed at the American Museum of Natural History using a Cameca electron microprobe equipped with three spectrometers employing a 15 kilovolts, 0.05 microamperes sample current, and a beam size of two to four um. Raw data was corrected for atomic number, fluorescence, and absorption effects.

1.5d Geo-chemical analysis: Two hundred and eighty samples were obtained for whole rock and trace element analysis. Activation Laboratories LTD. of Ancaster, Ontario, Canada, performed the geochemical analysis. The whole rock analysis was performed by Fusion-Inductively Coupled Plasma Emission Spectrometry (ICP) for the major oxides, all with 0.01% detection limits, plus trace elements Ba (2 ppm), Sr(2 ppm), Y(2 ppm), Zr(2 ppm), Sc(2 ppm), Be(1 ppm) and V(5 ppm). Research grade ICP/ Mass Spectrometry analysis was conducted for trace elements and REE for one hundred samples.

Several whole rock analyses were performed on a Phillips 1450 X-Ray Fluorescence machine following the procedures of Norrish and Hutton (1969) at City College for comparison to the values produced by the Fusion-ICP method. Several samples were analyzed for REE at Oregon State University by instrumental Neutron activation analysis and a comparison was made to the research grade ICP/MS analysis. No major deviations were noted for either method.

1.5e Age Dating: One sample of enderbitic gneiss was sent out to Geochron Laboratories of Cambridge, MA for U-Pb dating of three zircon grains. Mineral separation and preparation (air abrasion) was conducted by Geochron Labs. (Sam Bowring with three points on a Concordia diagram with air abrasion of minerals.) One sample of the dacite was also sent to out Geochron Laboratories of Cambridge, MA for a single zircon analysis.

2. ISOTOPIC DATING RESULTS

An unweathered massive enderbite (Sta. 280+50 at the Brooklyn/Queens border) with occasional angular mafic (1 meter meta-diorite) blocks, typical of the volcanic arc Ravenswood Granodiorite, as defined further along, was chosen as representative of the felsic intrusion that still incorporates oceanic mafics. A minimum age of the BIG Complex is thus obtained by employing U-Pb zircon analysis of non-regressed volcanic arc enderbite. However, as will be discussed further in Section 5.2b, the calc-alkaline signature of the enderbite may indicate that this intrusion, and thus zircon age date, may be an early differentiate associated with the initial development the Ravenswood volcanic arc.

The zircons extracted from the enderbite are elongate subrounded grains (Fig. 6). These physical features are common in metamorphic zircons and are similar to zircons found in deep seated mantle rocks of the Zabagard Basin (Seyler and Bonatti, 1988).

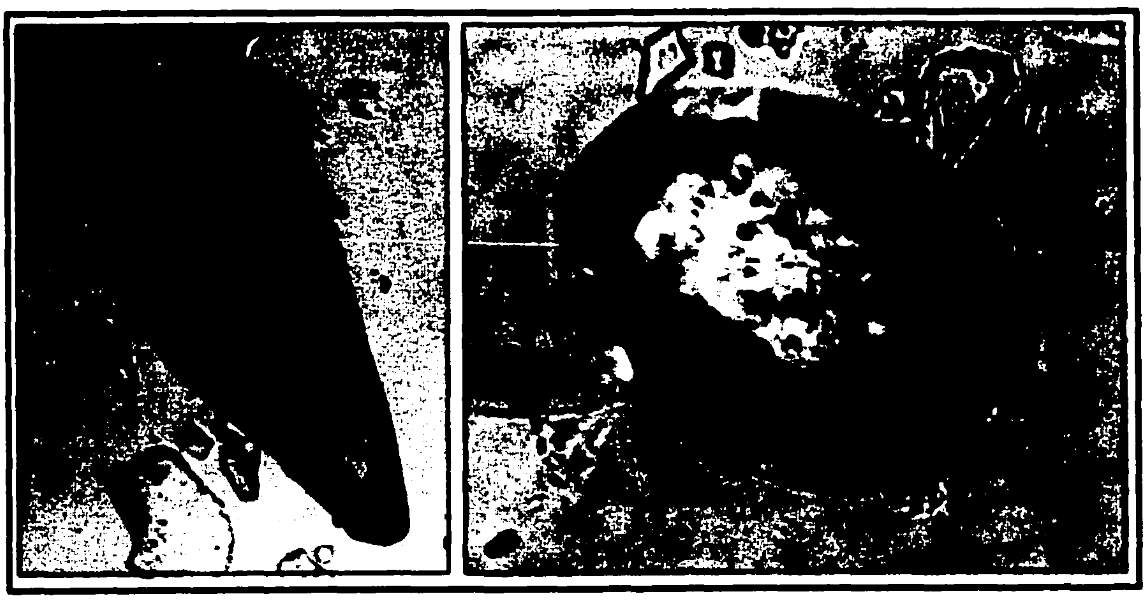


Figure 6. Cross-polarizer transmission (left) and plane-polarized transmission (right) of zircons from the BIG Complex enderbite.

Samples were separated, air-abraded and analyzed by Dr. Sam Bowsring at Geochron Laboratories, Massachusetts. Concordia diagram (Figure 7) shows the results for three zircons. The zircons have Pb^{207}/Pb^{206} values of 1032.8, 1060.4 and 1064.5. Bowsring considered Z_2, with~6.6% discordancy, as the best estimate for the calc-alkaline enderbite's age at 1032 ± 4My. Although not discussed in the analysis provided by Bowsring, the older Pb^{207}/Pb^{206} values may indicate inherited components.

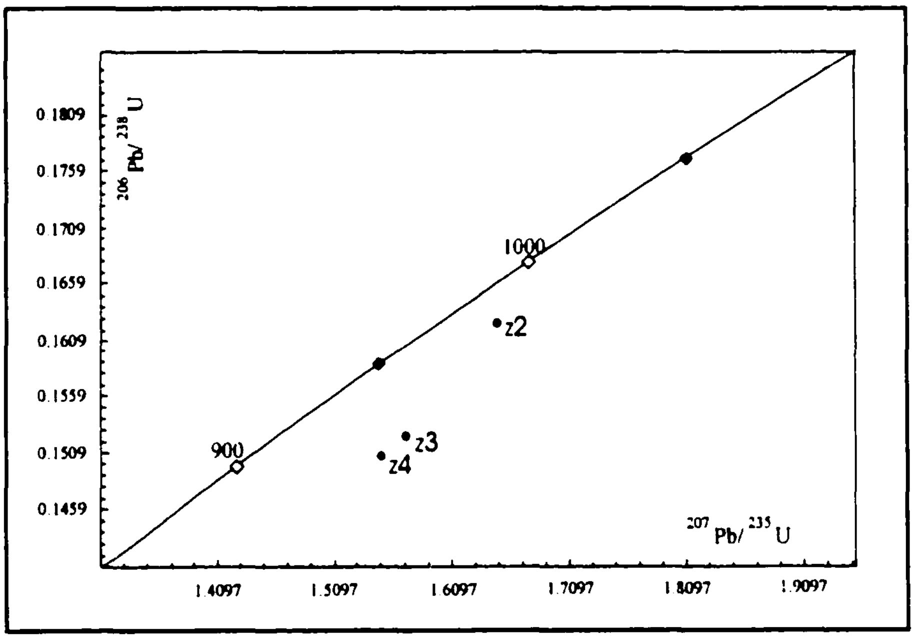


Figure 7. Concordia diagram for zircons from a BIG Complex calc-alkaline enderbite along the Brooklyn- Queens border.

Following the above dating, the tunneling contractor for the Queens section of CT#3 contracted the services of Professors P.C. Brock, P.W.G. Brock and C. Merguerian to select samples for dating in preparation of a geologic change of condition presentation (Grow-Perini-Skanska Joint Venture (GPSJV), 2000). Two samples of tonalitic gneiss were sent to S. Bowsring of MIT for separation and analysis of five zircon fractionations for each sample. Bowsring determined an inherited component for the samples ranging from 1070-1092 Ma. With one sample having three slightly discordant zircons ranging in age form 949 to 1007 Ma. Three zircons from the second sample had Pb-Pb ages ranging from 985-1018 Ma (GPSJV 2000; Brock et al., 2001).

Thus the above zircon dates support Berkey's (1933) original deduction that the Ravenswood is Precambrian in age and the present arguments that the Ravenswood cannot be assigned to the Hartland Formation. Although the Ravenswood intrusive suite proves to be Proterozoic in age, the Pb-Pb dates are younger than the calc-alkaline Losee Metamorphic Suite and the Byram-Lake Hopatcong Intrusive Suite of the New Jersey Highlands (Puffer and Volkert, 1991; Volkert 1995; Ratcliffe personal communication, 2000) as well as the Hudson Highlands' Storm King granite and Canopus pluton dated at ~1135 My (Ratcliffe and Aleinikoff, 1990, Ratcliffe personal communication, 2000). Similar metamorphic ages and offsets to those of the Highlands and the Ravenswood Suite are recorded across the Carthage-Colton shear zone, located near the eastern margin of the Grenville province in northern New York, which separates the Adirondack Lowlands of the Metasedimentary Belt from the Adirondack Highlands of the Granulite Terrane (Streepey et al., 2000). The age of the Ravenswood volcanic arc enderbite is consistent with the zircon U-Pb geochronology of the Ottawan Orogeny in the Adirondack Highlands (McLelland et al., 2001) and the intrusive Tyringham Gneiss of the Berkshire massif (Ratcliffe and Zartman, 1976).

Regional Proterozoic Y intrusives that are similar in age to the Ravenswood include the undeformed Mt. Eve granite in the New Jersey Highlands, which provides a zircon concordia upper intercept of 1,020 ± 4 (Drake et al., 1991) and a pegmatite within the Canada Hill Granite of the Hudson Highlands that yielded an U-Pb upper intercept of 1010 ± 6 (Aleinikoff and Grauch, 1990, reported in Rankin and others, 1991).

Thus, the age of the Ravenswood's volcanic arc suite as compared to other Proterozoic Y intrusives of the NJ and Hudson Highlands, along with its oceanic geochemical signatures (discussed below) and relationship to the Fordham gneiss provides important evidence for constructing a tectonic scenario for the Proterozoic Y BIG Complex of the Manhattan Prong. A petrographic investigation of the various rock types found in the BIG Complex will be presented first, followed by geochemical studies.

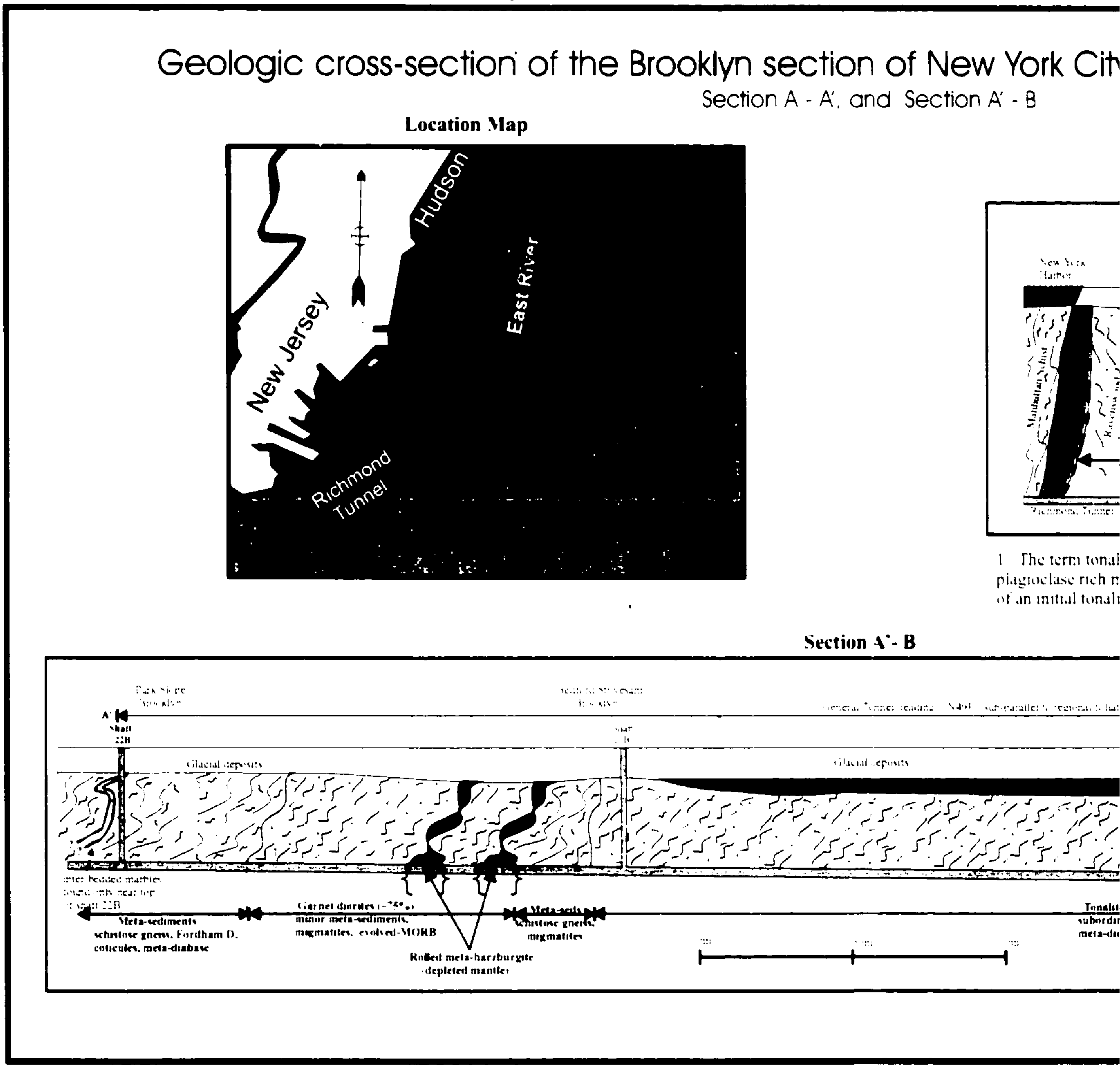
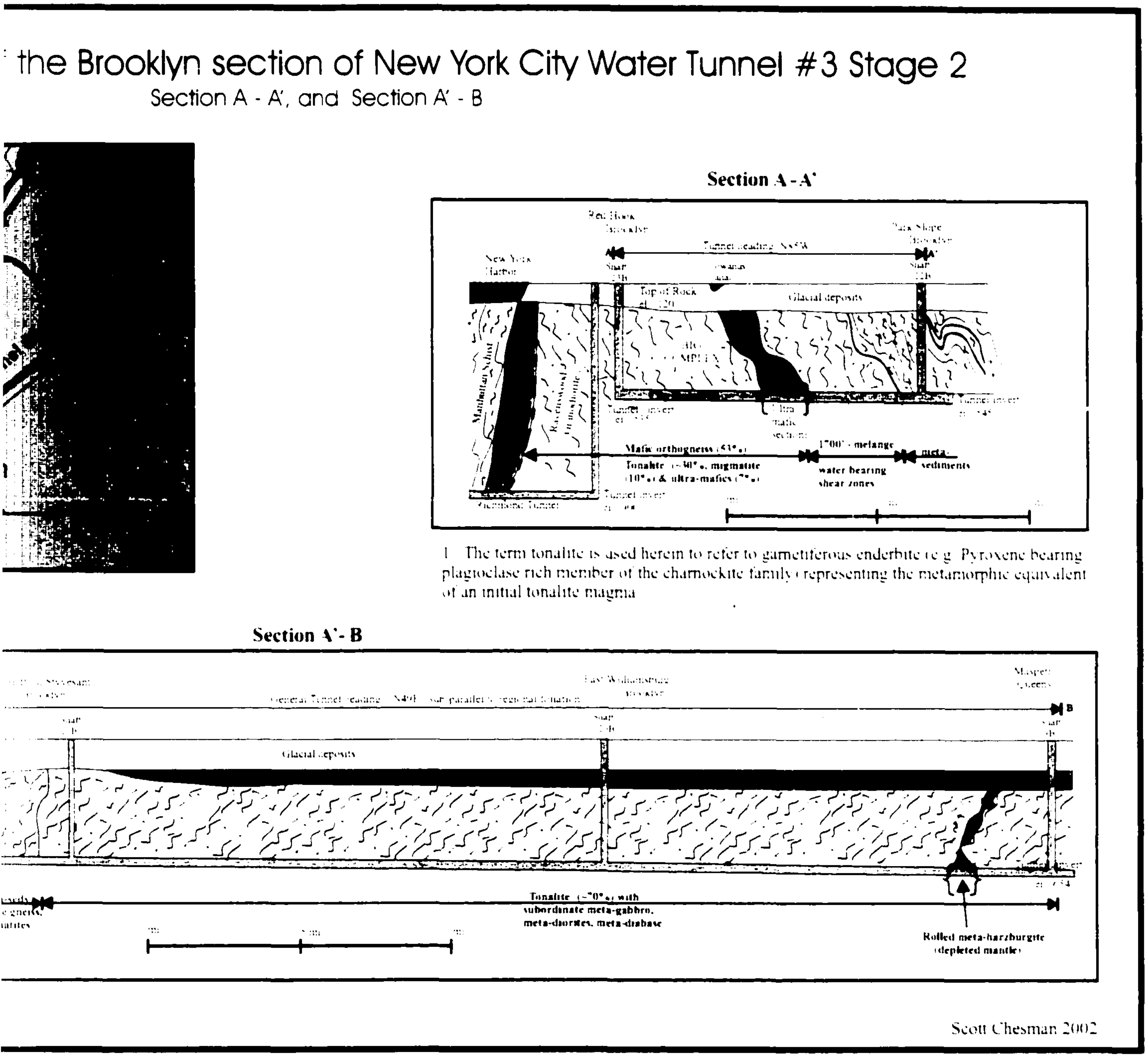


Figure 8. Generalized geologic cross-section maps of CT#3, Stage 2.



: cross-section maps of CT#3, Stage 2.

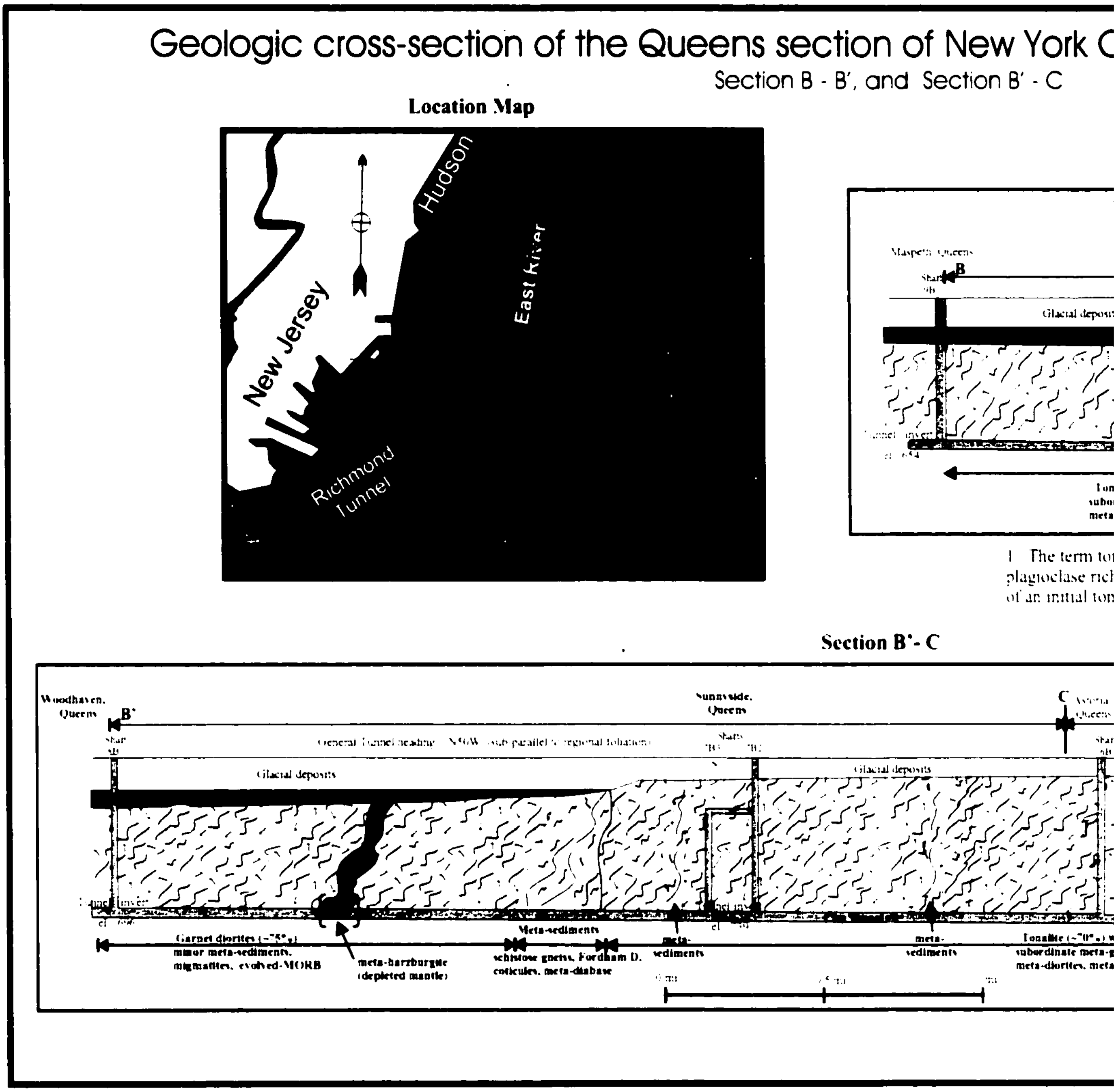
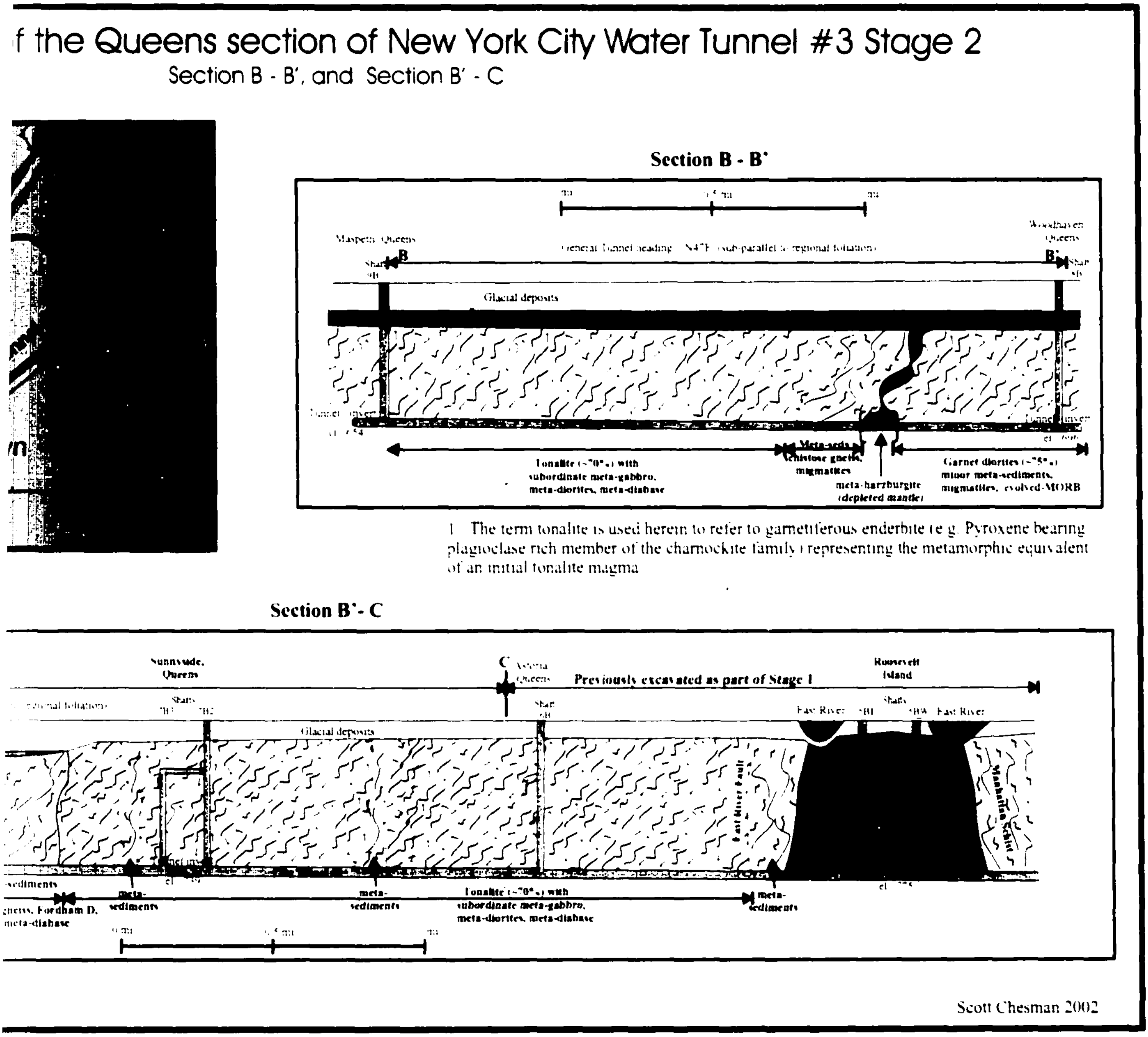


Figure 8 (continued). Generalized geologic cross-section maps of CT#3.



ılized geologic cross-section maps of CT#3. Stage 2.

3 PETROLOGY

The BIG terrane is a complex admixture of layered gneiss and migmatite. Gradations in gneissic/migmatitic layering occur every few hundred feet and distinctions between gneissic textural varieties are not made in the present study. This lithologic complexity, characterized by what are classically termed lit-par-lit injection phenomena, influenced Berkey (1931) to suggest the name Brooklyn Injection Gneiss (BIG) to encompass the entire region from western Long Island to the East River, Brooklyn and Queens.[7] The BIG migmatites (see Figure 9) are distributed reasonably uniformly in the CT#3 exposure, displaying at point-to-point all of the textural variability outlined by Mehnert (1968), as characteristic of migmatites as a class of rocks. As a possible exception, migmatites associated with metasediment sections tend to be represented by stromatic migmatite. Although no isotopic age dates are presently available for these latter migmatites it is considered here that emplacement of the Ravenswood differentiates resulted in hydration of the intrusions' borders and migmatization (partial fusion) of the Fordham sediments in this locality[8]. Additional migmatization may have occurred during the collapse and deformation of the Fordham back-arc basin at the end of the Grenville Orogeny.

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[7] In this study the Ravenswood suite is defined as any felsic or mafic subunit (diorite or tonalite) > 3 meter exposure along the tunnel line that exhibits a geochemical signature, discussed farther along in Section 5.2b, consistent with a volcanic arc suite.

[8] Meta-basites show a tendency to develop stromatic structures along borders and mafic sills within meta-sediment and tonalite often exhibit surreitic (dilation) structures; similarly, schlieren appears to be more common in the meta-sedimentary Fordham D units and ptygmaticd structures may be more-associated with K-metasimatized tonalite.

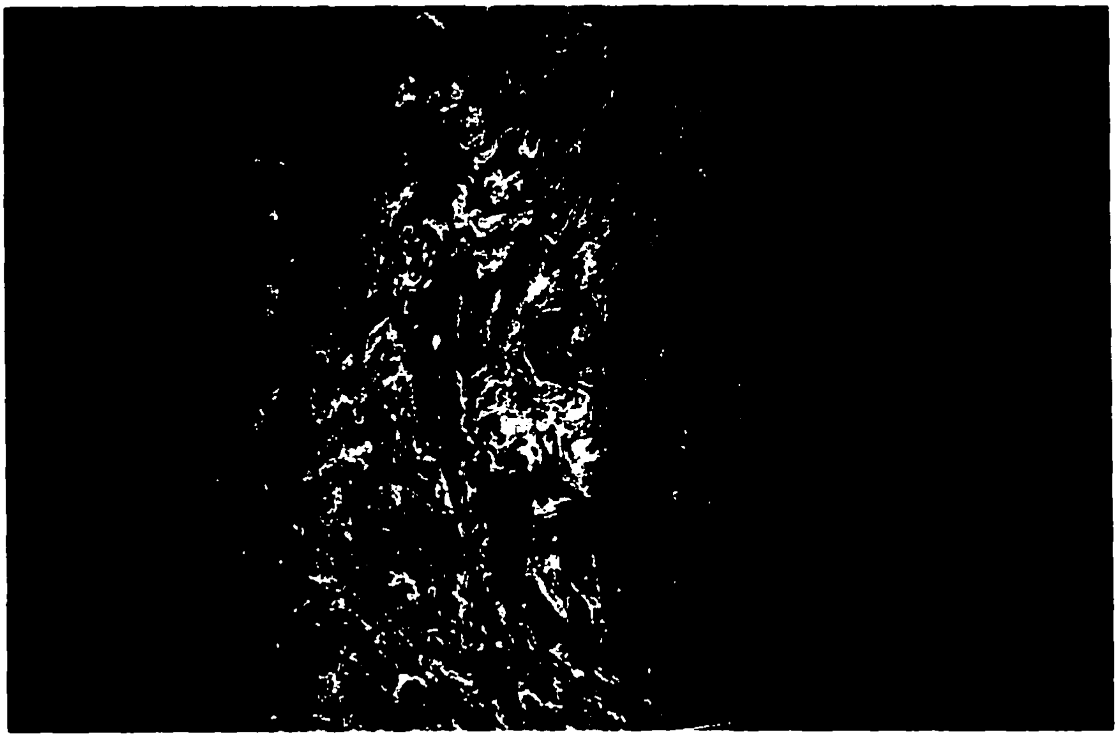

Figure 9. Complex structure of a migmatitic k-metasomatized gneiss beneath the Ravenswood section of Queens, where the introduction of potassium is associated with a pronounced reddish coloration of these units.

The percentages of BIG Complex rocks is estimated by recording percentages exposed along the ~20,000 linear feet of the tunnel line in the section which crosscuts the regional formation at close to ninety degrees (cross-sections A-A' and B'-C of Figure 8). By this method, approximately three percent of the BIG Complex comprises an ultra-mafic section (Figure 10) representing probable relict mantle; thirty percent is mafic orthogneisses (gabbro to diorite); twenty-six percent is tonalite[9]-[10] and eighteen percent is

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[9] Tonalite is used herein to refer to garnetiferous enderbite, e.g. metamorphosed tonalite - a pyroxene-bearing plagioclase-rich member of the charnockite family with 62<silica percent < 80 range.

[10] Diorite: the term is used here to signify metamorphosed garnet diorite with silica less than 52%

migmatite of uncertain origin. Both the mafic orthogneiss and tonalite are distributed throughout the tunnel as lenses, banded gneiss and mafic enclaves. These are classified as Ravenswood or Ravenswood Island arc suite provided that they exhibit a trace metal signature consistent with an island arc origin (see discussion, section 5.2b and 5.2c).

At this juncture, it is not possible to determine the extent to which meta-sediments (ca. 18 percent of the tunnel line) belong strictly to the older Fordham Gneiss versus later marginal basin and volcanoclastic deposits associated with the formation of the younger Ravenswood volcanic arc. Therefore, for the purposes of this study, all meta-sedimentary units are classified as Fordham Gneiss. The remaining five percent of the BIG Complex is comprised of various igneous intrusives, meta-diabase, porphyritic dacite and pegmatites, which cross-cut the regional foliation.

It is presently maintained that mafic lithologies may be included with the Fordham Terrane if they appear to represent remnants of depositional horizon onto which meta-sedimentary Fordham Gneiss units were deposited. Due to the intimate nature of the associations revealed in the CT#3 sequence, and the clear affinities of the mafic rocks with possible mafic basement material, it is presently assumed that this correlation is warranted and that certain mafics are basement associations of the Fordham Group in general and the BIG Complex in particular.

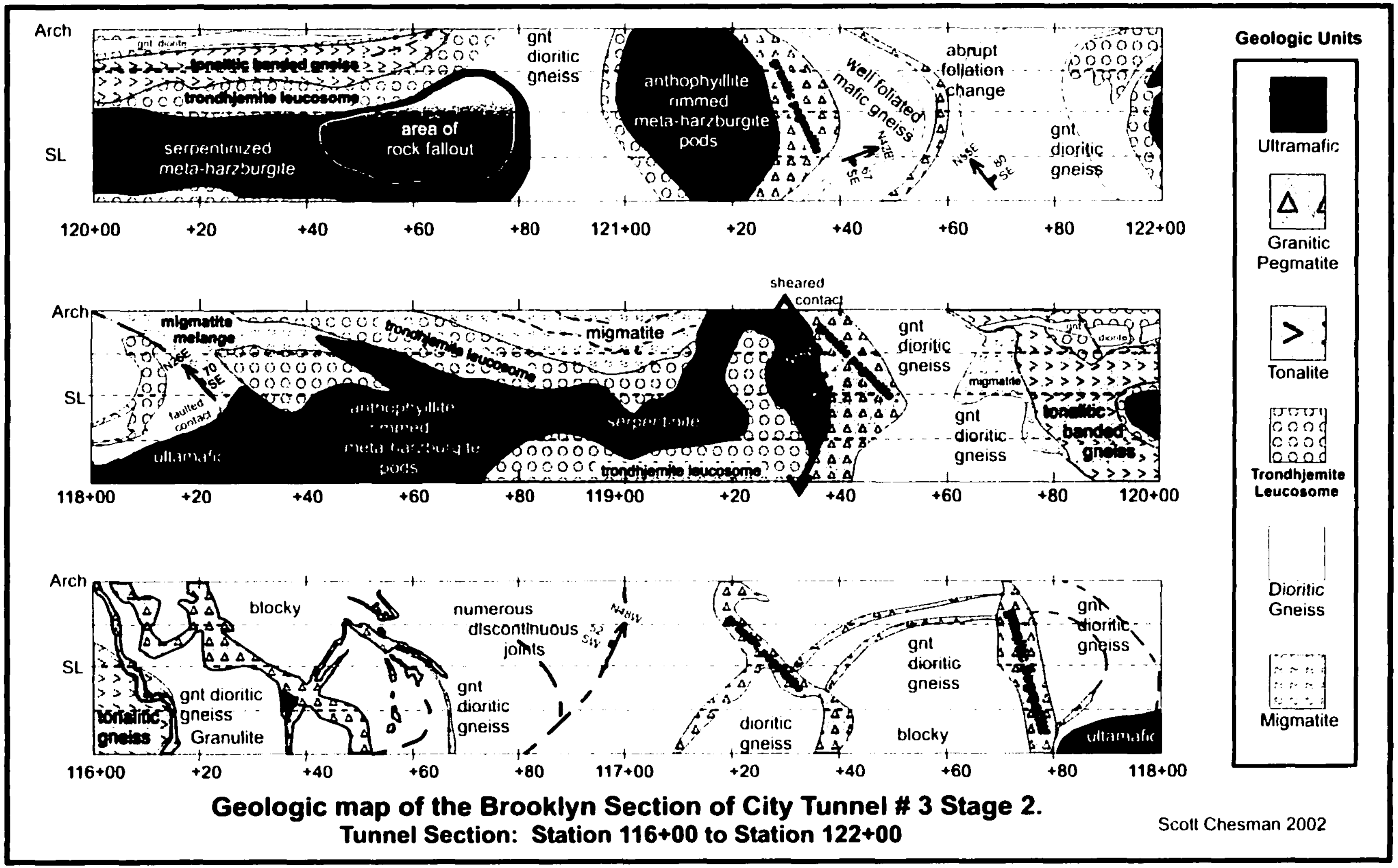

Figure 10. Geologic cross-section map displaying the relationship of meta-harzburgite (depleted mantle rocks) with other units of the BIG Complex.

3.1 Ultra-mafics and evolved MORB

Figure 11. Photograph of the tunnel exposure showing ultra-mafic xenoliths of the Fordham terrane that have been incorporated within a Ravenswood plutonic dioritic orthogneiss.

Mantle peridotite[11] is spatially associated with garnet meta-diorites within the tunnel, either as distinct bodies or xenolithic inclusions in various rock types (see Figures 10 map and 11). These units are interpreted as a mantle tectonites of metamorphic harzburgite with ophiolitic affinities (Steiner and Chesman, 1996). The presence of tectonites is typical of Ordovician ophiolite complexes along the eastern seaboard (Williams and Talkington, 1977). Steiner and Chesman (1996) considered the

[11] Using criteria established and discussed by Seyler and Bonatti (1988) to define mantle derived rocks.

deformation of the ultra-mafic bodies to have occurred in a manner similar to that proposed by Hooper & Hatcher (1989) for the Berner mafic complex (see Figure 50), in central Georgia, U.S.A. Hooper & Hatcher propose a complex distribution of plutons and pluton fragments that experienced varying degrees of hydration and internal strain during deformation with ultramafic pods displaying a preferred long dimension parallel to the local foliation trend.

The ultra-mafic BIG Complex tectonites are altered to serpentinites with cross-veins and borders consisting of anthophyllite. Most tectonites appear as rolled xenoliths with surface layers comprised of hydrated micaceous rims. Nevertheless many of the pods have cores of relatively unaltered harzburgite that contain olivine ($fo_{76}\,fa_{24}$) and enstatite ($en_{76}\,fs_{24}$) as well as granulite grade pargasite (Steiner and Chesman, 1996). Geochemical arguments related to partial melting will be discussed elsewhere (section 5.2e) that support the designation of these harzburgites as rolled melted residues. This depleted mantle model parallels that suggested by Seyler and Bonatti (1988). The Ravenswood associations, based on geochemical discussions in a later section, are probable derivatives of this melting event (hydrous melting of depleted wedges above subduction zones after the models of Seyler and Bonatti, 1988; and Davies and Stevenson, 1992). Details of the phase chemistry that establish the regional geothermometry are given elsewhere (Steiner and Chesman, in prep); the metamorphic grade based on mineral facies is discussed in section 3.2 (metasediment) and 3.3 (tonalite). From a petrographic standpoint, Spear (pg. 426, 1995) demonstrates the presence of OPX and CPX as an indicator of granulite facies conditions (using a quartzfayalite-magnetite (QFM) buffer) in metabasites although there is an ongoing discussion of the correct usage (see Newton 1992).

The granulite grade meta-diorites comprise highly deformed two-pyroxene garnetiferous gneisses that possess evolved MORB geochemical signatures believed to correspond with middle Proterozoic oceanic crust. Granulites (Figure 15B) exhibit mosaics of pyroxenes, plagioclase, in some instance brown amphibole and abundant almandine garnets. The pyroxene compositions are typically those of granulite grade lithologies (Figure 12). The Brooklyn tunnel exposure of this unit is shown in Figure 13.

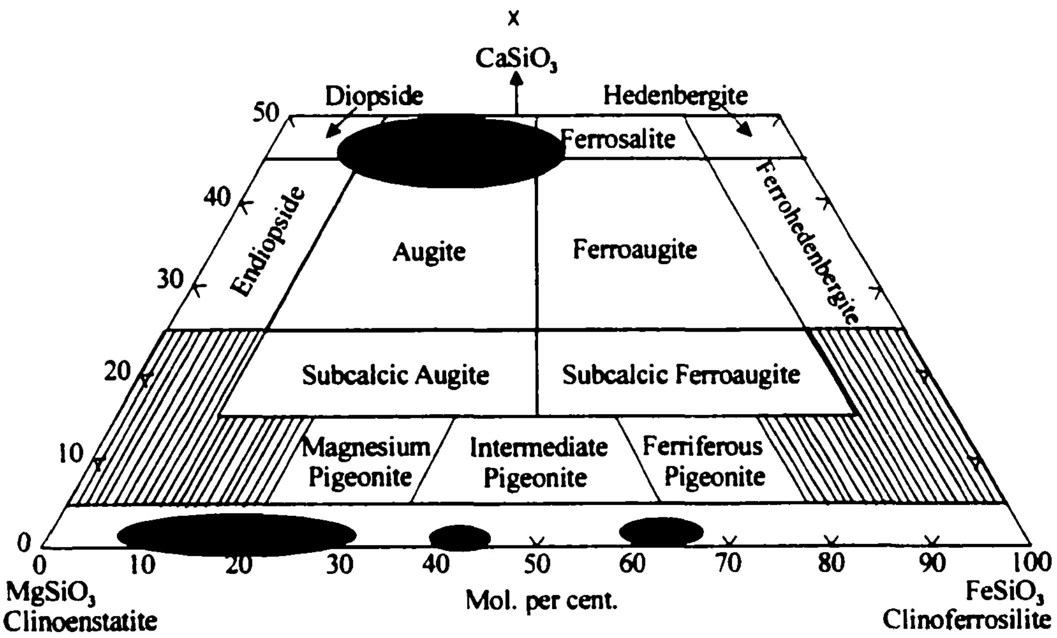

Figure 12. Pyroxene composition chart for Fordham terrane orthogneiss samples. General distribution of pyroxene compositions are typical of granulite grade associations (Spear 1995).

Figure 13. Exposure of highly deformed garnetiferous mafic granulite at tunnel Station 115+00 of the Brooklyn section of CT#3.

The relationship between the ultramafic horizon and the associated lithologies (Figure 10 map) shows that the assemblages are rolled (top panel), the regular bodies included within trondhjemite leucosome and this overall 'migmatite' is bordered by a subsidiary tonalite with a stromatic migmatite structure (migmatite: Figure 10 map). This harzburgite section is abutted by dioritic layered gneiss relative to other tunnel gneisses as highly garnetiferous, non-coronal, two-pyroxene granulite (see figures 13 and 15B). The dioritic gneiss is invaded by complex sets of pegmatite veins that appear to reflect structural controls as opposed to trondhjemitic leucosomes of the harzburgite section. As per the cross-section map, the diorite is defined herein as containing 20 to 40% hornblende as defined by IUGS classifications.

3.2 Metasediment

Figure 14. Complicated structure of the BIG Complex meta-sedimentary units exposed by CT#3 beneath Queens.

Extreme metamorphic grade and the compositional similarity, make it difficult to designate proximal well-foliated bearing units (see Figure 14) as meta-sediments (greywackies), in the absence of the development of aluminosilicate minerals, with well foliated Ravenswood. This is additionally complicated by the widespread migmatization of meta-sediments by the intrusion of the Ravenswood Suite.

In some sections, relationships are better established where two-pyroxene granulites appear to be conformable with garnet-sillimanite-k-feldspar-biotite gneisses of Hall's (1968) Fordham Gneiss Unit D. The garnets in particular have the distinctive lavender color used by Hall to distinguish this lithology although the garnets' major element chemistries are not markedly dissimilar to all other units in the Brooklyn Gneiss Complex. Sillimanite is fibrolitic in some thin sections. The K-feldspar –sillimanite – garnet – andesine – quartz – biotite – magnetite assemblage exhibited by the meta-sedimentary units (Figure 15C) are consistent with granulite facies for pelitic rocks of the Buchan Series (Raymond, pg. 553, 1995).

Several interbedded diopsidic marbles, each approximately 2 meters in width, are included within the paragneiss, common accessory minerals include diopside, scapolite and phlogopite (Figure 15D). Garnet rich bands are encountered in all units and the term coticules has gained acceptance for these units (Robinson 1976; Merguerian 1981; Ratcliffe, person comm.) despite the fact that the garnets are not spessartine as required by the initial definition (Baskerville et al., 1987).

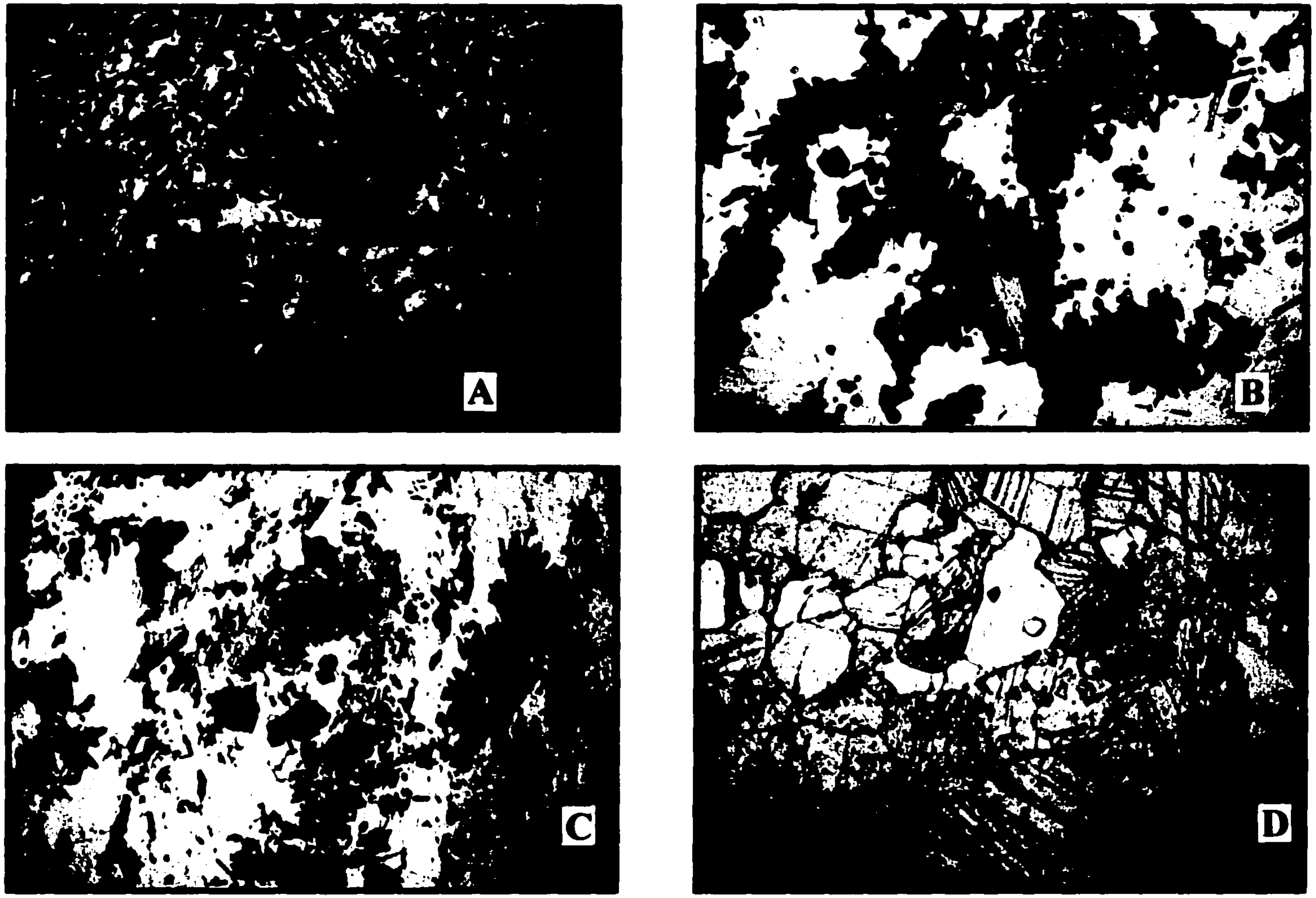

Figure 15. Microstructural textures of BIG Complex. A) Partially serpentinized depleted mantle sample showing abundant opaques; B) Two pyroxene gneiss; C) K-feldspar-sillimanite-garnet-biotite gneiss of the Fordham; D) Scapolite-diopsidic interbedded marble of the Fordham..

3.3 Ravenswood Volcanic Arc Suite

The definition of the classical Ravenswood Granodiorite (Zeigler 1911) is based on very areally restricted surface outcroppings in Long Island City, Queens. The pinkish K-feldspar bearing granodiorite falls within the Streckeisen (1976) definition, e.g. 10 to 35% modal K-feldspar and > 20% modal quartz. However the extensive CT #2 and CT#3 exposures show clearly that A) the suite is bimodal comprising gneisses and migmatites as well as mafic units, B) the primary (protolith) mafic rock is either gabbro, diorite (<52% silica) or tonalite. The pinkish cast derives from partial K-metasomatism that occurs throughout CT#2 and CT#3 and which happens to be accentuated in the East River section where the k-metasomatized rock is exposed at the surface.

As will be seen on evaluation of the trace metal signatures, the vast majority of the mafic and felsic lithologies possess a volcanic arc signature. If the term 'Ravenswood' is reserved for this suite of rocks, it therefore represents the general rock encountered in CT#2 and CT#3. A working definition is therefore proposed for Ravenswood lithologies within the BIG complex as illustrated in Table 1. Rocks that fail to exhibit an island arc signature as discussed below are termed 'BIG-diorite, -tonalite, etc., as are rocks that have not been analyzed for trace metals, as opposed to the term 'Ravenswood-diorite, -tonalite, etc. for analyzed rocks with an island arc signature.

The BIG complex is variably metamorphosed from upper amphibolite to granulite grade (Steiner and Chesman, 1996), but in places, the amphibolite represents retrograded granulite. The margins of the felsic orthogneiss bodies typically exhibit alternating bands of mafic and felsic compositions which mimic the classic appearance of the meta-sedimentary Fordham Gneiss (Berkey 1931; Prucha 1956). In some sections, it is apparent that the banding is due to penetrative injections of late stage fractioned tonalites within the margin of mafic bodies.

	< 20% > quartz	
Volcanic arc signature	Ravenswood Diorite	Ravenswood Tonalite
Non-volcanic arc signature	BIG Complex Diorite	BIG Complex Tonalite

Table 1. Definition of the Ravenswood members of the BIG Complex.

Although in a classic sense anorthosite is defined as plagioclase-enriched (> 90% plagioclase, or in other places as tonalite with <25% mafic minerals (IUGS classification, see, for example, Philpotts, 1990) most modern workers recognize tonalite, as distinct from anorthosite, as a class of large igneous rocks formed by the partial fusion of basaltic /gabbroic magmas "that pond and crystallize at the base of the crust (underplating)" pg. 335 in Winter (2001). Numerous experimental studies show that tonalite melts can in fact be produced in the laboratory by fusion under hydrous conditions (Green and Ringwood 1968, and references cited in Winter (pg. 335, 2001). It is therefore now customary to extend the usage of the term tonalite to silicic rocks (less than 25% mafic minerals) comprised almost entirely of plagioclase and quartz.

In other sections, the peripheral borders of both mafic and felsic plutonic rocks display well-developed penetrative foliations that are concordant with the local trends of the both the Fordham ultra-mafic series and Fordham meta-sediments (see also Blank 1973). Whether these fabrics were due to igneous injection (Berkey, 1931) or post emplacement heterogeneous deformation along the margins could not be determined due in part to the lack of relict contact aureoles. These features or lack of them creates difficulty in distinguishing banded Fordham Gneiss from injected Ravenswood leading Berkey (1931) to refer collectively to this migmatite as the Brooklyn Injection Gneiss.

Blocks of Fordham paragneiss and meta-harzbergites in some meta-gabbros and enderbites exhibit relict cumulate textures. In other blocks, leucosome segregations border the rafted meta-harzburgites. These features suggest that intrusion of the Ravenswood was facilitated by magmatic stoping as suggested by Berkey (1931). Alternatively, other plutonic bodies have elongated to angular inclusions of mafic rocks and rolled clasts of green pyroxenite (harzburgite). These rolled clasts do not demarcate any major tectonic marker[12], but are considered mantle clasts or older ocean crusts that were encapsulated by the volcanic arc differentiates during ascent (Figure 11).

3.3a Ravenswood Mafics: Coronitic meta-gabbros and meta-diorites.

Relict igneous textures are evident in less deformed gabbro bodies. Some sections exhibit relict cumulate textures with associated interstitial chromite (Figure

[12] Stanley and Ratcliffe (1985, citing Zen and others, 1983) consider serpentinized ultramafic fragments in the middle Proterozoic Berkshire massif of western Massachusetts as intrusive plugs, while elongated older ultramafics in the Cambro-Ordovician Rowe Schist are considered fragments of ocean crust that act as fault markers.

16A). Exsolution lamellae of orthopyroxene within clinopyroxene are also present. Deformed pyroxene grains indicate high strain rates after cooling.

Garnet coronas around pyroxene constitute a striking petrologic feature of a majority of the mafic orthogneiss (Figure 16C). The coronitic textures are petrographically similar to those found within the Adirondack Grenville terranes (Luther 1976; Whitney and McLelland, 1983; Kretz et al., 1989) and the Goochland Terrane (Farrar 1984). As in these provinces, retrograde green amphibole replaces the pyroxene along the garnet borders (Figure 16C). Replacement of the pyroxene by amphibole (amphibolization) is more pronounced in diorites often leaving only residual traces at amphibole cores. Garnets are chemically unzoned as is typical within other rocks of the BIG Complex (Steiner and Chesman, in prep). Occasional samples reveal biotite plus pyroxene associations indicating isolated cases of substantial retrograde conditions, perhaps along fracture systems.

3.3b Ravenswood Tonalite

The protolith for the felsic garnet enderbite orthogneiss is a presumed tonalitic fractional melt (see references in earlier section 3.3). In places it shows residual igneous texture but it predominantly exhibits more-or-less granoblastic texture. K-feldspar is abundant only in later pegmatite intrusions or where authigenic K-feldspar is introduced by metasomatic fluids (Chesman, 1997). Granulite grade myrmekite (Olsen and Kohlstedt, 1985) is often associated with strained and kinked plagioclase grains (Figure 16D). The granulite show ribbon quartz, common antiperthitic exsolution and other features of granulite grade rock.

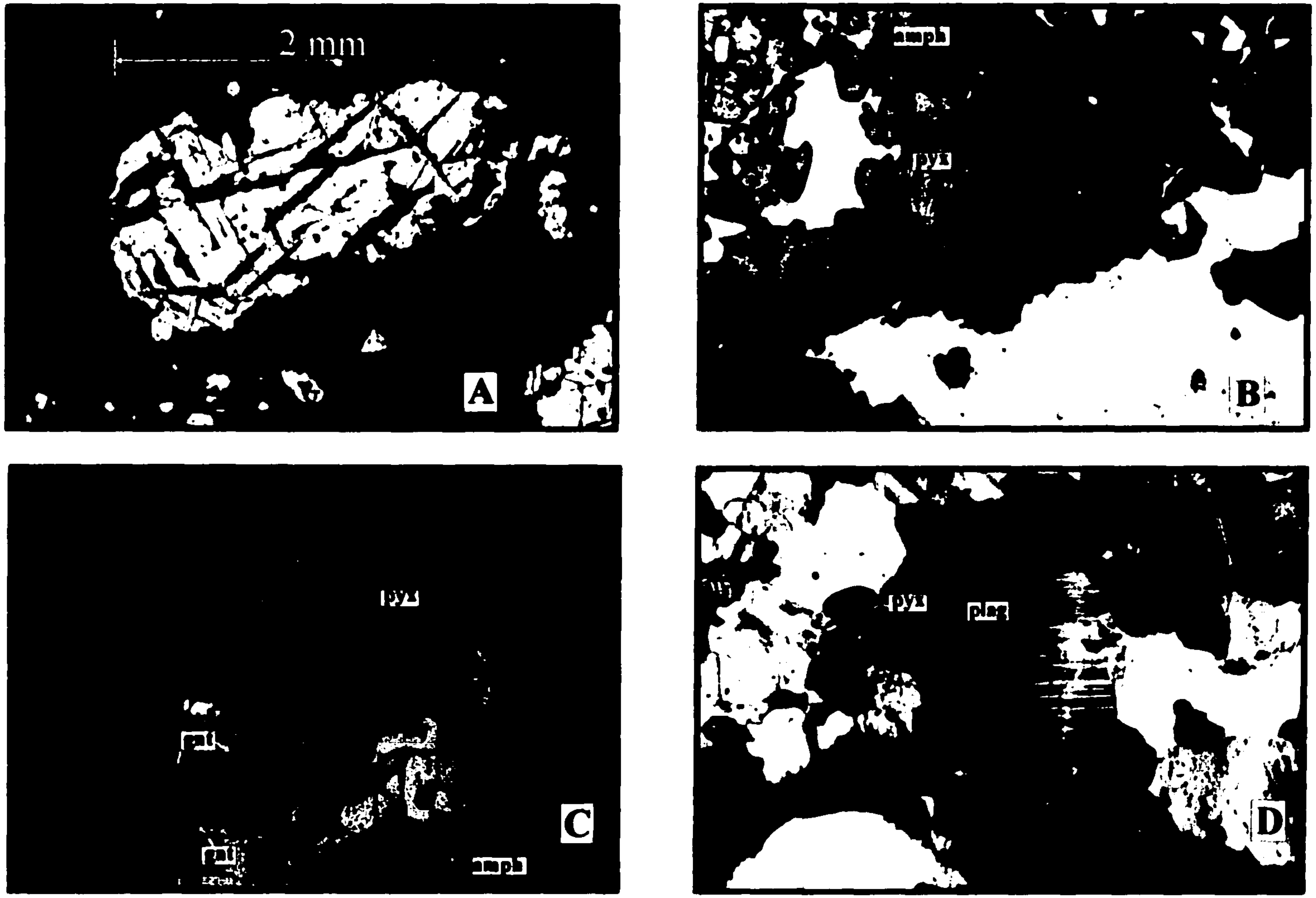

Figure 16. Microstructural textures of BIG Complex. A) SEM figure of chromite grain in a cummulate gabbro; B) Strained pyroxene grain in gabbro; C) Coronite texture of garnet rimming pyroxene and retrograde amphibole; D) Strained quartz grain in enderbite.

Enderbitic plagioclases exhibits reverse zoning and core sections that average An34 (Chesman 1997). A garnet bearing charnockitic granodiorite, a presumed calc-alkaline granodiorite protolith, proximal to the tonalites, exhibits unzoned more sodic plagioclase of average An28. This proximity indicates that compositional zoning is a relict igneous feature that resists homogenization during granulite grade metamorphism. Chesman (1997) uses this feature to discriminate among felsics in order to define different intrusive events. The trace element chemistry however, indicates that they are petrogenetically related to a similar source and most probably derived from partial melting that left the meta-harzburgite as residue (see below).

Quartz grains are typically sutured with unresolved elastic strain as indicated by underlatory extinction (Chesman et al., 1997). Ribbon quartz, also exhibiting unresolved strain is pervasive throughout the BIG Complex providing a mylonite-like foliation. However, due to the ability of quartz to anneal at <300C (Ehlers and Blatt, 1982), the unresolved strain is probably produced by a late Paleozoic or more recent tectonic event.

Pyroxenes within enderbites comprise principally metamorphic salite and hypersthene (see also Blank 1973). Many of the enderbites/tonalites have relict pyroxenes embayed by green hornblende indicating a hydrous retrograde reaction. Seritization in the plagioclase in most tonalites (non-pyroxene bearing) indicates pervasive hydrothermal alteration of many lithologies. As stated earlier, retrograde metamorphism has substantially reset enderbite granulite to amphibole plagioclase quartz fels and amphibole gneiss (Winkler 1974, pg 502 of Raymond, 1995).

The scale of these metasomatic retrograde effects is noticeable in the same thin section, with biotite and authigenic secondary K-feldspar forming interstitially between quartz and plagioclase grains (Chesman 1997a) on one corner of the slide and pyroxene retained on the other. Extensive K-metasomatism could even drive the classification to that of granite.

Garnets are pervasive through out the felsic formations and can be in high concentrations (Zeigler, 1911). The garnets examined in both pyroxene and non-pyroxene bearing felsic units are almandine-grossular.

A third set of felsic orthogneiss are granites and trondhjemites that are considered to be partial melts of deep seated rocks and are of uncertain age. Many of these units occur as thin, sill like, foliated intrusions that are more easily noticed where they penetrated the mafic granulites. These units are considered similar to the leucosomes generated around meta-harzburgites (Figure 10) and will be further discussed in the geochemical section.

3.4 Cross-cutting intrusives

The BIG Complex has been intruded by numerous cross-cutting units comprised of granitic pegmatite[13], diabase and porphyritic dacite. Although not associated with the development of the Ravenswood island arc, these intrusions provide structural and geochemical evidence for the Post-Grenville assemblage of the Manhattan prong.

[13] Several generations of late stage granitic pegmatite intrude the complex (Blank 1973; Chesman 1992) but are not discussed in this thesis.

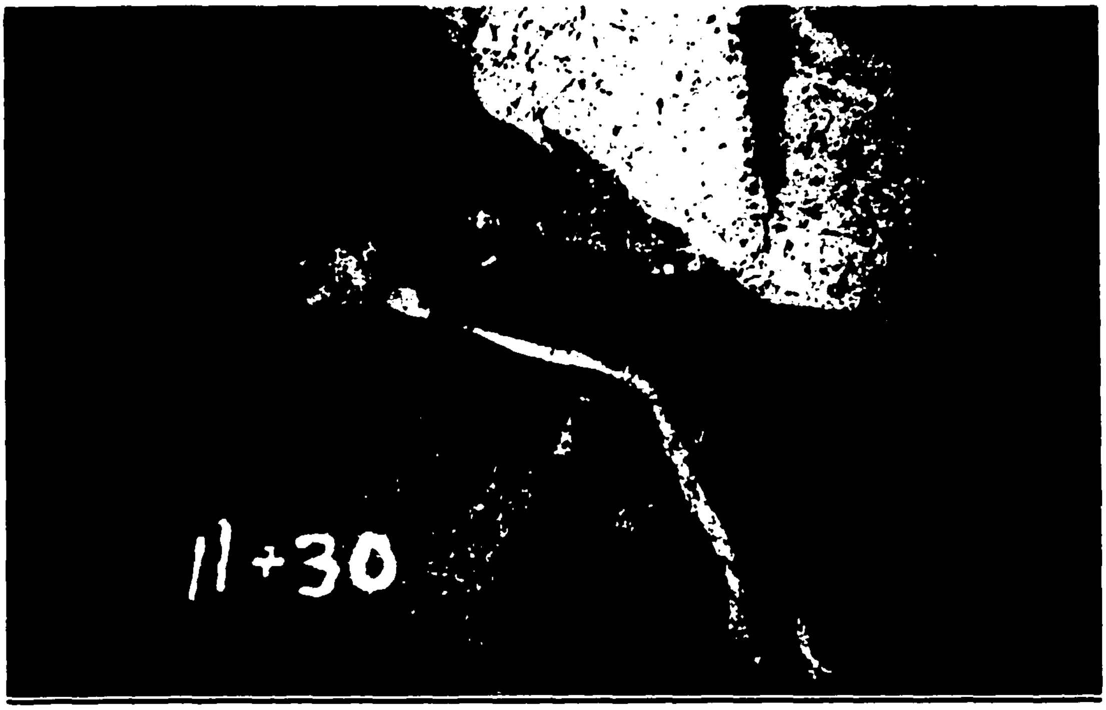

Figure 17. Meta-diabase dikes cut by aplite in the Brooklyn Tunnel.

3.4a Meta-diabase dikes

Numerous dikes and sills of meta-diabase that are commonly discordant to the regional foliation of the Proterozoic rocks are present throughout the BIG complex (also see Blank, 1973). Initial assumptions by the author were that these dikes were associated with the regionally important Palisades sill (Shirley 1987; Steiner et al., 1992). The dikes have a fresh igneous appearance in the tunnel cut (Figure 17), however when viewed in thin section the dikes widely carry metamorphic hornblende and exhibit an obvious metamorphic fabric (Figure 19A). Many of the thinner units have been thoroughly recrystallized to hornblende + plagioclase +/- garnet while the centers of the larger bodies (up to 30 m thick) have retained relicts of igneous pyroxene (Figure 19B). A number of the meta-diabase dikes are cross cut by later granitic pegmatite and aplite, which when considered with the amphibolization, indicate a pre-Palisades origin. Although not dated for this study, the meta-diabase dikes have a similar geochemistries and characteristics to those of Ratcliffe's (1987) Late Precambrian high TiO_2 meta-diabase dikes that crosscut the Hudson Highlands (discussed further in section 7).

3.4b Porphyritic Dacite

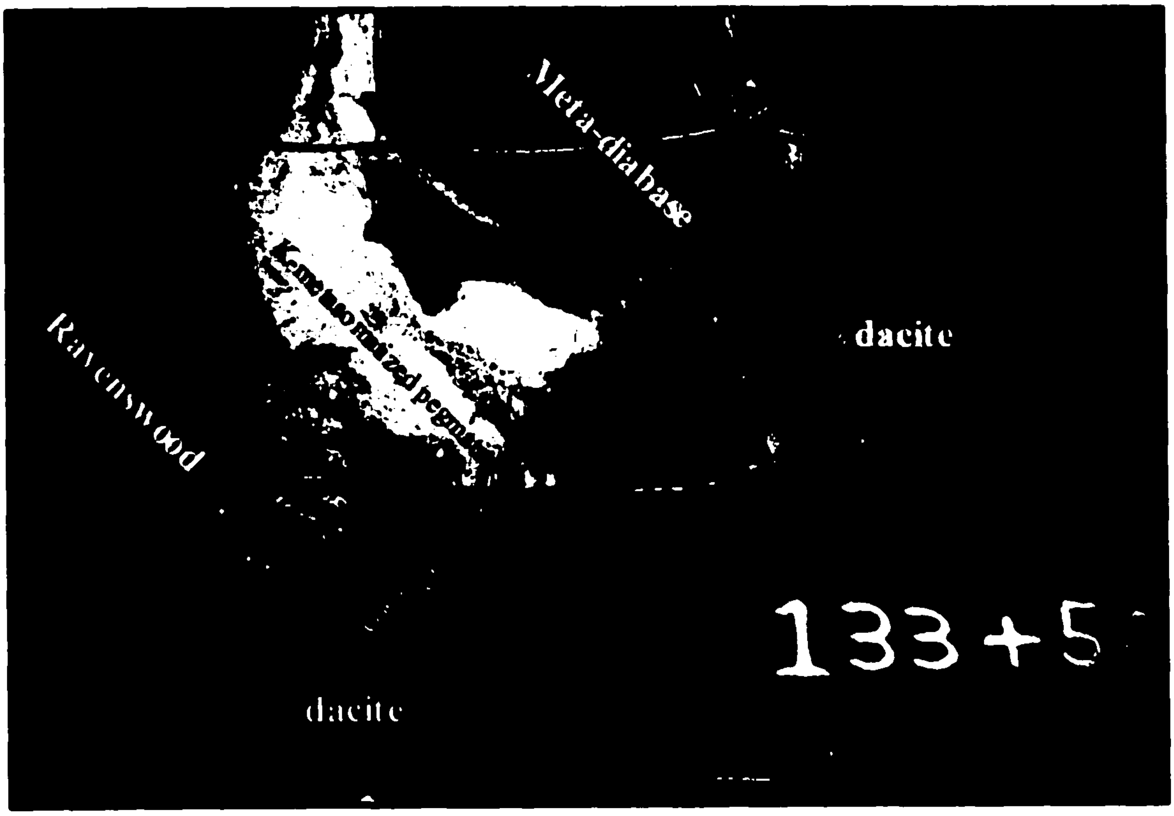

Figure 18. CT#3 tunnel exposure illustrating the cross-cutting relationship of the porphyritic dacite across foliation, pegmatite and meta-diabase of the BIG Complex.

The youngest rock type encountered within the BIG Complex are porphyritic dacite dikes intruded along NW striking reactivated faults in northern Brooklyn and Queens (Figures 18 and 20). A single zircon date of 298 Ma was obtained from the Brooklyn sample (Chesman, et. al., unpublished data). Thin section examination (Figures 19C and D) reveals the porphyritic nature of the intrusive, which consists of phenocrysts of biotite, plagioclase, orthoclase, hornblende, garnet and pyroxene in a fine-grained groundmass of potassium feldspar and quartz. The deuteric alteration products chlorite, sericite and calcite are abundant. The presence of vapor phase cavities within the rock suggests a hyperbyssal positioning of the intrusion.

Review of the City Tunnel #2 geology records indicates that several fine grained porphyritic intrusions were recorded in the Bronx.

> **"The dike of feldspathoid porphyry, which was passed through in the excavation of Shaft 3A, had become a series of dikes when crossed in the tunnel. The largest of these was eight to ten feet thick and four smaller ones varied in thickness from two inches to four feet."** [14]

Although published much later, Colony and Blank (1943) presented a thorough report on the petrology and geochemistry of the Bronx dikes classifying them as lamprophyres. Phenocrysts of phlogopitic biotite, kataphorite amphibole, minor pyroxene and orthoclase were identified in the Bronx samples but no feldspathoids were recognized, even though the dikes had very high alkaloid content. Colony and Blank (1943) examined a "purplish" garnet surrounded by a halo of chlorite and considered the garnet and many of the other minerals to be xenolithic, captured from the surrounding country rock.

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[14] Board of Water Supply, City of New York, Aqueduct Department Annual Report 1932, page 11.

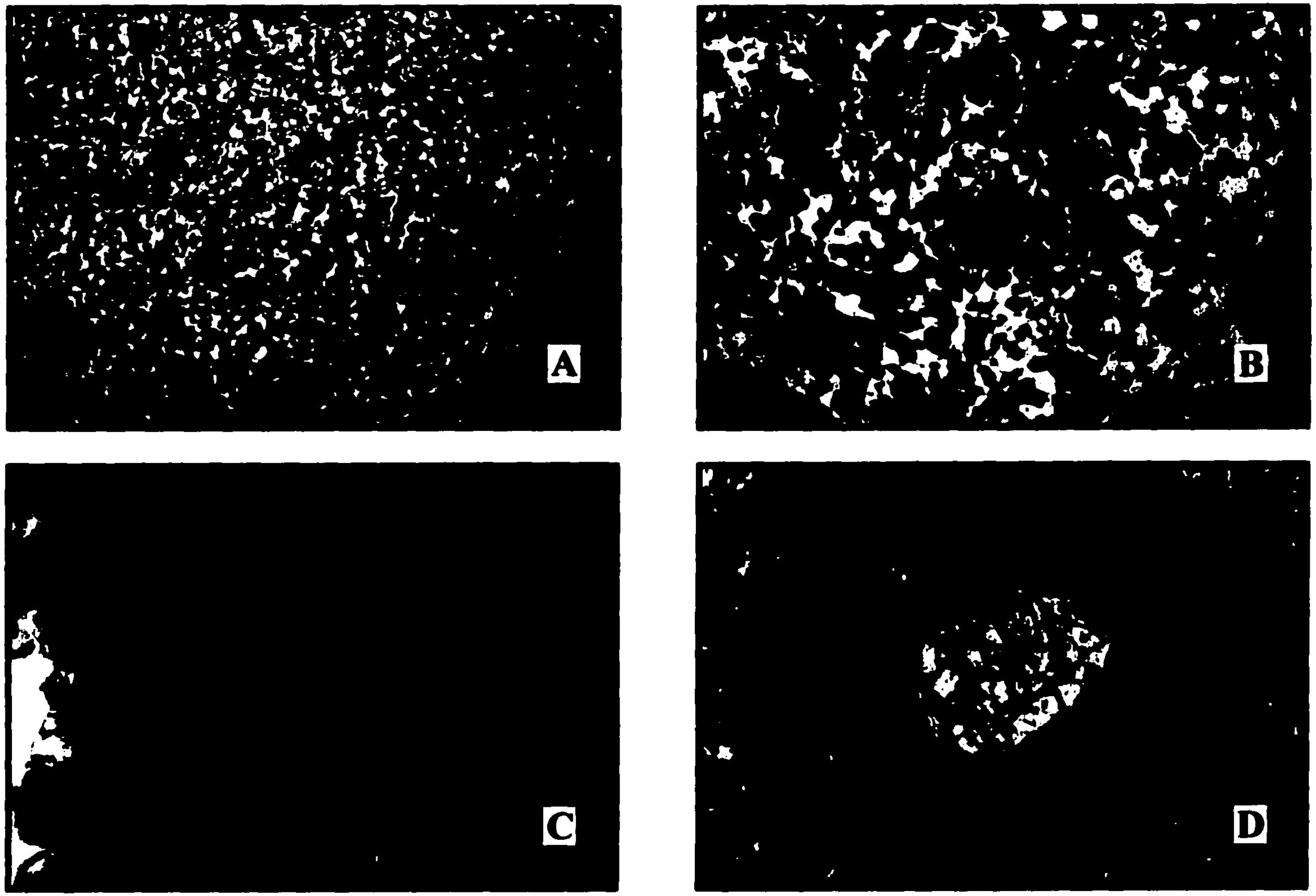

Figure 19. Microstructural textures of BIG Complex. A) Recrystalized amphibolite texture of thin meta-diabase dike; B) Recrystalized amphibolite texture of thick meta-diabase dike; C) Primary amphibole in porphyritic dacite; D) Chlorite corona around primary garnet in porphyritic dacite.

Petrographic studies of the Brooklyn and Queens dikes revealed igneous amphibole (Figure 19C) and pyroxenes possessing compositions different from those found in the BIG Complex (Chesman and Steiner, in prep). As perhaps the most telling mineralogic difference, the garnet within a corona of chlorite (Figure 19D) has a greater pyrope content. This indicates a possible mantle source, as does the overall zircon chemistry (Bowsring personal communication, 1996, to Reiss Institute).

The dacite intrusions (Figure 20) crosscut the regional foliation striking between N44W and N62W with an average dip of 60SW. These orientations are similar to the Bronx N60W occurrences (CT#2 records) and associated dike orientation trend within the Beemerville Complex (after Colony and Blank, 1943). Although requiring further research, the similar geochemistry, mineralogy and occurrence of barite rimmed orthoclase in the CT#3 (Chesman and Steiner, in prep) supports a petrogenetic relationship with the Bronx intrusives and a possible relationship with the Beemerville Complex.

The high K-content of the dacite, the presence of barium rims on feldspar, abundant calcite and large number of vesicles indicates that the dacite was rich in water. The intrusion of the dacite appears to be related to an influx of K-rich hydrothermal fluids that has produced at least one event of K-metasomatism in the area. The influx of metasomatic fluids associated with this intrusion as well as other earlier deformation events requires that special attention be paid to the possible alteration of the BIG Complex's geochemical signatures.

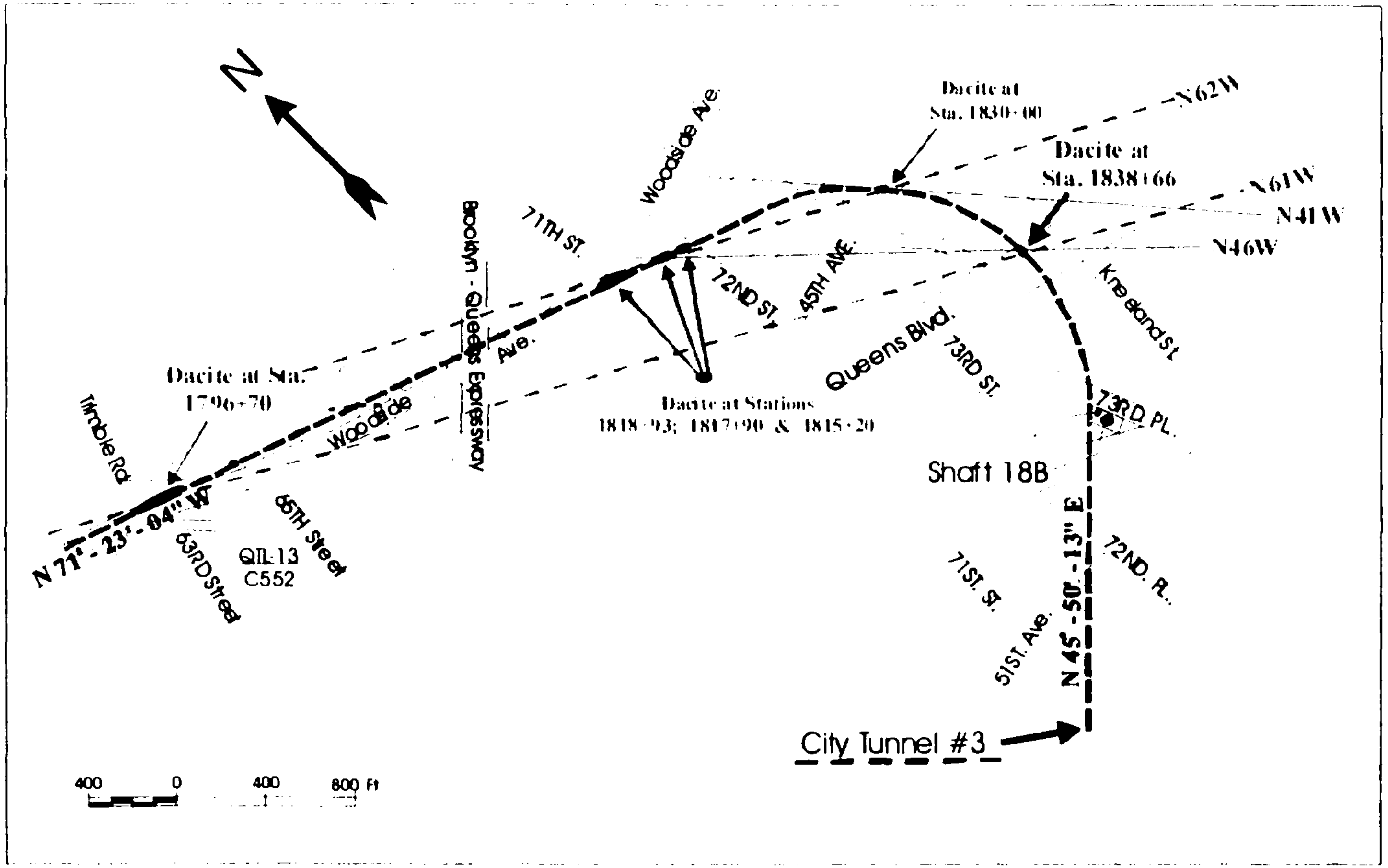

Figure 20. Location map of the Queens section of CT#3 depicting tunnel intersections with the dacite intrusions. Extrapolation indicates a ~N44W to N62W dike trend, which is consistent with the Brooklyn dacite (not shown) having a strike of N47W.

4 Metasomatic Fluids

Suprasubduction zone (SSZ) geochemical signatures discussed farther along and in Chesman (1997a) indicate that the BIG Complex formed at a destructive plate boundary. It is presently considered that metasomatic fluids generated within the subducting slab are responsible for the observed widespread incursion of K-Rb metasomatic fluids. These enriched fluids (K, Rb, and Ba) penetrate the overlying mantle wedge, and in places generated magmas with LILE-enriched profiles on MORB-normalized trace element variation diagrams. Therefore, LILE enrichment provides compelling additional evidence for a SSZ setting. However, the geochemical signature of the BIG Complex is complicated by its later association with destructive boundaries associated with Avalonian, Taconic, Acadian and Allegheny compressional orogenies. Metasomatic fluids from these later events upset the original Mesoproterozoic LILE geochemical signatures but the overall SSZ designation remains reasonably well established.

The BIG Complex's granulite grade complicates further geochemical interpretations. Granulites are commonly associated with the depletion of LIL elements Th and Sr (Weaver and Tarney, 1983). In view of this complex history, the effect of metasomatic fluids on the geochemical signature of the Brooklyn Injection Gneiss is presently addressed.

4.1 K-metasomatism

K-metasomatism is crucial to this study because; 1) it justifies the tonalite versus granodiorite definition; 2) it may be the underlying cause for RB-Sr age date variations; 3) may alter the LILE signature used in tectonic discrimination diagrams of Pearce (1982) and Pearce et al., (1984a and 1984b).

Petrographic examination of BIG felsic orthogneiss indicates that metasomatic K-feldspar commonly forms interstitially between plagioclase and quartz (Chesman, 1997; see also Barth 1936; Buyce and Friedman, 1974). Thus, although shear zones comprise perhaps the most effective pathway for metasomatic fluids (Andersen et al., 1991; Glassley 2001), fluids are well able to penetrate microfractures and grain boundaries away from the shear zones (see Harlov et. al., 1998). Metasomatic retrograde minerals in the same thin section with remnant orthopyroxene are commonly associated with K-metasomatism. Retrograde minerals are common in granulites (Shelly 1993) as is the possibility of secondary LILE geochemical signatures in highly recrystallized basement.

In several sections of the tunnel, extensive, fracture-controlled K-feldspar metasomatism converts meta-diorites via the alteration of plagioclase to salmon colored orthoclase. These metasomatic felsites are silica-enriched, LILE-charged potassic rocks that show Ba + Rb enhancements and minor Sr, Y and Yb depletion (Figure 21) when compared to relative to control samples, leucosome (219L) and paleosome (219P). REE profiles of the meta-diorite samples (218D and 218K) indicate slight depletion of mid-range REE's possibly related to cation-substitute or slight difference in clinopyroxene content between samples. This is in stark contrast to anatexis, as shown below which produces substantial light REE enrichment in the leucosome (Figure 22).

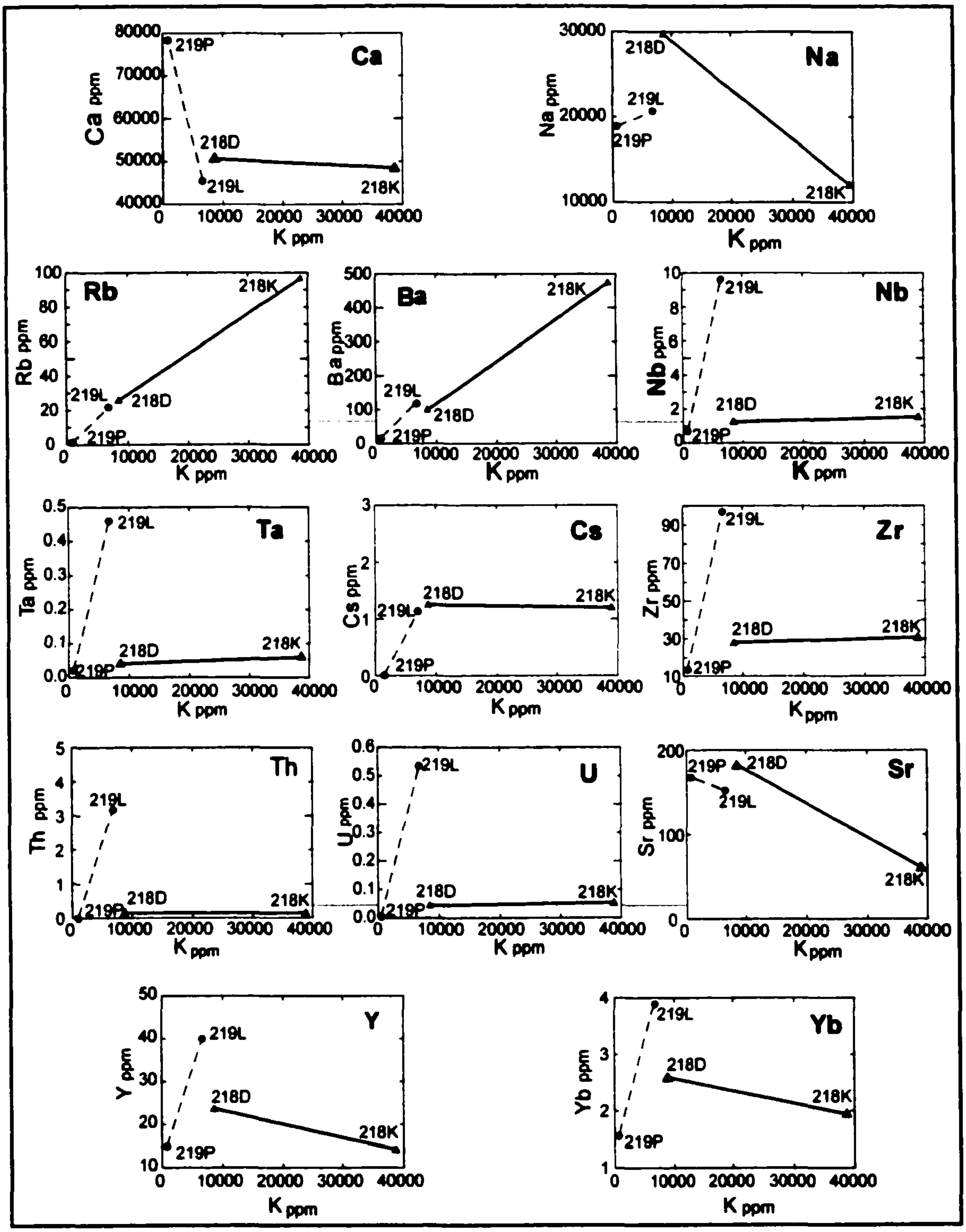

Figure 21. Analysis of trace element mobility during K-metasomatism. Comparison of increasing K due to migmatization of diorite, 219P and 219L (increase in Rb, Ba, K, and Na as well as the high-valent lithophile trace elements Zr, Nb, Y, Th, REE) vs. K rich fluids altering a nearby diorite, 218D and 218K (increase in only Rb, Ba and K with decreases in Y and Yb). A reduction of Sr occurred during both migmatization and K-metasomatism.

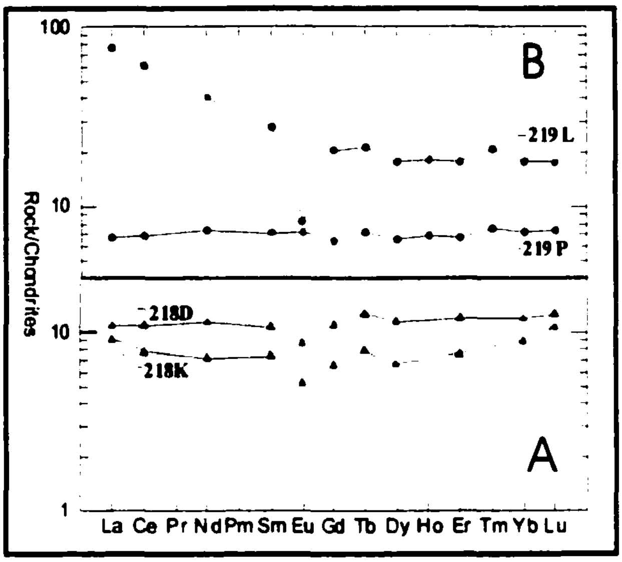

Figure 22. Comparison of (A) REE profiles of K-metasomatized meta-diorite verse an unaltered closely situated meta-diorite. Slight depression of mid-range REE during metasomatism may be associated with cation substitution or a difference in modal percent pyroxene between the two samples. (B) Partitioning of REE during anatectic melting of a BIG Complex diorite.

Sample	218D	218K	219P	219L
SiO[2]	52.47	54.54	50.94	62.12
TiO[2]	0.71	0.75	0.56	0.37
Al[2]O[3]	13.4	12.92	14.99	13.79
Fe[2]O[3]	12.08	10.59	12.42	7.76
FeO				
MnO	0.22	0.17	0.2	0.1
MgO	6.6	5.53	7.46	4.38
CaO	7.34	7.01	11.37	6.61
Na[2]O	4.01	1.67	2.55	2.79
K[2]O	1.02	4.66	0.09	0.8
P[2]O[5]	0.09	0.01	0.1	0.05
LOI	1.35	0.33	-0.48	1.44
Total	99.29	98.18	100.2	100.2

Table 2. A). Whole rock geochemical analysis of K-metasomatized sample (218K) and leucosome (219) with corresponding meta-diorite nonmetasomatized (218D) and paleosome (219P) samples.

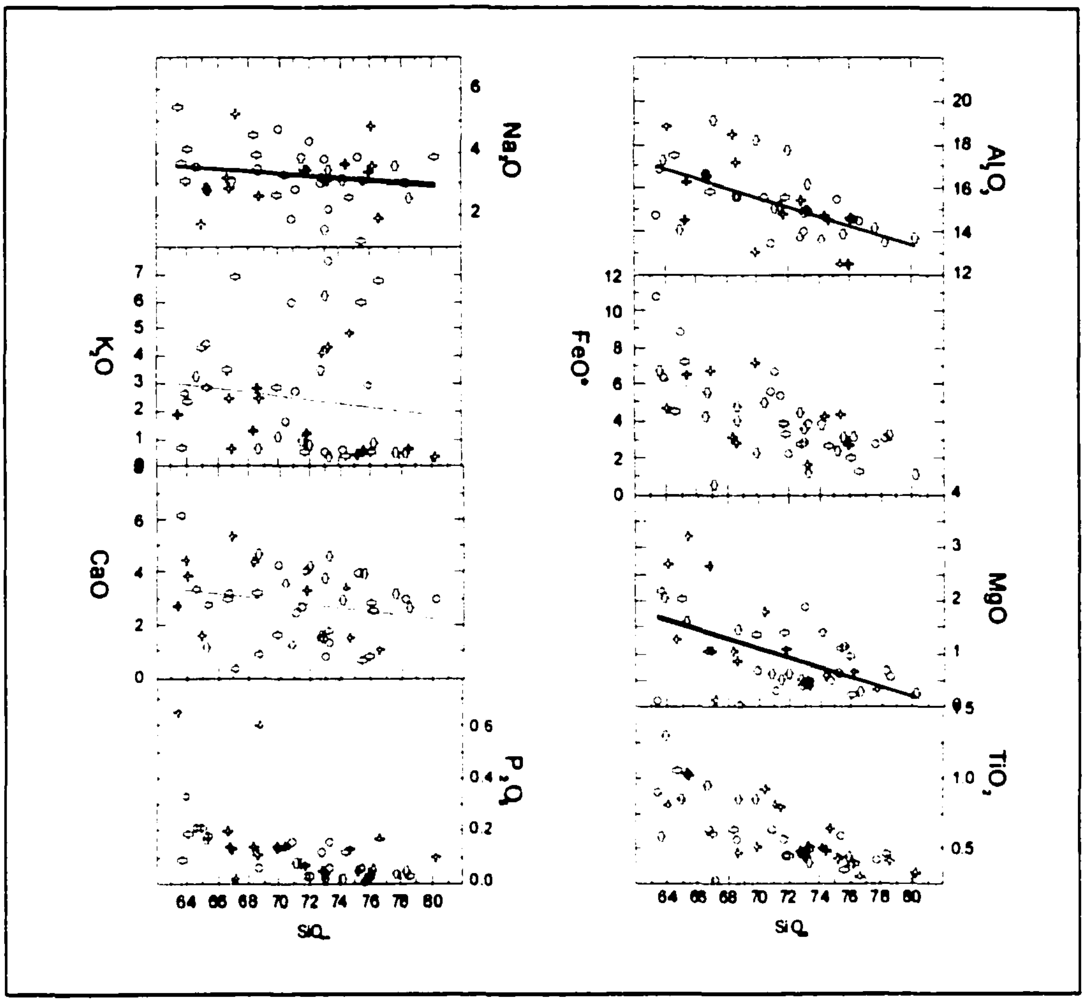

Figure 23. Variation diagram of the felsic orthogneiss samples of the Ravenswood Suite within the BIG Complex. The scatter of the K2O concentrations compared to other major elements indicates the effects of K-metasomatism.

Whole rock analysis of the unaltered diorite and K-metasomatized diorite (Table 2) shows almost even exchange between K-Na. The exchange of Na and K in the plagioclase can occur at relatively low temperature if accompanied by water (Mueller and Saxena, 1977), and this can lead to revision of the whole rock chemistry through potassium (K)-metasomatism. The variation diagrams of Figure 23 shows largely linear trends in major element content versus silica, associated with a significant amount of scatter. The relatively more pronounced variability of K_2O within the felsic varieties of the Ravenswood is consistent with the proposed metasomatic activity (Glassley 2001).

The surface outcrops of the Ravenswood Granodiorite, located near the East River, are proximal to the proposed ductile thrust fault, Cameron's Line, which is described in section 1.X as either a mylonitic contact between the between the Hartland Formation and the Fordham Gneiss (Merguerian, 1986) or alternatively as the possible intrusive contact between the Manhattan Formation and the Ravenswood Granodiorite (Baskerville, 1992). The presence of a major shear zone close to the Ravenswood Granodiorite sampling locations of Zeigler (1911) and Baskerville and Mose (1989) has contributed to the ongoing confusion over whether granodiorite is a key magmatic lithology or simply a spatially limited k-metsomatized variant on the principal BIG rock types. The present study shows that both the lithologic descriptions and possibly the Rb/Sr age dates have in fact been profoundly influenced by k-metasomatism.

Table 3
Data table presenting representative analyses for the Brooklyn Injection Gneiss Complex lithologies.

	Meta-harzburgite		Meta-basalts		Two-pyroxene gneiss		Meta-gabbros / diorites				
Sample	119+00	QTL-3	192+35	Q10+50	115+00	18B	144+72	Q78+20	Q134+20	218D	218K
SiO[2]	41.22	41.72	52.08	47.32	50.23	52.98	46.92	43.09	49.05	52.47	54.54
TiO[2]	0.02	0.02	1.31	0.96	2.76	2.59	2.39	0.47	0.4	0.71	0.75
Al[2]O[3]	0.6	0.46	11.59	14.57	13.04	13.74	11.98	12.85	14.12	13.4	12.92
Fe[2]O[3]	8.37	14.76	12.75	13.15	17.04	16.24	14.2	16.63	11.05	12.08	10.59
FeO											
MnO	0.1	0.17	0.15	0.22	0.24	0.26	0.19	0.26	0.19	0.22	0.17
MgO	34.57	35.31	7	9.12	3.9	4.66	10.83	12.7	11.44	6.6	5.53
CaO	0.16	0.59	10.91	11.26	7.78	6.54	7.45	4.86	11.17	7.34	7.01
Na[2]O	0.05	0.01	2.05	2.1	3.16	1.29	2.42	1.08	1.74	4.01	1.67
K[2]O	0.18	0.17	0.25	0.43	0.95	1.15	0.96	1.11	0.31	1.02	4.66
P[2]O[5]	0	0.01	0.1	0.08	0.3	0.4	0.69	0.05	0.06	0.09	0.01
LOI	12.53	6.62	1.38	1.08	0	0	1.4	6.76	1.23	1.35	0.33
Total	97.8	99.84	99.57	100.29	99.4	99.85	99.43	99.86	100.76	99.29	98.18
Sr	6.53	2	93.68	92.1	173.73	103	420.73	289	129	182.09	61.59
Rb	8.17	16	1.39	3.1	6.74	40	12.97	43	9	25.4	96.59
Ba	16.78	9.1	46.03	28	219.52	238	372.47	178	109	101.47	474.45
Th	0.648	0.08	0.113	0.29	0.317	0.2	0.597	0.49	0.13	0.155	0.106
U	0.211	0	0.012	0.07	0.178	0.27	0.445	0.49	0	0.045	0.056
Pb	0	0	0	6	7	0	0	5	0	10	0
Ta	0.041	0.06	0.108	0.06	0.969	0.79	0.553	0.24	0.02	0.04	0.06
Nb	0.41	0	2.47	1.4	17.21	16	10.12	3.6	0.7	1.34	1.61
P	0.00	43.64	436.39	349.11	1309.17		3011.10	218.20		392.75	44
Sc	3	8	52	48	44	49	29	20	41	46	36
Hf	0.16	0	1.45	1.1	4.7	6.6	4.8	1.3	0.6	0.96	1.09
Y	0.9	1.4	32.6	21	77	87	34.7	18	13	23.6	14.2
Zr	5.09	2.4	40.73	37	161.51	244	186.37	46	18	28.23	31
V	14	53	346	318	383	385	191	136	211	258	213
Cr	3607	3445	114	317	46	62	575	89	1140	25	48
Ni	2104	1177	31	89	13	26	195	235	351	53	55
La	0.69	0.13	2.63	2.47	17.18	13.9	23.81	13	1.94	3.55	3
Ce	1.24	0.58	9.27	6.97	46.81	36.4	54.65	33.5	4.44	9.42	6.68
Nd	0.53	0.41	9.32	6.09	33.48	26.5	33.28	17.8	3.34	7.1	4.49
Sm	0.13	0.12	3.27	2.24	9.87	8.02	7.29	4.17	1.09	2.15	1.48
Eu	0.121	0.016	0.971	0.975	2.294	1.738	2.239	1.524	0.414	0.669	0.399
Gd	0.09	0.16	4.34	3.01	11.46	9.44	7.31	3.6	1.56	3.02	1.81
Dy	0.12	0.22	5.42	3.79	13.67	14	6.22	3.28	2.06	3.88	2.26
Er	0.12	0.16	3.71	2.34	8.38	8.7	3.57	1.91	1.43	2.67	1.7
Yb	0.17	0.25	3.64	2.26	7.37	8.28	2.8	1.8	1.63	2.6	1.95
Lu	0.034	0.052	0.542	0.345	1.099	1.206	0.422	0.271	0.239	0.424	0.357

Table 3 (continued)

Data table presenting representative analyses for the Brooklyn Injection Gneiss Complex lithologies.

	Migmatite		Enderbitic gneiss				Meta-diabase		Dacite	West Side Granite	
Sample	209P	209L	B256+18	19B	TL-9	Q64+33	4+30	Q41+60	QNS	26B	Sample
SiO[2]	50.94	62.12	80.23	75.99	73.29	73.68	46.29	48.6	63.33	72.77	SiO[2]
TiO[2]	0.56	0.37	0.08	0.2	0.26	0.21	0.65	3.05	0.52	0.02	TiO[2]
Al[2]O[3]	14.99	13.79	11.68	11.73	12.9	13.1	13.93	13.44	14.27	14.36	Al[2]O[3]
Fe[2]O[3]	12.42	7.76	0.01	0.01	4.26	0.51	12.36	3.49	3.29	1.92	Fe[2]O[3]
FeO			1.11	2.79		2.99		10.45			FeO
MnO	0.2	0.1	0.02	0.04	0.07	0.05	0.17	0.22	0.07	0.02	MnO
MgO	7.46	4.38	0.25	1.44	0.47	1.25	11.08	5.73	1.59	0.15	MgO
CaO	11.37	6.61	2.99	3.11	4.59	3.82	10.76	9.83	4.31	1.47	CaO
Na[2]O	2.55	2.79	3.85	3.29	3.43	3.14	2.07	2.85	3.99	4.57	Na[2]O
K[2]O	0.09	0.8	0.37	0.97	0.35	0.86	0.72	0.85	4.05	3.75	K[2]O
P[2]O[5]	0.1	0.05	0.1	0.07	0.06	0.07	0.08	0.57	0.41	0.02	P[2]O[5]
LOI	0	1.44	0.26	0.63	0.48	0.5	1.39	0.34	3.27	0.3	LOI
Total	100.68	100.21	100.95	100.27	100.16	100.18	99.5	99.42	99.1	99.35	Total
Sr	168	152	145	680	179.84	244	76.45	391	1553	346	Sr
Rb	0.5	21	2.5	14	2.63	20	12.37	15	56		Rb
Ba	17	119	99	840	93.9	277	70.99	310	7370	577	Ba
Th	0	3.17	0.52	0.51	0.155	0.48	0.298	1.42	28.6		Th
U	0	0.53	0.13	0.28	0.12	0.19	0.072	0.45	4.59		U
Pb	9	17	0	0	0	12	5	0	56		Pb
Ta	0.02	0.46	0	0.05	0.023	0.12	0.055	1.26	0.59		Ta
Nb	0.7	9.6	0.8	0.9	1.04	1.6	1.38	19	12		Nb
P			436.39	305.47	261.83	305.47	349	2487		87	P
Sc	37	26	5	7	15	13	38	33		<1	Sc
Hf	0.6	4.1	1.7	3.2	0.99	0.9	1.29	4.8	4.9		Hf
Y	15	40	18	4.1	20.6	8.6	20.4	38	20	1	Y
Zr	13	97	58	111	30.99	33	43	195	175	22	Zr
V	242	172	5	25	18	64	202	360	57	3	V
Cr	545	139	0	52	0	36	642	68	28		Cr
Ni	53	33	0	69	0	12	243	48	16		Ni
La	2.22	25.4	11.6	9.21	5.15	5.37	3.92	24	183		La
Ce	5.94	52.3	21.6	15.6	10.67	10.3	8.46	53.3	349		Ce
Nd	4.61	25.5	11.1	6.61	6.97	5.25	5.19	33.5	141		Nd
Sm	1.44	5.57	2.82	1.3	1.94	1.32	1.69	8.11	19.2		Sm
Eu	0.555	0.632	0.743	0.966	0.811	0.541	0.636	2.83	4.121		Eu
Gd	1.77	5.71	3.1	0.95	2.59	1.37	2.33	7.76	7.48		Gd
Dy	2.25	6.14	3.13	0.73	3.25	1.5	3.08	7.33	4.25		Dy
Er	1.52	4.03	1.9	0.46	2.36	0.97	2.37	3.94	1.99		Er
Yb	1.57	3.9	1.59	0.54	2.53	1.06	2.26	3.31	1.62		Yb
Lu	0.251	0.6	0.232	0.094	0.387	0.163	0.38	0.476	0.257		Lu

5 Geochemistry of the BIG Complex:

The geochemistry sections justify the proposal that the BIG Complex evolved initially in a subduction related setting but also contains remnants of an oceanic setting, an island arc component and finally back-arc basin basalt.

This thesis relies heavily on trace element discrimination diagrams that employ HFS elements, and proceeds with caution in interpreting diagrams that rely on mobile LIL elements for tectonic discrimination. In the present interpretation the apparent depletion of Y and Yb are effects related to metasomatism, and are not primary features of these rocks.

5.1 Whole Rock Geochemistry

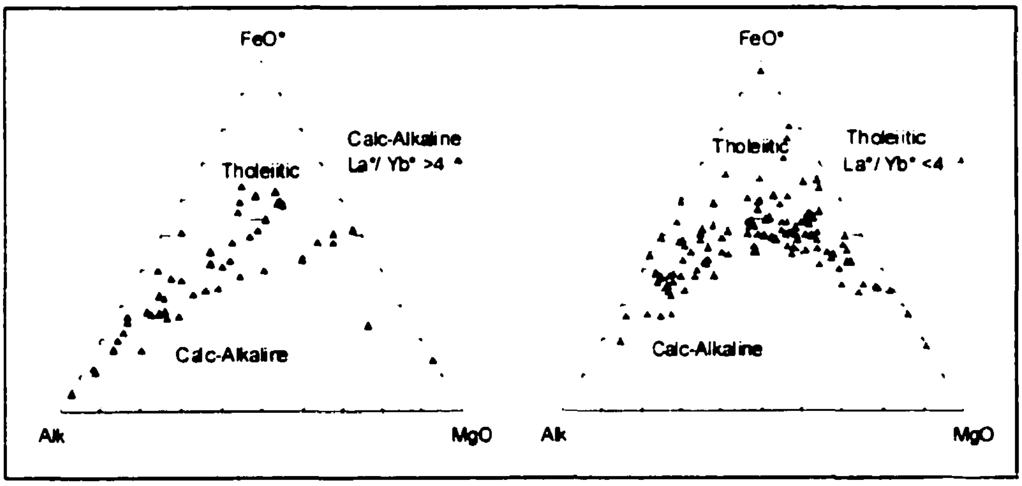

FIGURE 24. AMF diagrams showing calc-alkaline verse tholeiitic trends for the BIG Complex. Rocks with La*/Yb* >4 show a fair correspond with the calk-alkaline classification of Irvine and Baragar (1971) while samples with La*/Yb* <4, consistent with tholeiites, show a wider spread on the AMF diagram. This may indicate major element contamination of later stage tholeiitic intrusions with pre-existing calc-alkaline bodies and/or meta-sediments.

The Brooklyn bedrock was interpreted by Steiner and Chesman (1996) and Chesman (1997a) as representing a dismembered ophiolite and island arc complex

(Ravenswood Suite). Due to systematic resampling and additional geochemistry data, it now possible to more fully classify lithologies within the BIG Complex and provide additional support for the ophiolite/marginal basin and volcanic arc origin of the BIG Complex.

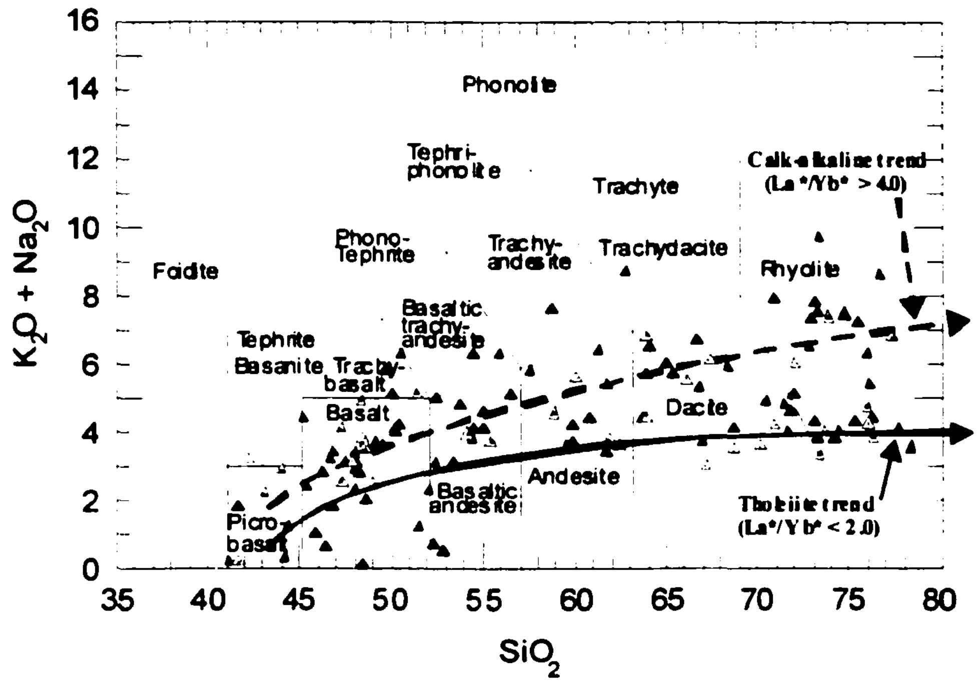

Figure 25. TAS alkalies-silica plot (after LeBas et al., 1986) for the Ravenswood Suite. As in figure 24, calc-alkaline trends are noticed for La/Yb > 4, with a tholeiite trend best established for rocks possessing La/Yb < 2.

Although the BIG Complex is made up of meta-ultramafics, meta-basalts, felsic gneiss and meta-sediments, the majority of the meta-basalts and felsic gneiss, approximately seventy percent, are interpreted as belonging to the Ravenswood Granodiorite Suite as originally defined by Zeigler (1911). In detail, geochemical evaluation finds the Ravenswood to be a fractionated tholeiitic bimodal (acid – basic) plutonic suite with transitional calc-alkaline affinities. A fair correspondence of samples possessing La*/Yb* > 4 with a calc-alkaline classification of Irvine and Baragar (1971) while tholeiitic samples with La*/Yb* < 4 exhibit a wider, cross classification range.

Despite uncertainty in the Na + K index related to variable K-metasomatism, meta-igneous rocks are best classified as a deep-seated plutonic assemblage that incorporates an early subordinate calc-alkaline suite with a later stage tholeiite suite. The calc-alkaline designation applies best to rocks with La*/Yb* ratios of > 4.0 as depicted on the TAS alkalis vs. silica diagram (Figure 25) of LeBas et al., (1986). Rocks designated as tholeiitic possess La*/Yb* ratios < 2.0 (Figure 25). The transition from low to high-K suites for intermediate volcanic rocks correlates simultaneously with increasing LREE and La/Yb ratio (Gill 1981). This trend generally holds true for the Ravenswood suite, with the upward shift of some tholeiitic samples off their La/Yb trends (Figure 25) due to incorporation of earlier calc-alkaline and/or meta-sedimentary xenoliths (Figure 26) as well as the proposed k-feldspar metasomatism event(s). All of which may have significantly affected the whole rock chemistry with little influence on the rare earth elements.

The presence of both tholeiitic and calc-alkaline differentiation trends as well as continuous gradations between the two types is a frequently observed characteristic of island arc systems (Ehlers and Blatt, 1982). In this interpretation, calc-alkaline and tholeiitic rocks are intimately intermixed throughout the tunnel along smeared tectonic and intrusive contacts that are difficult to detect, as is expected of rocks comprising the deep roots of island arcs. Although rock types ranging from picrites to rhyolites (dacites) are encountered in the tunnel, the exposed Ravenswood island arc suite is predominately variably foliated meta-gabbros (ca to 14%), meta-garnetiferous diorite and quartz diorite (ca 39%) and associated tonalite (ca 47%). These units invade and partially melt associated meta-sediment and, to a limited degree, commingle with earlier intrusives. Gabbro bodies, some showing residual cumulate textures, indicate magmatic stoping of ultramafics (harzbergite). Magmatic stoping and xenolith observations were described by Berkey (1931) in CT#2;

> **"In many varieties of rock are found within the principal granodiorite mass which are quite unlike the average Ravenswood itself. In occasional instances there are obscure traces of structure in compositions that suggest derivation from the original Fordham formation and it is believed that these represent xenoliths that have been almost completely destroyed by absorption and fusion in the magma. Still other places have lost all structural evidence and only retain somewhat abnormal composition. It is believed that these varieties are due to absorption of older material in the magmatic mass as it fused its way into the roof formations of a former time."** (Berkey 1931, pg. 6)

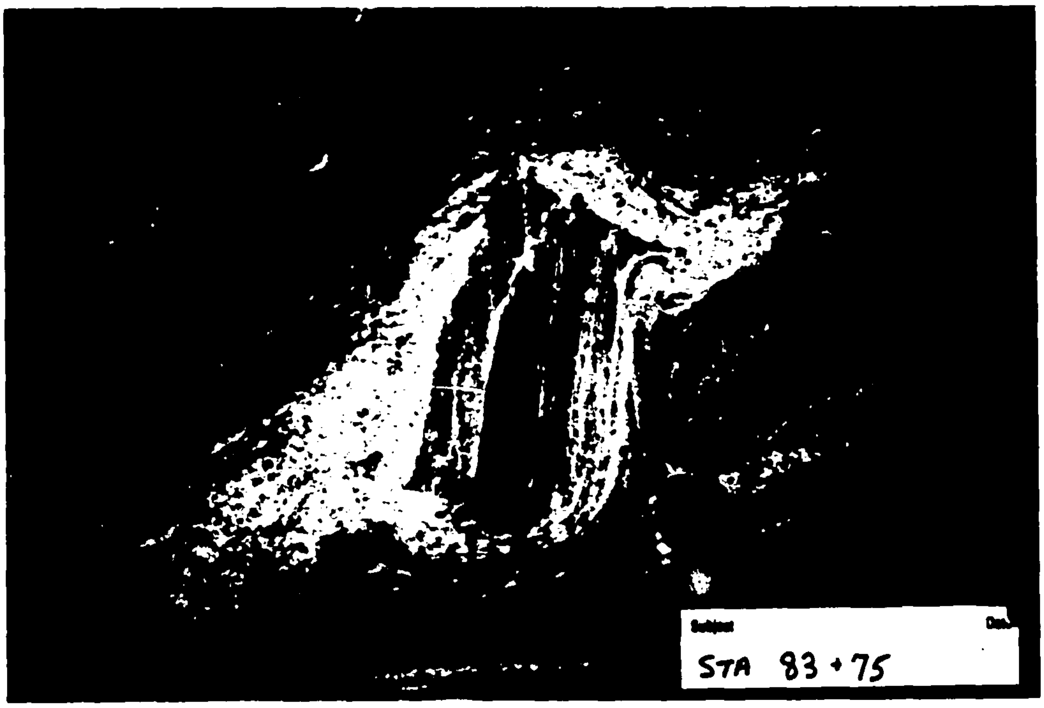

Figure 26. Clasts of an already metamorphosed Fordham Gneiss xenolith in tonalitic gneiss.

5.2 TRACE ELEMENT GEOCHEMISTRY

There is now a broad range of geochemical signatures from mafic terranes associated with subduction settings throughout the world. In addition there are numerous models for the effect of progressive melting that can either be used alone or used in conjunction with other diagrams to differentiate tectonic settings. As a key issue, Zr, Nb, Ba, and the LREE are thought to behave as incompatible elements in both tholeiitic and calc-alkaline sequences. In contrast, Y and HREE behave as incompatible elements in tholeiites but are partially or wholly compatible elements with the calk-alkaline series (Pearce 1982). Crystallization rates can also affect element ratios, particularly if Ti and V concentrations are observed to decrease relative to Zr and P due to early magnetite crystallization (Pearce and Norry, 1979). Tectonic process can also affect trace element concentrations. IAB, for example, are typically reported to be depleted in HFSE (Th, Nb, Hf, Ti, Zr, Y) but enriched in LIL cations (Cs, Rb, Ba, K, Sr) compared to MORB; Ti and Nb on the other hand tend to be magmatic enriched in WPB tholeiites relative to MORB (Pearce 1982).

Due to the very low abundances of Ta and Nb in volcanic arc rock, instrumental neutron activation analysis (NAA) and/or inductively-coupled plasma mass spectrometry (ICP/MS) are generally required to provide sufficiently accurate measurements for trace-metal tectonic interpretations (Saunders and Tarney, 1984). Ta and Nb are widely considered immobile during metamorphism and weathering and therefore valuable in assessing possible ratios for parental magmas (Saunders and Tarney, 1984).

The immobility of HFSE's contrasts markedly with the highly mobile behavior of the majority of the large ion lithophile elements (LILE), as discussed earlier in section

4.1. Nevertheless, the LILE's are commonly used as tectonic discrimination tools, particularly in the determination of SSZ environments (Pearce et al., 1984a). The influx of mobile LILE-rich fluids released from the subducting slab to the overlying mantle wedge, is thought to be a critical signature of the SSZ environment. Subduction zone magmas, generated by this fluxing, retain these LILE enrichment signatures (Pearce et al., 1984a and 1984b). Thus we presently use both LILE and HFSE to discern tectonic settings. However plots of the HFSE, Ti against Zr, is considered most effective for characterization of the primitive magma types.

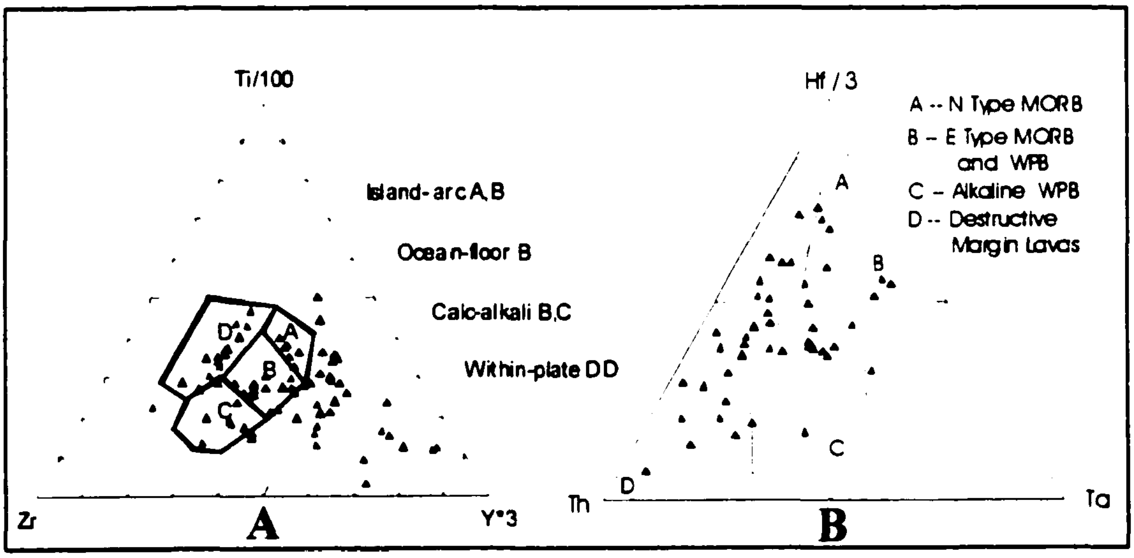

Figure 27. A) Tectonic Ti – Y – Zr discrimination diagram employing several HFSE after Pearce and Cann (1973) indicating the MORB and Arc signatures of the BIG Complex. The within plate samples represent back-arc basin basalts. B) Th, Hf, and Ta discrimination diagram for mafic rocks with boundaries for the different tectonic environment after Wood, 1980) indicating a predominance of BIG Complex subduction zone magmas (field D).

The Pearce & Cann (1973) tectonic Ti – Y - Zr discrimination diagram (Figure 27A) classifies the majority of BIG Complex mafics (SiO2 > 54%) as arc lavas and MORB. The Th – Hf - Ta diagram (Wood, 1980) supports the oceanic character, showing meta-basalts in both the N-MORB and E-MORB fields (Figure27B). Wood suggests that within plate basalts (WPB) can be distinguished from E-MORB with the combined application of the Th – Hf – Ta and Zr – Ti –Y discriminators. Applying these diagrams to the BIG Complex the Post-Grenville meta-diabase dikes can be differentiated from the E-MORB field of Wood (Figure 27B) however several MORB samples that plot within the E-MORB, but near the N-MORB boundary, possess evolved MORB geochemical signatures and will be discussed further in section 4.2.3a.

Although there are numerous diagrams that utilize the ratios of selected HFSE and/or LILE, the high degree of metamorphism and hydrothermal alteration imposed on the BIG Complex has also created a high degree of trace element variability. It has been found that granulite facies metamorphism can remove substantial quantities of K_2O, Rb, Th and U from rocks (Wood 1980; Weaver and Tarney, 1983; Taylor and McClennan, 1995). Another consequence of the granulite metamorphism is the ubiquitous presence, but with varying concentrations, of garnet in the BIG Complex. The incorporation of Y into the garnet structure (Gaines et al., 1997) may create significant shifts of Y levels.

Furthermore, the concentrations of these elements are also strongly dependent on the mantle chemistry where these magmas originated. Depleted or enriched mantles tend to produce magmas correspondingly enriched or depleted in the HFSE (Pearce 1982). Complicating this generality is the discovery that the mantle is heterogeneous not only on the km scale but also on the mm scale (Perfit et al., 1980) thus concentrations of these elements can be quite variable depending on where the melts are generated and the pathways taken (Davies and Stevenson, 1992). As a consequence of the above element variability, significant shifts may drive the implied setting of a sample from one implied tectonic subset to another. Thus, discrimination diagrams which employ only several elements should be used with caution.

Because of the above variations, it has become widely accepted to use discrimination diagrams that use a range of elements, including representative LILE, HFSE and REE subsets to provide an overall more focused geochemical discrimination. Trace element discrimination diagrams typically use the geochemistry of basalts erupted at mid-ocean ridges (MORB) as a normalizing composition for comparison with basalts generated in different tectonic settings such as ocean island basalts (OIB), island arc basalts (IAB) with-in plate basalts (WPB) or enriched (E)-MORB. The geochemical differences, depletions and/or enrichments, provide clues to various processes that may have contributed to the formation of these various magmas. It is the relative profiles or forms that are most indicative of a tectonic environment, with less emphasis placed on absolute abundance levels (Saunders and Tarney, 1984).

5.2a Mid-ocean ridge basalts (MORB)

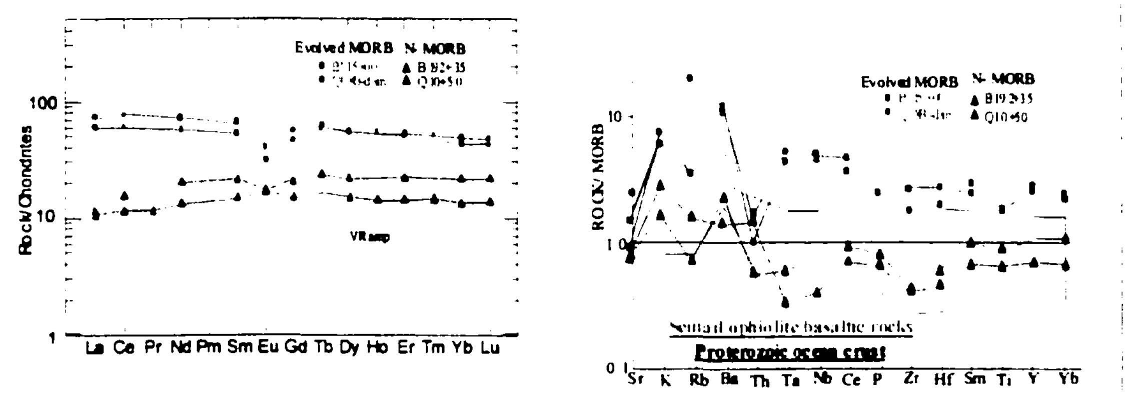

Figure 28. REE profiles and trace metal concentrations of the BIG Complex MORB associations. REE comparisons are made to the Ropes Creek Ophiolitic Assemblage (shaded area from Spell and Norrell, 1985). BIG trace metal concentration profiles show similar profiles to the Semail ophiolite (shaded field in B; after Pearce, 1982).

There is a growing consensus that perhaps the majority of ophiolite suites represent exhumed supra-subduction zone (SSZ) complexes rather than strictly ocean floor sections (Elthon 1991). Evidence in MORB signatures for proximity to subduction zones is the increase in LILE profiles. The HFSE values of the BIG Complex samples are broadly consistent with N-MORB values with a moderate increase in LILE concentrations, similar to other SSZ ophiolites such as the Semail ophiolite complex (shaded field in Figure 28B, data from Pearce, 1982) and the Ropes Creek Ophiolite Assemblage (Spell and Norrell, 1985). It is important to note that similar LILE-enriched MORB profiles from the Bransfield Strait, an ensialic back-arc basin off the southern coast of Chile, are interpreted as indicating the proximity of the MORB to a subduction zone (Karig 1971; Taylor and Karner, 1983; Alabaster and Storey, 1990).

The REE abundances of the BIG Complex N-MORB correlate well with the MORB REE signatures from the Ropes Creek Assemblage (shaded area of Figure 28A, data from Spell and Norrell, 1985). However the BIG Complex MORB can be divided into an evolved trace element abundance set (115+00 and 18B-DSM) and a typical tholeiite MORB set. The 'evolved' N-MORB's increased elemental abundances can be attributed to variations in the degree of fractional crystallization (see Pearce, 1983). The HFSE profiles of the evolved N-MORB are shown to have similar patterns but, increased elemental abundances to typical N-MORB (Figure 28B), consistent with systematic increases associated with the Pearce model.

According to Pearce (1983) evolved MORB may also reflect HFSE-enriched mantle source rather than strict fractionation. The BIG evolved N-MORB lithologies are more highly deformed and more garnetiferous than typical BIG N-MORB (i.e. in comparison to Q10+50 and 192+35). Increased deformation and the proximity to the meta-sedimentary Fordham Gneiss exposures indicates that the evolved N-MORB belongs to the older oceanic crust onto which the Fordham sediment was deposited and was therefore subjected to both pre- and syn-island arc evolutionary events.

It is possible that the evolved and typical N-MORB derived units, representing different mantle sources and times of generation, were incorporated and juxtaposed during the evolution of a back-arc basin. A mature back-arc basin allows for this duality due to slab rollback (Karig 1971) as the basin evolves. In this process, a back-arc spreading center (Jurdy and Stefanick, 1983) develops as a result of renewed upwelling of asthenosphere, which splits the older evolved N-MORB ocean crust and allows for the introduction of basalts with a normal tholeiitic MORB character (see Saunders and

Tarney, 1984). The proximity of two compositionally distinct mantle domains is discussed and reviewed by Hawkins (1997) to explain the presence of both Pacific-MORB and Indian-MORB signatures in the back-arc Lau Basin. The presence of variable mantle compositions has a major bearing on possible tectonic models for the Proterozoic Laurentian Margin.

5.3b Island Arc Basalts

It is now widely believed that dehydration of the subducted slab and possibly some accompanying sediments drives volatile rich fluids through fractures in the mantle wedge to generate melting within the wedge (Davies and Stevenson, 1992). The magmas generated above the subducting slab are thought to be associated with enhanced LILE and HFSE depletions. These geochemical anomalies are most pronounced within lavas of the volcanic arc and to a lesser extent in magmas generated in the back-arc basin.

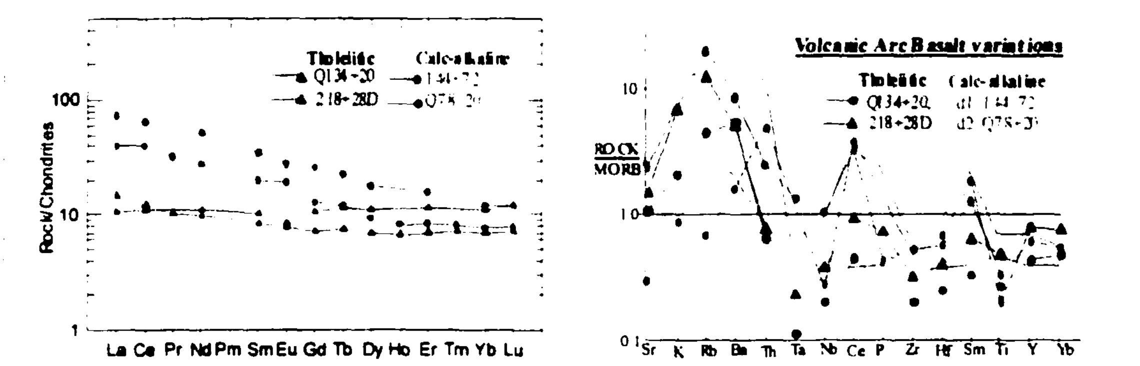

Figure 29. REE profiles for calc-alkaline and tholeiitic island arc basalts showing LREE enrichment in the calc-alkaline. The Ravenswood IAB discrimination diagrams show strong correlation with other volcanic arc basalt variations (shaded range) sampled by Pearce (1982).

IAB in general possess distinctive trace metal profiles due to LIL cation (Cs, Rb, Ba, K, Sr) enrichment and high-field-strength ions (Ta, Nb, Zr, Hf, Ti, Y) depletion relative to N-MORB. The HFS depletions are caused by their preferential removal from the mantle wedge rocks during subduction zone magma genesis (Wood et. al. 1979; Perfit et. al., 1980; Pearce 1982; Saunders and Tarney, 1984). The resulting complex convolutions in trace element profiles of both calc-alkaline and tholeiitic IAB volcanics (Pearce 1982) compare well with calc-alkaline and tholeiitic values of the Ravenswood basalts (Figure 29B). The HFSE depletions of the Ravenswood basaltic associations are paralleled by HFS depletions in the felsic units. This correlates with volcanic arc-granitic-SSZ signatures previously established by Chesman (1997a).

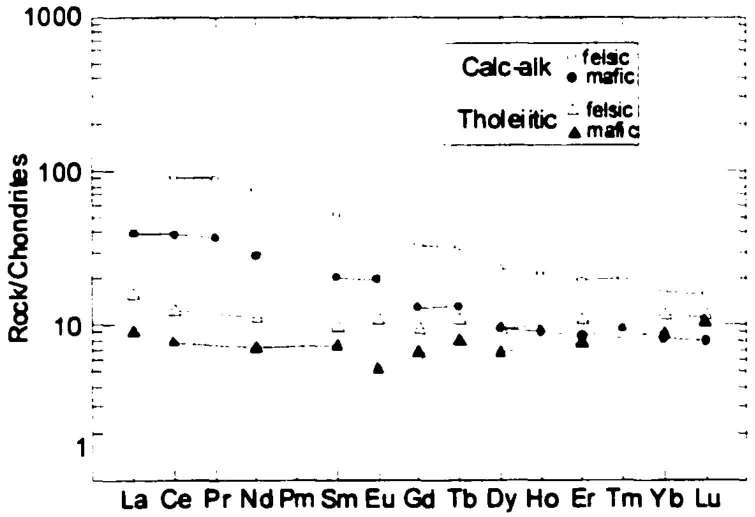

Figure 30. Comparisons of REE concentrations depicting a fractionation process for both calc-alkaline and tholeiitic suites of the Ravenswood island arc.

Comparisons of both calc-alkaline and tholeiitic REE profiles (Figure 30) indicates a separate fractionation event for each suite. In contrast to one another, Ravenswood calc-alkaline rocks shows stronger LREE enrichment (La/Yb > 4) and approximately ~10% increase in Th, Ce, and Sm levels (shown earlier in Figure 29B). These features may be due to the incorporation of subducted sediment into the melt phase

(Pearce 1983). However, it has also been proposed that calc-alkaline basalts are derived early in the formation of back-arc basins where the role of the previous calc-alkaline magmatism associated with the sub-continental lithosphere is still influential (Weaver et al., 1979; Pearce 1982). The latter scenario is presently considered to be the preferred consistent model for the BIG Complex, due to the proximity of older calc-alkaline rocks in the nearby New Jersey and Hudson Highlands.

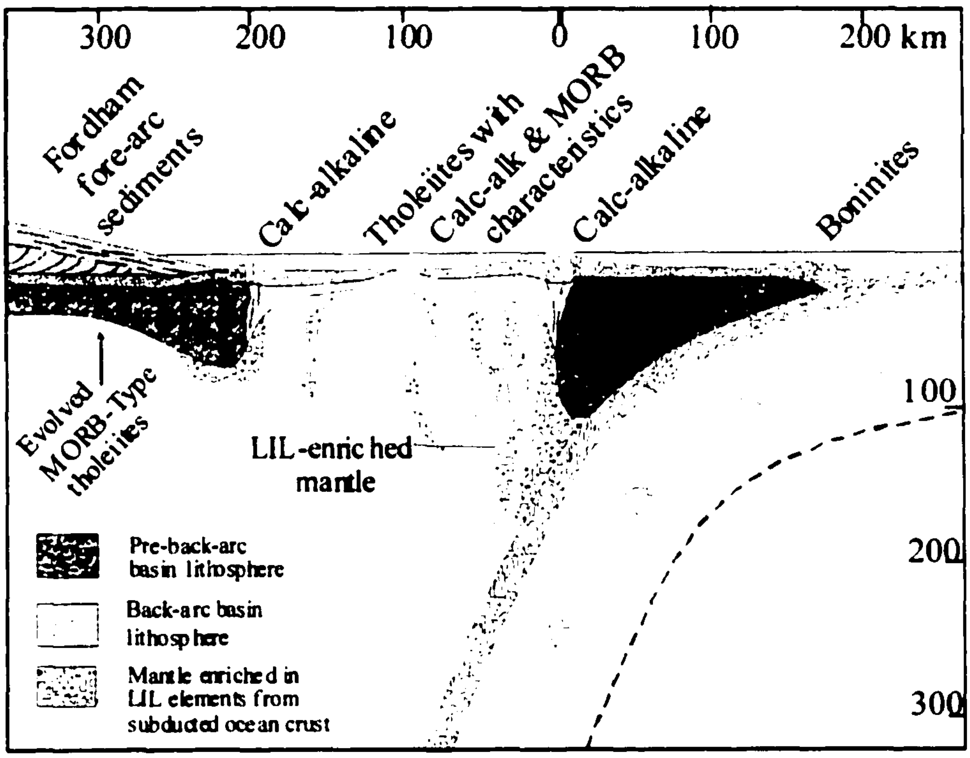

Figure 31. Schematic diagram (modified from Saunders and Tarney, 1984) depicting the BIG Complex island-arc-back-arc system. Initial island arc magmas are calc-alkaline with transitional back-arc tholeiitic basalts exhibiting N-MORB and calc-alkaline characteristics. Pre-back-arc basin lithosphere is represented by evolved N-MORB type tholeiites with overlying Fordham fore-arc sediments. The scale of the depicted basin is that from the Mariana and Parece Vela basin (Saunders and Tarney, 1984) and is used here as a reference model of the BIG Complex.

A juxtaposition of calc-alkaline and tholeiitic rocks, consistent with the proposed model for the BIG Complex, is shown in Figure 31 (modified from Saunders and Tarney, 1984). Saunders and Tarney describe an evolution of rock signatures related to the development of an arc/back arc basin in the Pacific. Here, calc-alkaline IAB evolve during initial stages of back arc basin formation. But, as the basin matures and widens uncontaminated tholeiitic N-MORB type basalts enter the basin and contribute to an alteration of signatures as lavas co-mingle (Figure 31). The Mariana and Parece Vels basin model of Saunders and Tarney (1984) appears to produce ratios of calc-alkaline to tholeiite that approximate those of the BIG Complex.

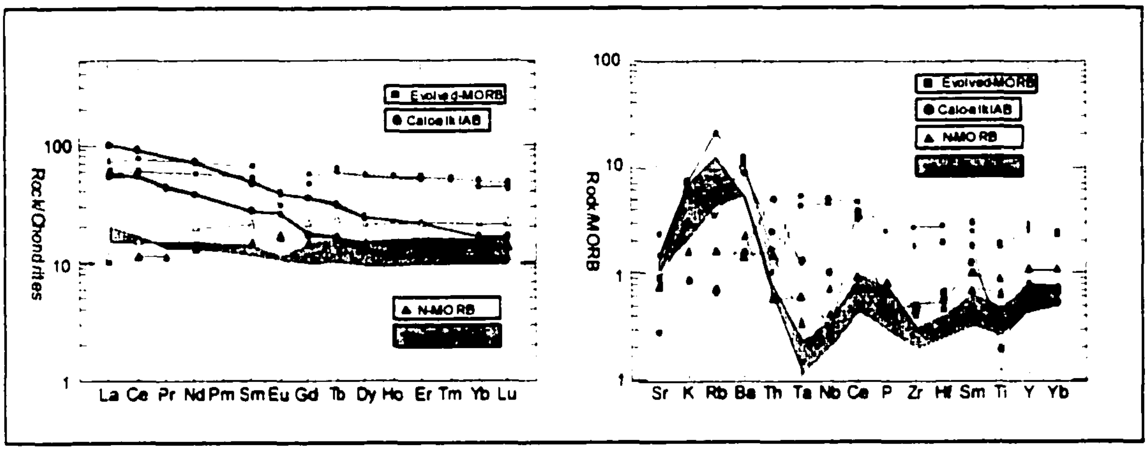

Figure 32. REE and trace element variations in the N-MORB and 'evolved' N-MORB, shown previously in Figure 29, compared to the calc-alkaline and tholeiitic IAB from the BIG Complex. The calc-alkaline IAB appears to be derived from incorporation of the pre-existing ocean crust (represented by the 'evolved' N-MORB) with newly introduced tholeiitic N-MORB. The calc-alkaline IAB exhibiting characteristic enrichment of Th, Ce, P and Sm (Pearce, 1982).

The REE and trace element profiles of both the calc-alkaline and tholeiitic IAB with the 'evolved' and typical tholeiitic MORB (Figure 32) show that the calc-alkaline

IAB have compositions intermediate between the evolved MORB and the typical MORB. The close association between the calc-alkaline IAB and the 'evolved' N-MORB therefore strengthens the argument for the evolution of the BIG Complex in a back-arc setting.

5.3c Volcanic Arc Granites Whole rock

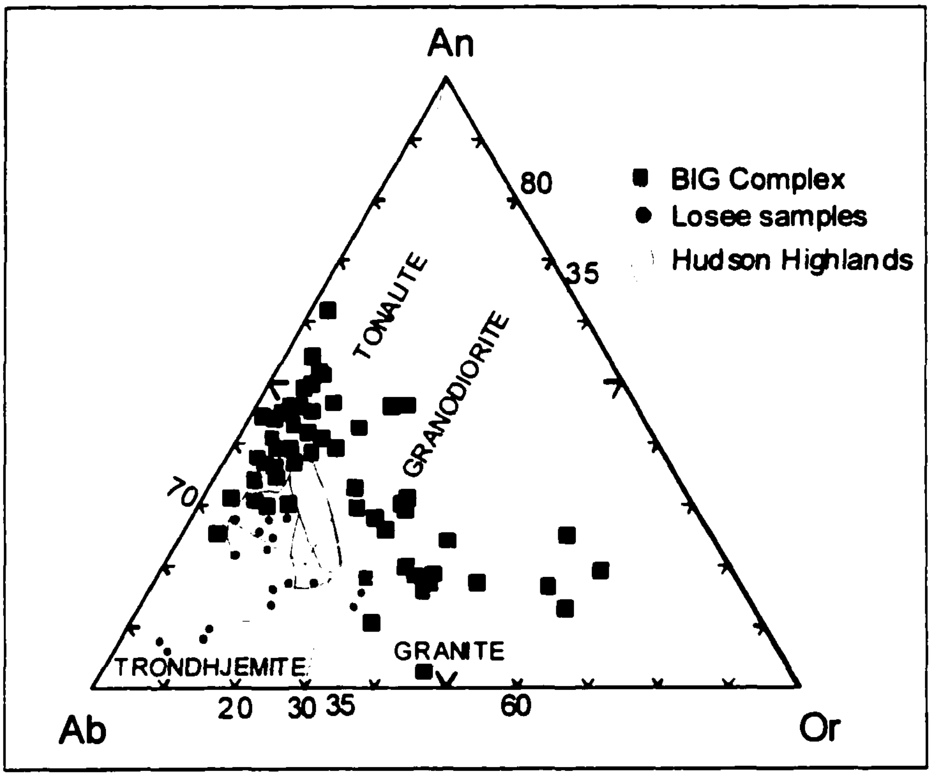

Figure 33. Comparison of BIG Complex feisics with the Losee samples of NJ and Hudson Highlands (see Puffer and Volkert, 1991).

Accompanying the IAB are a large percent of VAG that plot predominately as tonalites with some granodiorites and minor amount of granites on a normative feldspar plot (Figure 33). The overwhelming percentage of tonalites among the felsic rocks stand in marked contrast to the Middle Proterozoic Losee felsics[15] of New Jersey and the Hudson Highlands (shaded region in Figure 33) that plot predominately as trondhjemites (Puffer and Volkert, 1991).

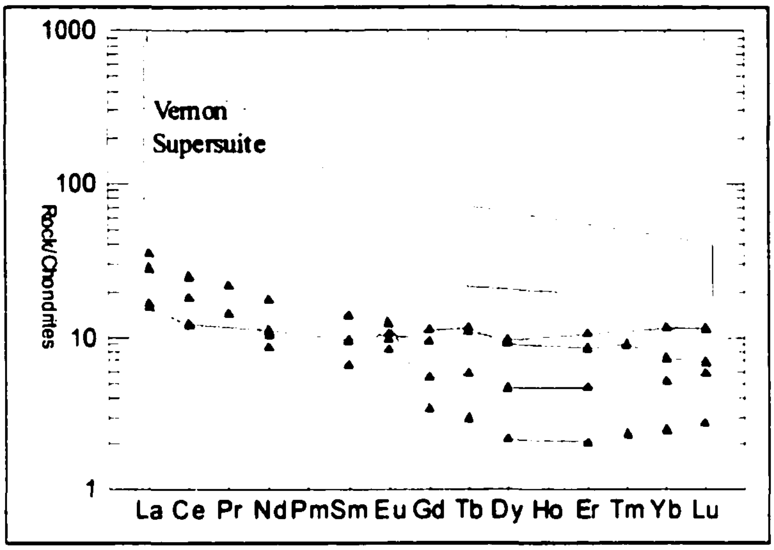

Figure 34. Comparison of REE concentrations between the Ravenswood tonalites and the Vernon Supersuite of the New Jersey Highlands of the Reading Prong (data from Volkert et al., 2000). All data normalized to Nakamura (1974).

Intrusive to the calc-alkaline Losee Suite is the ca. 1,110 Ma tholeiitic Vernon Supersuite (Volkert et al., 2000). The monzonite through granite composition and REE profiles of the Vernon Supersuite exhibit a marked difference from the Ravenswood felsics (Figure 34). The Vernon rocks display gently sloping HREE-depleted patterns and moderate LREE to HREE fractionation thought to be a result of partial melting of sub-continental lithospheric mantle and subsequent fractionation of alkaline magmas (Volkert et al., 2000). The Ravenswood calc-alkaline have similar REE patterns and may represent a transitional phase between the Vernon Supersuite and the Ravenswood

[15] The Losee Metamorphic Suite is interpreted as a ca. 1.3 Ga calc-alkaline continental magmatic arc assemblage composed of dacite, tonalite, trondhjemites and basalt (Volkert et al., 2000).

tholeiitic enderbites, which present flatter and lower REE concentrations and are more consistent with fractionation of the low-K tholeiite magma.

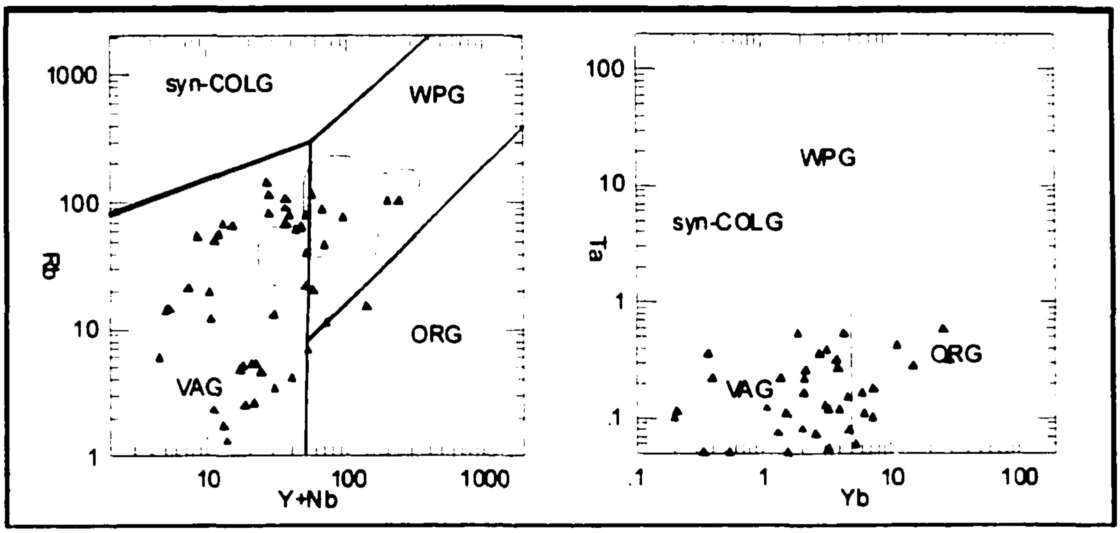

Fig. 35. A) Rb vs. Y + Nb graph of the BIG Complex felsics (SiO2 > 64%) which plot predominately in the volcanic arc granite (VAG) field compared to Volkert et al., (2000) Vernon Supersuite (shaded region) which plot predominately as within plate granites (WPG). B) Graph of BIG felsics using HFS elements Ta vs. Yb removes the effects of K-RB metasomatism indicating possible ocean ridge granites (ORG) with volcanic arc consistent with an ophiolite/volcanic arc complex determination (after Pearce and others, 1984).

Pearce et al.'s (1984b) Rb against Y + Nb discrimination diagram further underscores geochemical differences between the Ravenswood Granodiorite and New Jersey Highlands Vernon Supersuite (Figure 35). In this view, Ravenswood lithologies correspond best with the volcanic arc section as opposed to New Jersey felsics which plot as within-plate granites. Volkert and others (2000) suggest an interior origin for the Vernon Supersuite, citing conditions similar to those produced in ensialic back-arc basins[16], with magma emplacement occurring in extensional fault zones created during compressional tectonics.

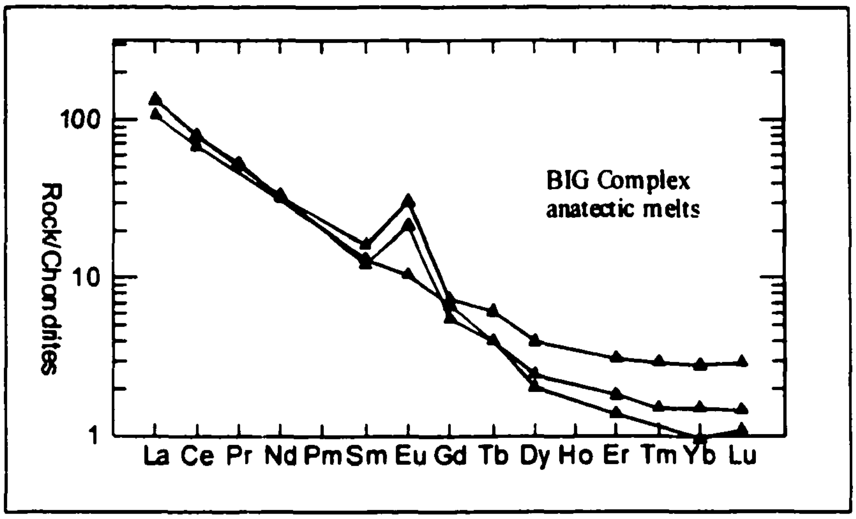

Figure 36. Subgroup of LREE enriched trondhjemites and granites representative of anatectic melts within the BIG Complex.

There is however a subgroup of the BIG Complex felsics, typically minor trondhjemite and granite sills, that exhibit LREE enriched profiles (La/Yb >6) comparable to tonalites and trondhjemites from the NJ Highlands Losee (Puffer and Volkert, 1991). As discussed by Puffer and Volkert (1991), the NJ tonalites are interpreted as melts extracted from an eclogite residue with the trondhjemites, in turn, a product of local anatectic melting of tonalite. The BIG Complex LREE enriched felsic subgroup has similar REE patterns with that of leucosomes sampled along the borders of BIG Complex gabbros and ultra-mafics xenoliths (Figure 36). A preliminary interpretation is that these units represent post island arc (i.e. Grenville or younger) anatectic melts derived from deeply seated eclogite residue that has migrated up sequence

[16] Volkert and others (2000) appear to describe an ensialic back-arc basin but do not use the term back-arc (see section 8.2 for tectonic discussion).

along the regional foliation. A corresponding leucosome generated granitic intrusion is the Canada Hill Granite with estimated age of 1,010 My (Aleinikoff and Grauch, 1990).

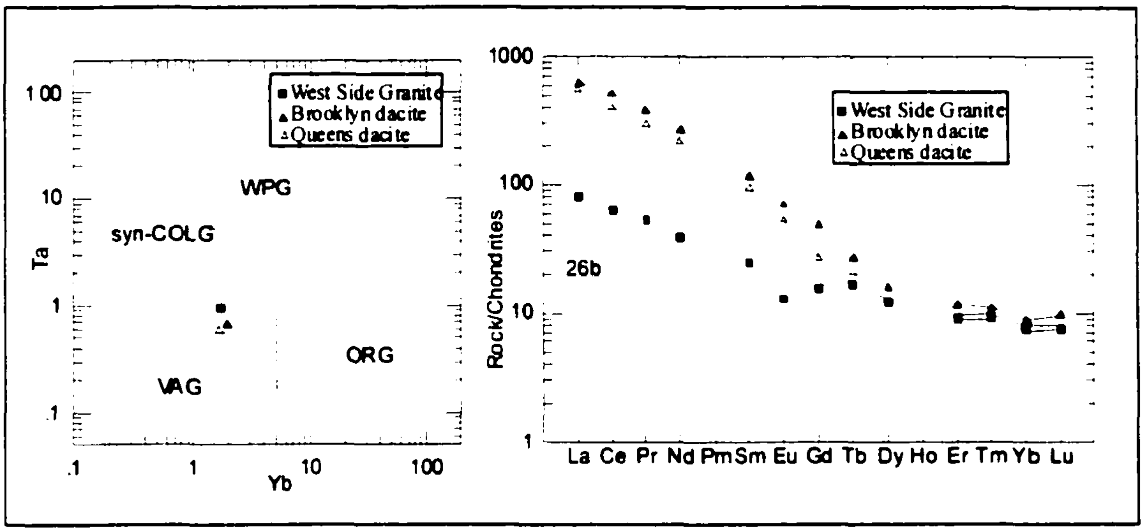

Figure 37. SSZ-VAG signatures of Paleozoic igneous intrusions within the New York City section of the Manhattan Prong.

A recently discovered, extensive trondhjemite body was mapped on the West Side of Manhattan during the sinking of Shaft 26B for the Manhattan spur of City Tunnel #3, Stage 2 (see Appendix II for shaft geology map). This trondhjemite (termed the West Side granite after Chesman 1997) is thought to be correlative to Westchester County Devonian age felsic intrusives also present within the Manhattan Prong (Mose and others, 1976). The West Side granite also exhibits a VAG geochemical signature (Figure 35) indicating continued subduction related events within the Manhattan Prong. No intrusions of Devonian trondhjemites were noticeable within the Brooklyn-Queens tunnel exposure of the BIG Complex; however the Alleghenian porphyritic dacite dikes of Brooklyn and Queens also exhibit a VAG trace element signature and SSZ discrimination signature (Figure 35) indicating the repeated subduction related events that have intruded and thus affected the Manhattan Prong rocks. REE concentrations of the dacite to the

West Side granite (Figure 37) show similar HREE abundances but a marked increase in middle to LREE. A petrogenetic correlation of these Paleozoic intrusives to an underlying mantle source is presently under study (Chesman and Steiner, in prep).

5.3d Back–Arc Basin Basalts

The geochemistry characteristics presented above for the Ravenswood Suit clearly agree best with a composite volcanic arc/ophiolite complex source region. The reverse zoning of plagioclase in the volcanic arc felsics, as stated in section 3.3b, is thought to be representative of decompression melting (Aldiss 1981). The buoyancy force associated with rising diapirs above subduction zones is considered a crucial factor in producing tensional surges in development of back-arc basins (Jurdy and Stefanick, 1983).

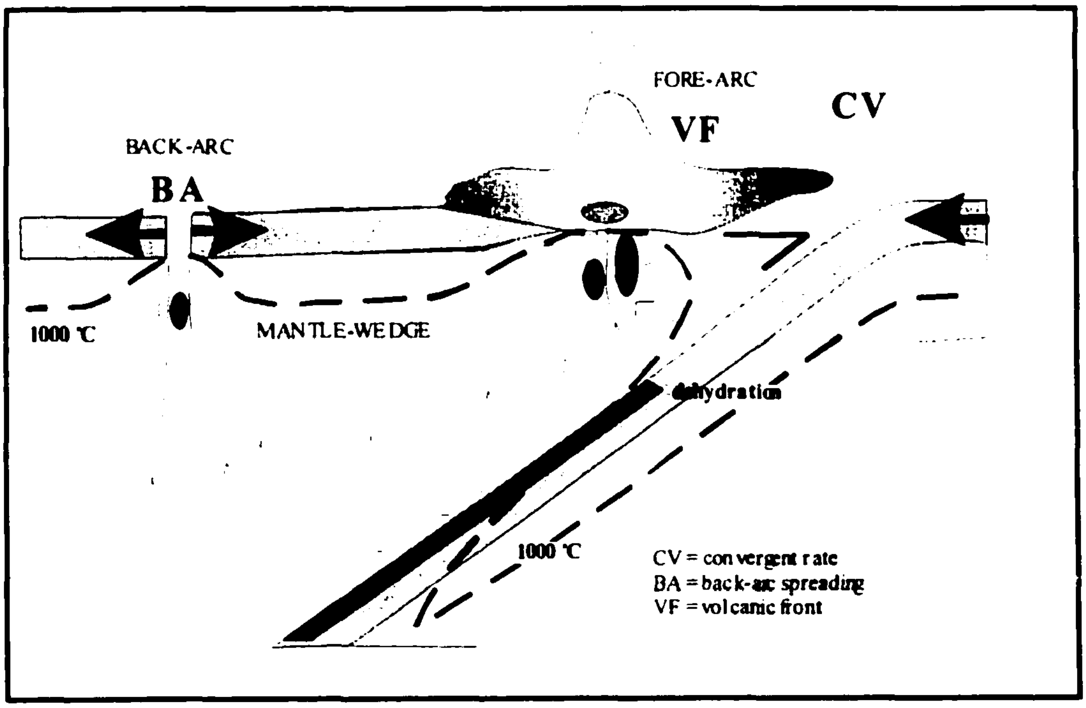

Figure 38. Schematic for the development of a back-arc basin (modified from McCulloch and Gamble, 1991)

Back-arc basin basalts (BABB) typically display trace element patterns that are transitional between MORB and IAB (Weaver et al., 1979; Saunders and Tarney, 1984; Woodhead, et. al., 1993). The transitional nature of the BAB signature is believed to be generated by partial melting of the upper mantle (Woodhead, et. al., 1993). Convective flow (Woodhead, et. al., 1993), assisted by pulling down of the mantle at the trench (Jurdy and Stefanick, 1983), causes mantle from the back-arc region to circulate into the wedge corner where IABs are generated as shown in Figure 38 (after McCulloch and Gamble, 1991). Melting and magma mixing decrease intermediate trace metal values and thus the transitional signature.

Repeated subduction-related fluxing and partial melting of the mantle wedge continuously depletes the mantle in Ta, Nb and other HFSEs. Initial back-arc basalts are marginally depleted in these elements but by the time the mantle makes it to the wedge corner where arc magmas are sourced, it is so depleted in Ta and Nb that there is very little left to contribute to IABs (Pearce 1983; Saunders and Tarney, 1984, Woodhead et. al. 1993). Thus the BABB are moderately depleted while the IAB's are severely depleted in these elements. However, the IAB generated early in the creation of an island arc may be only slightly depleted in the HFSE since the magma is produced from a mantle wedge has not been subjected to numerous partial melting events. Thus there can be some confusion in assigning moderately depleted basalts to either a BABB or IAB classification.

Woodhead et al., (1993) used HFSE ratios to alleviate uncertainty concerning the set of elements contributed by the subducted slab and/or mantle in the petrogenesis of arc magmas. Due to the lower mean abundances of incompatible HFSE (TiO2 and Zr) in IAB than in their BABB counterparts a ratio of V/Ti verse Ti/Zr is commonly used to differentiate between BABB and IAB.

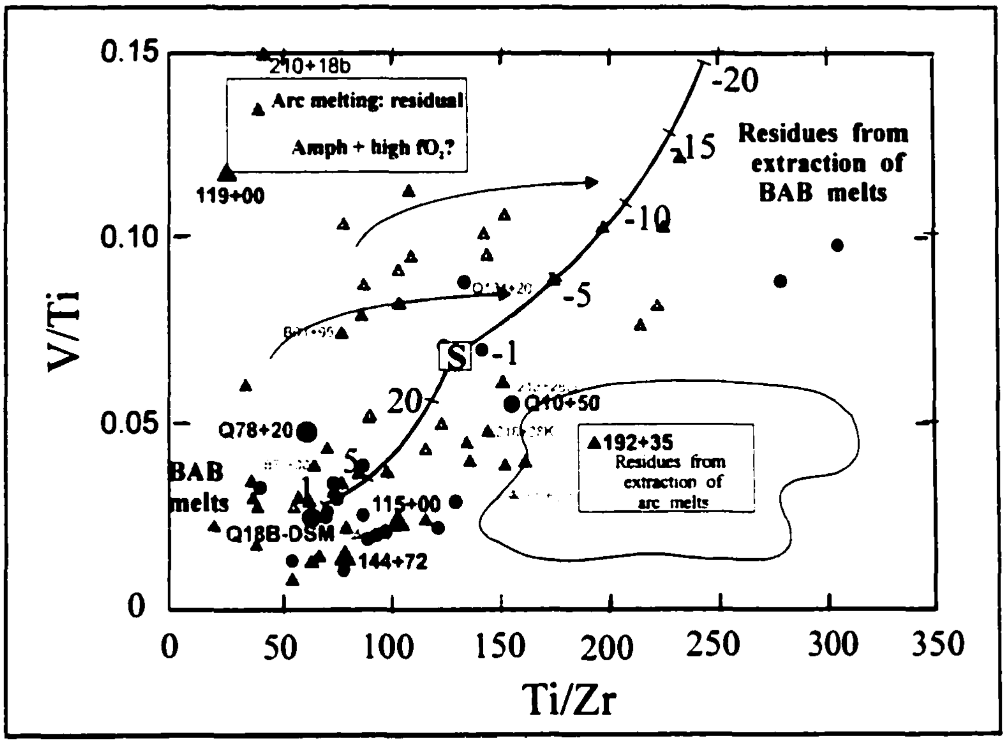

Figure 39. Plot of the BIG Complex mafics (<54% SiO2) on the Woodhead et al. (1993) diagram for generation of BABB from MORB source. A large percentage of BIG Complex mafics plot favorably in the field of BAB melts supporting the presence of a Middle Proterozoic Fordham back-arc basin.

BABB are assumed by Woodhead et al., (1993) to be generated from an N-MORB upper mantle composition source (S in Figure 39). The removal of BABB melts drives the residual source composition to higher Ti/Zr and V/Ti values (upper curve).

The Ravenswood's typical N-MORB composition, Q10+30 and 192+35, shows elevated Ti/Zr relative to Woodhead & others' N-MORB estimate. This variance from Woodhead et al. may reflect differences in the mantle source region or arise during regional metamorphism. Nevertheless, partial melting of these otherwise typical tholeiitic MORB appears to have generated a large group of Ravenswood mafics that plot as BABB according to the model of Woodhead et al.

Extraction of the BAB melts drives the residues toward higher Ti/Zr and V/Ti levels where batch melting may yield arc lavas of comparably higher Ti/Zr. Data also indicates that a number of Ravenswood tholeiitic basalts correspond to the Woodhead et al. field of IAB. This is consistent with the trace element discrimination diagrams discussed earlier. However, the calc-alkaline IAB, as earlier defined, plot as BABB. This is thought to be a consequence of derivation from the evolved MORB, rather than N-MORB. The evolved MORB plot within the Woodhead et al. BABB field in Figure 39, even though they have a MORB trace element profile (as discussed earlier). The close geochemical association of the evolved MORB with the calc-alkaline IAB (depicted earlier in Figure 32) may be explained as a derivation from a non-depleted mantle wedge during the initial formation of the volcanic arc. Such a setting would give the calc-alkaline units a moderately depleted HFSE signature, similar to a BABB, but with a more pronounced SSZ LILE enrichment.

The plotting of the evolved MORB in the BABB section (Figure 39) is most likely due to the HFSE ratios, expressed in the trace element discrimination patterns exhibited earlier in Figure 28. Woodhead et al.'s (1993) N-MORB averaged elemental data is different than the elemental abundance of the evolved MORB and thus the ratios would most likely plot differently in Figure 39. However, the possibility that these ratios may be due to the formation of metamorphic garnet and an abundance of ilmentite in the evolved MORB samples cannot be ruled out. Garnet preferentially incorporates V in its structure (Gaines et al., 1997) and the BIG Complex garnet rich samples have shown whole rock V enrichment. The increased presence of ilmentite, in what appears to be secondary precipitation of opaques in the garnet rich samples, likewise would increases the overall TiO_2 content. There is also increase in Zr, although less defined, with the BIG Complex garnetiferous samples. Magnetite is another opaque that is quite common in the Ravenswood mafics. As stated previously, crystallization rates can also affect element ratios where Ti and V concentrations tend to decrease relative to Zr and P due to early magnetite crystallization (Pearce and Norry, 1979). These variations can create differences in Ti/Zr and V/Ti ratios.

The incorporation of HFSE into refractory minerals, such as titanite and rutile, has been proposed as an explanation for the extreme depletions of Ta and Nb respectively in IAB (Saunders et al., 1980; Ringwood, A. E., 1990). Although recent studies have disputed this assumption (McCulloch and Gamble, 1991; Woodhead et al., 1993) others have suggested that amphiboles, stable in the hydrous mantle wedge may play an important role as depository for these elements (Bonatti and Michael, 1989; Woodhead et al., 1993). Fortunately, some meta-harzburgites, considered by the author to represent depleted mantle, were encountered in the tunnel and sampled to ascertain their possible petrogenetic relationship to the Ravenswood Suite.

5.3e Meta-harzburgites

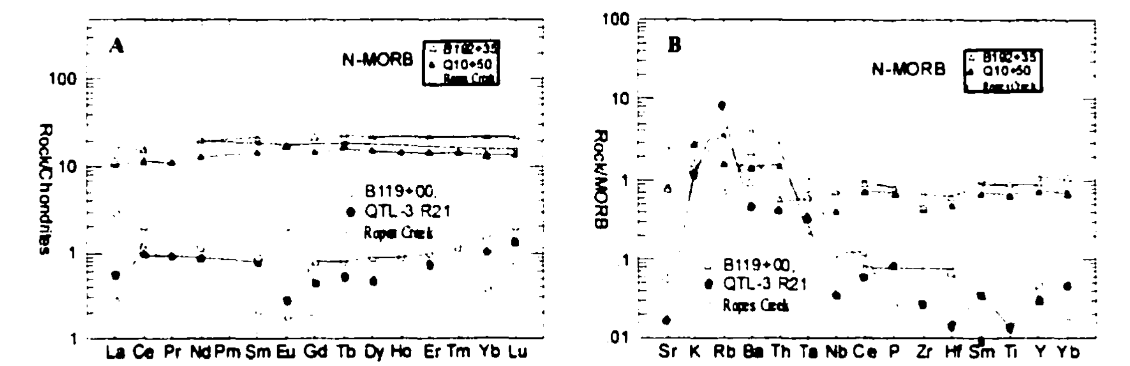

Figure 40. Chrondrite normalized REE concentrations and trace element discrimination diagrams comparing the BIG Complex with the Ropes Creek Ophiolite Complex (Spell & Norrell, 1985); A) Chrondite normalized REE concentrations of the BIG Complex meta-harzburgites compared to N-MORB type meta-basalts. B) Trace element discrimination diagrams (Pearce 1984a) of BIG Complex meta-harzburgites showing LILE enrichment and extreme HFSE depletion typical of a depleted mantle wedge. BIG Complex N-MORB and Ropes Creek samples shown for comparison.

Meta-harzburgites are presently interpreted as relic pieces of the Proterozoic depleted mantle wedge. The normalized REE concentrations and trace element profiles of the BIG Complex are similar to those of the Ropes Creek ophiolite assemblage (Spell and Norrell, 1985) and are shown for comparison in Figure 40b. Recent studies have indicated that these are SSZ-ophiolites (Spell and Norrell, 1985; Elthon 1991) using arguments advanced in Section 5.3a.

The subduction zone schematic (Figure. 38) depicts the widely held view that the mantle wedge is invaded by LILE enriched fluids that are driven off the subducting slab and sediments. These fluids are thought to enter the mantle via tensile cracks in the mantle wedge generating melts by fluxing (Davies & Stevenson, 1992). Differentiation of these produces mobile magmas ranging from gabbros to tonalities that generate a buoyancy force that decouples portions of the overlying mantle. The mobilized magma is also thought to stratify forming tonalite and cumulate gabbros with interstitial chromite. Fortunately, this decoupling allows for relict pieces of the mantle to be brought up with volcanic arc diapirs, as rolled xenoliths. Reversed zoning of plagioclase in the SSZ tonalites and enderbites (Chesman 1997a) is considered to form by incorporation of mafics perhaps accompanied by decompression melting during ascension of the tonalitic diapirs (Aldiss 1981). This constitutes additional evidence for the buoyancy effect during the development of the Ravenswood Island arc.

The metasomatic processes occurring within a subduction zone setting may also be responsible for the production of pargasitic varieties of amphiboles as typifies the mantle wedge (Davies and Stevenson, 1992) and mantle rocks (Seyler & Bonatti, 1988) in several localities. Steiner & Chesman (1996) support this idea based on a preliminary geo-thermometry calculations for the BIG amphibolites. The presence of residual amphiboles that retain Zr and Ti can be associated with the depletion of HFSE's in volcanic arc basalts derived from the partially melted mantle wedge (Woodhead et al., 1993).

The incorporation of the meta-harzburgites within plutonic IAB and VAG suggests that the units were part of the mantle wedge where generation of the Ravenswood island arc magmas was sourced. The extreme depletion of the HFSE, Ta and Nb, by realistic degrees of melting of an N-MORB alone cannot explain the low abundances in IAB (Pearce et. al. 1984a). Since the HFSE abundances of subduction zone magmas are thought to be a feature of the basalt's mantle source (Saunders and Tarney, 1984), the source region of IAB magmas should be infertile, depleted and refractory compared to the MORB source region (Davies & Stevenson, 1992). The geochemical profile of the BIG Complex meta-harzburgites is consistent with a depleted mantle source for the Ravenswood Suite which may have resulted through repetitive partial melting of the mantle. Portions of the back-arc volcanic pile may then have advected into the unstable mantle wedge and been involved with the melting regime that produced later arc volcanism. This type of repetitive melting has been proposed by Bonatti and Michael, (1989) to explain the geochemistry of mantle peridotites associated with subduction zones (see also McCulloch and Gamble, 1991; Davis & Stevenson, 1992; Woodhead et al., 1993).

In examining the geochemistry of volcanic rock suites it is sometimes assumed that the units with the most depleted trace element signature are source rocks for the remaining units. In order to evaluate this petrogenetic assumption, the REE abundances of the less depleted units of the BIG Complex are normalized to the most depleted unit. This comparison allows a magnified view of the variation from rock to rock than is possible on a chondrite normalized plot (Hanson 1989). Tholeiite whole rocks from the Ravenswood Suite (Table 3) are normalized to the averaged REE chondrite profile of the meta-harzburgite samples in Figure 40. Although the Eu anomalies in the two meta-harzburgite samples are severe, Sun & Nesbitt (1978) and others have argued that this feature may be caused by weathering, oxidation state or other factors; the averaged profile of the two depleted mantle samples tempers the Eu anomaly for a clearer overall comparison with the Ravenswood suite. The overall trends in Figure 41 are remarkably consistent showing pronounced increases in the LREE. With volcanic suites, it has been found that LREE concentrations typically increase more rapidly or to the same extent as HREE with increasing silica content (Gill 1981). Therefore fractionation is a viable explanation for some of the LREE variability. Although isotopic analyses are not yet available, the REE profiles are a strong indicator that the Ravenswood tholeiitic and calc-alkaline IAB are derived from a mantle source similar to the tectonically entrained meta-harzburgite.

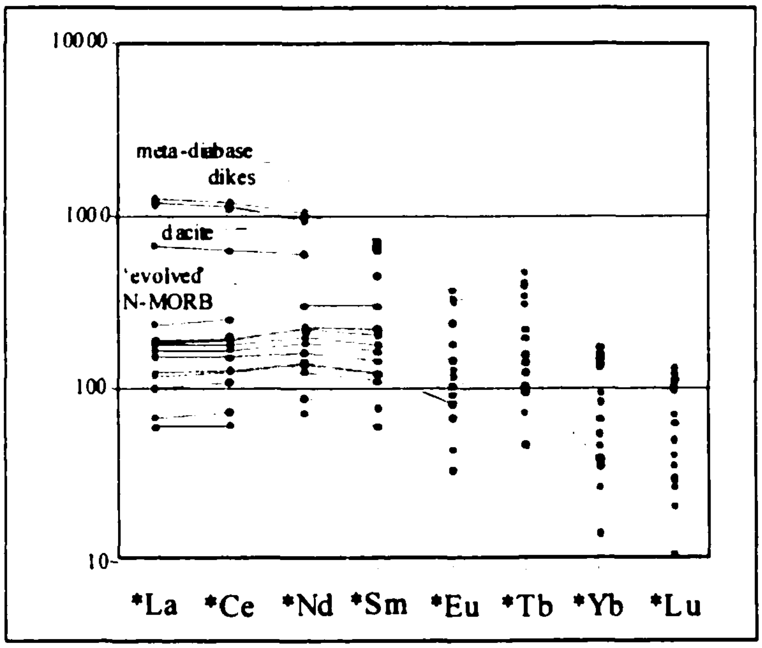

Figure 41. Whole rock REE profiles normalized to the depleted mantle meta-harzburgite samples. The typical Ravenswood and evolved MORB samples exhibit parallel patterns suggestive of a common parent.

An interesting association is presented in Figure 41 in regards to the later meta-diabase (Q41+60 and 4+30) and porphyritic dacite intrusions. Although there is a pronounced LREE increases the overall general pattern remains similar to those of the Ravenswood suite indicating a similar mantle source underlying the eastern seaboard from the Proterozoic through the Paleozoic.

6 Fordham Gneiss

A model has now been presented for the generation of the Ravenswood Suite in a volcanic-arc/marginal back-arc basin setting in full agreement with the early Berkey model (1931) for the partial assimilation of the Fordham Gneiss. The importance of the Fordham gneiss relationship to the other regional units was perhaps best stated by Prucha (1956);

> **"An understanding of the relationship of the Fordham gneiss to the other formations of the New York City group and to the Precambrian gneisses of the Hudson Highlands is a prerequisite to solving the stratigraphic correlation and structural problems of the extensive metamorphic terrane lying east and southeast to the Green Mountain anticlinorium axis."** (Prucha 1956, pg. 674)

The Proterozoic Y Fordham gneiss has been interpreted as a highly variable, predominately clastic and volcanic eugeosynclinal sequence, subjected to upper amphibolite metamorphism (Merrill 1890; Berkey 1910; Hall 1968; Aleinikoff 1985). Berkey (1907) detailed the high variability of the lithologies included within the Fordham Gneiss formation, which he considered to include banded granitic gneiss, hornblendic gneiss, micaceous and quartzose gneisses; mica-hornblende-chlorite-quartz- and epidote-schists; along with garnetiferous, pyritiferous, graphitic, pyroxenic, tremolitic and magnetic schists and gneisses; crystalline limestone, serpentinous limestones, ophicalcites, serpentine, tremolitic limestone and quartzite are also involved. Berkey (1907) considered the Fordham gneiss to be a tectonically deformed interbedded suite belonging to the Grenvillian Series (similar to those found in the Adirondacks and Canada). Berkey (1907), Balk (1936) and Robinson and Hall (1982) considered the Fordham Terrane to be coeval equivalent and lateral variant to the Hudson and New

Jersey Highlands, while Knopf and Jonas (1929) and Higgins (1972) broadened the regional correlation and considered the Baltimore Gneiss and Glenarm group rocks to be equivalents to the Fordham and other Manhattan Prong rocks.

These early studies established the existence of a Proterozoic eugeosynclinal sequence, correlative and possible inclusive with the Fordham sequence, extending laterally over much of the Laurentian margin. These interpretations have been supported by recent work on the Proterozoic Grenville front that assigned these rocks to an oceanic setting along the eastern margin of Laurentian (Rivers 1997). Thus the key issue of the BIG Complex verse related eugeosynclinal rocks relates to oceanward variations from the proto-Hudson Highlands to the proto-Ravenswood.

The highly variable stratigraphy of the Fordham, including serpentines and interbedded marbles (Berkey 1907), is consistent with volcanogenic sedimentation found in marginal basins (Karig and Moore, 1975; Carey and Sigurdsson, 1984). Given the presumed proximal relationship of the Fordham Gneiss to both the Laurentian continental based volcanics and the Ravenswood island arc, one would expect to find volcanic detritus as well as fore-arc terrigenous sediments within the greater Fordham gneiss set (i.e. Brooklyn to Queens). These arc and terrigenous sediments would collect preferentially in the Fordham basin adjacent to the Highlands, as compared to episodic volcanogenic sediments entirely toward the continental margin (see discussion relative to Figure 43).

Systematic facies transitions are rare in an arc environment due to the large influx of volcanoclastics and random sediment dispersion (Carey and Sigurdsson, 1984). As a back-arc marginal basin develops, large tectonic displacements create ridges and basins that result in a lack of vertical and horizontal sedimentary symmetry (Karig and Moore, 1975). Lithologically variable turbidite deposits as well as interfingered volcanic debris from possibly several volcanic islands and the continental margin may contribute to the sediment mix. As a basin subsides and widens Fe and Mn pelagic sediment, as well as biogenic calcareous ooze, are deposited directly on newly formed back-arc crust (Karig and Moore, 1975). Nevertheless, the Fordham's lithologic complexities (Berkey 1907; Prucha 1956; Hall 1968) are consistent with the sedimentary sequences recorded in modern back-arc basins (Karig 1971; Marsaglia and Devaney, 1995). Therefore the complex stratigraphy leads to difficulty in interpreting a depositional environment as well as a complicated geochemical signature.

6.1 Geochemistry of the Fordham Gneiss

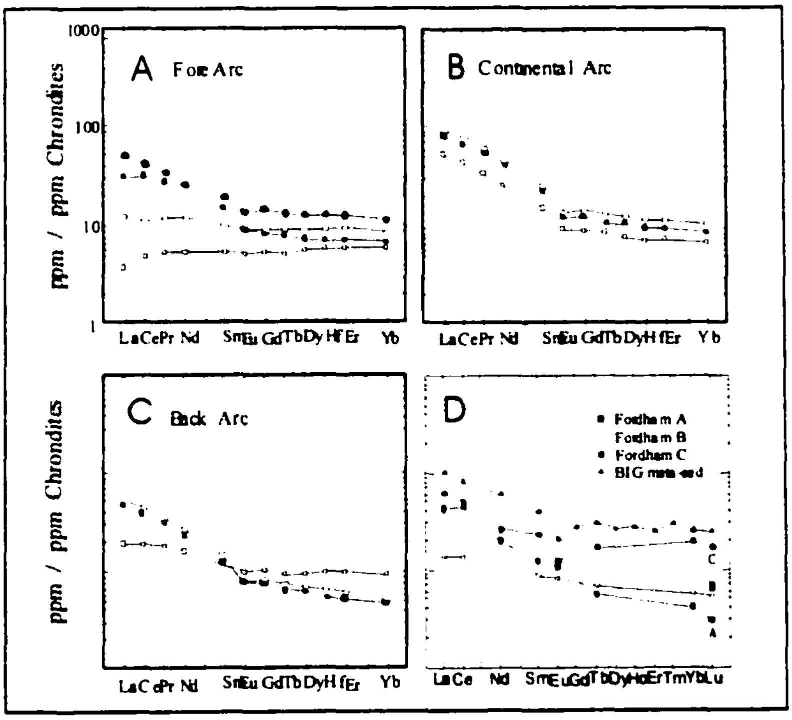

Figure 42. REE profiles of McLennan and Taylor's (1991) fore-arc (A); continental arc (B); and back-arc (C) sediments compared to the Fordham a, b, c and a BIG Complex meta-sediment (D).

Although the Fordham Gneiss exhibits a high level of stratigraphic variability it should retain geochemical signatures consistent with both fore-arc and back-arc sediments. In fact a comparison of the REE profiles from Hall's (1968) Unit A, C, and E (Steiner, unpublished data) and a sillimanite bearing garnetiferous gneiss (correlative to Hall's (1968) Fordham D, from Station 176+00 of the Queens tunnel) corresponds extremely well with McLennan and Taylor's (1991) examination of REE profiles from mixed fore-arc sediments with modern back-arc greywacke-shale turbidites (Figure 42).

The high degree of variability exhibited by the Fordham Gneiss (Berkey 1907; Hall 1968; Baskerville 1992), also explains the discordant age dating of the Fordham (Mose 1982; Aleinikoff 1985). U-Pb zircon date from a meta-igneous granitic sill within Hall's (1966) meta-sedimentary Member C[17] established a minimum age of 1,170 My for the Fordham (Aleinikoff 1985) with an earlier study yielding a Rb-Sr date of ca. 1,300 My for the same unit of the Fordham[18] (Mose 1982). The ca. 1,300 Ma date is considered here as representative of either volcanic detritus associated with the 1,400 to 1,230 Ma Laurentian continental magmatic arc (Rivers 1997), represented in the N.J. Highlands by the ca. 1,300 Ma Losee Metamorphic Suite (Puffer and Volkert, 1991; see also Volkert et al., 2000), or later terrigenous sediments derived from the Losee. Furthermore, Mose (1985) recognized a 1,100 Ma disturbance in the Fordham Gneiss that

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[17] There appears to be some confusion in recent studies suggesting that Member C of the Fordham is a meta-igneous unit. However, Aleinikoff (1985) is clear when he states, **"The Fordham Gneiss has been subdivided by Hall (1968) into five members all of which are metasedimentary."** The 1,170 Ma zircon age comes from an intrusive **"granitic gneiss within Member C"** and thus **"is regarded as a minimum age for the formation of the Fordham Gneiss".**

[18] There is some debate as to the accuracy of the Rb-Sr date of Mose (1985) due to later disturbances. This topic will be examined later in the thesis.

could correspond with disturbances associated with intrusion of the 1,110 Ma N.J. Highlands' Vernon Supersuite of Volkert et al., (2000). The Hudson Highlands have also yielded U-Pb zircon ages of 1,170 and 1,060 Ma (Tilton et al., 1960; Aleinikoff et al. 1982) and 1,150 Ma (Long and Kulp, 1962) from the Hudson Highlands.

Therefore, with consideration to the above range of igneous activity concentrated along this section of the Laurentian margin, it is likely that Fordham back-arc sediments as well as the associated Ravenswood volcanic arc/back-arc intrusives will contain inherited age components. In addition to inherited components associated with Fordham fore-arc or terrigenous sediments, it is suggested here that inherited age components may also be contributed by way of melting of the sub-continental lithosphere that was formerly associated with the Highlands' calc-alkaline magmatism. The sub-continental lithosphere is believed responsible for the calc-alkaline signature of the early stages of the Ravenswood volcanic arc and back-arc magmas (see section 5.3b).

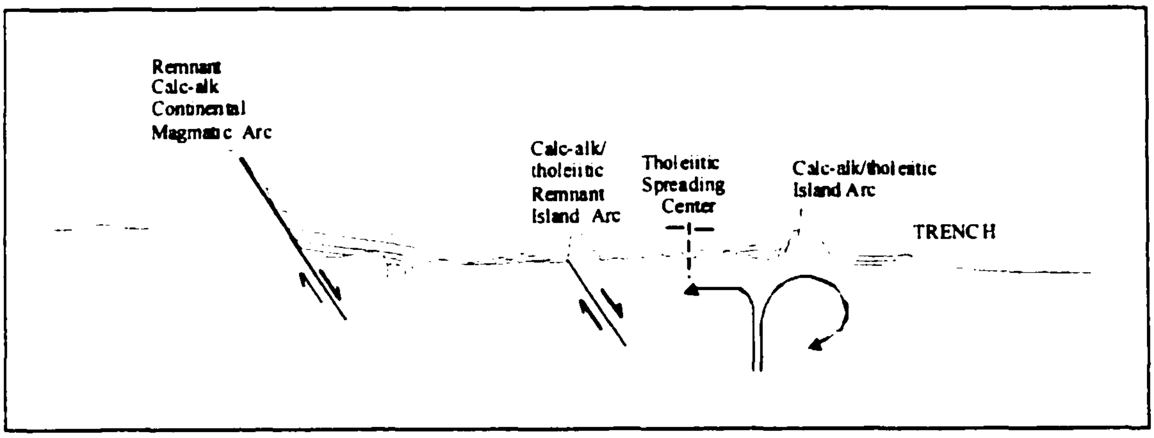

Figure 43. Tectonically controlled sedimentation features in marginal back-arc basins. (After, Karig and Moore, 1975; Bibee, et. al., 1980; Marsaglia and Devaney, 1995).

In the development of back-arc marginal basins, important age relationships are created by crustal extension occurring behind the arc. Karig and Moore (1975) point out that the older basins are adjacent to continental margins with other basins generally younger toward the arc (Figure 43). The estimated life of a back-arc has been calculated at 50-60 My after initial formation (Jurdy and Stefanick, 1983). Remnant arcs, which can be early formed calc-alkaline arcs closest to the continent, act as barriers to protect the younger basins and back-arc basins from overloading with terrigenous sediment. This relationship is shown in Figure 43 where physical obstacles are created to limit symmetrical sedimentation within the basin (Karig and Moore, 1975; Lewis and Pantin, 1984; Carey and Sigurdsson, 1984).

Figure 44. A) Tunnel section exhibiting deformed basalt originally injected into unconsolidated sediments. B) Contemporaneous deformation of the sediments and basalts sills.

Garnetiferous meta-sedimentary unit consisting of interbedded marbles, gneisses and schists, have been found to be intruded by basalts (Figure 44). It is interpreted that the basalts were injected into an unconsolidated lower section of Fordham back-arc sediment and subsequently deformed and stretched to form boudins with the sedimentary units[19].

The extensional setting and magmatic activity of a back-arc basin is generally recognized as a prime setting for hydrothermal activity (Ishibashi and Urabe, 1995). The pyrite rich meta-sedimentary units and the basalts from the BIG Complex, shown in Figure 44, are typical of hydrothermally affected sediments deposited in back-arc basins as interpreted by Carey and Sigurdsson (1984).

Figure 45. Coticule rocks in the meta-sedimentary units of the Brooklyn Tunnel.

[19] Basalt intrusions are found in the lower part of the Okhotsk and Japan Seas back-arc sedimentary sequence (see Sychev and Sharaskin, 1984).

The presence of coticules (Fig. 45; garnet, quartz, pyrite and biotite layers) is another stratigraphic feature generally considered to form as the result of hydrothermal activity (Spry 1990). The garnets in the Fordham coticules are almandine in composition, which has caused some local debate as to the correct classification and inferred depositional setting (Merguerian 1981; Baskerville et. al., 1987). Although coticules have classically been considered to contain manganese-rich garnet and quartz (Lamens et al., 1986; Spry 1990) the petrographical description of rocks consisting of primarily quartz and garnet is currently accepted as defining coticule, even though the garnets may be almandine in composition (Cornell et. al., 1996; Ratcliffe personal comm. 1996). The presence of almandine-bearing coticules is used by Cornell and others (1996) to assign a metamorphosed Mesoproterozoic sequence similar to the Fordham to a back-arc basin environment in the Natal Province of South Africa. The REE profiles of the Fordham coticules, one band found within the Fordham B of Baskerville (1992) on Roosevelt Island and the other in a metasedimentary section of CT#3 in Brooklyn, have identical abundances and profiles indicating the same lithology and are similar to the Natal Province REE profiles show in Figure 46.

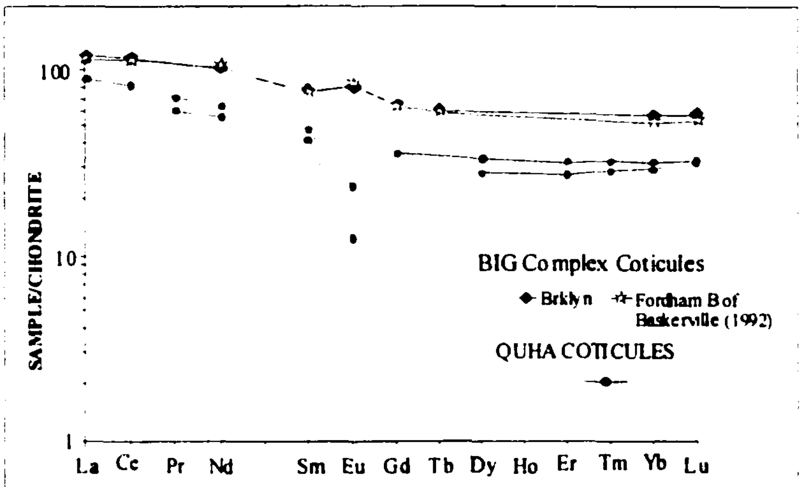

Figure 46. REE profiles from coticules from the Fordham Gneiss of Roosevelt Island and from Fordham metasediments from Brooklyn showing identical patterns. REE data from coticules of the Quha Formation, South Africa interpreted as Proterozoic back-arc deposits by Cornell and others, 1996) shown for comparison.

While mapping the tunnel exposure sedimentary horizons of the BIG Complex appeared to be laid down directly on the granulite basalts. Furthermore there is no sharp contact between the amphibolite to granulite grade Fordham gneiss meta-sedimentary units and Ravenswood granulite grade basalts. This contrast of metamorphic grade and resulting debate on the structural relationships between the Manhattan Prong rocks and the Highlands is a direct result of the unique combination of forces that are generated at subduction zones associated with back-arc marginal basins. The following section discusses the temperature and pressure environment associated with back-arc basins, the resulting petrographic features and the structural evolution of the basin.

6.2 The Fordham Marginal Back-arc Basin

The presence of granulite-grade mineral assemblages in an area that has supposedly undergone a regional Taconic age amphibolite grade event has been difficult to interpret (see section 3.3a). However, this dichotomy of metamorphic grade is now recognized to be typical of continental margins with complex tectonic histories (see Ellis 1987, Harley 1995, Gibson and Ireland, 1995). Martignole (1992) reviewed the exhumation model for the emplacement of high-grade terrane. In this model underplating during extension supplies the heat required for tectonic uplift resulting in a thickening of the crust from below. The underplating and uplift lead to the formation of a back-arc basin, and further extension allows for the eventual ascension and eruption of tholeiitic MORB into the basin relative to Figure 31, producing granulite grade temperatures at relatively shallow (<20 km) depths (Figure 47 after Gnibidenko et al., 1995).

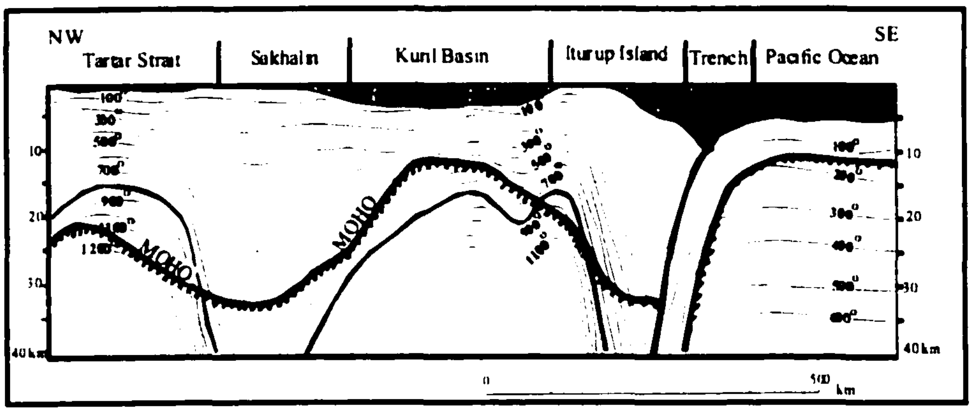

Figure 47. High temperatures at relatively shallow depths beneath the Kuril Arc and back-arc basin (after Gnibidenko et al., 1995).

Preliminary temperature and pressure study of the mineral pairs in the BIG Complex (Steiner and Chesman, in prep.) indicate conditions consistent with volcanic-arc/back-arc settings depicted in Figure 47 (Ernst 1974; Gnibidenko et al., 1995), as are geothermometry calculations based on amphibole data (Steiner and Chesman, in prep). The presence of high grade rocks this close to the surface allows for a shorter path-to-exposure as a result of tectonic forces commonly found in at destructive plate boundaries and thus the perseverance of granulite grade associates.

Jolivet and others (1994) describe the exhumation and juxtaposition of deep crustal metamorphic rocks via crustal extension in arc and back-arc regions. Jolivet and others proposed a change from HP-LT conditions during subduction which correlates to cool brittle-ductile transitions that may be sufficiently severe as to lead to deformation of pyroxenes (shown earlier in Figure 16B). The pressure and temperature drop during exhumation results in the formation of retrograde garnet coronas in HT-LP conditions (Jolivet et al., 1994) and agree well with present findings in the BIG Complex (shown earlier in Figure 16C).

Corona textures may also develop due to near-isothermal decompression and partial hydration of granulitic mineral assemblages in lower-crustal granulite and upper mantle rock (Muntener et al, 2000). The rapid exhumation of granulite grade rocks exhibiting corona textures is thus strongly indicative of decompression at the end of a prolonged period of progressive metamorphism, a model first proposed by Korja et al., (1996) to account of granulites preserved during thrusting of the Paleoproterozoic Lap Land belt. Extensional tectonics is also proposed to explain the exhumation of granulite grade rocks during continental extension in New Zealand (Gibson and Ireland, 1995). These processes are sometimes associated with concurrent migmatization, as detailed by Owen and Greenough (1997) for the formation of garnetiferous corona structures in amphibolite/granulite facies meta-basites from Grenville, Ontario. In this present instance exhumation and concurrent migmatization may also be associated with widespread metasomatic activity (section 4.1).

Clearly, further study of mineral associations is needed to clarify pressure-temperature paths for the BIG Complex granulites. This complex juxtaposition of rocks of variable metamorphic grade and the development of corona-relationships and other features is nevertheless symptomatic of the development and collapse of a back-arc basin (Bartholomew and Tarney, 1984; Jolivet et al., 1994).

As a consequence of the tensional and compressional forces described for the back-arc setting, the final collapse of the Fordham back-arc basin allows for reactivation of normal faults into thrust faults. These become detachments which unroof deep crustal HT-LP metamorphics (Jolivet et al., 1994). The reactivation of normal fault zones, especially the Ramapo Fault (Berkey 1907; Ratcliffe 1971) allows for upward thrusting of deep seated rocks. Multiple reactivations would generate repetitive openings and closures of the same basin which in turn creates a complex heredity for the BIG lithologies.

This scenario also supports the interpretation of the Fordham Gneiss as undergoing at least amphibolite grade metamorphism prior to the later incorporation in the Ravenswood and subsequent remetamorphism as proposed by Berkey (1931). Early metamorphism of older units of the Fordham Gneiss may be related to subduction at the base of a continental magmatic arc's fore-arc pile. Splitting of the fore-arc during step-back of the subduction zone, with subsequent volcanic arc and back-arc basin development is consistent with Berkey's (1931) observed intrusive relationship between the Fordham and the Ravenswood (shown earlier in Figure 26).

As noted, the Fordham gneiss units in Westchester and the Bronx are most likely derived predominately from the Highlands continental magmatic-arc, and to a much lesser degree from Ravenswood volcanoclastics. Sediments deposited eastward in the basin would eventually be isolated from terrigenous sources and reflect a Ravenswood-type signature and somewhat younger Proterozoic Y age dates. Grauert and Hall, (1973) found a wide spread of zircon ages in the Fordham Gneiss with some ages consistent with the presently proposed model for timing of the Ravenswood volcanic arc.

Berkey's (1933) interpretation that set Precambrian meta-sediments mostly to the east (as shown earlier in Figure 3) of the Ravenswood is supported by findings in the eastern most section of the CT#3 (approximately 1 km east of CT#2) of garnetiferous schist and gneiss associated with interbedded marbles (Figures 44 & 45). This stratigraphy, the presence of coticules and their association with granulite grade back-arc basalts suggest that this group of sediments formed in deeper portions of the Fordham marginal back-arc basin.

The presence of the Proterozoic ophiolite marginal back-arc/volcanic arc BIG Complex is consistent with proposals for exotic terranes existing east of the Laurentian margin (Farrar 1984; Rogers 1995). The recent discovery of Proterozoic basement in the Chesapeake Bay (Sheridan et al., 1999) further supports the age interpretation for the BIG Complex and supports the present suggestion for the evolution of Proterozoic rocks along the present eastern seaboard.

7. Geochemistry and timing of cross-cutting meta-diabase dikes

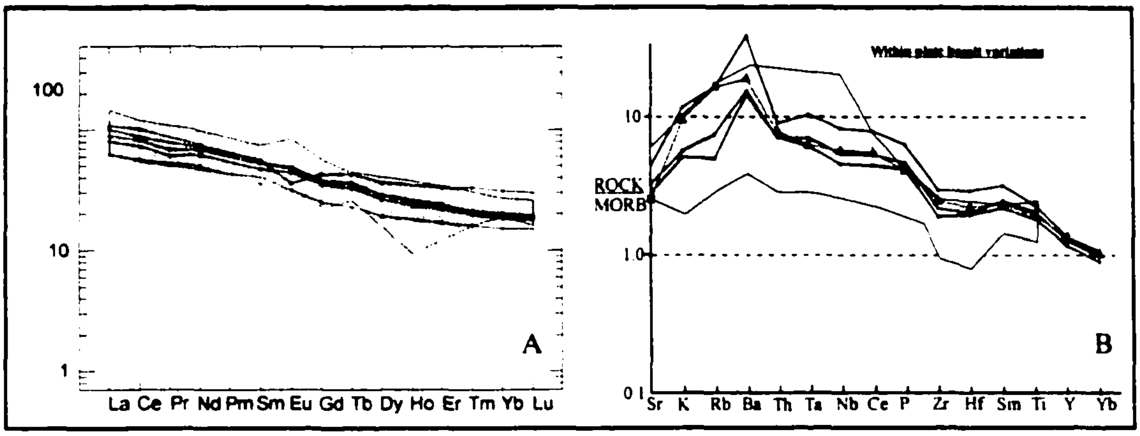

Figure 48. REE and trace element profiles of late Proterozoic diabase dikes from the BIG Complex compared to those found in the Hudson Highlands (Ratcliffe, 1987)

Plotting of the BIG Complex meta-diabase dikes on Pearce's (1982) multi-element discrimination diagram exhibits a pronounced within-plate basalt signature (Figure 48), which is in marked contrast to the IAB signature of the Ravenswood (shown earlier in Figure 29). The REE and trace element profiles of the BIG Complex diabase dikes are compared with those of the Hudson Highlands (shaded fields in Figure 48, data from Ratcliffe 1987) and exhibit almost identical profiles and abundances.

Ratcliffe (1987) considered the Hudson Highlands diabase dikes to be correlative to Vermont's dikes studied by Coish and others (1985) which although variable in chemistry are thought to represent late Proterozoic rifting events. Similarly, the 758 my Mt. Rodges metabasalts of Virginia are considered to represent an initial pulse of magmatism associated with rifting in the southern and central Appalachians which ultimately spanned 200 my (Aleinikoff, et. al., 1995).

The 564 my Catoctin metabasalts (Aleinikoff et al., 1995) represent the end stage of rifting and are considered by Aleinikoff and others (1995) to be representative of the numerous high TiO2 metadiabase dikes that intrude the Middle Proterozoic granitic basement in Virginia. Although no isotopic dating of the BIG Complex's metadiabase dikes were performed for this thesis, repeated injections are inferred by 1) cross-cutting of early diabase by later diabase injections (CT#3 mapping; see also Blank 1973) and 2) variable chemistries (Chesman, in-prep). Underplating of the BIG Complex during initial rifting events near the end of the end of the Late Proterozoic (ca. 750 Ma) may be recorded in some of the meta-diabase dikes while a Catoctin age relationship (ca. 564) with later cross-cutting dikes would correlate well with similar ages recorded for the Yonkers and Pound Ridge gneiss, 563 and 583 my, respectively (Long 1969; Mose and Hayes, 1975). Establishing a genetic relationship of the BIG Complex's within plate metadiabase dikes with these Avalonian age New York City area felsic orthogneiss bodies would be a significant contribution in completing a Proterozoic Z extensional setting for the region.

Although the occurrence of these dikes may not be unusual in Precambrian rocks (Aleinikoff, et. al., 1995) the petrographic fabric of the dikes holds important clues to the tectonic development of the Manhattan prong. But prior to a Paleozoic deformation discussion some interesting structural correlations should be made in regards to the onset of the Proterozoic Z extensional event.

8. Proterozoic Tectonics

Geophysical studies of deep seated suture zones is providing increasing evidence for repetitive accretions onto the North American Craton's since the Archean (Burke et al., 1976; see also Rivers 1997). Kearey and Vine (1996) cite evidence for a Wilsonian cycle of subduction and ophiolite obduction for Proterozoic mobile belts at plate margin settings. Although it is difficult to recognize ophiolite complexes greater than 1,000 Ma old (Moores 1986), Meso-proterozoic, and older ophiolite complexes involving back-arc basins are now being recognized around the world (Windley 1993; Keppie and Ortega-Gutierrez, 1999; Johnson and Oliver, 2000). It is presently suggested that the Mesoproterozoic rocks of the BIG complex represent a post-Ottawan pulse of a Wilsonian cycle for the eastern seaboard of the present day North American continent.

Recent studies on the Mesoproterozic southeastern Laurentian margin (Figure 49) indicate coeval cycles of arc formation and subduction-accretion from 1,750 to 1,200 Ma with a gross SE-younging of major lithologic units. Remnants of back-arc basins interrupt this sequence at random intervals (Rivers 1997). In this model, the present day New Jersey and Hudson Highlands characterize a calc-alkaline magmatic arc at the Laurentian continental margin (midrange of the northern stippled belt, Figure 49; after Hoffman 1991) within the Green sector of the Blue-Green-Long[20] axis (Rankin 1976). The NJ Highlands, Hudson Highlands and Reading Hills forms the longest uplift (>200

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[20] Rankin (1976 p. 5605) proposed the term "Blue-Green-Long axis" for the trace of the axes of three sectors of Grenville age anticlinoria (chain of massifs) along the western margin of the Appalachian metamorphic terrane. The Blue sector corresponds to the Blue Ridge anticlinorium in the southern and central Appalachians, the Green sector corresponds to the Green Mountains of Vermont in the central and northern Appalachians, and the Long sector of western Newfoundland (see also Rodgers 1995, p.464-465).

km) in the Green sector (Rodgers 1995, p. 467). Formations within the Highlands have metamorphic zircon U-Pb Grenvillian dates ranging between 1,170 Ma to 950 Ma (Tilton et al., 1960, Aleinikoff and Grauch, 1990; see Rankin and others, 1993 and references within) demonstrating a complex heredity.

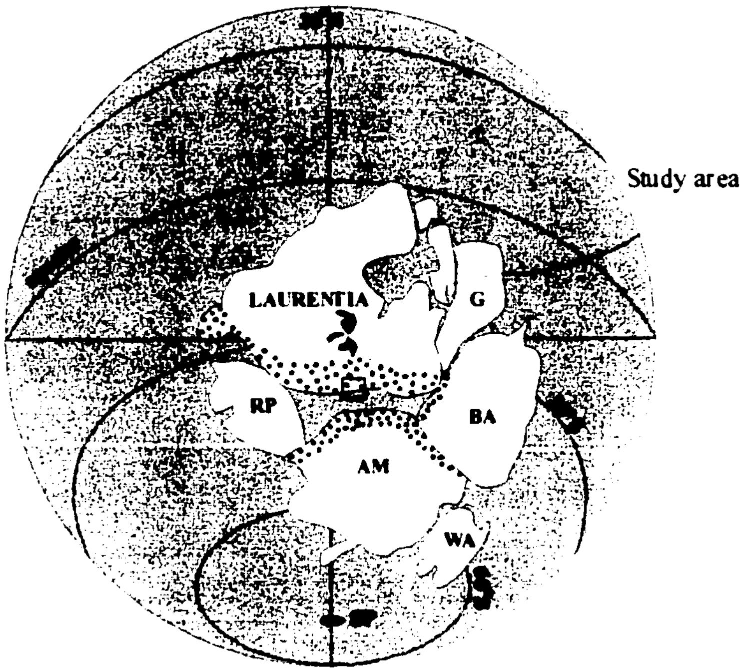

Figure 49. Reconstruction of the Rodinia supercontinent with cratons rotated to reflect a 1010 Ma Laurentian pole (after Weil, et. al., 1998). Grenvillian orogenic belts bordering a Grenvillian Ocean stippled after Hoffman (1991). Great Lakes on the Laurentian craton are shown for reference; present day eastern seaboard faces south along the Laurentian margin. AM = Amonzonia craton: BA = Baltaca (Fennoscandia): RP = Rio del La Plata craton: G = Greenland: WA = West Africa.

The Hudson Highlands can be separated into the present day western and eastern sections separated by the Proterozoic Canopus fault (Ratcliffe 1971). The zircon U-Pb upper intercept age of 1,135 ±11 (Ratcliffe and Aleinikoff, 1990, as reported by Rankin et al., 1993) for the Canopus Pluton of the eastern Highlands is similar to the ages interpreted for New Jersey Highland's meta-volcanics and meta-sedimentary rocks intruded by the Byram intrusive suite (Drake et al., 1991). However, Ratcliffe and others (1985, as reported by Rankin and others, 1993) used differences in the granitic intrusives to distinguish the western and eastern Hudson Highlands. A continuous correlation between the Fordham Gneiss of Westchester County with paragneisses of both the western and eastern Hudson Highlands has been proposed (see Rankin and others, 1993 and references within) suggesting the presence of a coeval isolated extensional basin between the sections of the Hudson Highlands. The southern reaches (present co-ordinates) of this isolated basin, demarcated by the southern limits of the eastern Hudson Highlands, transitions into pre-Ravenswood Fordham fore-arc continental margin sediments.

Further to the north, the Berkshire massif is included with the Grenvillian terrain (Rankin 1976). In parallel with the tectonic picture worked out in Figures 52-53, Rankin and others (1993) describes a felsic gneiss that rests unconformably on the 1040 – 1080 Ma Washington gneiss (Ratcliffe and Zartman, 1976). Rankin and others (1993) also relate the nearby Lee Gneiss of the Berkshire massif to a calc-alkaline Andean or ocean island arc setting. Associated shallow marine to marginal continental meta-sediments are considered by Rankin and others (1993) to suggest a possible a back-arc basin environment. We presently suggest that these Grenvillian units evolved in a tectonic setting similar or very closely parallel to that of the Fordham lithologies and provide support for the subduction-related petrogenesis of the Fordham-BIG complex.

Additional support can be found in studies of the 1 Ga basement rocks of South America, Middle America, Africa and Scandinavia (Sadowski and Betencourt 1996; McLelland et al., 2001; Keppi and Ortega-Gutierrez, 1999; Cornell et al., 1996; Romer 1996). Keppie and Ortega-Gutierrrez (1999) has suggested that these terranes originated as arcs in the Grenville Ocean between Laurentia, Baltica, and Amazonia (Figure 49). These authors related the various arcs to an ocean margin, but are generally non-specific as to the continental versus island arc setting for specific terranes. The present model attempts to distinguish the New Jersey and Hudson Highland gneisses as rocks with continental affinities (Volkert et al., 2000) as opposed to the BIG complex with more island arc/oceanic rocks.

8.1 BIG Complex: A Proterozoic Y Dismembered Ophiolite

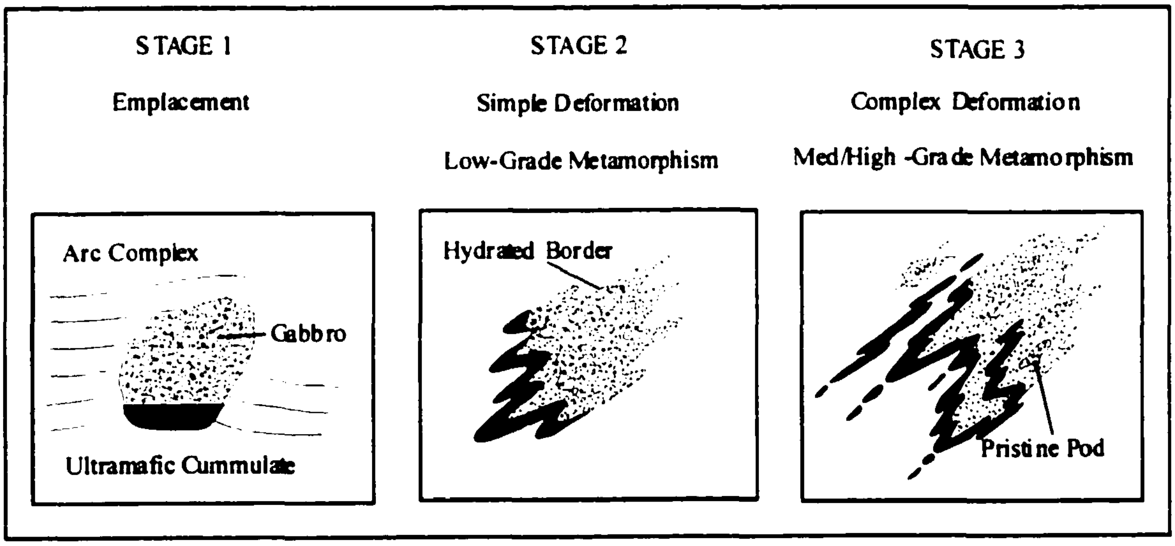

Figure 50. Hooper and Hatcher (1988) model for deformation of gabbro.

The Penrose Conference on Ophiolites (Anonymous, 1972) recognizes oceanic crust, MORB, as an essential criterion for ophiolites, along with pillow lavas, sheeted dikes, gabbros, and etc. (see the review by modern petrology text Ehlers and Blatt, 1982). These standard associations are disrupted and dismembered in high grade metamorphic terranes making the recognition of former ophiolite associations somewhat problematic. Workers such as Hooper & Hatcher (1989) show that the cumulate and other horizons become strung out and rolled during tectonic events (Figure 50), but suggest that their trace element signatures, such as those discussed at length in chapter 5, remain viable tracers for ophiolites. It was suggested earlier that MORB, IAB, some meta-sediments and depleted mantle rocks in fact retain their geochemical signatures and can be used to reassemble ancient tectonic configurations. The overall characteristics of the BIG tonalites and associated rocks are typical of back-arc basins with basalts possessing transitional geochemical characteristics between MORB and SSZ- IAB (Saunders and Tarney, 1984). The transitional geochemistry of the BIG Complex and associated evolved MORB, the Ravenswood volcanic arc suite and the tholeiitic MORB, are consistent with a marginal back-arc basin. This combination of dismembered ophiolite, marginal BABB and volcanic island arc rocks with SSZ signatures are now known to characterize a substantial number of recognized ophiolite complexes (Alabaster et al., 1982; Pearce et al., 1984a; Elthon 1991). This growing consensus underpins the following model for the evolution of the BIG complex.

Although most, if not all, recognized ophiolite complexes, especially those along the eastern seaboard are Phanerozoic or younger (Williams and Talkington, 1983), the present study shows that the geochemical and lithologic arguments are presently considered strong enough to characterize the Mesoproterozoic BIG Complex as an ophiolite complex. If this interpretation is correct, it represents the first recognition of a Mesoproterozoic oceanic back-arc basin - island arc composite in the Grenvillian Province of the Central Appalachians and is herein termed the Brooklyn Injection Gneiss (BIG) Complex after Berkey (1931).

8.2 Limited Wilsonian Cycle

It has been proposed that an Andean style[21] calc-alkaline continental magmatic arc was developed along the Laurentian margin between 1,400 to 1,230 Ma (Rivers 1997). The Laurentian magmatic arc is represented by the ca. 1.3 Ga Losee Metamorphic Suite (Puffer and Volkert, 1991), which comprised the root zone of the New Jersey Highlands (Figure 51a). Unconformably deposited on the Losee and the Mesoproterozoic ocean crust were sedimentary units consistent with the earliest units of the Fordham Gneiss[22]. Intrusive to the correlative Fordham units are various Hudson Highland plutonic rocks with zircon upper intercept ages of ca. 1,150 Ma (see Rankin et al., 1993) but with no clear tectonic setting having been defined.

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[21] For Andean margin studies that describe setting most consistent with the findings of this study please refer to; Bartholomew and Tarney, 1984; Storey and Macdonald, 1984; Alabaster and Storey, 1990.

[22] Rb-Sr and U-Th-Pb ages of the Fordham range from 1,300 Ma to 980 Ma (see Rankin et al., 1993).

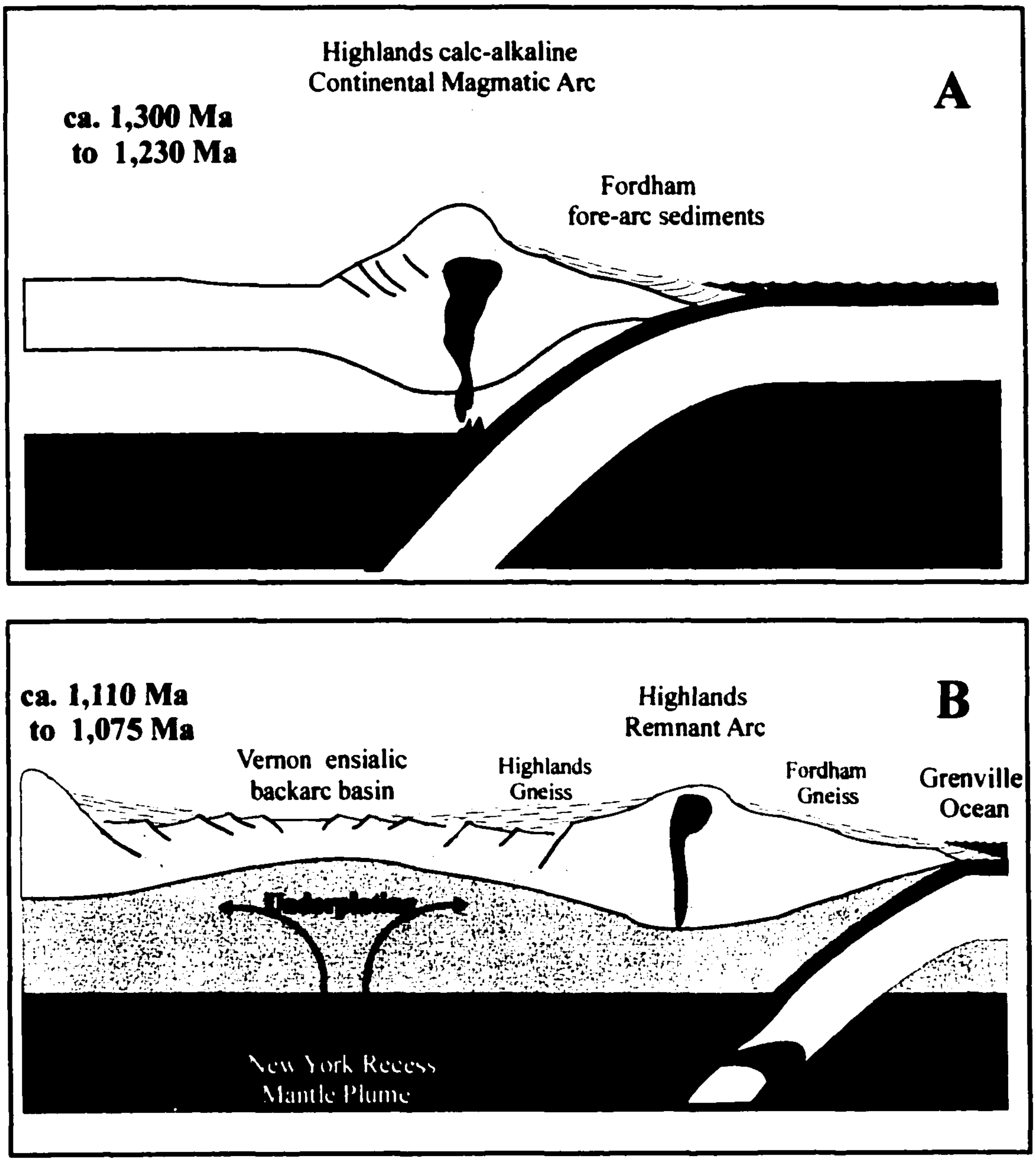


Figure 51. Interpretive north-south section across the Laurentian units (New Jersey and Hudson Highlands) prior to the development of the BIG Complex. A) Ca. 1,300 Ma to 1,230 Highlands calc-alkaline continental magmatic arc, source of the Fordham fore-arcsediments. B) Ca. 1,110 Ma to 1,075 Ma. Development of an aborted ensialic back-arc basin forming inland of the Laurentian margin (see Volkert et al., 2000). The time periodof 1,230 Ma to 1,110 Ma, although producing active plutonic intrusions into the Highlands is not depicted in this study and is awaiting further data, but is thought to be associated with continued subduction.

A recent study by Volkert et al., (2000) describes the emplacement of the A-type granitoids of the Vernon Supersuite, N.J. Highlands at ca. 1,110 to 1,074 Ma (abraded zircon Pb-Pb ages) as occurring during conditions of regional contraction. The alkaline geochemistry and within plate granite signature (see section xx) lead Volkert and others (2000, pg. 344) to reflect a origin, as a result of regional compression with a **"combination of crustal extension cratonward of the plate margin, delamination of the continental lithospheric mantle or slab detachment, and the melting of the lithospheric mantle by rising asthenosphere."** The Vernon granitoids are interpreted by Volkert et al. as being emplaced into extensional fault zones created during a compressional event[23]. Although not specifically stated by Volkert et al. it appears that the setting described is consistent with an aborted ensialic marginal basin (Figure 51b). This type of setting is consistent with that described by Aberg et al., (1984) for the early the Cretaceous Andean margin of central Chile. Alberg et al. cite the coupled action of plate subduction and ensialic spreading-subsidence along the margin, with spreading initiated by upwelling mantle material. The process described by Volkert et al. (2000) for generation of the Vernon Supersuite and a proposed presence of a Laurentian hot spot for mantle upwelling may have a profound affect on the later formation of the Fordham marginal back-arc basin.

[23] Volkert et al., (2000) consider the possibility of a Vernon Supersuite correlation with the 1,150 Ma Storm King Granite but require more petrologic and isotopic studies.

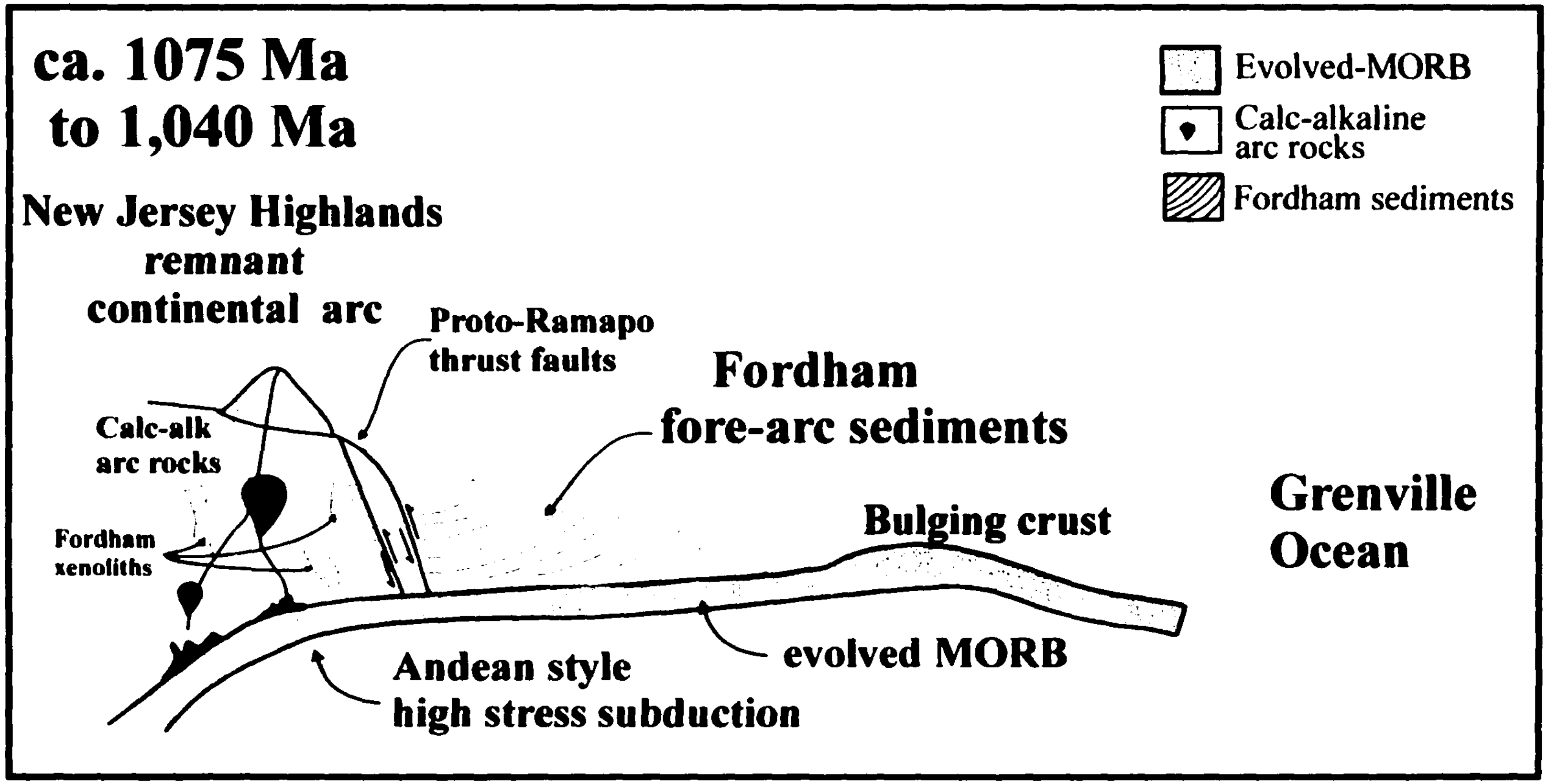


Figure 52. *North-south section across the Laurentian units (New Jersey and Hudson Highlands) showing a north-dipping Andean style subduction zone with volcanic arc (New Jersey and Hudson Highlands) and fore-arc sediments (Fordham sediments); drag produces deformed bedding in sediments and a fore-arc-bulge that leads in Figure 47 to a step-back subduction zone. Compression in the fore-arc leads to thrust faulting associated with the development of an early stage Ramapo fault during the Ottawan pulse of the Grenville Orogeny..*

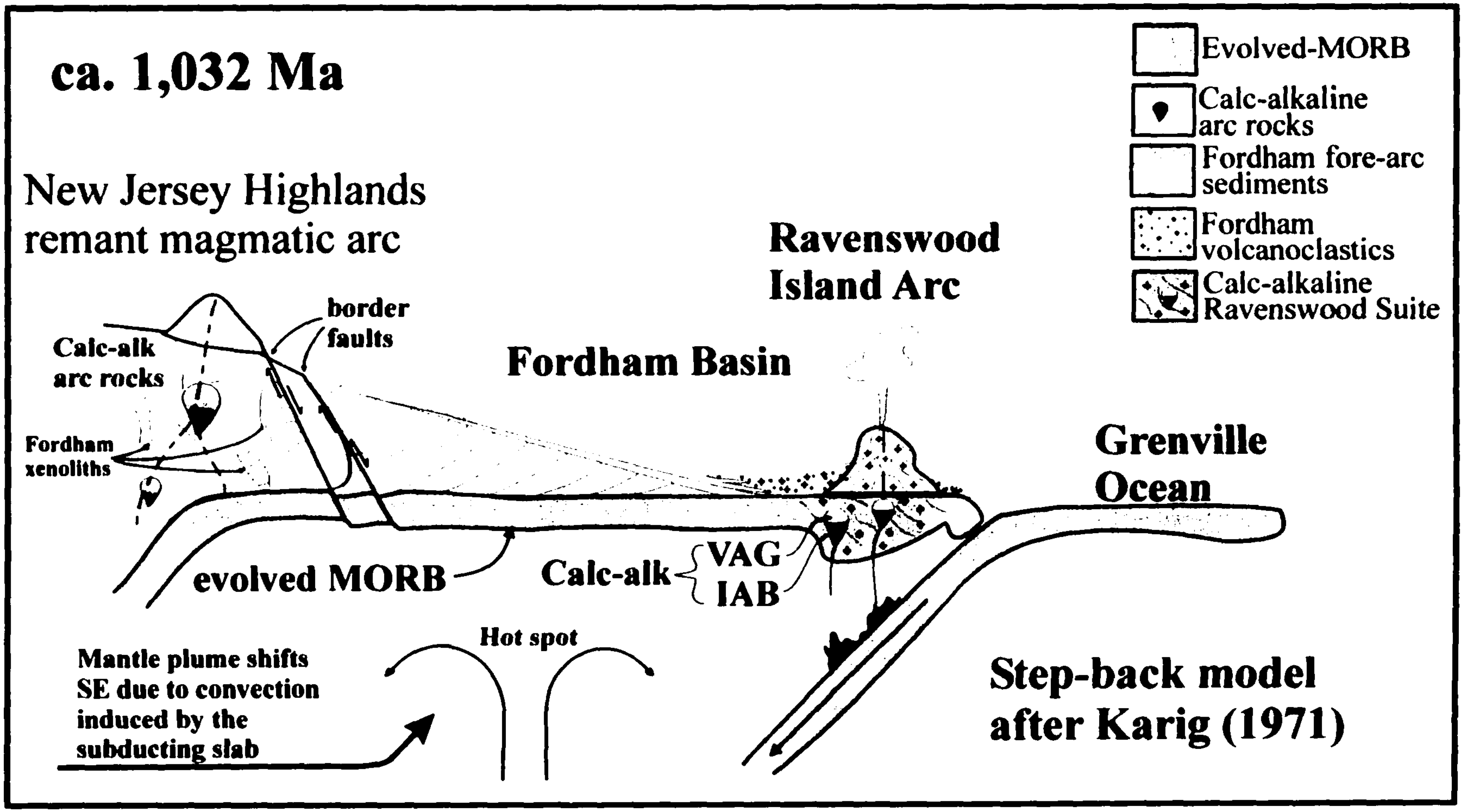


Figure 53. *North-south section across the Laurentian units (New Jersey and Hudson Highlands) showing a step back in underthrusting to form the Ravenswood Island arc. Calc-alkaline signatures are dominant in fractionated IAB and VAG. Fordham sediments include fore-arc and volcanoclastics deposited in the newly created Fordham Basin. Subduction and igneous activity has halted within the New Jersey and Hudson Highlands arc with normal displacement along the Ramapo fault..*

Volkert et. al. consider the Vernon Supersuite to have been emplaced prior to granulite grade conditions generated during the Grenvillian compressional event that occurred between 1,080 and 1030 Ma[24]. Although it is generally considered that the Grenville orogeny destroyed the intervening Grenvillian Ocean during a continent-continent collision in this section of Laurentia (Drake et al. 1991b) this study's findings of the BIG Complex MORB-IAB-BABB signature rocks with 1,032 Ma to 980 Ma Pb-U ages requires a different tectonic scenario.

The initial stage of the evolution of the BIG Complex involves a northward-dipping Chilean style subduction zone along the southeast margin of Laurentia (Rivers, 1997) (study area of Figure 49 enlarged in Figure 51 to 55). It is proposed that the compressional regime active at this time (ca. 1,080 to 1,030 Ma) generates a series of first-generation thrust faults, including the early thrust-fault stage of the Ramapo Fault, along the eastern margin of the Highlands (after Ratcliffe, 1971). Frictional drag is thought to produce downbuckling of deep-seated sediment, a fore-arc bulge, and eventually buckling of the ocean crust and a "step-back in underthrusting" (compare to Figure 53, see Karig, 1974). Compression of the deeper sections of the Fordham-forearc sediment pile (Figure 52 probably lead to an initial metamorphism of the Fordham Group (Berkey, 1907) that is superimposed by later metamorphic events related to back-arc spreading and other tectonic configurations. The level of this early metamorphism is presently held to have been completely over-printed by later metamorphic events. This

[24] The upper bound for the Grenvillian compressional event is constrained in the New Jersey Highlands by the undeformed Mt. Eve Granite, which has an upper intercept Pb-U age of 1,020+-4 (Drake et al., 1991a), although zircon overgrowths in the Canada Hill Granite indicate temperatures as high as 600^{O}C as late as 950 to 960 Ma (Aleinikoff and Grauch, 1990).

produces a minimum age for the Fordham meta-sedimentary Unit C at 1,170 My (Aleinikoff 1985) with a possible 1,300 My component (Mose 1982). In the widely accepted Bird-Dewey model for the Paleozoic margin (see discussion, Figure 58), compression of this sort also leads to basin formation and thus to a possible second-stage of deposition of the Fordham Group, illustrated for the ca. 1,080 to 1,030 Ma model in Figure 53 ('Fordham Basin' onlap set). This scenario is consistent with the proposed continuous correlation between the Fordham Gneiss of Westchester County with paragneisses of both the western Hudson Highlands and the eastern Hudson Highlands (see Rankin and others, 1993). Although lithologically similar to the tholeiitic Ravenswood suite, the zircon U-Pb upper intercept age of 1,135 ±11 (Ratcliffe and Aleinikoff, 1990) for the Canopus pluton and the distinct calc-alkaline association of the Wiccopee pluton of the eastern Hudson Highlands (see Rankin and others, 1993) indicates that the Ravenswood developed as a later, distinct Middle Proterozoic island arc.

Shifting of the Highlands subduction zone to the east, transformed the New Jersey Highlands into a remnant arc setting while initiating the development of the Ravenswood island arc to the south (east in Figure 53). The proposed tectonic scenario is supported by the differences in the Vernon Supersuite's youngest zircon age of 1,074 Ma, that may indicate timing for the collapse of the ensialic back-arc basin, with the initial calc-alkaline pulse of the Ravenswood volcanic arc at ca.1,032 Ma (see section 2). The inherited components of ca. 1,064 Ma for the Ravenswood enderbite may be indicative of the shifting of the mantle plume and underplating the marginal crust in the yet to be formed Fordham back-arc basin.

In the present model (after Saunders and Tarney, 1984), the initial phase of island arc magma genesis produces calc-alkaline volcanics due to the geochemical influence of the continental margin and possible subducted sediment. Volcanoclastic-derived sediments are deposited along the eastern portion of the Fordham back-arc basin; the western edge of the basin is bounded by normal faulting produced by reactivation of the proto-Ramapo fault (Figure 53). The Jurdy & Stefanick (1983) induced flow model, where a cold subducting slab produces forced convection to initiate a back-arc spreading center after a gradual stress buildup of 5-10 m.y. from the onset of subduction, is proposed for the development of the Fordham back-arc basin. The Fordham back-arc basin is initially floored by older oceanic crust or pre-back arc basin lithosphere, considered an essential condition for the occurrence of back-arc spreading (Molnar and Atwater, 1978 as reported in Jurdy & Stefanick, 1983), represented by evolved MORB (Figures 52 & 53 and discussed in geochemistry section 5.2a) situated between the Laurentian margin and the newly created Ravenswood volcanic arc. Although not presently evaluated, the evolved MORB of this study may be correlative to the New Jersey Hexacopf Formation mapped at the base of the New Jersey Highlands calc-alkaline rocks and considered by Drake (1985) to be pre-Highlands oceanic crust.

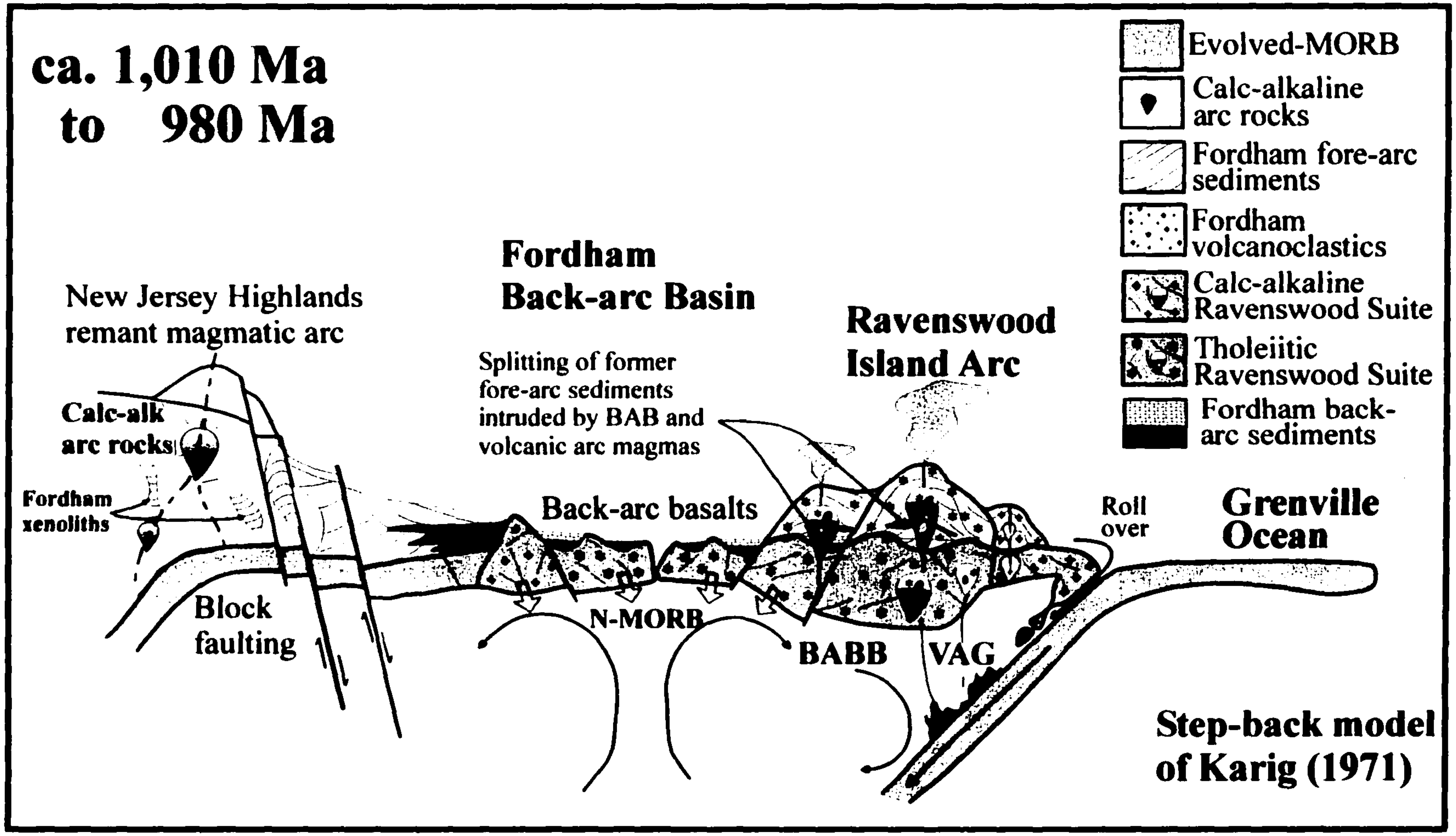


Figure 54. *North-south section across the Laurentian units (New Jersey and Hudson Highlands) showing the development of the Ravenswood Island arc and Fordham back-arc basin. Initial calc-alkaline chemistries has yielded to N-MORB and tholeiitic signatures in the IAB and BABB. Extension and settlement of the Fordham basin produces normal faulting along the borders and within the basin producing isolated sub-basins. Fordham back-arc marine sediments includes coticules and limestones.*

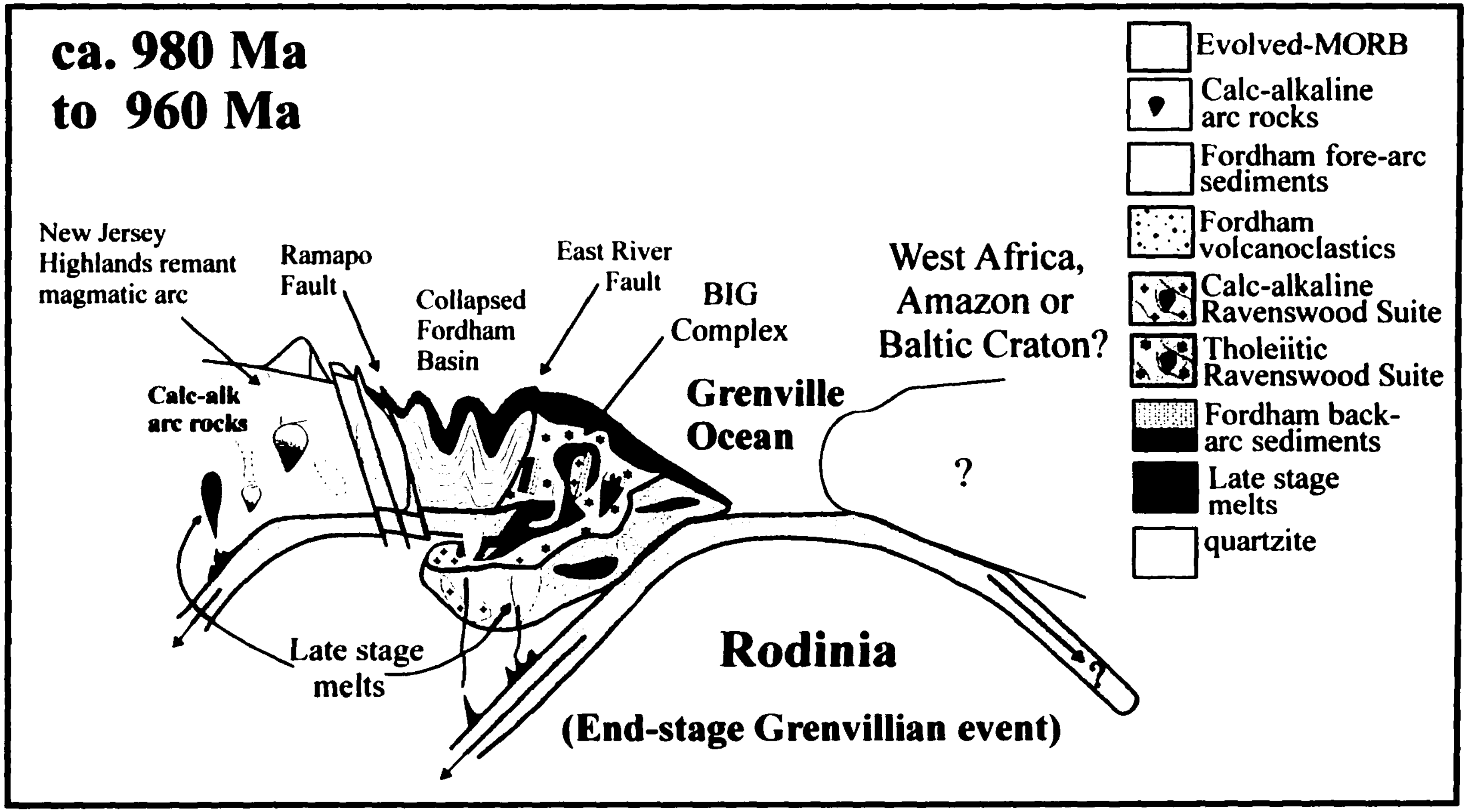


Figure 55. *North-south section across the Laurentian units (New Jersey and Hudson Highlands) showing the collapse of the Fordham back-arc basin reactivating both subduction zones to generate minor late stage melts. Thrusting along reactivated normal faults exposes deep-seated granulite grade rocks. The Ravenswood island arc acts as a ram to deform and isoclinally fold the Fordham basin sediments against the NJ and Hudson Highlands. The border faults are the Ramapo and East River faults. Later reactivation of these faults are responsible for the opening and collapse of the Paleozoic Manhattan basin.*

The roll-over of the overriding plate into the subduction trench created tensional forces within the Fordham back-arc basin, which split the initial calc-alkaline dominant volcanic arc and allowed for the underplating and eruption of tholeiitic N-type MORB into the back-arc basin (Figure 54). Extension of the basin facilitated the sliding back of crustal blocks into the region of HT-LP metamorphic conditions of the back-arc basin (large arrows in Fig. 54), resulting in the formation of the garnet coronite textures[25] within the BIG Complex basalts. Subsidence of the basin floor into the deeper realms of the back-arc basin allowed for the incorporation of pre-metamorphosed sections of Fordham Gneiss to be unroofed within the Ravenswood magma chambers (small insets within the Ravenswood arc, Fig. 54; see Berkey quote, section 5.1). The differential subsidence within sections of the Fordham back-arc basin created isolated sections for deposition of the younger Fordham Gneiss units and the development of hydrothermally derived coticules.

Insofar as presently determined, the width of the Fordham marginal basin has not been estimated. Jurdy and Stefanick (1983) show that the extent of the tensile region of a back-arc basin is directly dependent on the estimated depth to the tip of the underlying convection zone. Shallow convection assumption of 300-400 km (Toksoz and Hsui, 1978) lead Jurdy and Stefanick (1983) to estimate a corresponding theoretical back-arc basin width of 400 km from the trench. Although different assumptions may be warranted with depths of the Proterozoic age convection, the 300 - 400 km estimate will be used as a model for the Fordham back-arc basin (see also Figure 31).

[25] The majority of garnet coronite textures are restricted within the Ravenswood meta-gabbro as opposed to tonalite units.

The Ravenswood's range of calc-alkaline to tholeiitic island arc units[26], sections of depleted mantle, and the exposure of plutonic units indicate that the Ravenswood volcanic arc was relatively mature at the time of final collapse. The time span between initial arc formation at ca. 1,032 Ma (calc-alkaline enderbite zircon Pb-Pb date) and subsequent final closure of the basin, in order to be consistent with the estimated 10-20 My typical life span for back-arc spreading (2-3 cm/yr) and a subsequent convergence rate of 6-10 cm/yr (Jurdy and Stephanick, 1983), is interpreted to be on the order of 50 Ma. (consistent with the average Mesoproterozoic back-arc basin life span noted by Rivers 1997, pg. 144). Thus cessation of arc/back-arc basin would thus be ~980 Ma, also consistent with the ~980 Ma end stage Rigolet Pulse of the Grenvillian Orogeny (Rivers 1997).

During the collapse of the Fordham back-arc basin (Figure 55), high grade metamorphics were unroofed and transported westward to compress and isoclinally fold the Fordham basin sediments (Hall 1968). The collapsed Fordham basin was bounded by the Ramapo fault to the west and the East River fault demarcated by the western extent of the Ravenswood granodiorite (initially identified as Cameron's Line by Merguerian and Baskerville, 1982; Baskerville and Mose, 1985, Baskerville 1992). The Fordham basin paired border faults were reactivated, in a heredity model similar to that proposed for the Ramapo fault by Berkey (1907) and Ratcliffe (1971). Although further petrographic research is required it may be possible that the water laden younger Fordham basin

[26] Island arc and associated plutonic units are mineralogically reconstituted by metamorphism and are not presently distinguished; recall that the BIG Complex includes the Ravenswood and Fordham lithologies.

sediments depressed the formation of pyroxene during collapse of the back-arc, thus preserving predominately amphibolite grade metamorphic assemblages.

The final stage of the Grenvillian Orogeny is typically considered a continent-continent collision (Bond et al, 1984; Drake et al., 1991b; Rankin et al., 1993). Rivers (1997) cites evidence for closure of the Elzevirian oceanic back-arc basin (~1190 Ma) to be immediately followed by a continent-continent collision with cessation of subduction-related magmatism and initiation of within-plate plutonism between 1190 and 980 Ma for the Grenville Province of the Adirondacks and the southern Canadian Shield. However this study uses geochemical methods to indicate that the end stage collision in this section of the Reading Prong (south of the eastern Hudson Highlands) may be marked at ~960 Ma by an island arc-continent collision resulting in Proterozoic crustal growth via arc-accretion.

These findings support the increasing body of evidence for a narrow Proterozoic Ocean basin (Grenville Ocean, Fig. 49) with the possibility that at least a portion of this Grenvillian Ocean may never have fully closed. The continued existence of a Grenvillian Ocean at this location would have been facilitated by an offset in the original Laurentian margin of North America and therefore would indicate the New York recess had been established before Cambrian time (Rodgers 1995) and is perhaps a pre-Grenville feature. Findings of anomalously old oceanic crust at the Mid-Atlantic Ridge (Pilot et. al., 1998, Pratt 2000) may indicate that orogenies do not necessarily have to completely close out ocean basins and that blocks of older oceanic crust can be left isolated during subsequent rifting (Bonatti and Crane, 1982).

Furthermore, the upwelling associated with the development of a the aborted Vernon ensialic back-arc basin (of Volkert et al., 2000) and subsequent Mesoproterozoic Fordham oceanic back-arc basin may indicate an inherited location for plume generated triple junctions associated with the Late Precambrian opening of the Iapetus Ocean (Rankin 1976) and perhaps the present day Atlantic Ocean and Palisades sill.

The incomplete destruction, then later reactivation or recreation of the Fordham Basin, as defined by border faults, would create several generations of basins with similar sedimentary environments. In fact there have been numerous debates about whether there is a continuous sedimentary profile between the Fordham Gneiss and the Inwood Marble - Manhattan Schist (Berkey 1907; Balk 1936; Prucha 1956; Scotford 1956). Prucha (1956) saw a continuum between the Fordham and the Inwood-Manhattan series while Berkey (1907) recognized a relatively short hiatus.

The next part of this thesis presents evidence on how the reactivation and collapse of the former Fordham basin contributed to the above difficulties in structural interpretations and proposes that the BIG Complex has had a crucial role in determining the structural geology of the Manhattan Prong. The first phase of this argument starts with the Proterozoic Z extension of the Laurentian margin.

9 Late- Protoerozoic Extension of the Fordham Basin

The first leg of the Wilsonian model for the Proterozoic basin now leads to a consideration of its influence on the evolution of a Paleozoic basin. The stratigraphic succession in the Proterozoic model (Figure 55) ends with the deposition of the Precambrian coarse fluvial deposit at the onset of the development of an erosional plane. The evolved erosional (peneplain?) along the now passive margin of the North American continent as it may have involved the BIG-Fordham succession is represented in Figure 56.

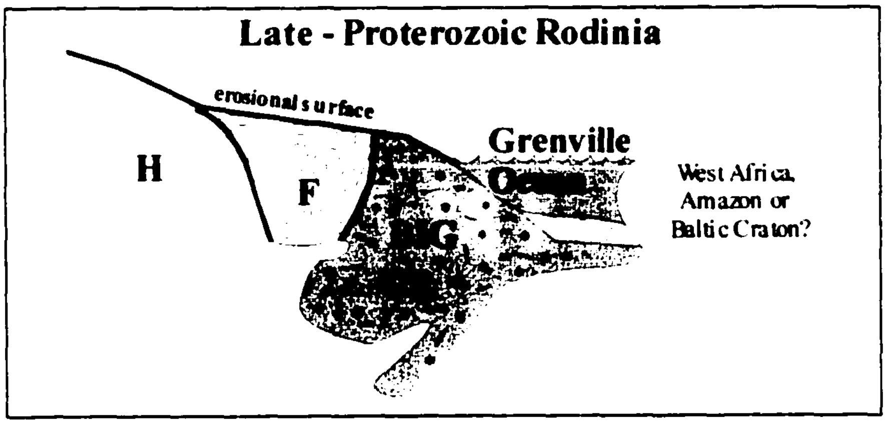


Figure 56. Development of the Late- Proterozoic Rodinian passive margin marked by the extinction of the Highlands and Ravenswood subduction zones leading to the creation of erosional plain.

Extension of the Late Precambrian passive margin has been proposed by Ratcliffe (1971) to have involved the reactivation of the Ramapo fault system along the eastern border of the present day western Hudson Highlands. As discussed in section 8.2 the Ramapo fault system is the continental side of the collapsed Fordham back-arc basin

(Figure 55). Most previous workers, beginning with Berkey (1907), postulate a down-drop along the Ramapo fault produced a half-graben that received Phanerozoic units of the New York City Group.

> "**Present conditions of the strata preserved along this line** (e.g. Ramapo Fault) **indicate two separate movements: first, block faulting and tilting by which the south wall**[27] **was dropped probably 2000 feet or more, carrying down into the trough thus formed all the overlying Cambro-Siluric formations that at that time covered the Highlands; later a thrust from southeast closely folding the sediments**[28] **entrapped in this trough and in places thrusting the gneisses upon them." (Berkey, pg. 375. 1907).**

This model is proposed also by Hall (1968). The present study allows for the possibility instead of an inter-fingering of the Inwood and Manhattan lithologies adjacent to the Ramapo fault as shown in the present revision (Figure 57) of Hall and other workers.

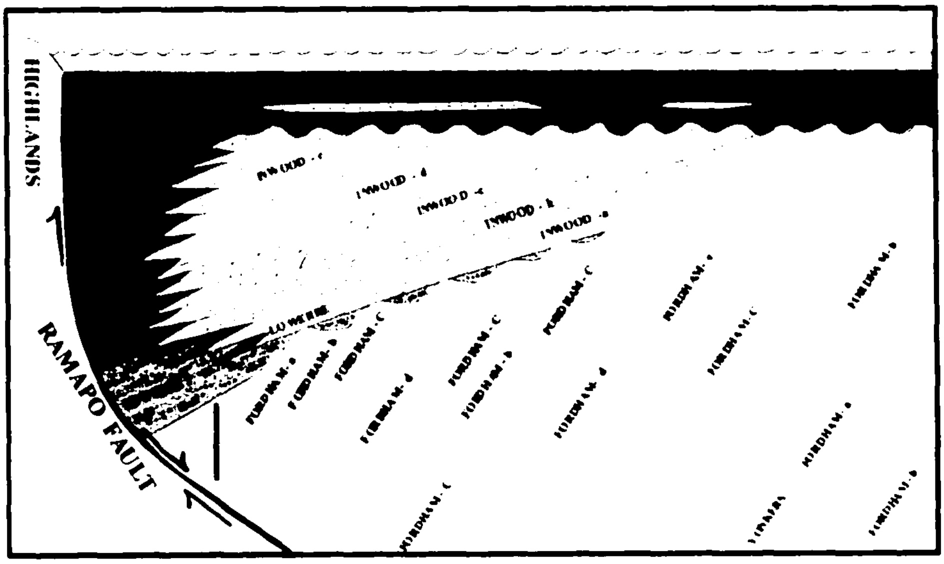


Figure 57. Reactivation of the Ramapo fault system into a half graben for the deposition of the Lowerre-Inwood-Manhattan sequence and proposed interfingering between Manhattan and Inwood (after Hall 1968; Berkey 1907, Ratcliffe 1971).

[27] Relative to the present day orientation of the Ramapo fault, the south side is the oceanward section at that time.

[28] Refers to the NYC rock groups.

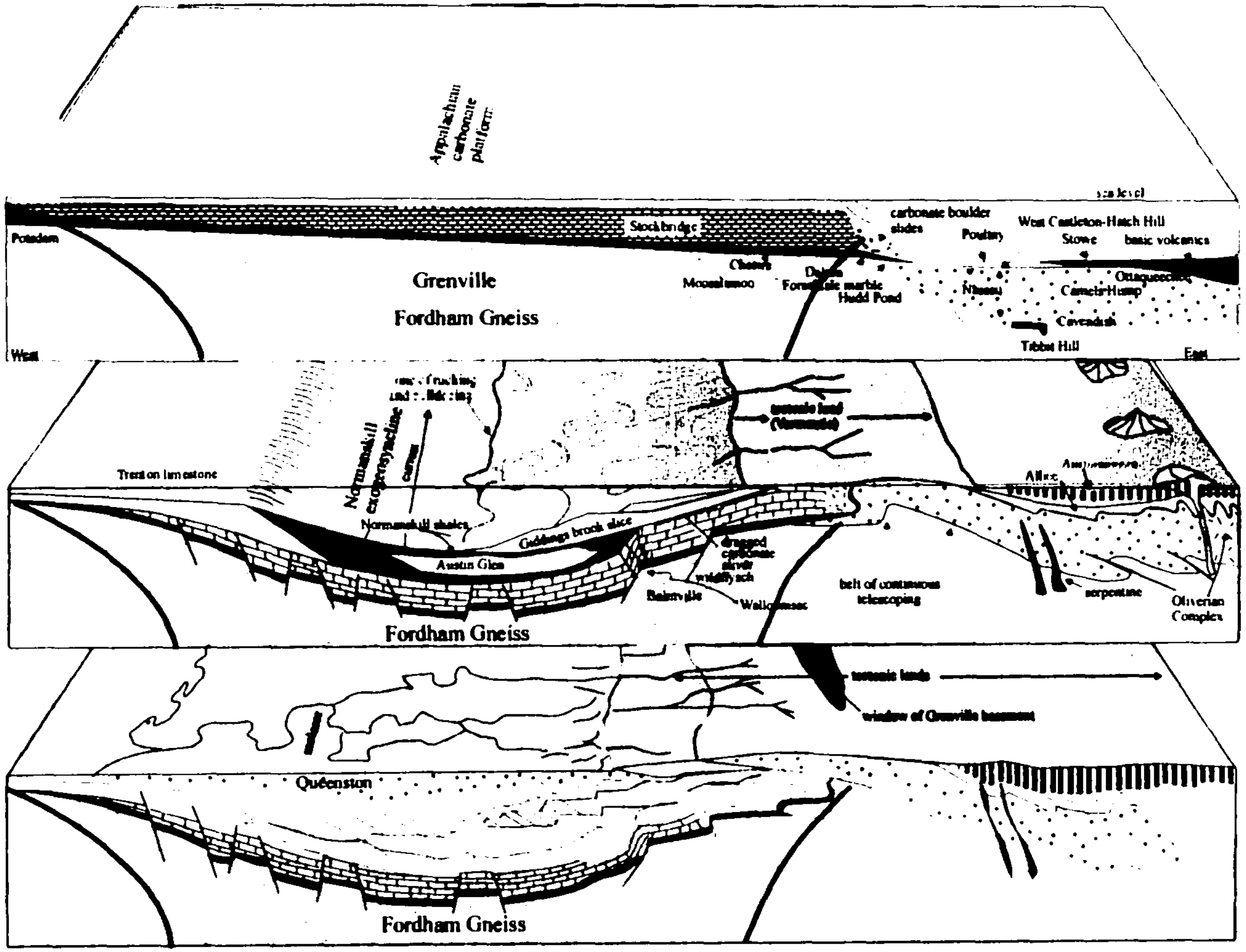


Figure 58. Modification of Bird and Dewey's (1970) diagram for the North American continental margin in western New England depicting late Precambrian to Ordovician expansion and Ordovician to Devonian contraction of the proto-Atlantic ocean basin. Hatch marks represent the schematic location of the Fordham gneiss proposed for this study.

The late Precambrian passive margin is associated with the development of a Proto-Atlantic Ocean depicted in Bird and Dewey's (1970) model for the origin of the northern Appalachians in New England (Figure 58). Page (1976) considers the Chesire-Stockton-Normanskill series of New England (Fig. 58) to be the northern equivalent to the Cambro-Ordovician Lowerre-Inwood-Manhattan sequence, both of which formed in an offshore depositional basin that extended from Long Island north into Canada. The Bird-Dewey model illustrates the development of the Cambrian-Silurian basin that, in the present model, overlies the Proterozoic basement (Fordham Gneiss) as indicated by hatch marks (Figure 58). Although the Bird and Dewey model sequence shows the collapse of the basement as a consequence to Taconian age westward thrusting, the half-graben proposal for the development of the Manhattan section of this marginal basin (after Berkey 1907; Hall 1968; Ratcliffe 1971, see Figure 52) is revised here to involve paired border faults (Figure 59a) consistent with an extensional setting rather than compressional setting for development of the Manhattan Basin.

The consequence of adopting the Hutchison and Klitgord model (Figure 59a; Hutchison and Klitgord, 1988) to present day Long Island platform is shown in Figure 59b. The graben-model is clearly linked to well-defined modern fault systems; the Ramapo border fault to the west and the East River fault to the east. Hutchison and Klitgord propose that the west dipping border fault could be either reactivated backthrusts or extensional faults newly created during stretching. In the context of the BIG Complex's newly recognized volcanic-arc / back-arc setting the west dipping border fault formed during late Proterozoic extension of the Fordham back-arc basin (consistent with Hutchison and Klitgord's (1988) model for extension). This fault, the East River fault, is a probable reactivated backthrust associated with the Grenville age collapse of the Fordham marginal back-arc basin, rather than a structure that separates Paleozoic terranes, e.g. Cameron's Line Merguerian and Baskerville, (1982); Baskerville (1992 and 1994).

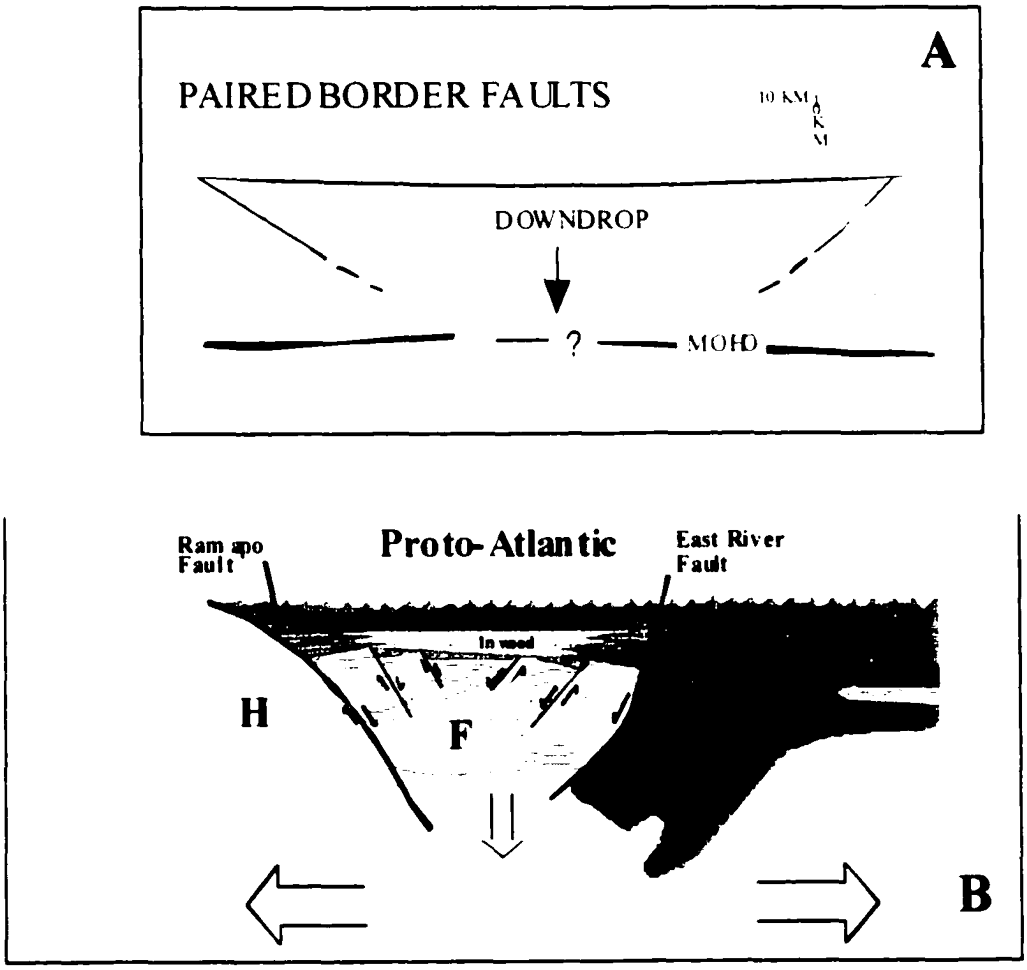


Fig. 59a-b. Extensional model employing paired border faults in A (from Hutchison and Klitgord, 1988) with that proposed in this study for the development of the Manhattan marginal basin (B).

In summary, this study supports both Berkey's (1907) and Ratcliffe's (1971) argument for reactivation of a late Precambrian fracture system but suggests an earlier, Grenville age activation of these systems in establishing the tectonic framework for subsequent compressional and extensional events along the Atlantic Margin. The presence in the BIG Complex of cross-cutting meta-diabase dikes, with geochemical signatures almost identical with other regional Late-Proterozoic dikes (see section 7, Ratcliffe 1987; Aleinikoff et al., 1995), lends additional credence to the model due to the basaltic intrusions and their consistent orientation and nature of these structures relative to the Ramapo fault fracture system (see Ratcliffe, 1971). The unique combination of metamorphic fabrics within the BIG Complex meta-diabase dikes with undeformed cross-cutting intrusive contacts provides an important clue to the subsequent collapse of the Manhattan marginal basin (see latter portion of Bird and Dewey's tectonic synthesis model in Figure 58). The following section details a structural framework for Paleozoic deformation of the Manhattan Prong rocks.

10 Paleozoic Deformation

Ratcliffe (pg, 139, 1971) observes that multiple Paleozoic metamorphic events and related structural deformation prominent in the Manhattan Prong adjacent the border fault (Ramapo fault), are difficult to detect in the Precambrian rocks of the Hudson Highlands west of the Hudson River. This is demonstrably true of the BIG sequence. Moreover, Paleozoic "structural overprinting" is also minimal. Nevertheless, the enclosure of the Manhattan basin by these major Proterozoic blocks indicates that the BIG Complex was actively involved in the collapse of the Manhattan basin. This agrees with the isoclinal folding of the Lowerre-Inwood-Manhattan sequence in a psuedo-conformable relationship with the Fordham Gneiss (Berkey 1907; Balk 1936; Prucha 1956).

In a seminal paper on the Manhattan Prong rocks in Dutchess County, New York, Balk (1936) proposes that large felsic orthogneiss bodies may constitute leading blocks that sandwich and deform (i.e. squeeze) sedimentary layers during compressional events. The Balk model is applied here to account for the deformation of the Lowerre-Inwood-Manhattan sequence within the New York City area.

At the extreme western tip of Long Island, the Ravenswood Granodiorite of the BIG Complex apparently acted as a buttressing block with its present day leading edge situated near the East River in western Queens and Brooklyn (see Berkey's x-section, Figure 4). The Ravenswood block thus participated in isoclinally deforming the Fordham Gneiss units during the Grenville closure of the back-arc basin. Subsequent to Late

Proterozoic extension, Taconic thrusting mimicked the Grenville closure deforming the Lowerre-Inwood-Manhattan formations, again squeezing the units against the Hudson Highlands block. This repeated deformation created a psuedo-conformity (Balk 1936) between the Proterozoic and Paleozoic members of the New York City Group that has created so much debate about the age of the New York City Group (Berkey 1907; Balk 1936; Scotford 1956; Prucha 1956).

This repeated Wilsonian opening and closing was facilitated by continued reactivations of the Ramapo fault - East River fault systems. Although Cameron's Line demarcates the Taconic thrusting of Cambro-Ordovician age sediments over autochthonous formations this study supports the position that Middle Proterozoic units were also thrust into their present allochthonous position during the Taconic and possibly later orogenies by an underlying detachment fault.

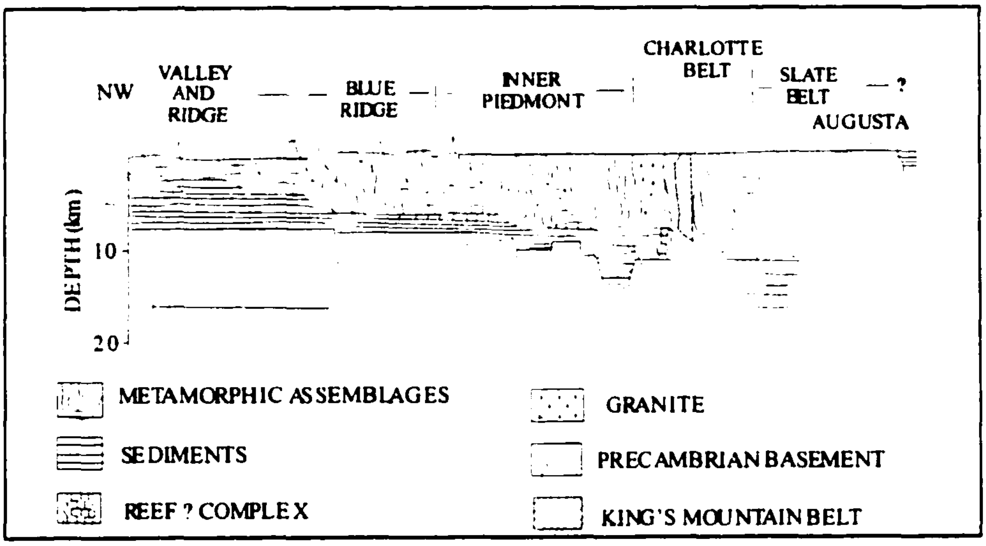


Figure 60. Structural interpretation of the southern Appalachians from Cook et. al., (1979).

The COCORP seismic studies along the Appalachians has provided much needed support for the possible allochthonous nature of numerous formations exposed up and down the eastern seaboard. Based on the COCORP seismic studies Cook et al., (1979) proposed a structural interpretation for the southern Appalachians (Figure 60) suggesting upwards of 260 km of thrusting.

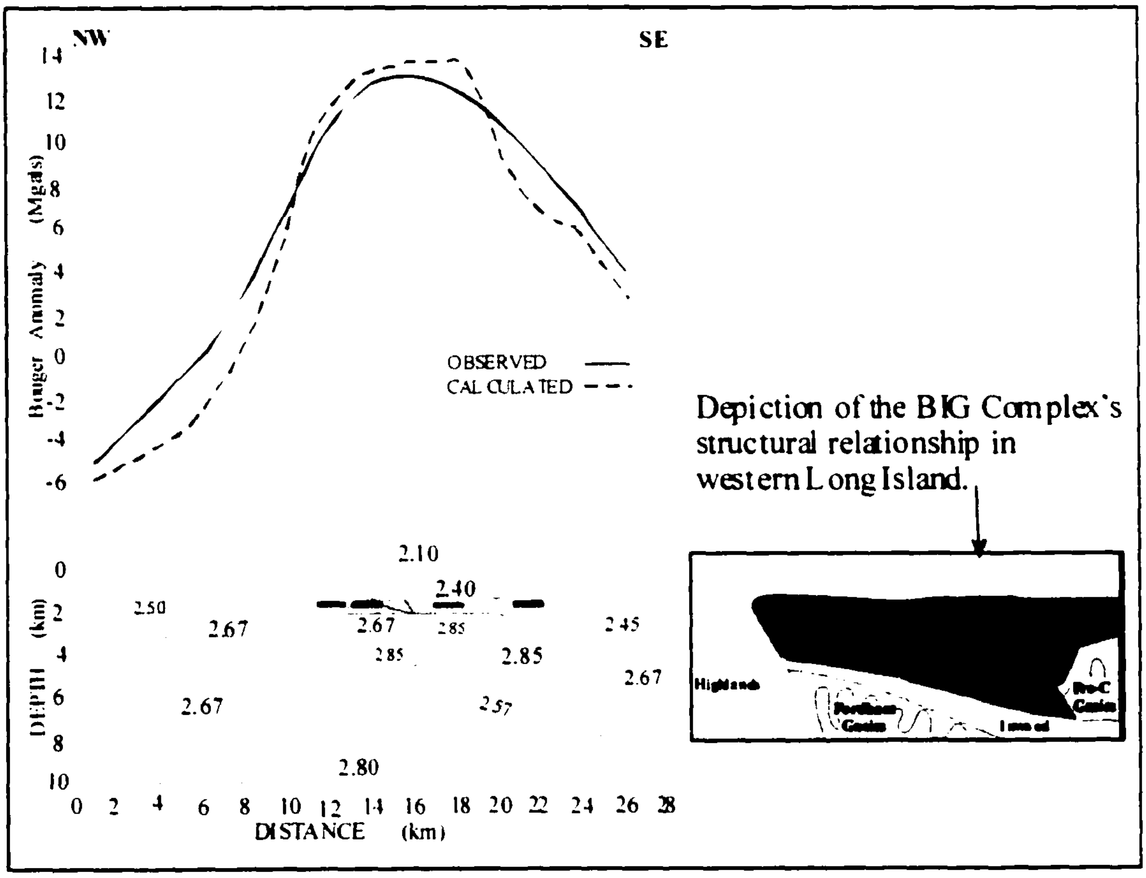


Figure 61. Sheridan et al., (1991) Salisbury Bouguer gravity anomaly measurements along Buena vibroseis line across the New Jersey coastal plain 150 km south of New York City. Although extrapolated north to central Long Island, the interpretation of ophiolite fragments (represented as 3.10 g/cm^3 density blocks), the obducted nature of the Proterozoic Y Goochland Terrane and the general fault geometry is consistent with the lithologies and proposed structural relationship of the BIG Complex in western Long Island (right side of figure).

Based on Salisbury Bouguer gravity anomaly measurements along Buena vibroseis line across the New Jersey coastal plain 150 km south of New York City Sheridan and others (1991) presented evidence for thrusting of Paleozoic rocks and Proterozoic Y rocks of the Goochland Terrane over unmetamorphosed Cambro-Ordovician drift-stage sedimentary rocks (Figure 61). The present reconstruction supports the model of Harris and Bayer (1972) that extends this overthrusting into southern New England along a major detachment. A proposed depiction of the BIG Complex's structural relationship to the Manhattan Prong rocks and the New Jersey Highlands is presented on the right side of Figure 61 and presents an overthrust model for the New York City area during the Taconic Orogeny with similar structural and lithologic characteristics to Sheridan and others' model.

The interpretation in the Sheridan et al., (1991) study of ophiolite fragments within the overthrust units (Figure 61) has an important relationship to this thesis in that Sheridan et al., (1991) consider low-angle thrust faults as Paleozoic and are related to Ordovician age ophiolite fragments. However, it now must be considered that the ophiolite fragments may in fact be Proterozoic and not Ordovician in age with the low-angle faults having originally formed during the obduction of Proterozic back-arc basin/ophiolite complexes. The rock types, ages and petrographic features of the Goochland Terrane (Farrar 1984) are very much similar to and is considered coeval with the BIG Complex.

10.1 Summary model for the Paleozoic overthrusting of the BIG Complex

Taconic age thrusting of eastern sedimentary units over Ordovician carbonates has been known for some time north of New York City (Bosworth and Rowley, 1984, Selleck and Bosworth, 1985, see also Stanley and Ratcliffe, 1985). Although proposals exist for the allochthonous nature of portions of the Reading Prong during the Acadian and Alleghanian (Isachson 1964; Rogers 1995), schemes employing the overthrusting of crystalline Precambrian rocks in New York City have been harder to justify. This proposal assumes that the New Jersey and Hudson Highlands were rooted during the collapse of the Manhattan basin, acting as a solid mass against which the remaining members of the New York City group were compressed and folded during the Taconic orogeny (e.g. application of the Balk model).

The model for the Taconic development of the Manhattan Prong differs from the Bird and Dewey (1970) model for western New England, in that reactivation of the Ramapo fault – East River fault system during Late Precambrian extension (Figure 62a) was responsible for down-dropping the newly created Manhattan basin. Downdropping of the basin caused block faulting of the Lowerre-Inwood units, depicted as a half-graben by Berkey (1907) and Hall (1968) but now envisioned as a paired Ramapo – East River border fault system.

Initial compression of the region during the Taconic orogeny initiates collapse of the basin. A low angle thrust fault formed in a weak basal zone of the BIG Complex island arc/ophiolite, established a decollement surface on which the BIG Complex pushes a wedge of Fordham-Inwood-Manhattan forward (Figure 62b). Overthrusting along this surface allows for a buttressing block of Ravenswood Granodiorite of the BIG Complex

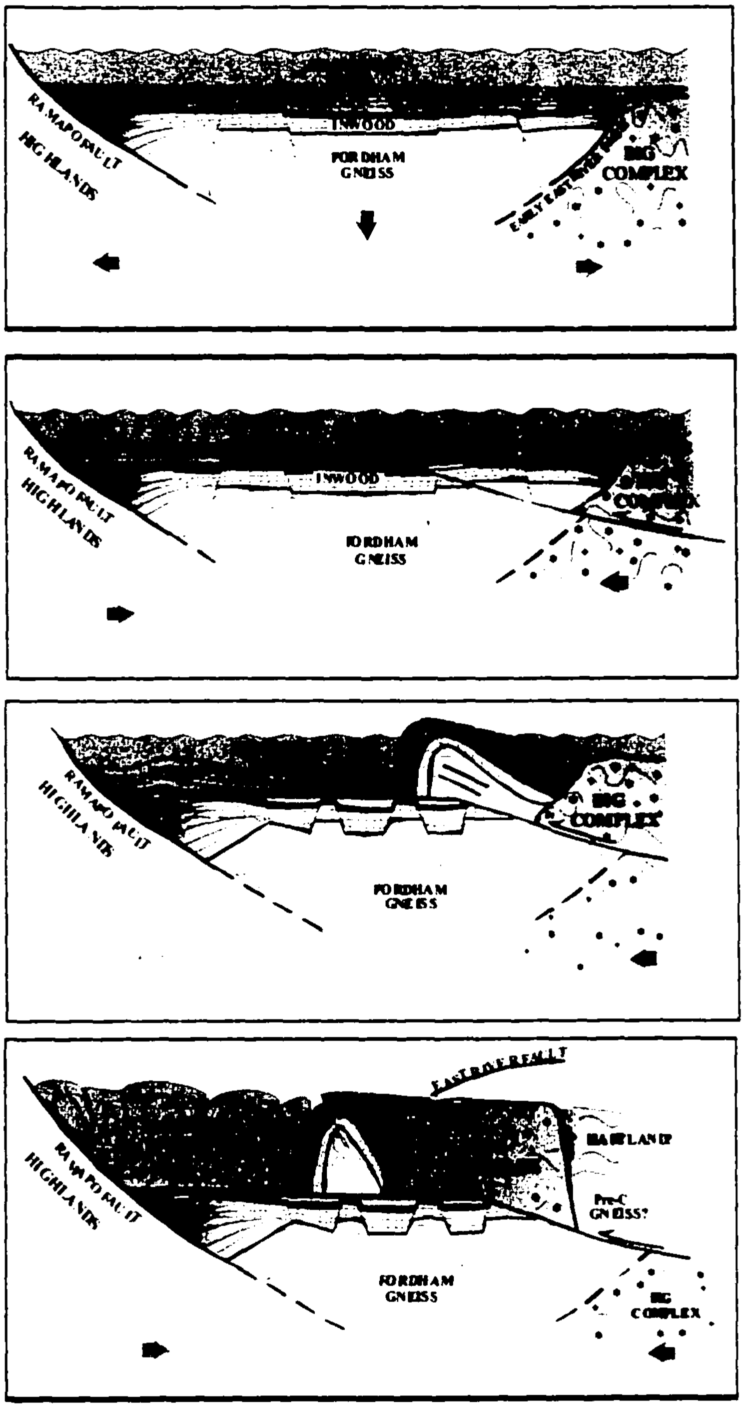


Figure 62. Proposed Taconic overthrust model for the New York study area. A) Formation of the Manhattan during Late-Proterozoic extension based on paired border faults (Hutchison and Klitgord, 1988). B). Initial phase of compression. C) Development of overthrust faulting. D) Collapse of Manhattan Basin with the BIG Complex compressing the Manhattan Schist as it rides on detachment surface above sediments.

to plow the Manhattan schist units (Balk model, 1936) ahead of a Fordham – Inwood – Manhattan anticline. This surface is proposed to be along the contact of the Inwood-Manhattan sequence (Figure 62c) in following of regional Taconic overthrusting models (Bosworth and Rowley, 1984; Selleck and Bosworth, 1985; Sheridan et al., 1991).

Consistent with other models for Appalachian Overthrusting (Cook, et. al., 1979), movement of the BIG Complex was facilitated by sliding over water saturated Cambro-Ordovician sediments. Thrusting of large blocks of the BIG Complex generated significant stress within the rocks, enough to recrystallize the thinner Late Proterozoic meta-diabase dikes and sills (Figure 16B) however not enough strain was created to totally recrystallize the thicker dikes or deform the dikes contacts with the BIG Complex. It is proposed that during the collapse of the marginal basin, the induced strain was accommodated by the reactivation of Proterozoic shear zones and sliding along the fluidized decollement zone.

The collapse of the Manhattan basin is marked by the anticlinal folding of the Fordham-Inwood-Manhattan (after Baskerville 1994) ahead of the Ravenswood/BIG Complex sliding block (Figure 62d) and the complex folding of the Manhattan Schist.

Although final emplacement of the BIG Complex results in the amphibolite grade metamorphism of the Manhattan Schist units within New York City (Langer and Bowes, 1969), no Paleozoic overprinting of the BIG Complex is noticed. This is somewhat consistent with Ratcliffe's (1975) model for buoyant ejection of slices of deeply subducted Grenvillian rocks that were thrust westward over carbonate cover, yet thrust far enough to be protected from later Acadian and Alleghenian metamorphic overprinting (as described in Robinson and Hall, 1980). The BIG Complex was also subjected to cold thrusting along a detachment fault and has been able to basically float coldly through later Acadian and Alleghenian events that have affected other sections of the Manhattan Prong (Brock and Mose, 1979).

11. Conclusions

The Brooklyn Injection Gneiss (BIG) Complex represents a diverse suit of rocks associated with the stepping back of the Grenville subduction zone and subsequent development of a volcanic arc/ back-arc basin situated along the Laurentian margin at ca. 1,035 to 980 Ma. A depleted mantle model of partial melting and subsequent fractionation producing Ta, Nb, Zr, and Ti depletions similar to that of modern island arcs is responsible for generating the Ravenswood bimodal suite. Magmas with calc-alkaline signatures were generated during initial arc/back-arc development where the role of the previous New Jersey Highlands calc-alkaline magmatism associated with the sub-continental lithosphere was still influential. As the basin matured and widened uncontaminated tholeiitic island arc and N-MORB type basalts entered the basin via mantle upwelling and contributed to an alteration of signatures as lavas co-mingled. The transformation of the Laurentian margin from a passive continental margin to a back-arc depositional setting is reflected in the variability and character of the Fordham gneiss lithologies as well as their associated REE profiles. Emplacement of the Ravenswood differentiates resulted in the hydrated borders of the intrusions and migmatization of the Fordham sediments.

The establishment of a volcanic arc/back arc ophiolite complex along this section of the Laurentian margin during the Grenvillian orogeny (ca. 1,080 – 1,030 Ma), suggests that the end stage Grenvillian event did not involve a continent-continent collision but an island arc-Laurentia collision at ca. 960 Ma. This would imply that the Grenvillian Ocean may never have been totally eliminated. The continued existence of a Grenvillian ocean at this location would have been facilitated by an offset in the original Laurentian margin of North America and therefore would indicate that the New York recess of Rodgers (1975) is a pre-Grenville feature. Furthermore, the up-welling associated with the development of a Mesoproterozoic back-arc spreading center may indicate an inherited location for plume generated triple junctions associated with the Late Precambrian opening of the Iapetus Ocean (Rankin 1976) and perhaps the present day Atlantic Ocean.

The border faults of the Fordham back-arc basin may also have contributed to the repeated cycles of compressional orogenics and continental breakups centered on the New York recess. The margins of the Fordham basin were initially delineated by normal faulting typically associated with extension of the back-arc. The western border fault is considered to be the present day Ramapo fault system forming at the eastern margin of the New Jersey and Hudson Highlands (see Ratcliffe 1971). The eastern margin of the basin is delineated by the last stages of the Ravenswood volcanic arc. Collapse of the Fordham basin, ca. 960, involved the unroofing and transportation westward of high grade metamorphics to isoclinally fold the Fordham sediments and penetratively deform the Ravenswood plutonics in a manner consistent with that outlined by Hooper and Hatcher (1989) for island arc plutonic bodies. The collapsed Fordham back-arc basin was bounded by the Ramapo fault to the west and the East River fault demarcated by the western extent of the Ravenswood "granodiorite", formally identified in this area as Cameron's Line. The lack of deformation in the 1,020 Ma Mt. Eve granite indicates that the New Jersey and Hudson Highlands were stationary and were not affected by the collapse of the basin.

A third generation of reactivation for the Fordham basin border faults occurred during the late Precambrian breakout of Laurentia. The transformation of the Laurentian margin to passive margin is marked by the intrusion of high TiO2 diabase dikes (spanning from ca. 760 to 565 Ma, see Aleinikoff et al., 1995) that exhibit within-plate geochemical signatures, yet appear to be generated from the same mantle source as that of the BIG Complex. The newly created marginal (Manhattan) basin was formed via extension and downdropping along the preexisting Ramapo fault (Ratcliffe 1971) and the East River border fault. The Manhattan basin is the site of deposition for the Cambrian-Ordovician Lowerre quartzite-Inwood Marble-Manhattan Schist series of the New York City Group.

Subsequent collapse of the Manhattan basin during the Taconic orogeny involved reactivation of both the East River fault and the Ramapo fault in a structural process similar to the original collapse of the Fordham Basin. In an application of the Balk (1936) model, the leading edge of the BIG Complex acted as a buttressing bock to compress and fold the remaining members of the New York City group up against a stationary New Jersey and Hudson Highlands. The enclosure of the Manhattan basin by these major Proterozoic blocks indicates that the BIG Complex was actively involved in the collapse of the basin. Taconic thrusting mimicked the Grenville closure deforming the Lowerre-Inwood-Manhattan sequence, creating a psuedo-conformable relationship with the Fordham gneiss which led to early debates as to the correct age of the New York City Group. Furthermore, the Taconic compressional orogeny introduced subduction related Rb-K-metasomatic fluids, that disrupt the Sr-Rb systematics of the BIG Complex leading to an inaccurate Sr-Rb age estimate (Baskerville and Mose, 1989) and produced a secondary potassium feldspar mineralization masking the original tonalite composition of the Ravenswood "granodiorite".

Although Cameron's Line demarcates the Taconic thrusting of Cambro-Ordovician age sediments over autochthonous formations this study supports the position that Middle Proterozoic units were also thrust into their present allochthonous position during the Taconic and possibly later orogenies along an underlying detachment fault. This interpretation is supported by Sheridan and others' (1991) geophysical studies across the New Jersey coastal plain in which evidence for thrusting of Paleozoic rocks and Proterozoic rocks of the Goochland Terrane over unmetamorphosed Cambro-Ordovician drift stage sedimentary rocks was presented. Sheridan and others (1991) interpreted high density blocks as Ordovician age ophiolite fragments within the overthrust units. The Ordovician age designation by Sheridan et al. is probably due in part to the fact that most, if not all, recognized ophiolite complexes along the eastern seaboard are Phanerozoic or younger. This study suggests that eastern seaboard complexes, with volcanic arc/back-arc ophiolitic affinities that have not been carefully dated, may need to be reevaluated for a possible Middle Proterozoic association.

12 Appendix I

MEMORANDUM

ON THE FORDHAM-RAVENSWOOD TRANSITION

IN THE VICINITY OF SHAFT 9A

By

Charles P. Berkey

NEW YORK CITY BOARD OF WATER SUPPLY

New York, March 28, 1931.

MEMORANDUM

ON THE FORDHAM-RAVENSWOOD TRANSITION

IN THE VICINITY OF SHAFT 9A

One of the strictly Geological questions that is puzzling to a Geologist trying to map the ancient rock floor on the Long Island side of the East River is that of the relation between the Fordham gneiss on one hand and the Ravenswood diorite on the other. Earlier interpretations based on the exploratory borings and on the very few out-crops that are to be seen on the Long Island side emphasize the injection phenomena and assume that the diorite has invaded the older banded gneiss in such a manner that the result is a mixed rock carrying both sets of constituents, so that the change from one formation to the other is transitional rather than abrupt and sharp contact. It was believed that the transition might be so gradual that one would find it difficult to draw a line between the one formation and the other. Indeed it was assumed in one of the earlier reports made to the board that there are so many places of mixed type and such variable character on this account that a new name was suggested for the whole system, namely, the Brooklyn Injection Gneiss.

The new tunnel has now exposed a larger amount of the underground structure involving these ancient rocks than has ever been exposed before. Dr. Blank has been in constant touch with the Geologic features and on the lookout for unusual or

-2-

special evidence on this as well as other questions. The difficulty in determining the real nature of these rocks and accounting for the differences that are seen in them has led him to suggest a field inspection to make special study of one of the stretches of ground where the change from one formation to the other ought to be seen.

Accordingly, Friday, March 13th was set aside for this inspection, and the whole afternoon was spent in the tunnel between shafts 9A and 10A. This memorandum is intended to record the nature of the observations and a part of the discussion.

Observations

The inspection began in the immediate vicinity of Shaft 9 and preceeded systematically toward shaft 10. The reason for beginning at shaft 9 is because some of the rock in the tunnel at this location is evidently of metamorphic type entirely unlike the other rock usually classified as Ravenswood, but very similar to certain varieties of Fordham Gneiss which it is believed to represent. This condition continues for some distance beyond the shaft toward the south where there are other varieties including garnetiferous schist and granitic gneiss more or less strongly banded and consistent in all essential respects with the Fordham Gneiss formation.

It was noted by Dr. Blank that a fairly good Fordham type could be traced southward at least to this position, but that a change to a more massive quality and a somewhat more basic composition comes about gradually. With this perfect gradation there was doubt whether any new formation was represented. The inspection was made therefore, to check this point and determine if possible whether or not there is any true Ravenswood type.

In going southward only a short distance from the shaft dark bands of rock were noticed that have a composition essentially that of a hornblend diorite. But the rock has a somewhat streaked structure resembling in that respect the bands and streaks that are always characteristic of the Fordham formation. Southward still farther granitic bands were noted and these alternate with darker rocks in alternating banded structural arrangement. It was observed also that the dark rock gradually increases in total quantity until, in certain places, very wide bands appear in which the structure is quite massive.

There seems to be no possible doubt of the igneous character of some of this material. The composition is that of igneous rock and the structure thoroughly characteristic. Thus far no one has succeeded in showing how a rock of this character could be formed in any other way than by crystallization from a magma. In this portion of the ground, therefore, it is confidently believed that the dark massive portions of the rock are igneous. Furthermore the structure is about what should be expected if Ravenswood injections penetrated the Fordham formation in lit-par-lit structure in sufficient amount to develop these bands.

It was noted also that the dark bands are seldom sharply separated from the lighter colored portions taken to be representatives of the older formation. Indeed the gradation from one to the other is so regular that one cannot tell where the change from one type to the other takes place. It is confidently believed that this gradational condition represents the mixing of material. It is assumed that the fused Ravenswood material was in such fluid condition that it could saturate and soak into and penetrate the older formation and mix with it in all sortsof proportions. Thus it developed a new composition in many places, partly destroying the original rock, but seldom completely destroying the foliated structure. Thus it happens that the new rock is streaked and banded like the original and maintains the same general trend and structure of the original formation.

It was also noticed in going southward that the massive character of the rock increased materially where for long distances the streaked and banded habit is very obscure. For considerable distances the tunnel walls show so little of this structure that it has virtually no influence on the general breakage of the tunnel. The breakage is not very different from that of a simple gneissoid granite. This portion is confidently believed to represent the Ravenswood type. The fact that it has a gneissoid structure is judged to be in part due to the incorporated material from the Fordham with the preservation of traces of its structure, and, in part also due to movement of the mass during cooling with development of streakedness and slight foliation under these conditions.

This explanation is further supported by the existence of pegmatitic streaks and segregation bands within the mass which are clearly end-stage products representing development as a very last step in solidification of the mass. It was also noted that in many places there has been fracturing in a late stage with development of pegmatitic quartz along the fractures.

It was noted also that in places very locally the bands in the rock are crumpled and twisted showing movement in sufficient amount to develop these structures. It must mean that the rock was in a certain stage plastic enough, even some of the banded portions that are believed to represent remnants of the older Fordham, to bend and crumple in much the same manner that would be expected of a plastic mass.

In some places the rock is cut by large pegmatite dikes with very coarse structure and a composition which suggests relation to the Ravenswood formation. It is believed that these are largely end product solutions rejected from the cooling mass, which in their escape through proportions already cooled have developed as dikes.

In many places varieties of rock are found within the principal granodiorite mass which are quite unlike the average Ravenswood itself. In occasional instances there are obscure traces of structure in compositions that suggest derivation from the original Fordham formation and it is believed that these represent xenoliths that have been almost completely destroyed by absorption and fusion in the magma. Still other places have lost all structural evidence and only retain somewhat abnormal composition. It is believed that these varieties are due to absorption of older material in the magmatic mass as it fused its way into the roof formations of a former time.

There is great variety of structure and composition represented in this tunnel, but all of them appear to be perfectly understandable on the basis of such an origin as is here assumed in a mixed origin produced by the invasion of an original schistose formation by a dioritic magma. Despite the difficulties of finding a point where one formation stops and the other begins it is perfectly clear, it seems to me, that there are always the elements of at least two formations, one an older banded metamorphic rock that existed long before the other one came, and a later igneous mass that came in as an invading magma penetrating the other, absorbing it, mixing with it,further metamorphosing,it, and in some places entirely taking its place. The older formation is Fordham gneiss. The younger igneous one is the socalled Ravenswood granodiorite. Between typical outcrops of these two is a wide transition zone in which the two rocks are mixed. In this zone the structures and compositions are very varied, almost every possible mixture being represented. This appears to me to account for the difficulty in drawing a line between them, it is not a contact line but a transition instead.

As a matter of fact, I am inclined to believe that there is very little Ravenswood diorite that is not modified somewhat by material absorbed or incorporated in partially fused condition from the older Fordham that formed the roof above the invading magma at that time. Here and there for short distances there is massive rock of very simple composition that may be regarded as essentially the Ravenswood type, but, throughout much the greater distances in the tunnels on the Brooklyn side, the rock is obviously mixed. It is therefore a reasonably fair proposal to speak of it as the Brooklyn Injection Gneiss as was done in one of the earlier reports.

SUMMARY

I have not attempted to give a detailed description of the great variety of rock composition and structure seen on this inspection trip. I am convinced by the inspection that the explanations given in the earlier reports in this series of studies in which the Ravenswood granodiorite was represented as an invading igneous mass is correct. I am also convinced that the invasion was so thorough and so permeating that the change from one formation to the other is a gradual transition rather than a sharp contact, and the final product in most places is a mixed rock that is neither exactly like the original gneiss nor exactly like the invading magma. Rather it is a resultant product made by the mixing of the two. In many places, the mixing results from soaking and injection, and in others it is due to absorption and incorporation of the older material into the magma itself, often with almost complete destruction of the original composition and structure of the older rock.

I regard the stretch of ground between Shafts 9A and 10A as a typical transition from Fordham gneiss on the one side to typical Ravenswood granodiorite on the other.

I understand from Dr. Blank that essentially the same quality of rock that is so well developed in the vicinity of Shaft 10 and for at least half the distance between 9 and 10 extends much farther to the south, how far, I am not prepared to say from personal observation, and that still farther to the south the rock becomes much more varied. Again this agrees reasonably well with the interpretation given in the report made on the original series of borings in 1927 or 1928, and I am convinced that the general structural features there assumed are essentially correct.

Charles P. Berkey

Charles P. Berkey

Geologist

New York City
March 20, 1931

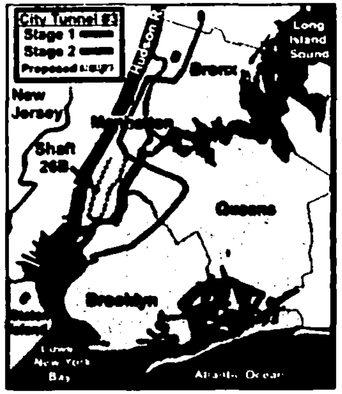
CITY OF NEW YORK
DEPARTMENT OF ENVIRONMENTAL PROTECTION

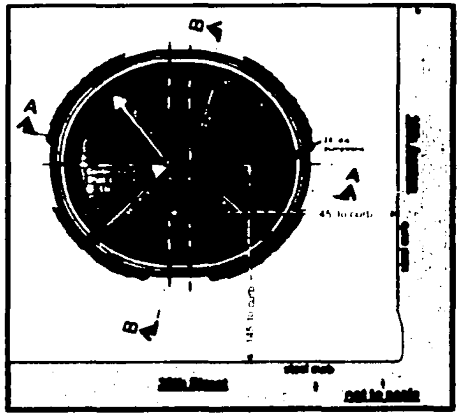
CITY TUNNEL NO. 3
STAGE 2
SHAFT 26B
GEOLOGIC FEATURES

20 0 10 20 30 40 feet

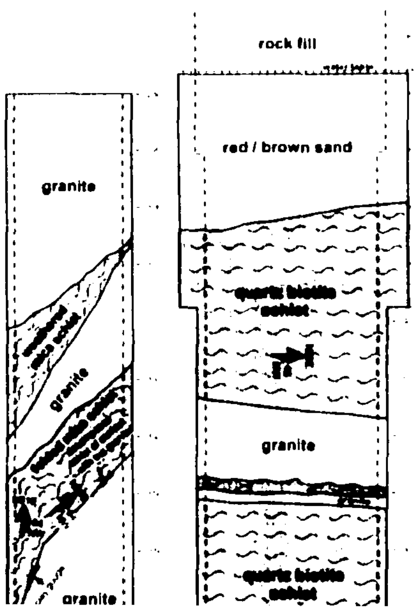
Scale: horizontal = vertical

Mapped and drawn by: Scott Chesman June 28, 1997





VERTICAL SECTION B-B



City Water Tunnel No 3 Stage 2, Shaft 26B

[illegible]

Regional Geology

[illegible]

Geology of Shaft 26B

Overburden

[illegible]

Rock Type

[illegible]

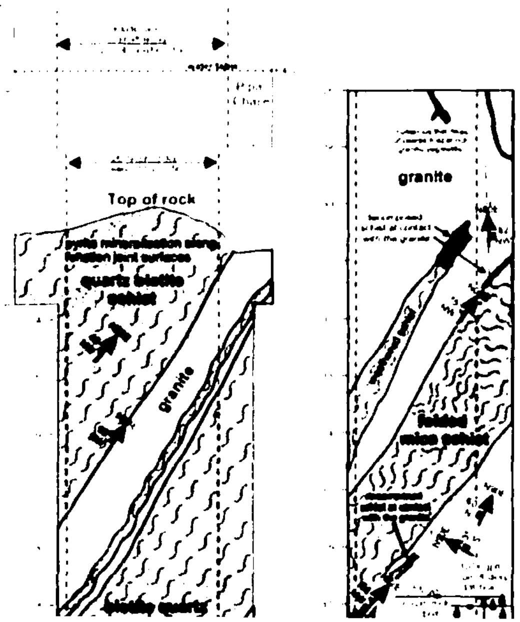
Rock Structure

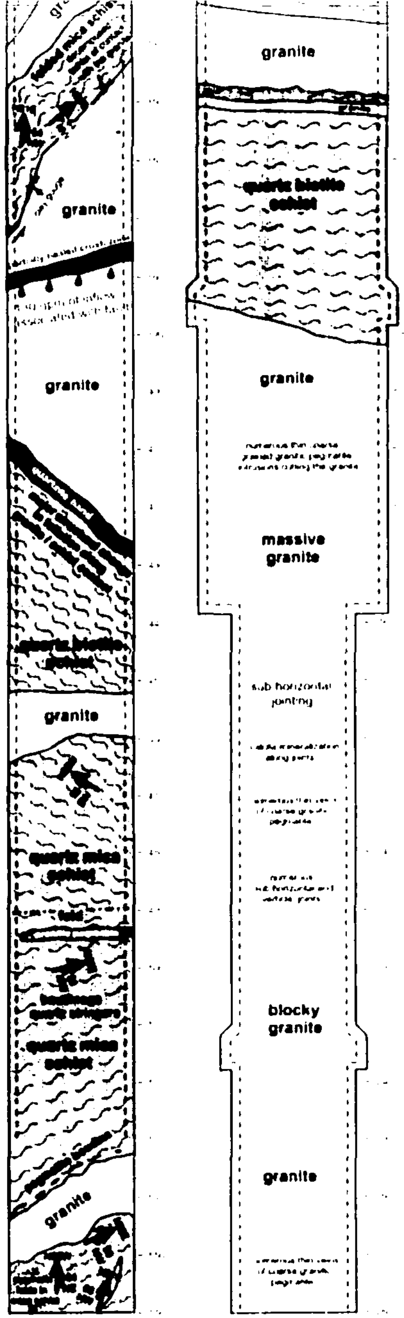
[illegible]

Faulting and groundwater inflows

[illegible]

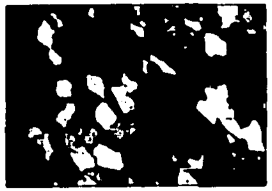
VERTICAL SECTION A-A





Rock Structure

Faulting and groundwater inflows



Shaft sample (elev - 145)
mica schist

SiO_2	[illegible]	% Quartz	15
TiO_2	[illegible]	% Plagioclase	25
Al_2O_3	[illegible]	% Microcline	
FeO	[illegible]	% Mica	55
MnO	[illegible]	% Chlorite	<1
MgO	[illegible]	% Amphibole	
CaO	[illegible]	% Garnet	5
Na_2O	[illegible]		
K_2O	[illegible]		
P_2O_5	[illegible]		
LOI	[illegible]		
Total	[illegible]		

Shaft sample (elev - 450)
granite (trondhjemite)

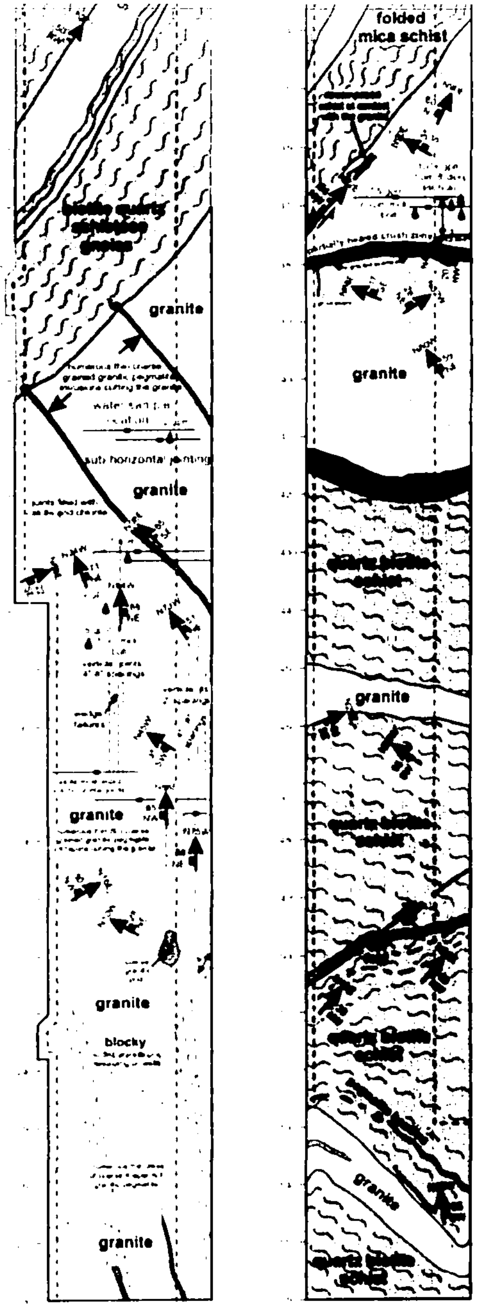
SiO_2	[illegible]	% Quartz	[illegible]
TiO_2	[illegible]	% Plagioclase	[illegible]
Al_2O_3	[illegible]	% Microcline	[illegible]
FeO	[illegible]	% Mica	[illegible]
MnO	[illegible]	% Chlorite	[illegible]
MgO	[illegible]	% Amphibole	
CaO	[illegible]	% Garnet	
Na_2O	[illegible]		
K_2O	[illegible]	TBM Excavation Indices	
P_2O_5	[illegible]	Cutter Life Index (CLI)	[illegible]
LOI	[illegible]	Drilling Rate Index (DRI)	[illegible]
Total	[illegible]	Density (g/cm3)	[illegible]

Borehole sample (elev - 575)
26B-CS-A run #59
schistose gneiss

SiO_2	[illegible]	% Quartz	17
TiO_2	[illegible]	% Plagioclase	54
Al_2O_3	[illegible]	% Microcline	
FeO	[illegible]	% Mica	27
MnO	[illegible]	% Chlorite	<1
MgO	[illegible]	% Amphibole	
CaO	[illegible]	% Garnet	
Na_2O	[illegible]	Cutter Life Index (CLI)	[illegible]
K_2O	[illegible]	Drilling Rate Index (DRI)	54
P_2O_5	[illegible]	Density (g/cm3)	[illegible]
LOI	[illegible]		
Total	[illegible]		

Borehole sample (elev - 570)
26B-CS-B run #57
schistose gneiss

SiO_2	[illegible]	% Quartz	2?
TiO_2	[illegible]	% Plagioclase	45
Al_2O_3	[illegible]	% Microcline	
FeO	[illegible]	% Mica	28
MnO	[illegible]	% Chlorite	<1
MgO	[illegible]	% Amphibole	
CaO	[illegible]	% Garnet	
Na_2O	[illegible]	Cutter Life Index (CLI)	[illegible]
K_2O	[illegible]	Drilling Rate Index (DRI)	58
P_2O_5	[illegible]	Density (g/cm3)	[illegible]
LOI	[illegible]		
Total	[illegible]		



Groundwater analysis
October 18, 1995

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