IGNEOUS ROCKS OF THE NEWARK BASIN:
PETROLOGY, MINERALOGY, ORE DEPOSITS AND GUIDE TO FIELD TRIP

EDITED BY

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Geological Association of New Jersey
Ist Annual Field Conference

AT

KEAN COLLEGE, UNION, NEW JERSEY

OCTOBER 19—20, 1984

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Program

Friday, October 19:

5:00 to 6:00 P. M. – Business Meeting, and election of officers.

6:00 to 7:00 P. M. – Dinner

7:00 to 9:00 P. M. – Symposium – "Petrology, Geochemistry, Mineralogy, and Stragigraphy of the Igneous Rocks of the Newark Basin"

Hosted By – Lee Meyerson, Dept. of Earth & Planetary Environments, Kean College.

7:00 – 7:20 – John H. Puffer, Geology Dept., Rutgers Univ., Newark. " Jurassic Eastern North American Tholeiites"

7:20 – 7:40 – Michael J. Hozik, Geology Program, Stockton State. "Paleomagnetism in the Central Newark Basin"

7:40 – 8:00 – Jonathan M. Husch, Dept. of Geol., Rider College. " Mesozoic Basaltic Rocks from West-Central New Jersey and Pennsylvania: Major and Trace Element Geochemistry of Whole Rock Samples"

8:00 – 8:20 – Warren Manspeizer, Geology Dept., Rutgers Univ., Newark. " Strike-Slip Newark-Type Basins (Triassic-Jurassic) Along the Atlantic Passive Margins of Eastern North America and Northwest Africa"

8:20 – 8:40 – Joseph J. Peters, American Museum of Natural Science. " The Minerals of Bergen Hill, New Jersey"

8:40 – 9:00 – Jack Troy and Nicholas Facciolla " The Minerals of the Paterson Area Trap-Rock Quarries"

Saturday, October 20:

8:00 A.M. to 4:00 P.M. – Field Trip.

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CHAPTER ONE

EARLY JURASSIC EASTERN NORTH AMERICAN THOLEITTES

by

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Geologic Setting

Extrusive Rocks - Eastern North American (ENA) extrusive rock exposures of early Jurassic age are confined to the Mesozoic basins located between Nova-Socotia and Virginia (Fig. 1) but drill core samples of ENA basalt have been found in Georgia (Gottfried and others, 1983), South Carolina (Gottfried and others, 1977), and at some offshore Atlantic coastal plane locations (Hurtubise, personal communication). The extrusive rocks typically occur as multiple thick and widespread flow units interbedded with non-marine red mudstones, sandstones, and conglomerates. The extrusive rock units are in each case tholeiitic flood basalts. The Mesozoic basins that contain volcanic rocks (Table 1 and Fig. 1) include the Fundy Basin (the North Mountain Basalt), the Hartford-Deerfield Basin (the Talcott, Holyoke, and Hampden Basalts), the Newark Basin (the Orange Mt., Preakness Mt., Hook Mt. Basalts) and the Culpeper Basin (the Mt. Zion Church, Hickory Grove, Sanders, and Millbrook Basalts).

Intrusive Rocks - Eastern North American intrusive rock exposures of late Triassic to early Jurassic age are irregularly distributed from northern Newfoundland to Alabama (Fig. 1). The ENA intrusive rocks typically occur as thin diabase dikes south of Virginia and as thick diabase sills north of Virginia but there are many exceptions to this pattern. The diabase intrusions are concentrated within the Mesozoic basins but are also commonly found within Appalachian Paleozoic and Precambrian terrain several kilometers from the nearest Mesozoic sediments (Fig. 2).

Inter-basin distribution - Most of the early Jurassic dikes and flows occur independently from any of the major border fault systems associated with Mesozoic basin development. The intrusive rocks are instead distributed throughout the Mesozoic basins, and the flows are confined to portions representing the uppermost layers of basin filling. The Mesozoic basins were well established and largely filled before early Jurassic tholeiitic igneous activity was initiated. The border faulting and basin subsidence processes (Faill, 1973) that controlled

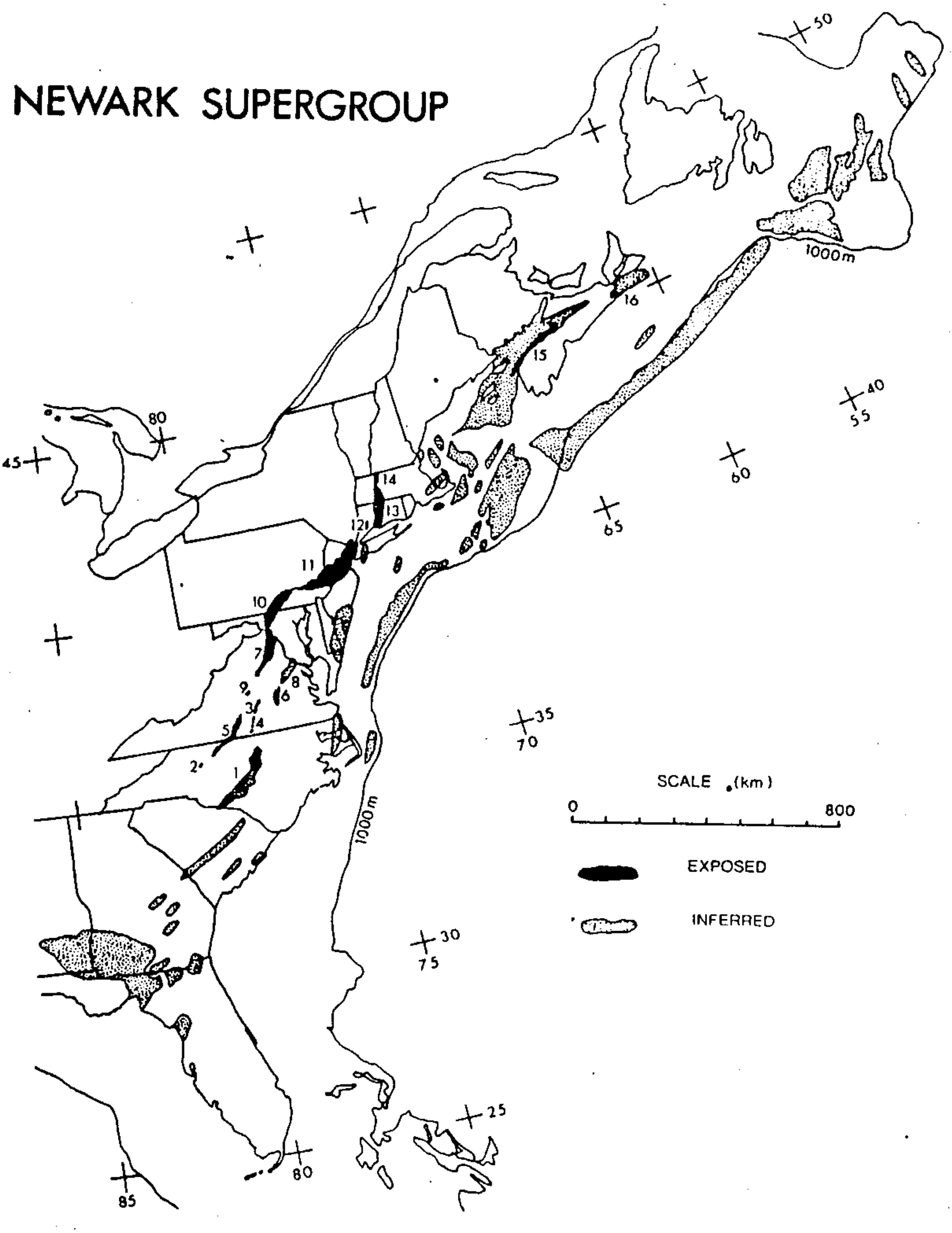

Fig. 1 Newark Supergroup of eastern North America. Key to numbers given in Table 1. The Newark Basin is 11. Data from Olsen, 1978.

Table 1

Key to Figure 1	Rock-stratigraphic term	Basin name	Age range
1	Chatham Group	Deep River Basin	Carnian-?Norian (Late Triassic)
2	undifferentiated	Davie County Basin	?Late Triassic
3	undifferentiated	Farmville Basin	?Carnian (Late Triassic)
4	undifferentiated	4 small basins south of Farmville Basin	?Carnian (Late Triassic)
5	Dan River Group	Dan River and Danville Basins	Carnian-?Norian (Late Triassic)
6	Tuckahoe and Chesterfield Groups	Richmond Basin and subsidiary basins	Carnian (Late Triassic)
7	none	Culpeper Basin	Norian-?Sinemurian (Late Triassic-Early Jurassic)
8	none	Taylorsville Basin	Carnian (Late Triassic)
9	undifferentiated	Scottsville Basin and 2 subsidiary basins	?Late Triassic-Early Jurassic
10	none	Gettysburg Basin	Carnian-Hettangian (Late Triassic-Early Jurassic)
11	none	Newark Basin	Carnian Sinemurian (Late Triassic-Early Jurassic)
12	none	Pomperaug Basin	?Late Triassic-Early Jurassic
13	none	Hartford Basin and subsidiary Cherry Brook Basin	Norian-?Bajocian (Late Triassic-?Middle Jurassic)
14	none	Deerfield Basin	?Norian-?Toarcian (Late Triassic-Early Jurassic)
15	Fundy Group	Fundy Basin	?Middle Triassic-Early Jurassic
16	Chedabucto Formation (=Eurydice Formation?)	Chedabucto Basin (=Orpheus Basin?)	?Late Triassic-Early Jurassic

(after Olsen, 1980)

Mesozoic igneous rocks of the Appalachian region.

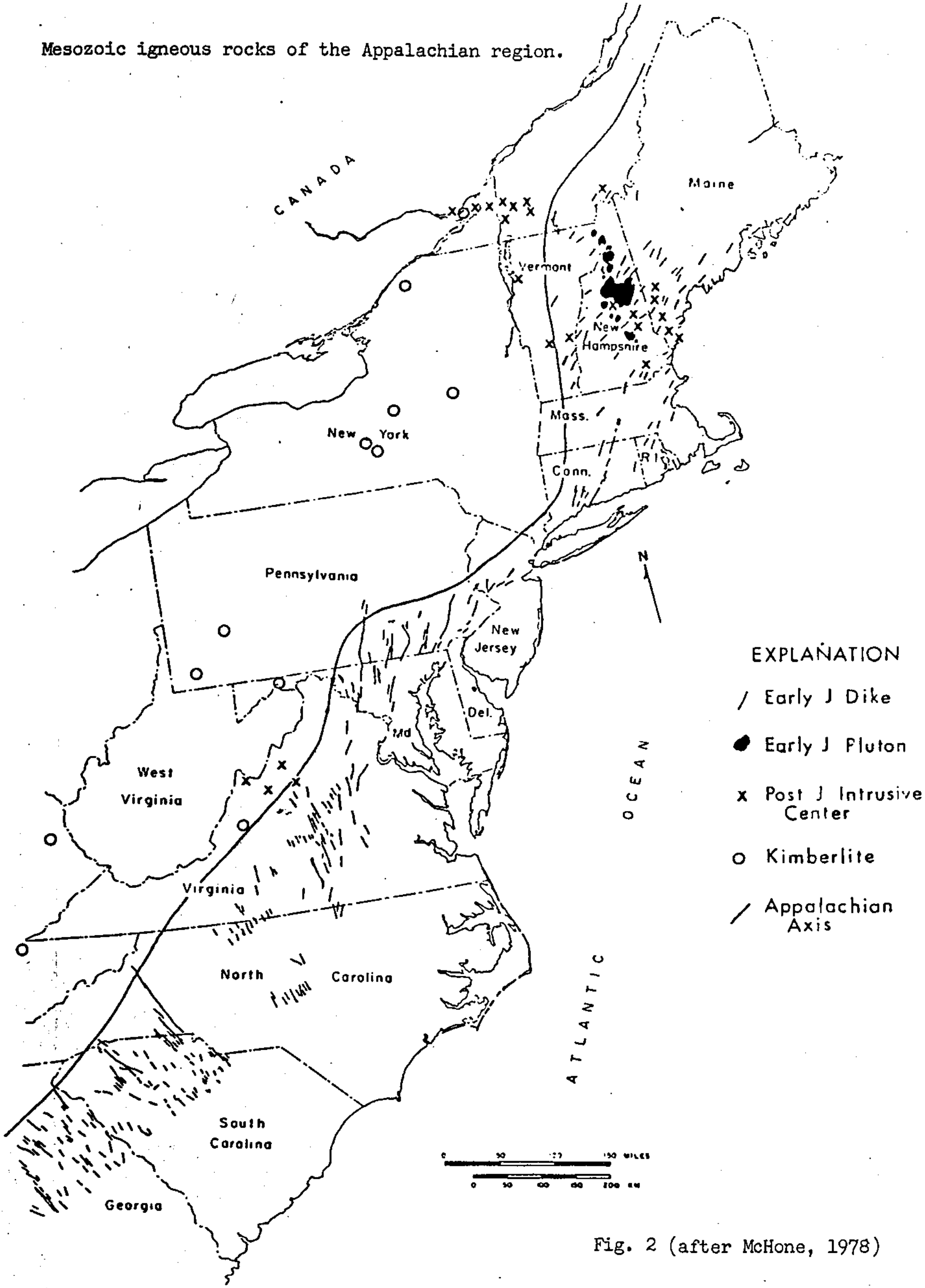

Fig. 2 (after McHone, 1978)

the initial stage of basin development apparently did not coincide with any ENA tholeiitic activity.

Geochemistry and Petrology of early Jurassic igneous populations

The petrology and geochemistry of the early Jurassic ENA province is surprisingly uniform and predictable when compared to the highly varied compositional range of most volcanic suites associated with island-arc or compressional settings. Despite the large size of the province only four major and one minor geochemical populations as recognized by Weigand and Ragland (1970) and Ragland and Whittington (1983a) account for the great majority of the igneous occurrences. The uniform geochemical distribution of the province implies that the prevailing tectonic processes were acting uniformly throughout the entire province or at least widespread portions of it. The geochemical populations include two olivine normative types (high LIL and low LIL), two quartz normative types (high-Ti and low-Ti) and a minor alkali-olivine type. Weigand and Ragland (1970) also recognized an additional quartz-normative type (the high-Fe type) that is now recognized as a variety of the high-Ti type. The normative groups are described in Table 2.

(a) Olivine normative types

The olivine-normative types contain low concentrations of incompatable trace elements but are enriched in compatable elements. The major element chemistry of the olivine normative tholeiites is similar in most respects to mid-ocean ridge basalt. The olivine normative tholeiites, however, unlike "normal" MORB display a slight chondrodite normalized enrichment of light REE relative to heavy REE (Ragland and others, 1971; Bryan and others 1977; Philpotts and Schnetzler, 1968; Gottfried and others, 1977) suggesting a mantle source that was less depleted than the MORB source.

The two olivine normative types are distinguished from each other on the basis of differences in large-ion lithophile trace elements (Ragland and Whittington (1983).

Table 2

Early Jurassic ENA Igneous Rocks

	High-Ti Quartz Normative							Comparable Tholeiites	
	dikes and sills			basalt					
	high-Ti	York-Ha	Palisade Chill zone	Mt Zion	Orange Mt	Talcott	North Mt.	Karroo	"average"
No. of analyses	20	30	7	7	11	7	53	21	1228
SiO_2	51.1	51.84	51.98	51,37	51.45	51.16	52.16	51.8	51.2
TiO_2	1.12	1.09	1.22	1.18	1.02	1.06	1.06	1.13	1.6
Al_2O_3	14.2	14.34	14.48	14.24	14.34	14.08	14.29	14.8	15.9
Fe_2O_3	11.6	1.18	1.37	1.58	1.48	1.65	10.35	3.92	2.9
FeO	-	8.75	8.92	9.28	8.88	9.22	-	7.26	8.0
MnO	0.19	0.20	-	0.17	0.15	0.16	0.16	0.17	0.17
MgO	7.41	7.72	7.59	7.58	8.19	7.87	7.05	7.1	6.2
CaO	10.66	10.73	10.33	10.78	10.86	11.09	10.35	10.57	9.9
Na_2O	2.12	1.96	2.04	2.05	2.10	2.03	2.39	2.4	2.4
K_2O	0.66	0.60	0.84	0.21	0.54	0.49	0.60	0.74	0.7
P_2O_5	-	0.12	-	0.13	0.13	0.13	0.16	0.13	0.21
H_2O^+	-	0.23	1.04	0.66	0.48	0.93	0.69	-	0.81
H_2O^-	-	-	-	-	0.29	0.45	0.27	-	-
TOTAL	99.06	98.76	99.81	99.43	99.96	100.32	99.53	100.02	99.99
Trace elements (ppm)									
Ba	-	160	-	145	182	174	-	256	250
Co	49	47	53	61	45	47	-	34	38
Cr	277	302	315	282	260	322	-	317	153
Cu	111	121	110	104	127	123	114	-	141
Ni	81	89	95	79	61	72	82	73	77
Rb	21	25	-	11	37	22	21	30	33
Sr	186	187	175	191	183	186	244	190	471
V	-	310	-	273	272	270	296	300	266
Zn	84	77	-	-	96	92	-	-	-
Zr	92	115	120	99	116	87	112	85	111
Reference	1	2	3	4	5	6	7	8	9

Early Jurassic ENA Igneous Rocks (cont.)

High-Fe Quartz Normative

	Cu-depleated					Cu-enriched		
	High-Fe	Sanders	Preakness	Holyoke	Pomperaung	Pal-int.	Ladentown	Cushetunk
No. of analyses	15	14	11	10	19	7	5	2
SiO_2	52.69	52.26	52.24	51.79	51.35	51.70	51.69	51.74
TiO_2	1.14	0.99	1.20	1.06	1.17	1.58	1.26	2.24
Al_2O_3	14.21	14.39	13.70	14.16	14.04	14.08	14.24	11.75
Fe_2O_3	13.87	12.82	2.15	1.59	14.20	2.51	2.35	2.49
FeO	-	-	9.43	10.10	-	9.18	9.28	12.84
MnO	0.22	0.20	0.21	0.19	0.21	-	-	0.22
MgO	5.53	5.66	6.62	5.98	5.38	6.63	6.23	4.07
CaO	9.86	10.04	9.83	10.44	9.95	9.86	9.72	8.42
Na_2O	2.51	2.51	2.64	2.49	2.33	2.49	2.63	3.09
K_2O	0.64	0.54	0.65	0.57	0.39	0.82	0.61	1.67
P_2O_5	-	0.12	0.12	0.14	0.12	-	-	-
H_2O^+	-	-	0.82	0.77	0.63	-	0.75	1.08
H_2O^-	-	-	0.21	0.50	0.20	-	0.29	-
TOTAL	100.67	99.53	99.82	99.78	99.97	98.85	99.05	99.61
Trace elements (ppm)								
Ba	-	141	160	130	-	220	149	-
Co	52	60	45	51	-	55	48	62
Cr	94	123	38	29	33	206	-	6
Cu	74	74	81	84	54	143	150	581
Ni	34	58	31	42	46	68	55	24
Rb	22	21	40	28	24	-	-	79
Sr	178	136	139	159	317	188	222	206
V	-	302	333	317	335	350	282	270
Zn	99	-	104	106	-	-	-	-
Zr	94	95	94	87	103	133	-	198
Reference	1	4	5	6	10	3	11	12

Early Jurassic ENA Igneous Rocks (cont.)

	Low-Ti Quartz Normative		Hook Mt. - Hampden		Olivine - Normative	
	low-Ti	Rossville	Hook Mt.	Hampden	ol-norm	Quarry
No. of analyses	37	20	6	6	60	15
SiO_2	51.66	50.56	49.08	49.40	47.90	46.60
TiO_2	0.76	0.74	1.38	1.41	0.59	0.43
Al_2O_3	14.95	16.56	13.72	13.55	15.26	15.45
Fe_2O_3	11.77	1.07	4.23	3.76	12.10	1.66
FeO	-	9.02	10.10	10.44	-	8.42
MnO	0.20	0.18	0.23	0.23	0.18	0.17
MgO	7.44	6.79	5.93	5.63	10.52	13.10
CaO	10.80	10.81	10.36	10.68	10.75	10.55
Na_2O	2.23	1.95	2.21	2.22	2.00	1.57
K_2O	0.48	0.39	0.37	0.43	0.29	0.35
P_2O_5	-	0.09	0.16	0.19	-	0.07
H_2O^+	-	0.46	1.16	1.07	-	1.15
H_2O^-	-	-	0.32	0.72	-	-
TOTAL	100.29	98.62	99.25	99.74	99.59	99.52
Trace elements (ppm)						
Ba	-	115	110	140	-	97
Co	53	46	56	53	65	67
Cr	218	205	62	63	766	1020
Cu	68	66	188	187	108	102
Ni	48	63	51	50	308	455
Rb	15	21	34	24	8	22
Sr	127	137	96	163	115	136
V	-	-	385	355	-	200
Zn	86	79	125	138	84	71
Zr	60	66	94	108	50	25
Reference	1	2	5	6	1	2

Reference: Table 2

1. High-Ti, low-Ti, high-Fe, and ol-norm tholeiites (Weigand and Ragland, 1970).
2. York-Haven, Rossville, and Quarryville tholeiites (Smith and others, 1975).
3. Palisades Sill (7 chill zone and 7 interior zone, int. samples (Walker, 1969).
4. Mt. Zion Church, and Sanders Basalts of Culpeper Basin, Virginia, (Puffer, Hurtubise, and Lievy, in preparation).
5. Orange Mountain, Preakness, and Hook Mountain Basalts of Newark Basin, New Jersey (Puffer and Lechler, 1980).
6. Talcott, Holyoke, and Hampden Basalts of Hartford Basin, Connecticut (Puffer, and others, 1981).
7. North Mountain Basalt of Fundy Basin, Nova Scotia (Puffer, Hurtubise and Olsen, in preparation).
8. Karroo Basalts of South Africa (Cox and Hornung, 1966).
9. Average of 1228 tholeiitic basalts and dolerites, world-wide (Manson, 1967; and Prinz, 1967).
10. Pomperaung Basalt of Pomperaung Basin, Connecticut (Hurtubise and Puffer, in preparation).
11. Ladentown Basalt of Rockland County, New York (Puffer and others, 1982).
12. Cushetunk Mountain dolerite, New Jersey (Puffer and Lechler, 1979).

TABLE 3. RARE EARTH ELEMENT CONTENT OF FIRST AND THIRD WATCHUNGS

	First Watchung	Third Watchung
	GE-1662	F1
La	9.8	8.39
Ce	22.9	22.2
Nd	13.0	11.8
Sm	3.34	3.66
Eu	1.07	1.21
Dy	4.10	5.93
Er	2.39	4.14
Yb	2.18	4.38
La/Yb	4.50	1.92
La/Sm	2.93	2.29
$(La/Sm)_{E.F.}$*	2.03	1.59

Note: Data from Kay and Hubbard (1978)

*$(La/Sm)_{E.F.}$ = chondrite-normalized ratio of La to Sm; E.F. = enrichment factor.

The texture of the low-LIL olivine normative diabase is typically subophitic with very few phenocrysts. Typical low-LIL diabase is composed of approximately 55 percent plagioclase (An_{70}), 30 percent augite, and 15 percent olivine (Fo_{80}). Accessory and trace minerals include approximately 2 percent ilmeno-magnetite and traces of ilmenite, iron sulfide and chromite. Modal data pertaining to the recently recognized high-LIL dike population are not yet available.

(b) High-Ti quartz tholeiites

Weigand and Ragland (1970) found that the ENA quartz-normative tholeiites may be subdivided on the basis of titanium content. The high-Ti type plots onto a TiO_2 versus mafic index diagram as a distinct cluster of points separated from a low-Ti cluster. The high-Ti type also plots close to a high-Fe cluster that is now recognized as a variety of the high-Ti type. The high-Ti type as originally defined (Weigand and Ragland, 1970) contains from 0.95 to 1.25 weight percent TiO_2 with a mafic index of 57 to 65. The incompatable element content of the high-Ti type is higher than the low-Ti type and much higher than the olivine normative types (Table 2),

The chemical composition of the high-Ti tholeiites falls within the chemically diverse group of continental tholeiites and are particularly similar to the basalts of the Basutoland subprovince of the South African Karroo province (Cox, and Hornung, 1966). On a world-wide basis high-Ti tholeiites are somewhat lower in Na_2O, Al_2O_3, TiO_2, and Sr than an average of 1228 tholeiitic basalts and dolerites (Table 2), and contain less normative albite and ilmenite.

The high-Ti type occurs as both intrusives and flows. Where it occurs as diabase such as the Palisades sill, it is intergranular to subophitic and consists of approximately 43 percent plagioclase (An_{65}), 50 percent clinopyroxene, 2 percent hypersthene, 2 percent olivine, 2 percent ilmeno-magnetite, and 1 percent ilmenite. Accessory minerals include apatite, biotite, pyrite, and chalcopyrite.

Where the high-Ti type occurs as basalt flows it consists of approximately 35 percent plagioclase (An_{65}), 35 percent pyroxene (augite, pigeonite and hypersthene), 28 percent glass, and 3 percent opaque iron-titanium oxides. Accessory and trace minerals include apatite, biotite, K-spar, and pyrite. Phenocrysts typically include olivine, plagioclase, glomeroporphyritic clusters of augite and plagioclase, and less common orthopyroxene rimmed by augite and olivine. Philpotts and Reichenbach (1983) experimentally found that olivine and plagioclase appear on the liquidus simultaneously and suggest that the olivine and plagioclase phenocrysts formed upon extrusion and cooling of the Talcott basalt. Augite and pigeonite formed 15°C below the liquidus.

(c) High-Fe quartz tholeiite

The high-Fe quartz tholeiites are characterized by a high mafic index ranging from 65 to 75. Compared with other ENA tholeiites, the high-Fe type contains high incompatable trace element concentrations and low compatable trace element concentrations. The high-Fe tholeiites may be further subdivided into a Cu, and Ti/Fe, enriched group and a depleated group (Table 2). The major and trace element range of the high-Fe tholeiites is greater than any of the other ENA groups largely because of fractionation relationships.

The high-Fe type occurs both as diabase intrusions and basalt flows. In both cases there is a wide textural and mineralogical range. Where the type occurs as diabase it typically consists of 59 percent plagioclase (An_{55}), 33 percent augite, 7 percent pigeonite, 2 percent hornblende, 3 percent opaque oxides, and commonly about 3 percent quartz - micropegmatite intergrowth. Accessory minerals include biotite, chlorite, apatite, sphene, and occasionally some olivine. Some high-Fe diabase contains up to 25 percent orthopyroxene (En_{60-80}) with less augite and plagioclase and no pigeonite.

Where the high-Fe type occurs as basalt flows it is typically much coarser grained than high-Ti basalt. The high-Fe type of basalt is composed of approximately 55 percent plagioclase (An_{55}), 40 percent pyroxene (augite, pigeonite, and minor orthopyroxene), 5 percent glass, 3 percent iron-titanium oxides and traces of altered olivine. Accessory K-feldspar, quartz, biotite, apatite, and pyrite are found in much larger quantities than in high-Ti basalt flows. The basalt is approximatly aphyric but a fine grained generation of plagioclase and pyroxene is mixed with glass as a mesostasis that comprises 25 to 35 percent of the rock.

At the Southbury (Pomperaung) basin of Connecticut, Philpotts (1979) found good evidence of liquid immiscibility in high-Fe basalt that takes the form of glass spheres surrounded by a less reflective glass. The immiscible globules are absent from the lowest 5.5 m of the flow, but are abundant in the basalt above 9 m from the base. The lower portion contains magnetite as an early phase and crystallized at oxygen fugacites close to the NNO buffer while the upper portion contains late dendritic magnetite and crystallized near the QFM buffer (Philpotts and Doyle, 1983).

(d) Low-Ti quartz tholeiites

On a mafic index versus TiO_2 diagram the low-Ti type constitutes a distinct cluster of points ranging from 0.6 to 0.9 percent TiO_2; much lower than the high-Ti type (Table 2). The low-Ti type contains relatively low concentrations of both compatible trace elements and incompatable trace elements.

The low-Ti type occurs in eastern North America predominately as diabase intrusives. The intrusions are intergranular to subophitic typically consisting of about 47 percent plagioclase (An_{60}), 50 percent pyroxene, 2 percent iron-titanium oxides, and 1 percent olivine (Fo_{85}). Pyroxene phynocrysts are augite and groundmass pyroxene in a mixture of pigeonite and hypersthene. Accessory and trace minerals include apatite, pyrrhotite, and chalcopyrite.

(e) Alkali olivine diabase

A group of Late Triassic or Early Jurassic alkali-olivine diabase intrusions have been found in eastern New England, from Rhode Island northward through New Hampshire and southern Maine (McHone and Trygstad, 1982; Pierce and Hermes, 1978). Stoddard (1983) has also described a Late Triassic-Early Jurassic alkali-rich suite from the eastern North Carolina Piedmont, but very little petrologic or geochemical data are yet available.

The New England intrusions are generally olivine-bearing augite plagioclase dolerites, somewhat altered, and are porphyritic with plagioclase phenocrysts. They tend to be slightly nepheline normative to transitional between alkalic and tholeiitic basalts, and are marked by high TiO_2 values (from 1.6 to 2.2 weitht percent). Rare-earth elements and mineral compositions indicate a lack of substantial contamination or crystal fractionation in their petrogenetic history (Hermes and others, 1984). Although most likely members of the ENA dolerite/basalt province, these alkalic intrusions remain poorly understood, and are not well dated.

Hurtubise and Puffer (in prep.) are examining all known eastern North American occurrences of alkali-olivine diabase and basalt and find that most such occurrences belong to an early Cretaceous population.

CHAPTER TWO

RELATIONSHIPS AMOUNG ENA THOLEIITTES

by

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1. Fractionation relationships

The degree to which the various rock populations are related by means of fractionation processes is a subject that is undergoing considerable current investigation: Ragland and Whittington (1983), Philpotts and Reichemback (1983), Puffer and Hurtubise (1983) Gottfried (1983), Pegram (1983), Husch (1983). Some evidence suggests that fractionation processes were of minor importance in the geochemical development of the various rock. populations whereas other evidence suggests that it was a major factor.

Olivine normative-quartz normative fractionation:

Weigand and Ragland's (1970) original fractionation concept operated under the assumption that quartz-tholeiites probably developed out of olivine-normative tholeiitic magma by fractionation at shallow depth ($<$ 15 km) as implied by the work of Green and Ringwood (1967). Weigand and Ragland (1970) suggested that some ENA olivine - normative magma was ponded in the crust where it fractionated into the quartz-normative types. Smith and others (1975) agreed with their concept that suggested that some olivine-normative diabase magma (the Quarryville diabase of Pennsylvania) fractionated to form both of the quartz normative diabase types of Pennsylvania (the York Haven and the Rossville). Where the olivine normative magma rose rapidly into the upper crust it formed the Quarryville diabase; where it rose more slowly through the mantle it crystalized olivine and assimilated orthopyroxene then differentiated at shallow depths to yield the York Haven diabase (a high-Ti type). Crustal assimilation processes were suggested as having increased the silica and incompatable element content of the magma. Smith and others (1975) suggested that the Rossville magma (a low-Ti type) underwent a sequence of events similar to the York Haven but with little to no crustal assimilation.

The work of Smith and others (1975) and Ragland and others (1971) point out the inadequacy of simple crystal fractionation as a complete explanation of the high degree of contrast in the REE content of the olivine-normative versus the quartz-normative magmas. Ragland and others (1971) also point out that one problem with selective contamination as an alternative explaination is the chemical uniformity of the high-Ti tholeiites. The high-Ti tholeiites are virtually identical from Nova Scotia to Georgia (Table 2).

The uniformity in composition of the high-Ti tholeiites also led Bryan and others (1977), deBoer and Snider (1979) and Puffer and others (1981) to conclude that is is highly unlikely that fractionation would have occurred under identical conditions throughout eastern North America and then progressed to exactly the same point before the resulting high-Ti magma was intruded and extruded. As stated by Weigand and Ragland (1970) "Several episodes of diapir ascent, partial melting, magma segregation and fractionation, each resulting in extraordinarily similar chemical types is difficult to envision".

We are, therefore, faced with two difficulties in explaining olivine-normative→quartz normative fractionation amoung ENA tholeiites: 1) the highly contrasting trace element contents and 2) the high degree of compositionaal uniformity of the high-Ti type.

The recent recognition of the high-LIL olivine-normative population (Ragland and Whittington, 1983) solves the trace element gap problem as applied to the low-Ti population. It appears that fractionation of high-LIL magma could have generated low-Ti magma. The generation of high-Ti magma, however, remains a difficult problem.

Cox (1980) also recognises the "uniform magma" argument as applied to the Karroo basalts of South Africa but points out that when examples of "uniform" continental flood basalt are examined in detail, fractionation trends consistent with deep crustal pressures (up to 12 kb) are evident. Cox (1980) shows that crystallization of olivine, clinopyroxene, and plagioclase at deep crustal pressures may buffer chemical changes (particularly silica) thus partially accounting for the apparent uniformity. Cox (1980) suggests that extensional tectonic settings are consistent with the implacement of deep picrite sill complexes at the base of the crust. Fractionation of these deep sill complexes yield magmas of decreasing densities that are transmitted to the surface as soon as the densities are sufficiently deminished.

Stolper and Walker (1980) find that the densities of fractionating magmas decrease until pyroxene and plagioclase join the crystallization sequence. Further fractionation causes density to increase. They find that density is largely a function of Fe/Fe+Mg (mol) which is at a minimum within a 0.3 to 0.6 range which they refer to as a "window of eruptibility". Of the various early Jurassic magma types, the high-Ti magma plots closest to the density minimum (0.44).

An alternative to the high-Ti population as a fractionation product is the possibility that it may be a primary magma. Carmichael and others (1974) suggest that continental flood basalts of uniform composition are probably unfractionated primary melts of mantle peridotite. Ragland and Whittington (1983) suggest that if the high-Ti magma was primary it must have formed under higher P_{H_2O} conditions than the olivine normative types. dePaolo (1979) has shown that with 2.5 percent water a quartz tholeiite could equilibrate with the mantle above the pyrolite solidus at pressures up to about 15 kb. Such a wet magma sourse would be unlike that of the MORB source but the trace element data of Pegram (1983) indicates that the sourse of quartz normative ENA magma was clearly distinct from the MORB source and was instead chemically similar to an island arc source.

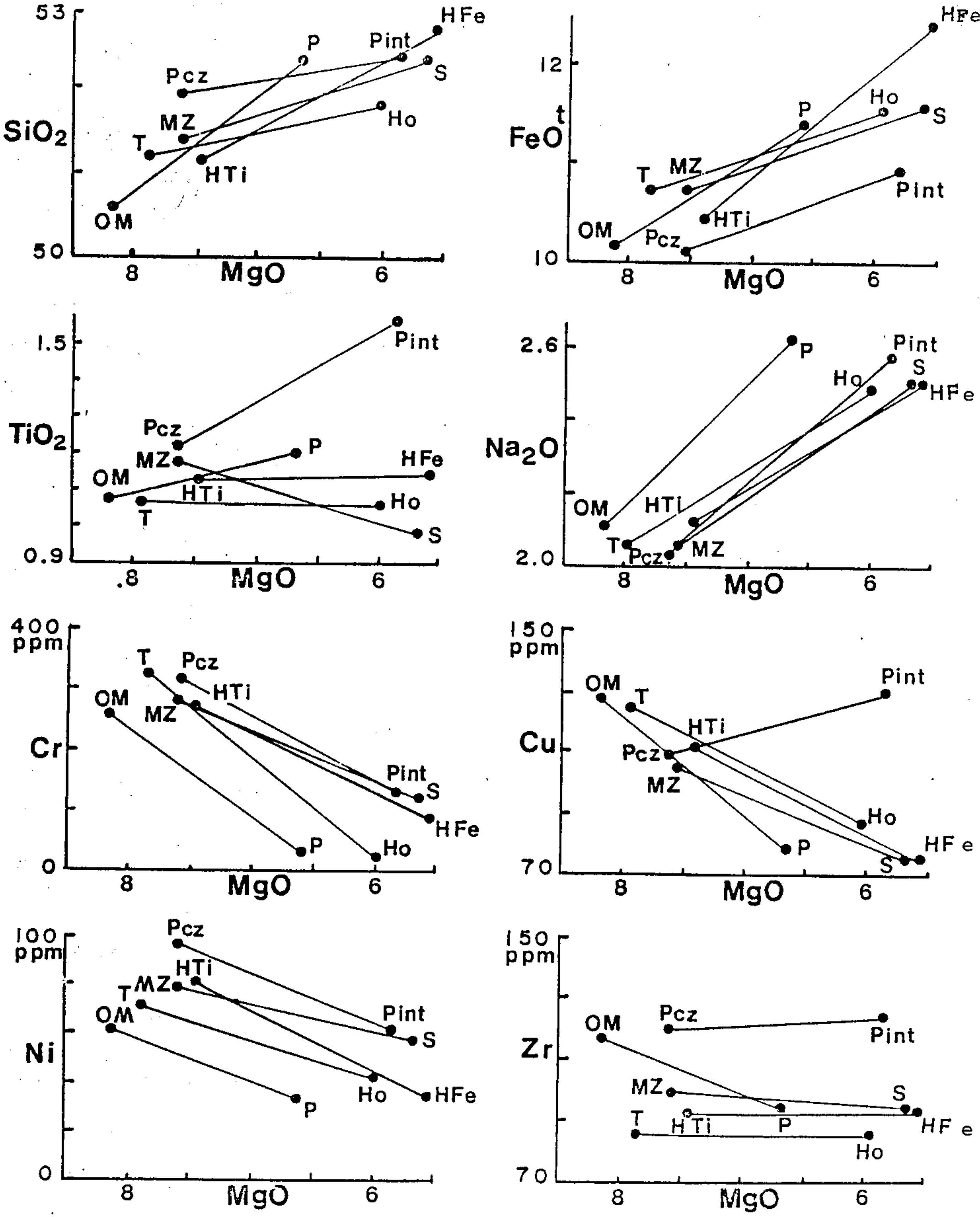
High-Ti → High-Fe fractionation.

Weigand and Ragland (1970) first suggested the probability that high-Fe tholeiites are fractionation products of high-Ti tholeiites. Support for this suggestion was offered by Puffer and Lechler (1980), who demonstrated that the Second Watchung flow of New Jersey (a high-Fe type) is a fractionation product of the First Watchung (a high-Ti type). A magma of approximately the composition of the Second Watchung can be calculated by means of Wright and Doherty's (1970) modeling by separating augite, plagioclase and olivine from First Watchung magma. Puffer and others (1981) also indicate that the Holyoke flows of Connecticut (high-Fe) could have been derived from Talcott magma (high-Ti) through similar fractionation.

Reichenbach and Philpotts (in press) agree that this is a feasible mechanism of differentiation but interpret the orthopyroxene as a refractory residue from an upper mantle source, some of which was assimilated. Their calculations call for the assimilation of 7.8 percent orthopyroxene into Talcott magma and crystallization of 7.9 percent olivine, 15 percent clinopyroxene, and 13.3 percent plagioclase to yield Holyoke magma. The orthopyroxene assimilation mechanism is similar to that suggested by Smith and others (1975) as applied to olivine-normative (Quarryville) → high-Ti (York Haven) fractionation.

Similar fractionation processes appear to explain the relationship of the Hickory Grove, Sanders and Millbrook basalts of Virginia (high-Fe types) to the underlying Mount Zion Church basalt (a high-Ti type). More recently Husch (1983) has shown that four high-Fe diabase bodies from western New Jersey are probably fractionation products of high-Ti magma as represented by three other New Jersey diabase bodies located near by.

Figure 3


53
50
SiO2
MgO
8
6
HFe
P
Pint
S
Pcz
MZ
T
Ho
HTi
OM
12
10
FeO t
HFe
Ho
P
S
T
MZ
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OM
Pcz
1.5
0.9
TiO2
Pint
Pcz
P
MZ
HFe
OM
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.8
2.6
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Na2O
P
Pint
S
Ho
HFe
OM
HTi
T
Pcz
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400 ppm
0
Cr
T
Pcz
HTi
OM
MZ
Pint
S
HFe
P
Ho
150 ppm
70
Cu
OM
T
Pint
HTi
Pcz
MZ
Ho
P
HF e
S
100 ppm
0
Ni
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HTi
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150 ppm
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Zr
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Figure 3.

Fractionation of high-Ti type ENA quartz tholeiites. MgO (weight percent) variation diagrams illustrating fractionation of high-Ti tholeiites into high-Fe tholeiites:

High-Ti (No of samples)	High-Fe (No of samples)	ENA Basin	References
Talcott Basalt, T, (7)	Holyoke Basalt, Ho, (10)	Hartford CO	Puffer and others, 1981
Orange Mt Basalt, OM, (8)	Preakness Basalt, P, (11)	Newark NJ	Puffer and Lechler, 1980
Palisades dolerite, chill zone, Pcz, (7)	Palisades dolerite interior, Pint, (5)	Newark NJ	Walker, 1969
Mount Zion Church Basalt, MZ, (7)	Sanders Basalt, S, (14)	Culpeper VA	Puffer, Leivy, and Hurtubise, in prep.
High-Ti dolerites, HTi, (20)	High-Fe dolerites, HFe, (15)	ENA	Weigand and Ragland, 1970

Note: 1) the tight but geographically random clustering of high-Ti tholeiites.

2) the divergent TiO_2 and Cu fractionation trend of the Palisades Sill is probably due to shallow, in-situ fractionation of Palisades magma in contrast to much deeper seated fractionation of more typical high-Ti tholeiites. Other shallow high-Ti sills such as Dillsburg, PA (Hotz, 1953) and Cushetunk Mountain, NJ (Puffer and Lechler, 1979) display TiO_2 and Cu fractionation trends similar to Palisades dolerite and presumably crystallized too rapidely and at confining pressures too low to permit gravitational separation of dence titanium oxides or immisable Cu sulfide liquid phase.

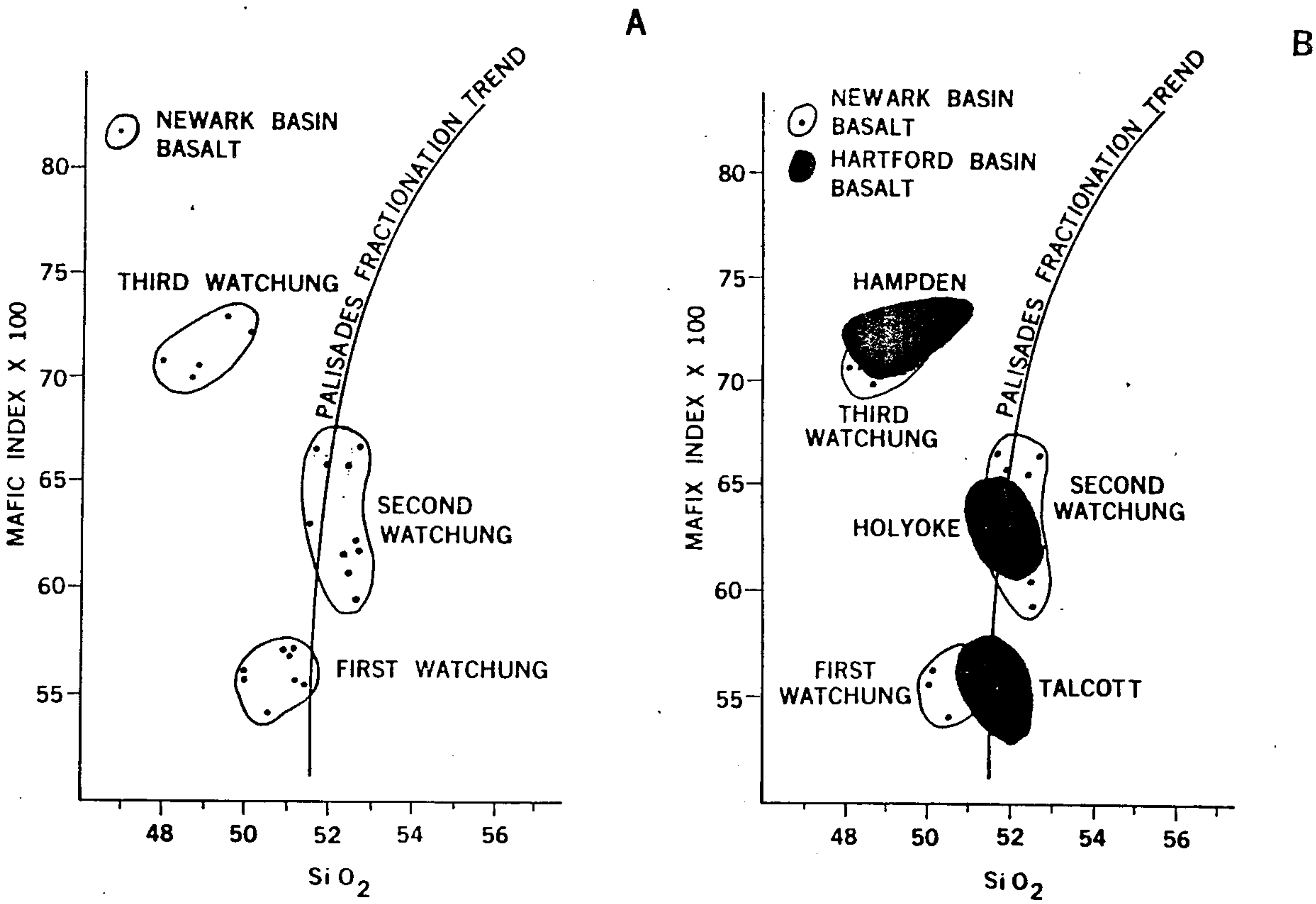


FIGURE 4. SiO_2 VS MAFIC INDEX $(FeO+Fe_2O_3/FeO+Fe_2O_3+MgO)X100$ (NEWARK BASIN DATA FROM PUFFER AND LECHLER, 1980 AND HARTFORD BASIN DATA FROM HURTUBISE, 1979).

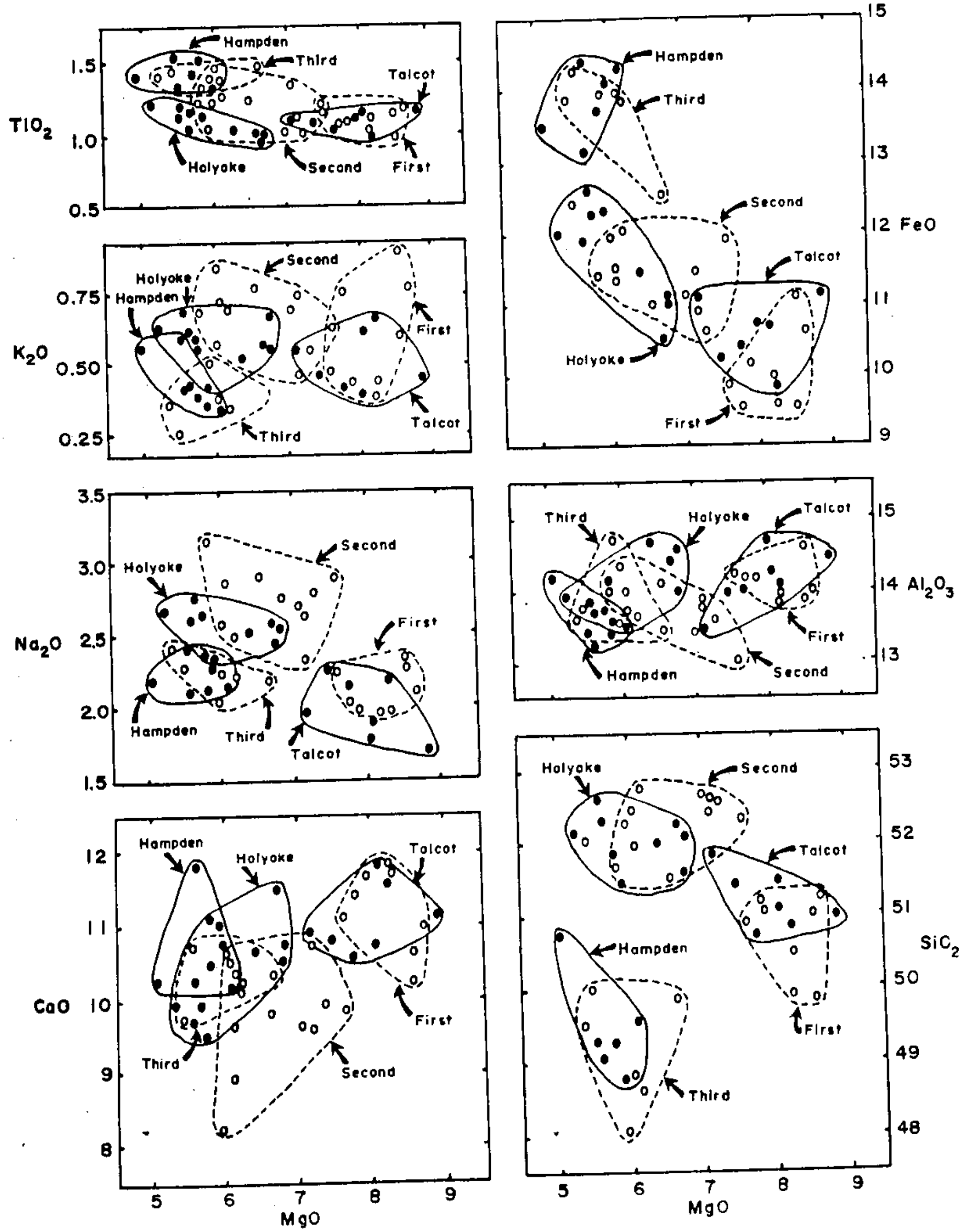


Fig. 5 MgO variation diagram comparing the chemistry of the Watchung basalts (first - Orange Mountain, second - Preakness, third - Hook Mountain) with the basalts of the Hartford Basin.

In each case the major element chemistry of the fractionated high-Fe tholeiites plots directly on the Palisades fractionation trend as described by Walker (1969). The trace element chemistry of the high-Fe tholeiites, however, falls into two groups, only one of which plots onto the Palisades trend: (1) a Cu, and Ti/Fe, enriched group and (2) a Cu, and Ti/Fe, depleted group (Table 2).

During Palisades trend fractionation, Cu and Ti become enriched. Examples of other high-Fe type tholeiites that display similar Palisades-type enrichment include the Dillsburg Sill of Pennsylvania (Hotz, 1953), the Ladentown Basalt of New York (Puffer and others, 198), and the Cushetunk Sill of New Jersey (Puffer and Lechler, 1979). Most high-Fe dikes (Weigand and Ragland, 1970) and flows (Puffer and others, 1981), however, display deminished Cu and Ti/Fe values (Table 2). Gottfried (1983) suggested that the contrasting trends in Cu are due to the strong partitioning of Cu into an immicible sulfide phase that separated from most high-Fe magma during ascent. A concurrent study by Puffer and others (1983) observed contrasting Cu and Ti/Fe trends among some high-Fe tholeiites associated with the Palisades Sill and also suggested separation of an immicible sulfide phase and ilmenite from the Cu and Ti/Fe depleated rock.

Puffer and others (1983) suggested that the difference between the rocks displaying the enrichment trend vs. those displaying the depletion trend was the depth at which ponding and fractionation was occuring. McBirney and Noyse (1980), found that gravitational settling is unlikely to occur in magmas unless the separating phase is considerably denser than the silicate magma phase (such as a sulphide or oxide phase) and unless crystallization is occurring slowly enough (as in a large deep magma chamber) to permit effective separation. The in-situ fractionation that was occurring within the Palisades and Dillsburg Sills after they were emplaced into shallow crustal sites was presumably occurring under oxygen and sulfur pressures, volatile migration rates, and crystallization rates that contrast with the deep crustal levels of most high-Ti fractionation. Philpotts and Reichenbach (1983) suggests that fractionation of Talcott magma (a high-Ti type) occurred at deep crustal levels, perhaps the base of the crust, which was presumably an environment more condusive to gravitational separation than the shallow crustal environment of the Palisades.

Third Watchung-Hampden Fractionation

The uppermost flow unit of the Newark Basin (the Third Watchung or Hook Mt. Basalt) and the Hartford Basin (the Hampden Basalt) are chemically the same as each other (Table 2) but are unlike any of the underlying flows. The Si, Na, Rb, Ni, Cr, and Sr contents of the Third Watchung-Hampden flows are intermediate between those of the lowermost flow units and the middle flow units. Puffer and others (1981), and Reichenbach and Philpotts (in press), therefore, point out that the upper flows could not have developed as a continuation of the same fractionation trend that yielded the middle flow units (the Preakness - Holyoke Basalts). Reichenbach and Philpotts (in press), have suggested that the uppermost flow unit may have developed out of high-Ti magma via an independent parallel fractionation route. They suggest that a 31 percent crystallization of Talcott magma (a high-Ti type) to form plagioclase (An_{72}), augite ($Ca_{37}Mg_{58}Fe_5$), orthopyroxene (En_{87}), and olivine (fo_{90}) will produce a residue of Hampden composition. Their fractionating magma was at least temporarily ponded in a shallow crustal environment where presumably late crystallizing olivine could become involved in the fractionation process and where high-f_{O2} conditions (equivalent to the Ni-NiO buffer) could generate the high ferric/ferrous ratio characteristic of the Hampden. The enriched Cu and Ti/Fe of the Hampden and Hook Mt. (Table 2) is consistent with shallow crustal fractionation.

2. Spatial relationships

(a) Low-LIL olivine tholeiite-dikes are dominate in the Carolinas and strike NW. They occur exclusively south of the Newark Basin as far as the southeastern edge of Alabama, (Fig.2). The dikes of South Carolina are exclusively olivine-normative but to the north, in Virginia, North Carolina and Pennsylvania they are closely associated with high-Fe, high-Ti and low-Ti dikes. To the south, in Georgia and Alabama they are associated with low-Ti dikes.

(b) High-LIL olivine tholeiites are less common than the low-LIL dikes but have been recognized in the Carolinas and Virginia (Ragland and others, 1983) where they occur as a major N-S trending dike swarm that converges within about 30° of arc upon an area between Charleston and Georgetown, South Carolina.

(c) High-Ti quartz tholeiites are the most widespread and common ENA type and are probably the most volumetrically abundant. The type may also include many of the quartz-tholeiite occurrences of Morocco (Manspeizer and others, 1978) and other African locations. The type typically occurs as thick sills and flows from Newfoundland to Virginia (Fig. 2) and as dikes from Pennsylvania to Virginia. Where the type occurs as dikes they generally trend to the NE or N-S. Where the high-Ti tholeiites occur as basalt flows they are consistently the earliest member of any flow sequence and are typically overlain by high-Fe basalt flows.

(d) Low-Ti quartz tholeiites - are common throughout the Paleozoic and Precambrian terrain of Georgia and Alabama where they occur together with low-LIL olivine normative tholeiites but are also found in the Messozoic Gettysburg and Culpeper Basins. They occur exclusively as dikes and rare sills and generally trend to the NW. The distribution of the low-Ti population is not well documented throughout New England where many of the ENA intrusives have not been chemically analyzed.

(e) Alkali olivine diabase - is found as dikes in southeastern and northeastern New England, especially from Rhode Island to Southern Maine. The dikes average about N 40° E in strike and occur outside but marginal to the rift basins in the area (Hermes and others, in press).

(f) High-Fe quartz tholeiites - do not constitute a separate group according to the recent classification scheme of Ragland and Whittington (1983) but are a differentiate of other types, most commonly the high-Ti type. The high-Fe tholeiites occur either as extrusive flows, as the interior portion of thick high-Ti sills or as independent high-Fe dikes (Table 2). Wherever high-Fe tholeiites occur as part of a composite or fractionated sill the remaining portion of the sill is a high-Ti tholeiite. Wherever the high-Fe tholeiites occur as flows, they are the second member of the flow sequence and overlie a high-Ti basalt member.

3. Temporal Relationships

(a) Field evidence.

Crosscutting field relationships among the ENA dolerites are rare but a few have been observed. Smith and others (1975) report that the Rossville dikes of Pennsylvania (a low-Ti type) crosscut the York Haven dikes (a high-Ti type). Lanning (1972) also concluded that the Quarryville dikes (a low-LIL olivine normative type) are cut by Rossville dikes.

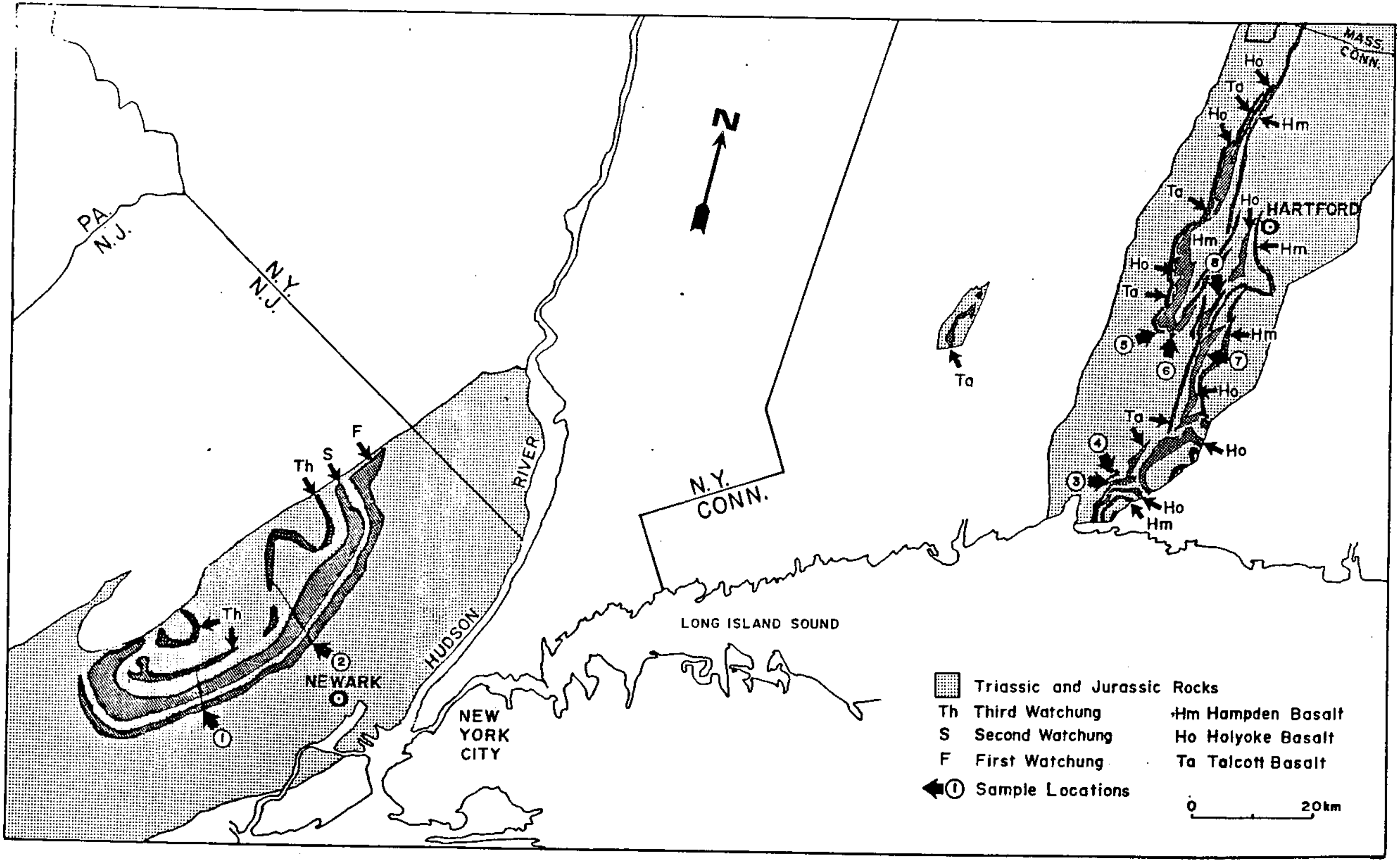


Fig. 6 Basalts of the Newark and Hartford basins, with sample locations (Puffer and others, 1981).

(b) Radiometric evidence

East-Central Alabama - Deininger and others (1975) report ages of 184 to 193 Ma for the olivine-normative (low-LIL) Salem dike and 161 to 168 Ma for the low-Ti quartz normative Auburn dike.

Georgia Piedmont - Smith and Dooley (1983) report that conventional K-Ar ages of NW trending Georgia diabase intrusions (principally low-LIL olivine normative) are highly variable (190 to 1620 Ma). Most dates, however, range from 190 to 195 Ma.

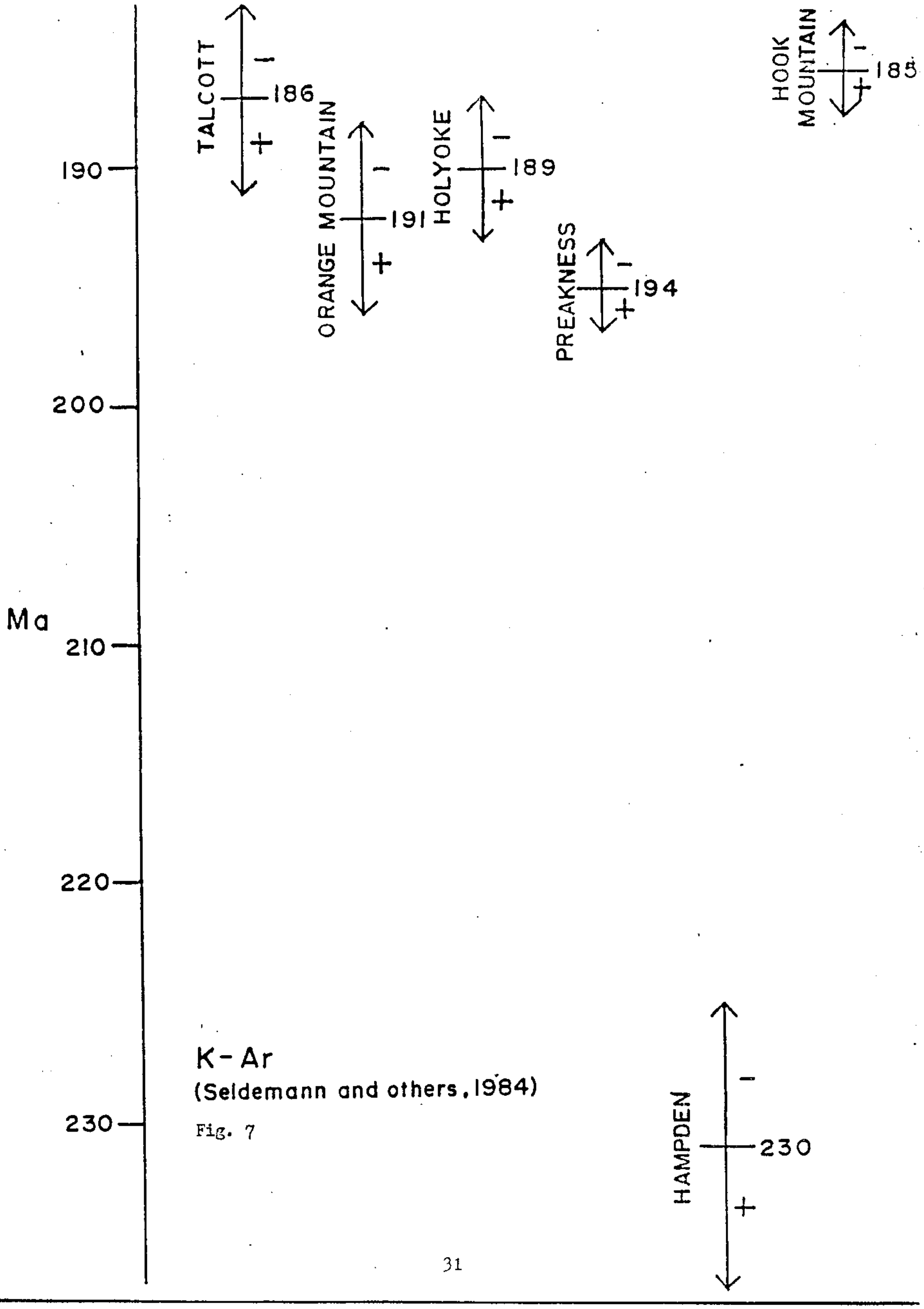
The Carolinas - Smith and Dooley (1983) report that preliminary K-Ar ages of 195 Ma have been determined for several diabase dikes from South Carolina (principally low-LIL olivine normative).

Culpeper Basin - Sutter and others (1983) report that diabase sills intruded into the Culpeper Basin yield total-gas $^{40}Ar/^{39}Ar$ apparent ages from 187 to 206 Ma and plateau ages that range from 192 to 200 Ma that connot be distinguished from one another at the 95 percent confidence level. The Culpeper diabases yield a mean age of 197 ± 4 Ma and include high-Ti, low-Ti and low-LIL types.

Gettysburg Basin - Sutter and Smith (1979) report an average $^{40}Ar/^{39}Ar$ plateau age for the Gettysburg sill of 177 ± 6 Ma (a complex multiple intrusion of high-Ti and low-Ti diabase) and for the Frederick dike of 177.5 ± 3.4 Ma which is a low-Ti type intrusion.

Newark Basin - The Palisades Sill (a high-Ti type) has yielded a 190 ± 5 Ma K/Ar range (Lambert, 1971) and a 186 to 192 Ma $^{40}Ar/^{39}Ar$ range (Dallmeyer, 1975). Seidemann and others (1984) report K-Ar dates of 191 ± 8, 194 ± 4, and 185 ± 4 Ma for the Orange Mountain, Preakness, and Hook Mountain Basalts (Fig. 7).

Hartford Basin - Sutter and Smith (1979) presented radiometric data pertaining to five diabase intrusives in Connecticut. They report an average $^{40}Ar/^{39}Ar$ age of 183.8 Ma (172 to 194) and an average $^{40}Ar/^{36}Ar$ isocron of 175 Ma (168 to 180). Most of these intrusions are high-Ti types. Seidemann and others (1984) report K-Ar dates of 186 ± 8, 189 ± 6, and 230 ± 12 Ma for the Talcott, Holyoke, and Hampden Basalts (Fig. 7). The sequence of age data, however is the reverse of the stratigraphic sequence. Seidemann and others (1984) attribute difficulties



K-Ar
(Seldemann and others, 1984)

Fig. 7

in obtaining accurate K-Ar dates to the presence of excess radiogenic ^{40}Ar that may have been added during hydrothermal alteration of the basalt.

Fundy Basin - Papezik (1983) reports that the quartz-normative basalt and diabase of the Fundy Basin has been dated at 190 to 210 Ma.

Most radiometric data, therefore, suggests the existence of at least two igneous events during the Jurassic: 1) a 185 to 195 Ma event dominated by low-LIL olivine normative tholeiites in the Southeast and dominated by high-Ti tholeiites in the Northeast; a 175 Ma event dominated by quartz-tholeiite intrusions (principally low-Ti ?).

(c) Paleomagnetic evidence

Southeastern U.S.A. - Smith and Dooley (1983) report that paleomagnetic data suggest that the NW suite of diabase dikes typical of the southeast (the low-LIL olivine normative type) may be older that a less common N-S trending suite recognized in North Carolina (the high-LIL olivine normative type).

Culpeper Basin - Sutter and others (1983) suggest a paleomagnetic pole near the apparent polar wandering path for North America passed through the Culpeper sills at about 200 Ma.

Gettysburg Basin - Volk (1977) determined that the Rossville and York Haven dolerites are distinctly different both in age and in magnetic domain structure. She places the Rossville-type diabase at about 185 Ma and suggests that it is the youngest diabase in the Newark Supergroup.

Newark Basin - Hozik and Colombo (1984, this guidebook) determined that the Byram, Sourland Mountain and Rocky Hill intrusions (high-Ti and high-Fe types) and the Sand Brook Basalt (a high-Ti type) are about the same age and were emplaced at the same time as the Palisade Sill, Orange Mountain Basalt and possibly the Preakness Basalt. Their data does not correlate with either of the 175 or 190 Ma events suggested by Smith and Noltimier (1979) but does plot near the polar wandering path of Harrison and Lindh (1982) for the time period

180 to 200 Ma. Hozik and Colombo (1984) conclude that it now appears likely that there was only one major period of igneous activity in the Central Newark Basin (a high-Ti - high-Fe event) in agreement with Puffer and Hurtubise (1983).

McIntosh and others (in press) have determined that there is virtually no paleomagneticaly measurable time gap separating the Orange Mountain (First Watchung) from the Preakness (Second Watchung) basalts in good agreement with Puffer and others (1981). The Feltville Formation, therefore, was probably deposited rapidly, well within one million years.

Hartford Basin - Smith and Noltimer (1979) presented new paleomagnetic results for 20 sites on dikes from the Hartford Basin and reviewed previously published data. They conclude that the paleomagnetic data support two periods of igneous activity: (1) a 190 Ma group that includes the Talcott, Hampden, Holyoke Basalts and the East Rock, West Rock, and Mt. Carmel diabase; and (2) a 175 Ma group including the Barndoor, Bridgeport, Buttress, Cross Rock, Fairhaven, and Higganaum dikes. Their data, however, is inconsistent with the work of Martello and others (1984) who demonstrate a genetic relationship of most of the Connecticut dikes to the 190 Ma flows.

De Boer and Snider (1979) found that inclination variation curves along the Appalachians for North American poles pass through the Talcott (218-244 Ma), the Holyoke (192-218 Ma) and the Hampden (168-192 Ma). Their data suggests that the Talcott is older than any of the Newark or Gettysburg rocks most of which plot within the Holyoke range. The radiometric, paleontological, geochemical, and petrographic data, however, suggests that the Talcott is a time stratigraphic correlative of the Orange Mountain (First Watchung) and Jacksonwald (York Haven) basalts.

The paleomagnetic data, therefore, supports the existence of at least three igneous events during the Jurassic: 1) an early low-LIL olivine normative event in the southeast followed by 2) a high-LIL olivine normative event in the southeast that was presumably concurrent with the high-Ti event of the northeast followed by 3) a quartz-normative (principally low-Ti ?) event.

(d) Paleontological evidence

Olsen and others (1982) assessed all the available paleontological and palynological data and found that the time span over which basalt flows were deposited is limited to the Hettangian and Simemurian of the Early Jurassic. Their new fossil fish data suggested that the Towaco Formation above the Preakness Basalt (Second Watchung) of the Newark Basin and the lower Turner Falls Sandstone above the Deerfield Basalt of the Deerfield Basin are older than the Shuttle Meadow Formation above the Talcott Basalt of the Hartford Basin and the sediments above the Mt. Zion Church Basalt of the Culpeper Basin.

Important new discoveries of fossil fish in the Haymarket Fish Bed of the Culpeper Basin, however, have been interpreted by Olsen (1983) as indicating more overlap amoung fish groups than previously thought, this strongly suggests ... " the appealing assumption of a one to one correlation of Jurassic formations of the Newark and Hartford Basins may be correct after all - as suggested by the geochemical data of Puffer, Hurtubise, Geiger, and Lechler (1981)." Unfortunately Olsen's new data have recently been overlooked (Seidemann and others, 1984 and McIntosh and others, in press, for example).

The paleontological evidence of Olsen and others (1982) and the sedimentological evidence of Heubert (1983) indicate that the lakes that occupied the Newark Basin probably did not connect with the lakes that occupied the Hartford Basin but this does not perclude the possibility that the Basalt flows were comagmatic.

(e) Geochemical and Petrographic Evidence

There are at least three applications of geochemical and petrographic data to the assessement of temporal relationships:

1) Two rocks that are chemically and petrographically the same are more likely to be stratigraphic correlatives than two rocks that are chemically different. It is tempting, for example, particularly in the absence of any hard evidence to the contrary, to suggest that the various ENA high-Ti rocks are chemically and petrographically the same (Table 2 & 4) because they were generated together in the same environment and were emplaced together when prevailing tectonic events permitted. The remarkable uniformity of the high-Ti basalts is observed at three levels:

(a) The chemical composition (Table 2) of the high-Ti flows maintains a particularly uniform composition from the Culpeper Basin of Virginia (Mt. Zion Church) to the Hartford Basin of Connecticut (Talcott). The North Mt. Basalt of Nova Scotia still qualfies as a high-Ti type (Papezik, 1981; and Puffer and Hurtubise, 1982) but it is slightly more felsic than the others.

(b) Trace element ratios plotted along the eastern edge of North America (Fig. 8) do not indicate any particular north-south trend. The trace elements chosen for Fig. 8 were recommended by Anderson (1981) as sensative to hot-spot petrogenesis and should increase as any hot-spot is approached. The absence of any north-south trend does not support the placement of early Jurassic hot-spots located in the Carolinas or in New England.

(c) The petrology and mineralogy of the high-Ti basalts are also surprisingly uniform. Each of the ENA high-Ti flows are finer grained, more porphyritic, and thinner than the overlying high-Fe flows. The phenocryst compositions are particularly uniform and distinctly different than those of the overlying flows (Table 4). Typical augite phenocrysts are restricted to a narrow range close to $Ca_{35}Mg_{54}Fe_{11}$ and plagioclase although somewhat more wide ranging averages about An_{65}.

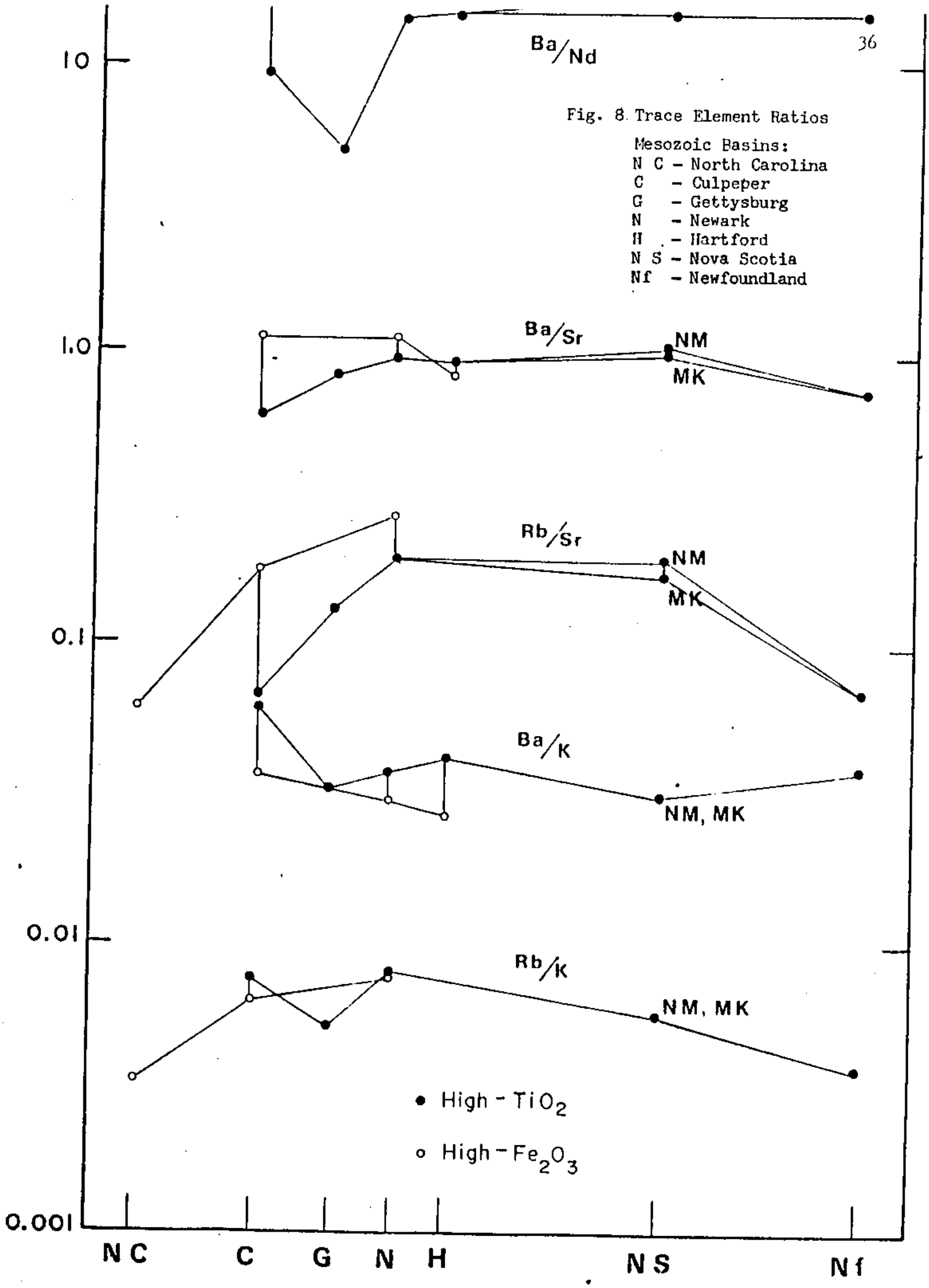


Fig. 8 Trace Element Ratios

Mesozoic Basins:
N C – North Carolina
C – Culpeper
G – Gettysburg
N – Newark
H – Hartford
N S – Nova Scotia
Nf – Newfoundland

Table 4

Augite and Plagioclase Compositions of Jurassic E.N.A. Basalts

Basin	Culpeper		Gettysburg		Newark		Hartford		Nova Scotia	
	Augite	Plag.	Augite	Plag.	Augite	Plag.	Augite	Plag.	Augite	Plag.
Third Flow	–		–		Hook Mountain		Hampden		–	
SiO_2					52.49	52.03	52.70	52.12		
TiO_2					0.44	0.05	0.39	0.03		
Al_2O_3					3.64	30.40	2.09	30.73		
FeO					11.63	0.75	13.17	0.60		
MgO					17.57	0.11	16.15	0.09		
CaO					14.73	13.21	15.79	11.86		
Na_2O					0.22	3.50	0.17	3.40		
Total					100.71	100.05	100.45	98.84		
Second Flow	Sanders		–		Preakness		Holyoke		–	
SiO_2	52.81	51.35			52.45	54.59	52.62	53.43		
TiO_2	0.41	0.00			0.32	0.00	0.22	0.02		
Al_2O_3	1.71	30.10			1.69	29.14	1.40	30.74		
FeO	12.54	0.68			12.65	0.59	12.07	0.64		
MgO	14.91	0.08			15.76	0.14	16.62	0.10		
CaO	16.75	12.83			16.85	11.02	15.71	12.31		
Na_2O	0.55	3.61			0.17	4.54	0.11	3.67		
Total	99.69	98.66			99.89	100.02	98.75	100.90		
First Flow	Mt. Zion Church		Jacksonwald		Orange Mountain		Talcott		North Mountain	
SiO_2	52.27	52.35	53.30	52.45	52.63	53.28	53.53	52.59	54.00	52.80
TiO_2	0.38	0.00	0.25	0.03	0.29	0.05	0.26	0.13	0.27	0.02
Al_2O_3	3.14	30.95	3.07	30.66	3.36	29.23	3.17	30.01	2.01	30.45
FeO	6.89	0.48	6.05	0.71	6.67	1.06	6.71	0.80	8.23	0.89
MgO	19.43	0.20	19.04	0.16	19.34	0.22	19.22	0.11	19.36	0.07
CaO	16.17	12.40	18.03	13.22	17.13	11.85	16.87	11.65	16.61	10.84
Na_2O	0.17	3.10	0.20	3.14	0.22	3.66	0.26	3.59	0.19	3.48
Total	98.45	99.48	100.22	100.37	99.65	99.35	100.02	98.97	100.86	98.55

2) The chemistry of rocks that are part of a unique or irreversible sequence of geochemical events should be consistent with those events. Tholeiitic trend fractionation, for example, predictably generates increasingly felsic rocks. The high-Fe flows overly and are slightly younger than the high-Ti flows because they are the products of high-Ti magma fractionation. High-Ti flows do not occur in any of the stratigraphic sections above high-Fe flows and high-Ti dikes have not been found that cross-cut high-Fe dikes. Minor local reversals, however, may have occurred due to rapid fractionation in shallow magma chambers resulting in some high-Fe magma emplaced before the final high-Ti activity. The high-Fe magma developed at shallow depth, however, is enriched in Cu and is distinguished from the more common Cu depleated rocks.

3) Magma compositions are influenced by several variables that are unlikely to be held constant throughout geologic time. Unchanging chemical compositions therefore, require special sets of conditions. If high-Ti magma was extruding intermittently throughout the 35 my range between 210 and 175 ma as suggested by the widest ranging radiometric data, one plausable explanation is that "steady state" conditions as described by O'Hara (1977) were responsible for magma development. O'Hara's (1977) steady-state conditions refer to open-system fractional crystallization in a magma chamber fed with batches of parental magma which mix with residual magma already there. The characteristics of steady state magma are: uniform composition, large differences in the composition of lava and parent, and strong control of composition by low pressure phase equilibria. These characteristics are met by high-Ti tholeiites, assuming a high-LIL parent magma, but some important prerequisites are missing.

O'Hara (1977) specified that steady-state magma should be uniform in composition "provided the thermal insulation of the magma chamber, the rate of supply, the amount and nature of assimilation, and the composition of the parental magma are maintained constant." Eastern North America was undergoing rapid and permanent tectonic change during the early Jurassic including crustal thinning (decreased thermal insulation), rifting, and extention (increased rate of supply). A much narrower time range (perhaps 190 to 195 my as indicated by the large majority of the radiometric data) may, therefore, more appropriately apply to the high-Ti magma (Fig. 9). Subsequent tectonic changes may be responsible for the development of the low-Ti magma which post-dates the high-Ti type by several million years. O'Harra (1977) predicts that if insulation and supply rates change over time then large variations in incompatable elements with small variations in compatable elements should occur. The incompatable element content is clearly the chief difference in high-Ti vs. low-Ti magma (Table 2).

The time span represented by the high-Ti magmatic event, however, was apparently of sufficient duration to account for the extrusion of multiple flows and the deposition of considerable thicknesses of non-marine sediments. The high-Ti (Mt. Zion Church) basalt of the Culpeper Basin of Virginia is overlain by a sequence of at least five high-Fe basalts (the Hickory Grove, Sanders 1,2, and 3 and the Millbrook Basalts) that are each interbedded by layers of early Jurassic Sediment. The compositions of the high-Fe basalts remained uniform throughout this sequence but rapid deposition of sediment is indicated by the lithologic and paleontological evidence (Olsen and others, 1982). The steady-state conditions of O'Hara (1977) probably pertains to the high-Fe basalt of the Culpeper Basin. High-Fe magma was probably repeatedly displaced from a chamber located near the base of the crust (perhaps 8kb, Philpotts and Reichenbach, 1983) by renewed supplies of parental magma (probably high-LIL).

Period	Stage	Georgia Alabama	Carolinas	Culpeper Basin	Gettysburg Basin	Newark Basin	Hartford Basin	Deerfield Basin	Fundy Basin	Newfoundland
Juras.	Sinemur.	low-Ti		low-Ti ?	low-Ti					
	Hettang.	low-LIL	high-Fe high-Ti high-LIL low-LIL	high-Fe high-Ti low-LIL ?	high-Fe high-Ti low-LIL	Hook Mt. high-Fe high-Ti	Hampden high-Fe high-Ti	high-Ti	high-Ti	high-Ti
Trias.	Rhaetian									

Fig. 9 Correlation chart for early Jurassic ENA tholeiites based on summary of all available stratigraphic data as of July 1984.

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CHAPTER THREE

VOLCANIC ROCKS OF THE NEWARK BASIN

by

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The Watchung Basalts are clearly the most widespread basalt units of the Newark Basin but some other basalt units are also exposed within the basin. From north to south the basalt or basalt sets are known as the: 1) Ladentown Basalt, 2) Union Hill Basalt, 3) Watchung Basalts, 4) New Germantown Basalts, and 5) Sand Brook Basalts (Fig. 10).

1) The Ladentown Basalt

The Ladentown Basalt occurs at the northern end of the Newark Basin in Rockland County, New York (Fig.10). The Ladentown flows form most of a northeast-trending escarpment which projects 30 m above the adjacent Mohawk River Valley occupied by the Ramapo Fault. The escarpment is underlain by the coarse conglomerates of the Passaic Formation.

The Ladentown Basalt is 61 to 91 m thick near the western exposed edge of the unit but drilling data indicates that it extends southwest at depth to the Ramapo Fault where it is more than 137 m thick (Ratcliffe, 1980). The basalt occupies a shallow basin formed by the intersections of a NE trending syncline with a NW trending syncline (Ratcliffe, 1980).

The basalt displays pahoehoe flow structure, and pillow structure. Three vesicular layers indicate that there were at least three separate flows. Kummel (1900) proposed that the Ladentown lavas resulted from fissure eruption of Palisades magma. The interconnection with the Palisades is confirmed by aeromagnetic, gravity and drilling data (Frimpter, 1967; Koutsomitis, 1980; Kodama and others, 1981) and geochemical data (Puffer and others, 1982).

The Ladentown Basalt chemically resembles other high-Fe ENA basalts but there are some distinct petrographic and chemical differences. The glass component of the Ladentown Basalt greatly exceeds typical values for high-Fe extrusives and opaque oxides, in contrast to typical high-Fe extrusives, occur as quench dendrites in the glass comprising an average of 4 volume percent of the basalt. The Cu content (150 ppm) and TiO_2 content (1.26 percent) of the Ladentown Basalt is higher than the Cu and Ti content of typical high-Fe basalt (Table 2). The chemical

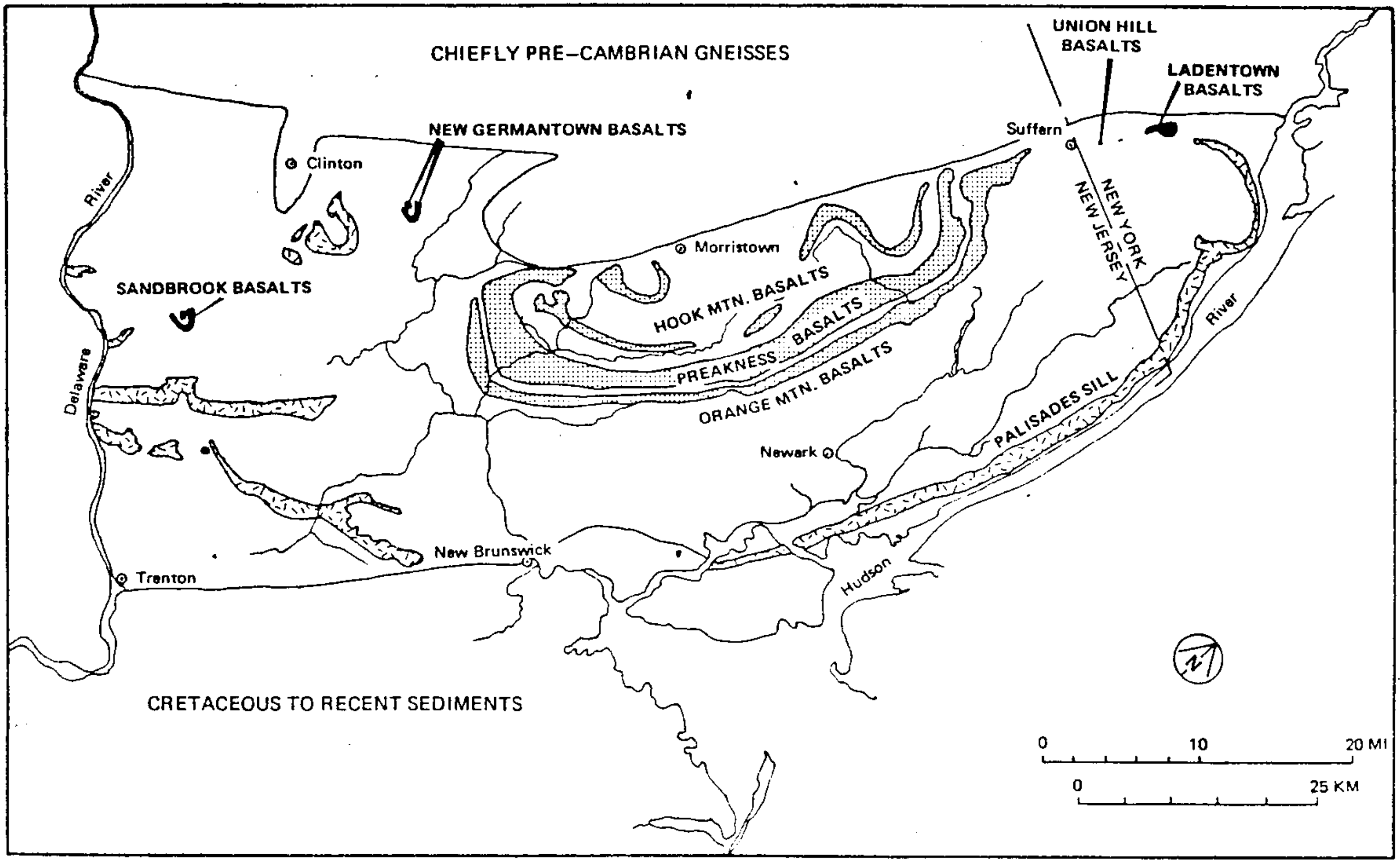


Figure 10Igneous Rocks of the Newark Basin in New Jersey and New York (after Mason, 1960) composition of the Ladentown Basalt is, however, virtually equilivent to the composition of the interior zone of the Palisades Sill (Walker, 1969). Puffer and others (1982) therefore, suggest a sequence of events that involves three magmas:

1) A high-Ti magma intruded as the first pulse of the composite Palisades Sill and extruded simultaneously as the Orange Mountain (First Watchung) basalt onto the late Triassic Passaic Formation.

2) A second pulse of high-Ti magma intruded onto the interior of the Palisades Sill. Shallow in-situ fractionation within the Palisades Sill developed a Cu enriched magma some of which extruded onto the Passaic Formation as the Ladentown Basalt.

3) A Cu depleated magma slowly fractionated out of high-Ti magma at a deep level then extruded onto the early Jurassic Feltville Formation as the Preakness (Second Watchung) basalt. The Cu depletion of the third magma was probably due to gravitational separation of an immisable sulfide liquid phase.

2) The Union Hill Basalt

The Union Hill Basalt is located within the village of Suffern, New York 6 km south of the Ladentown Basalt of Rockland County. The basalt forms a prominant circular knob with 31 m of relief. Kummel (1900) considered the Union Hill rock a diabase based on its coarse grained size but pillow structure, pahoehoe texture, and an absence of contact metamorphism along its contact confirms an extrusive origin.

Although the Union Hill Basalt is exposed only 6 km south of the Ladentown Basalt it chemically and petrographically more closely resembles typical high-Fe basalt. The Preakness Basalt, a typical high-Fe type, is exposed about 15 km south of Union Hill and may have been displaced from Union Hill by the Ramapo Fault system.

3) The Watchung Basalts

The basalts of the Newark Basin are exposed as a set of three N-E trending ridges that dip to the west at about 15°. The three basalt units were known as the First, Second, and Third Watchungs until they were renamed by Olsen (1980) as the Orange Mountain, Preakness, and Hook Mountain Basalts respectively. The geochemistry of the flows is described by Puffer and Lechler (1980) and Puffer and others (1981).

The Orange Mountain Basalt (a high-Ti type) averages 183 m thick (Faust, 1978) and consists of at least two flows. The upper flow is characteristically fine grained pillowed, and amygduloidal. The lower flow is also fine grained but displays well developed Tomkeieff structures (lower colonnade, entablature, and upper colonnade).

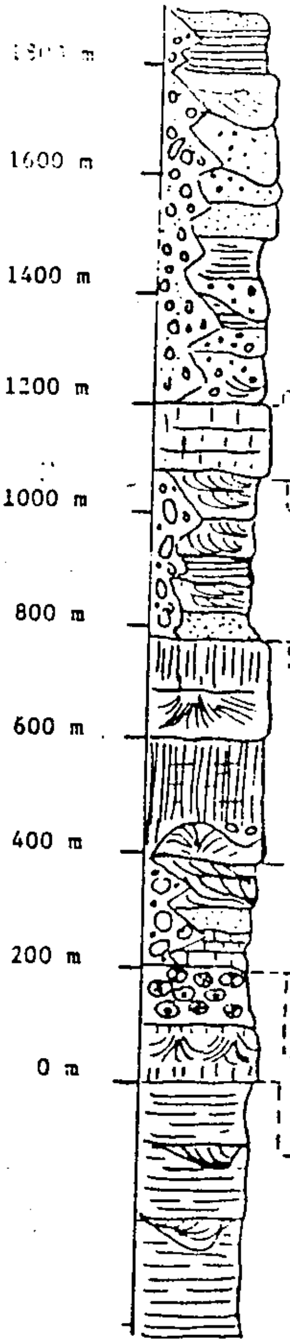
The Preakness Basalt (a high-Fe type) extruded onto 170 m of Feltville Formation. The Preakness averages 229 m in thickness (Faust, 1978) and consists of three flows. The basalt is unusually coarse grained and also displays well developed Tomkeieff structures.

The Hook Mountain Basalt (chemically equilivent to the Hampden Basalt of the Hartford Basin) extruded onto 340 m of Towaco Formation and flowed toward the southwest unlike the north-east flow direction of the underlying flows (Manspeizer, 1969). The basalt averages 91 m in thickness (Faust, 1978) and consists of at least two amygduloidal and deeply altered flows.

One deeply altered exposure of basalt near the Ramapo Fault may represent a flow younger than the Hook Mountain but my preliminary chemical analyses suggest that it is probably a displaced fault slice of Hook Mountain or possibly Preakness basalt.

4 and 5) New Germantown and Sand Brook Basalts

The New Germantown and Sand Brook Basalts are located in Hunterdon County, 10 km and 29 km southwest of the termination of the Orange Mountain Basalt. The New Germantown and Samd Brook Basalts are both folded into northwest plunging synclines and are each composed of two distinct basalt units. The lower unit of each group forms an arcuate ridge which encloses a smaller upper unit separated by Jurassic redbeds. The lower flow units are both typical high-Ti type basalt whereas the upper flows are distinctly coarser grained and are high-Fe type basalts (Geiger, in prep.).



BOONTON FORMATION: alternating sequences of red-brown conglomerate with cyclical sheets of fining-upward structures and poorly sorted graded beds, and black-to-gray thinly-laminated dolomitic siltstones with fish fossils.

HOOK MOUNTAIN BASALT: fine-to-moderately coarse-grained, abundantly vesicular and amygdaloidal basalt in 3 distinct flow units with poorly developed columnar structures.

TOWACO FORMATION: interbeds of laterally persistent finely-laminated gray calcareous siltstone and black-to-olive green shale, and cross-bedded gray sandstone and conglomerate in fining-upward sequences.

PREAKNESS BASALT: moderately coarse-grained with little glass; closely spaced, non-curvicolumnar joints of the entablature; and tumuli, pillow lava, palagonite forset bedding; 3 clearly-defined flow units.

FELTVILLE FORMATION: buff-to-gray feldspathic sandstone with thin stringers of coal; underlain by gray-to-black limestone (biopelsparite and intrasparite) with plant fragments and fish fossils; interfingers westward with massively-bedded dark-gray border conglomerate.

ORANGE MOUNTAIN BASALT: fine-grained with typical Tomkeieff structures; pillow lava, flow lobes and fissure eruptions.

PASSAIC FORMATION: red-brown sandstones in fining-upward sequences with cut-and-fill channels, calcrete, worm burrows, and dinosaur footprints; sequence becomes coarser-grained to the northeast.

Fig. 11 Schematic composite column for the latest Triassic-Jurassic rocks, Newark Basin. (from Manspeizer, 1980).

The New Germantown Basalts form a horseshoe shaped ridge truncated by a normal fault that is part of the Ramapo border fault system along the northwest margin of the Newark Basin. The lower flow unit is about 120 m thick and is separated from the upper flow unit by about 600 m of red-beds. Erosion has removed all but the lower 75 m of the upper flow unit (Cornet, 1977).

The Flemington Fault truncates the Sand Brook Basalts. The lower unit of the Sand Brook is 137 m thick (Bascom, 1909) and is separated from the upper flow unit by about 300 m of red-beds. The upper flow unit forms an isolated knob along the synclinal axis that has been eroded to a thickness of about 75 m (Cornet, 1977).

Figure 12. Proposed time-stratigraphic correlation chart. *Stratigraphic nomenclature proposed by Olsen (1981). The time-stratigraphic correlation between the Talcott Basalt and First Watchung Basalt is based on strong geochemical and paleontological evidence; the stratigraphic positions of the Holyoke Basalt and Second Watchung Basalt are approximations based on geochemical, depositional rate, and petrologic (cooling rate) evidence; the time stratigraphic correlation between the Hampden and Third Watchung is based on geochemical and depositional rate evidence.

Figure 12 appears on the following frame.

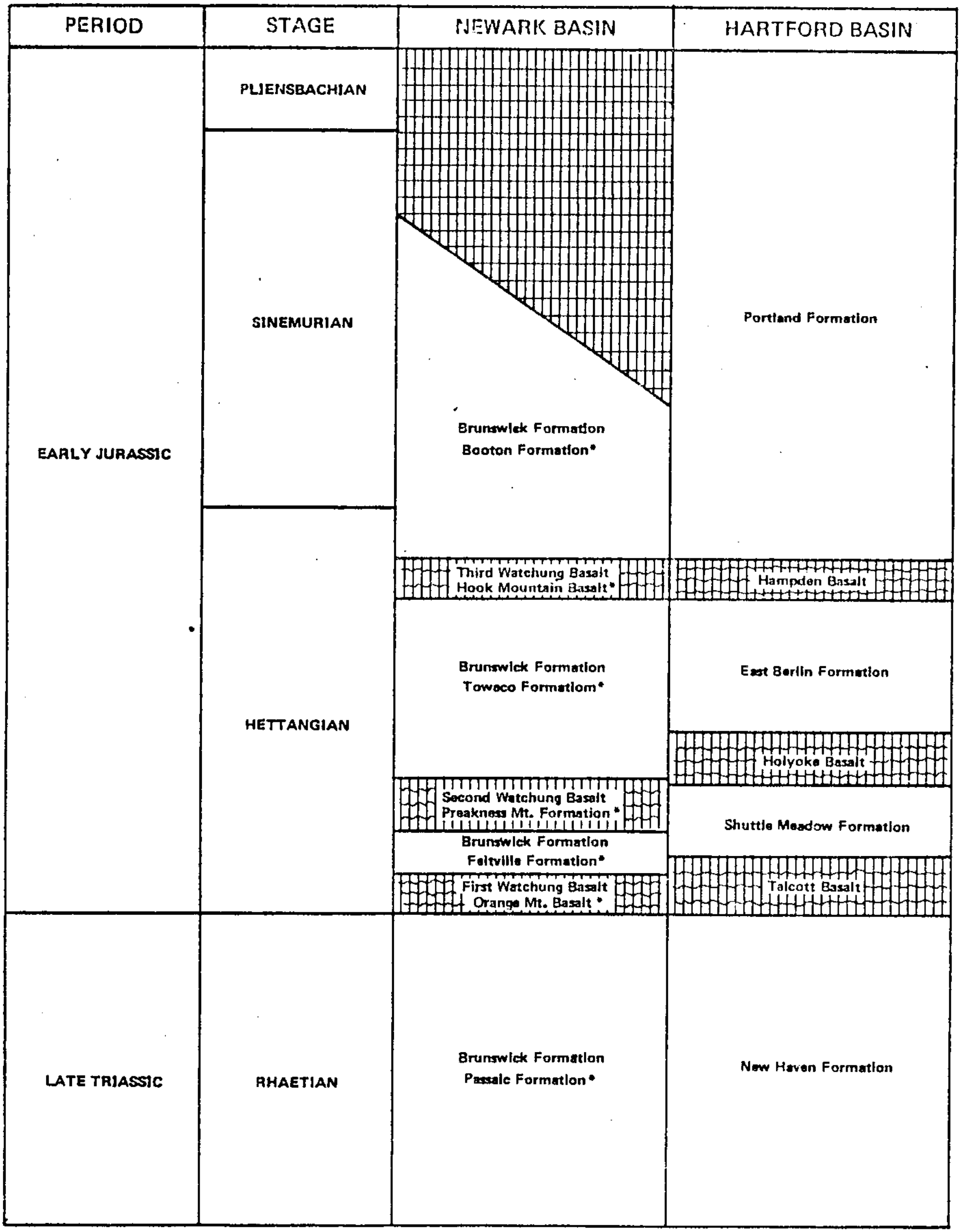


Fig. 12 after Puffer and others (1981).

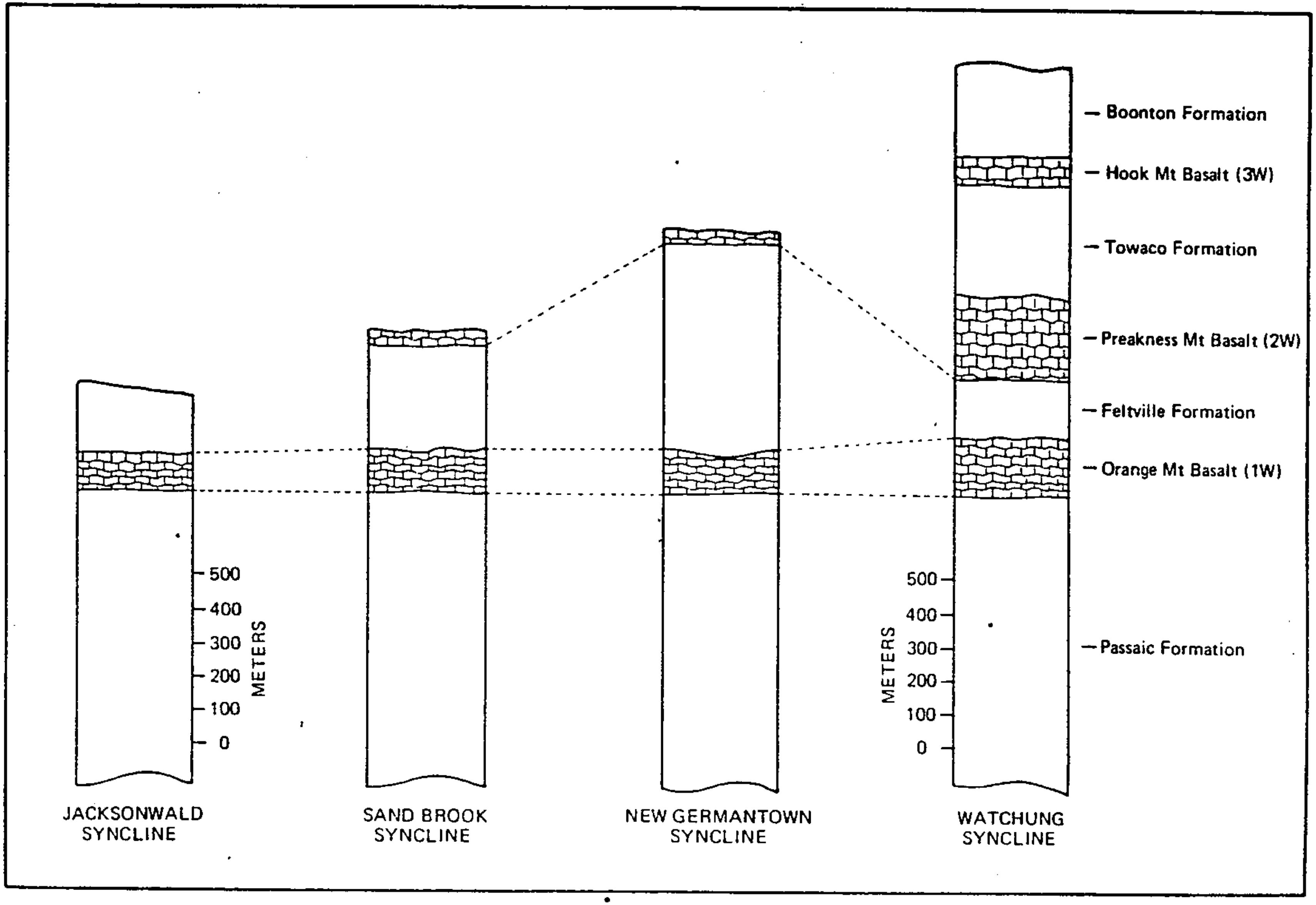


Fig. 13 Correlation of Jurassic basalts of the Gettysburg and Newark basins (after Cornet, 1977; and Geiger, 1984).

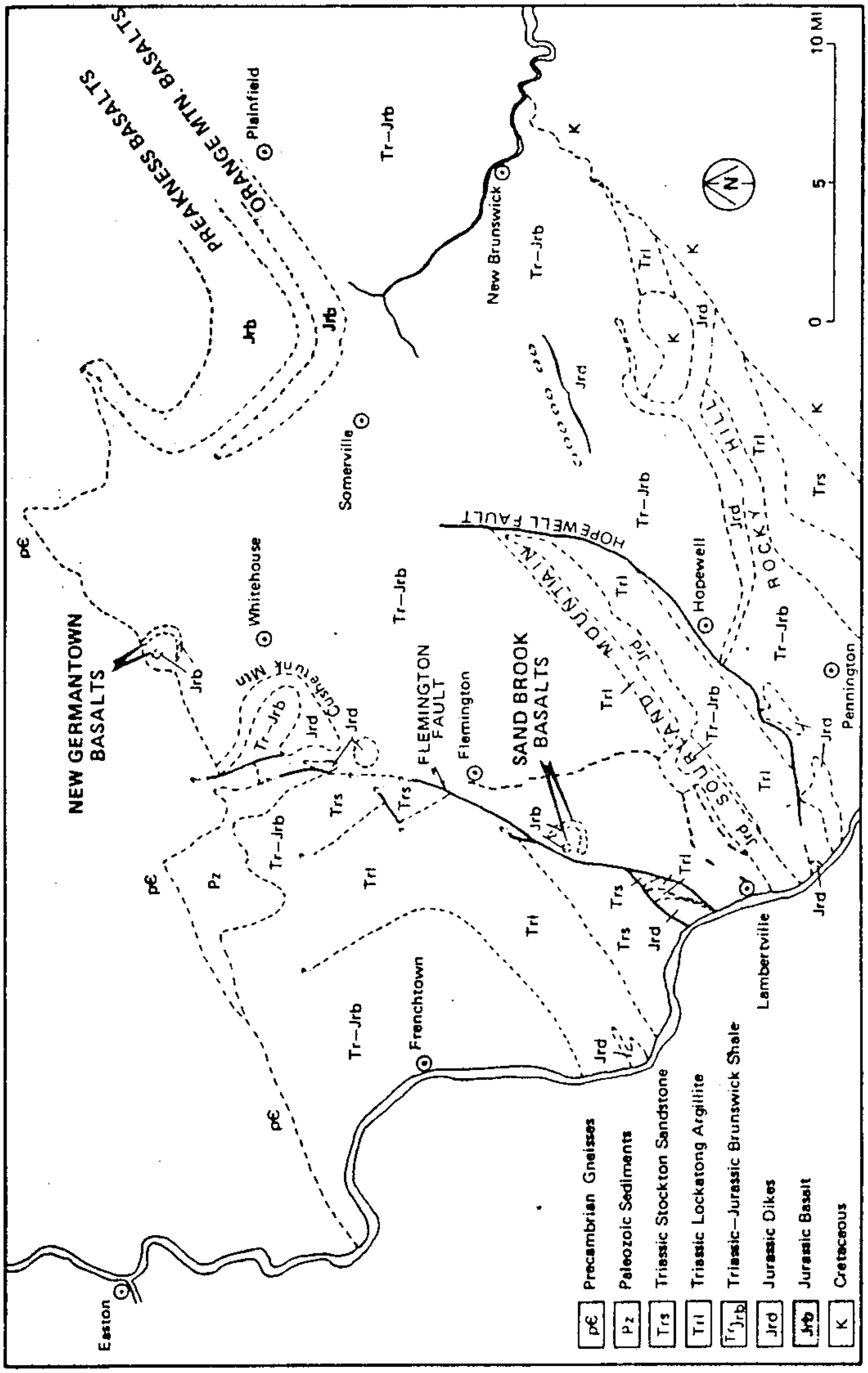


Figure14 The New Germantown and Sand Brook Basalts (after Kummel, 1897)

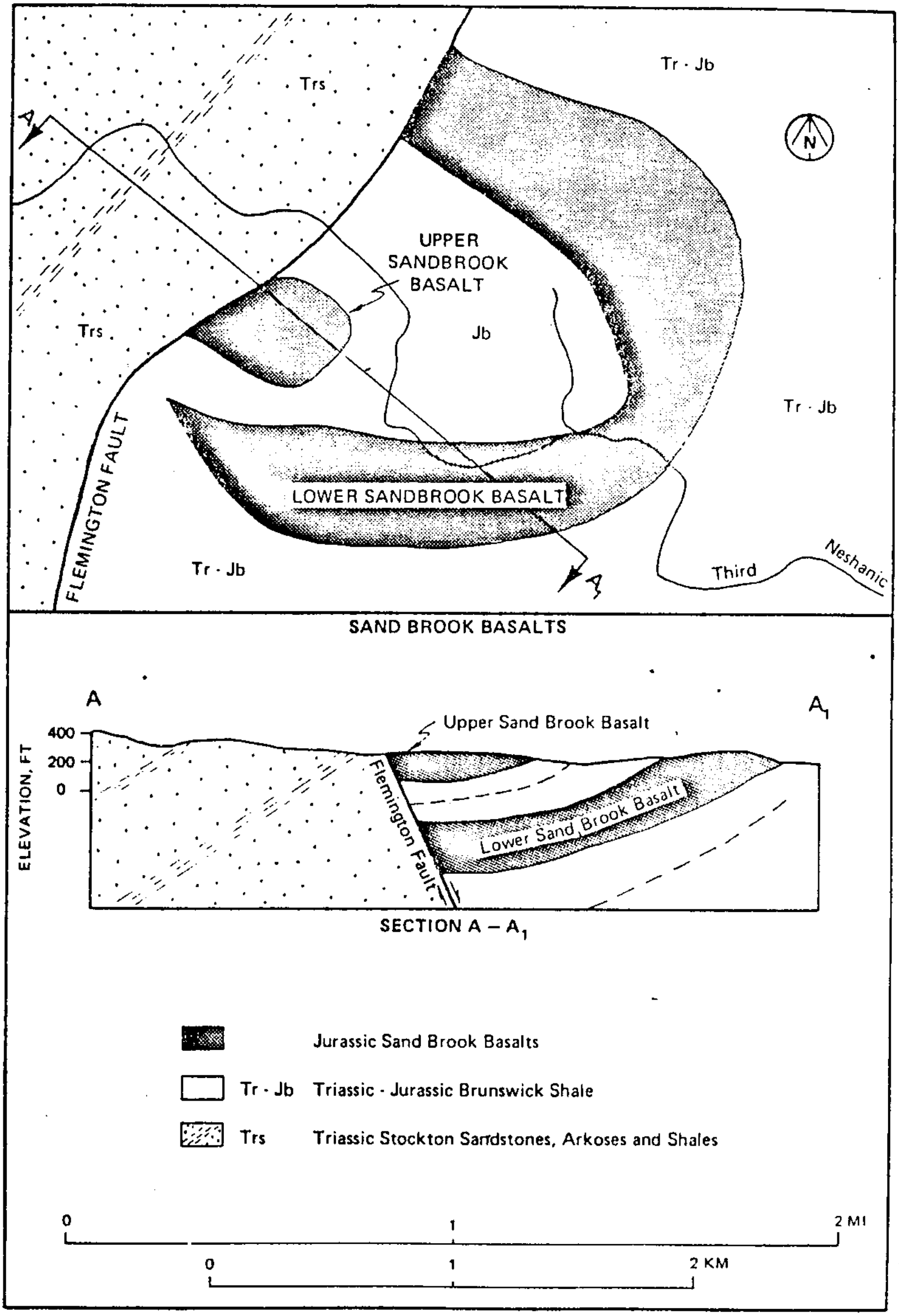


Fig. 15 after Geiger, in prep.

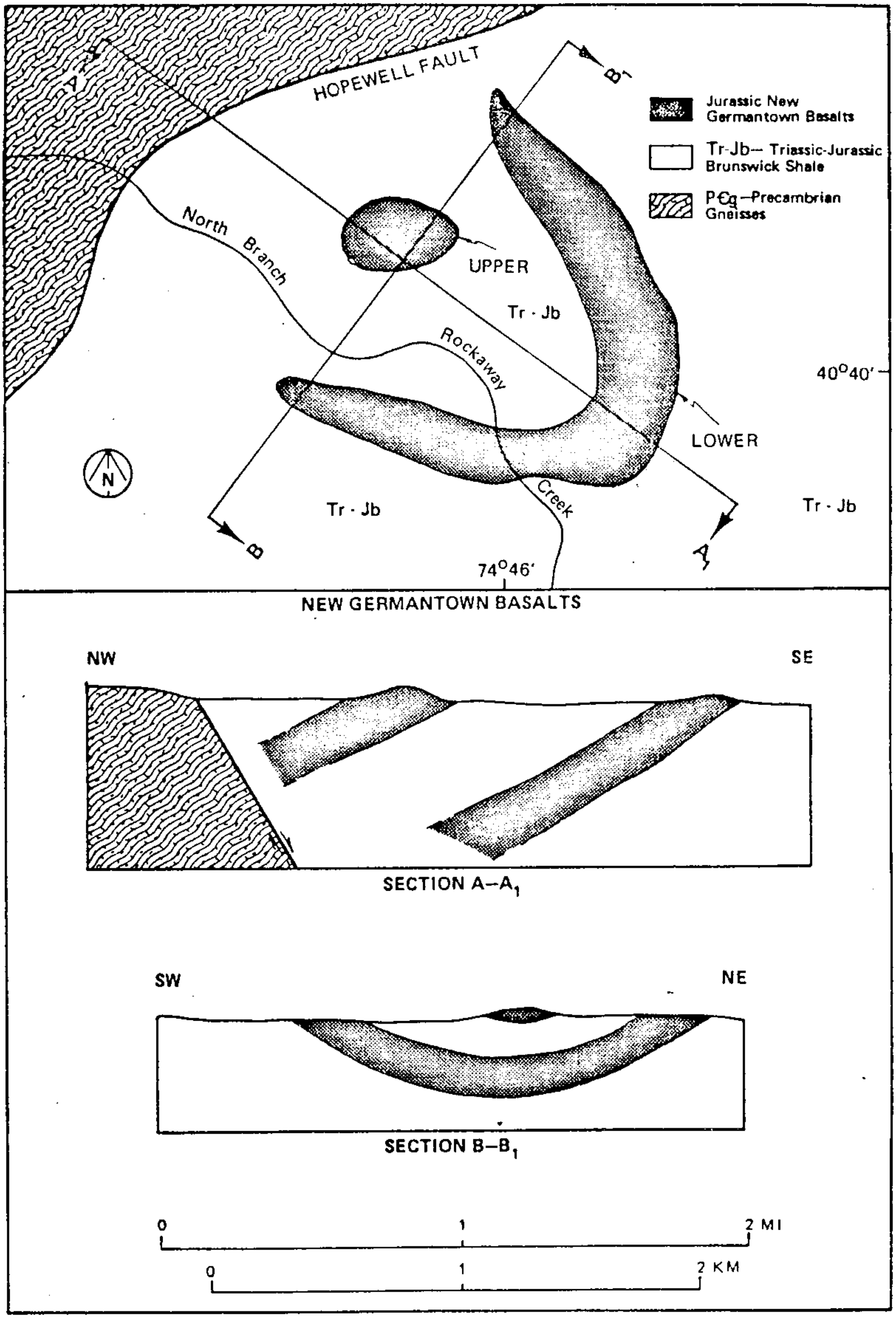


Fig. 16 after Geiger, in prep.

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CHAPTER FOUR

THE PALISADES DIABASE INTRUSIVE*

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* Modified after Puffer, Geiger, and Caamano (1982) "Mesozoic igneous rocks of Rockland County, New York:" Northeastern Geology, v. 4, p. 121-130.

INTRODUCTION

The Palisades diabase magma intruded approximately parallel to the bedding planes of the Mesozoic sedimentary rocks of northern New Jersey (fig. 17) but is discordant throughout at least some of the segment located north of Nyack, New York (Darton, 1890; Lowe, 1959; Koutsomitis, 1980 ms). The development of any cryptic layering such as that observed along the New Jersey portion of the Palisades, may be related to the structural attitude of the implacement of the diabase.

Since the Palisades intrusive cannot be accurately referred to as a sill throughout its entire length, the northern portion of the Palisades diabase intrusive located north of Nyack, New York (fig. 17) will be referred to as the "northern Palisades", whereas the portion located south of Nyack will be referred to as the "southern Palisades" throughout the remainder of this paper. The term New City Park Dike follows the usage of Dames and Moore (1977) who use the term to refer to the north-south-striking diabase intrusive that branches off of the Palisades intrusive at Nyack, New York (fig.17).

GEOCHRONOLOGY

Erickson and Kulp (1961) report a 190 ± 5 m.y. radiometrically determined (K-Ar) date for the intrusive which agrees with the 186 to 195 ± 5 m.y. K-Ar range of Lambert (1971), the 186 m.y. and 192 m.y. Ar^{40}-Ar^{39} dates of Dallmeyer (1975) and the 196 to 194 m.y. paleomagnetically determined range of deBoer (1968). Most of the samples dated by the above authors were taken from the lower chill zone of the Palisades intrusive. Walker (1969), however, has determined that the Palisades is a composite intrusive consisting of two magma pulses. The diabase magma emplaced by the first pulse had probably not completely crystallized before the second pulse penetrated rather passively through its interior. A still-liquid interior would presumably afford the most penetrable avenue of intrusion. The second magma pulse was therefore probably implaced quickly after the first pulse, well within the range of age data determined for the first pulse. Walker (1969) also found that the second magma phase was more fractionated than the first phase.

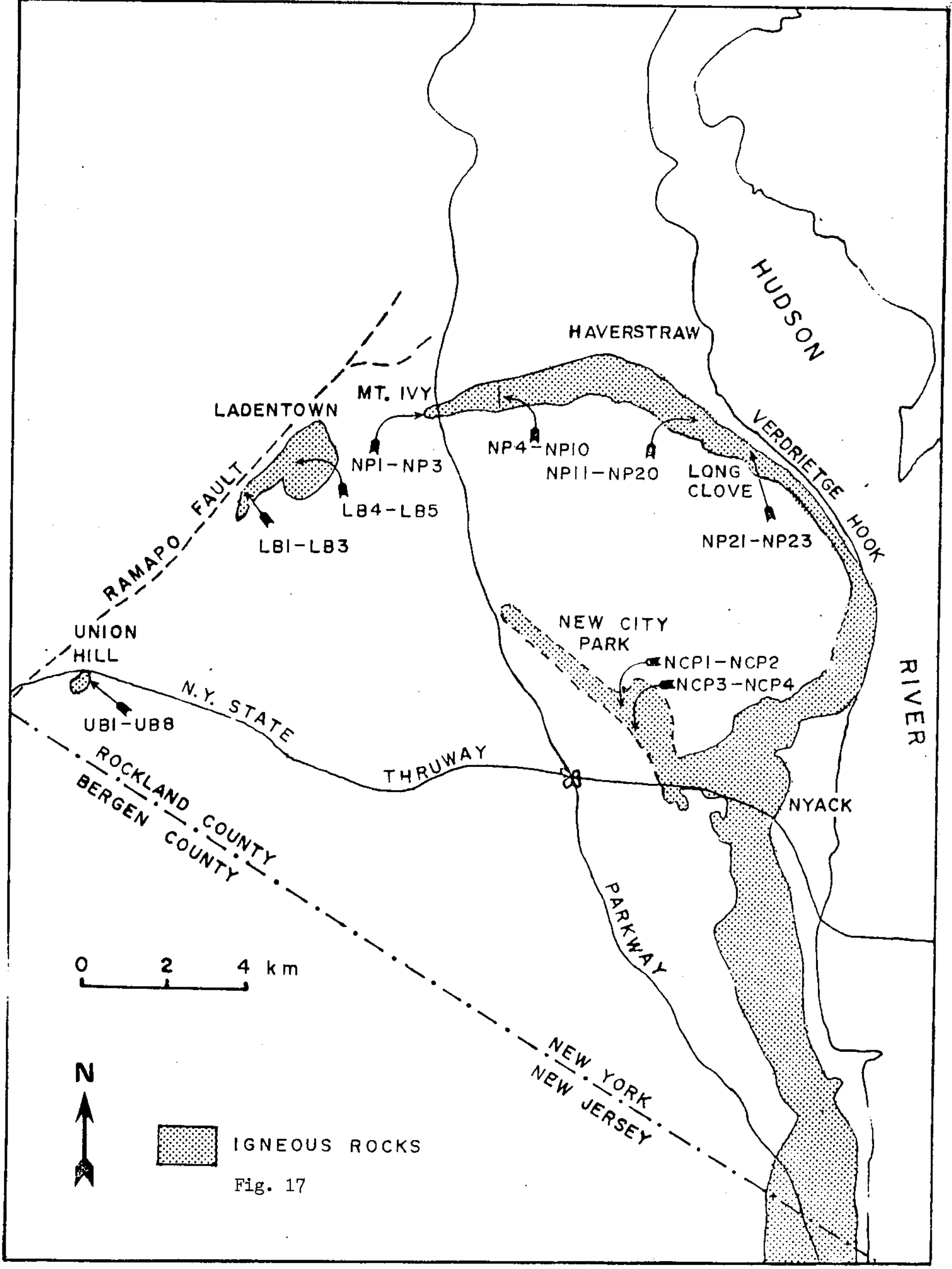
HUDSON
RIVER
HAVERSTRAW
MT. IVY
LADENTOWN
NP1-NP3
NP4-NP10
NP11-NP20
LONG CLOVE
VERDRIETGE HOOK
NP21-NP23
LB4-LB5
LB1-LB3
RAMAPO FAULT
UNION HILL
UB1-UB8
N.Y. STATE THRUWAY
NEW CITY PARK
NCP1-NCP2
NCP3-NCP4
NYACK
PARKWAY
ROCKLAND COUNTY
BERGEN COUNTY
NEW YORK
NEW JERSEY
0 2 4 km
N
▨ IGNEOUS ROCKS

Fig. 17

The Northern Palisades and New City Park Dike

At Nyack, the northern Palisades is exposed about 0.6 km west of the Hudson River. North of Nyack the intrusive strikes north-east toward the Hudson River along Verdrietege Hook and then curves toward the west from Haverstraw to Mt. Ivy, thus forming a sickle-shaped arc (fig.17). Within the sickle-shaped arc the dip of the sediments intruded by the northern Palisades varies from 10 to 15° west-northwest near its southern end to 45° south to southwest along its northern portion, (Kummel, 1900; Sanders, 1974).

Although there is little doubt that the southern Palisades is a sill, there is good evidence (Darton, 1890; Lowe, 1959; Koutsomitis, 1980 ms) that at least part of the sickle-shaped portion of the intrusive is a S-W dipping dike. Darton (1890) based his interpretation on field observations made along the inner side of the northern portion of the sickle (upper contact) from Long Clove to Mt. Ivy and in particular, at a contact exposed at the south portal of the West Shore Railroad tunnel located southeast of Long Clove (fig.17). Lowe (1959) interpreted the northern Palisades as a dike on the basis of drill-hole data and columnar structures at quarries near Long Clove. Koutsomitis' (1980 ms) structural interpretation based on geophysical gravity modeling studies suggests that the Palisades is approximately concordant south of Verdreitege Hook, but northward becomes increasingly discordant particularly along its westward extension from Long Clove to Mt. Ivy. Koutsomitis (1980 ms) also demonstrated with gravity models that the New City Park intrusive (approximately 185 m thick, fig.17) is a dike that dips approximately 60 degrees to the southwest.

Sanders' (1974) structural interpretation agrees, in general, with that of Koutsomitis (1980 ms). Sanders (1964), however, would confine the appreciably discordant portion of the Palisades intrusive to its occurrence west of Haverstraw. Sanders (1974) suggested that the apparent discordance southeast of Haverstraw is actually an outcrop pattern due to post-intrusion folding. An anticline oriented transverse to the strike of the Newark Basin (the Danbury anticline) is largely responsible for the local NW-SE strike and SW dip of the Palisades located SE of

Haverstraw (Sanders, 1974).

PETROGRAPHY

Northern Palisades

The modes of the Northern Palisades samples (table 5) closely resemble Walker's middle fractionation stage of Palisades magma as represented by Walker's samples W-R-60 and W-N-60. The rock is holocrystalline and sub ophitic, composed of plagioclase and clinopyroxene (augite-pigeonite), and minor amounts of opaque oxides, hornblende, biotite, sphene and apatite. The plagioclase laths range in length from 0.17mm to 1.40 mm with a few as long as 2mm. The plagioclase composition ranges from An_{50} to An_{60}, averaging An_{54}; it is slightly more sodic than the plagioclase in Walker's (1969) samples W-N-60 and W-R-60, located approximately 112.3 and 175 m respectively above the base of the 308 m thick sill.

Clinopyroxene grains are subhedral and range in length from 0.5 mm to 4 mm. Bladed crystals are common and show the herringbone structure typical of augite. Pigeonite has been identified in these samples based on its low positive axial angle and is commonly intergrown with augite.

New City Park Dike

The modal composition of the New City Park Dike (table 5) compares well with that of Walker's (1969) lower chill zone, (W-889LC60), both have a plagoioclase to pyroxene ratio of less than one. Samples NCP1 and NCP2 were collected at the chill zone where the diabase has crystallized into a very fine-grained homogeneous rock, which is holocrystalline and composed of clinopyroxene (augite and pigeonite), plagioclase (An_{60-65} and minor amounts of orthopyroxene and biotite. Microphenocrysts of augite that range from 0.40 to 1.5 mm in length and plagioclase laths that range from 0.1 to 0.4 mm in length comprise approximately 4 percent of the rock.

Samples NCP3 and NCP4 are from the interior of the New City Park Dike. The modal composition of these samples correlates fairly well with Walker's (1969)

TABLE . MODAL ANALYSES

	Ladentown	Union Hill	N. Pal.	N.C.P.D.	Orange Mt.**	Preakness**
No. of Analyses:	5	5	19	4	8	11
Plagioclase	37	53	51	44	35	54
Pyroxene	36	39	41	48	35	40
Glass	20	tr	-	-	25	5
Opaque oxides	4	5	5	4	4	3
Olivine	-	-	-	-	3	-
Quartz	tr	tr	1	tr	tr	tr
Other*	3	3	2	2	tr	tr
Plag. Comp.	An_{60}	An_{56}	An_{55}	An_{63}	An_{66}	An_{55}

* Other mineral components include orthoclase, biotite, sphene, chlotite and apatite.

** Orange Mt. and Preakness modal data after Puffer and Lechler (1980).

sample W-889LC-60. The New City Park Dike samples NCP3 and NCP4 are composed of clinopyroxene (augite and pigeonite) and plagioclase (An_{59-62}) with trace amounts of orthopyroxene and biotite. Pyroxene grains range from 0.4 to 1.5 mm in length but are much wider than the pyroxene found in samples NCP1 and NCP2. Plagioclase laths range from 0.1 to 0.7 mm in length.

GEOCHEMISTRY

Chemical analyses of northern Palisades and New City Park Dike samples (Table 6) plotted onto MgO variation diagrams (fig. 18) indicate two separate and distinct fields. The chill zone of the southern Palisades at Edgewater, New Jersey (Walker, 1969) plots close to the field represented by the New City Park Dike. Walker (1969) interprets the chill zone as representative of the first magma phase of the composite Palisades diabase intrusive. The exposed portion of the New City Park Dike was probably also emplaced by the first pulse of Palisades magma.

The composition of the New City Park Dike magma is slightly more fractionated than the Orange Mountain (First Watchung) Basalt of New Jersey (table 6) but there is a high degree of composition similarity. The first magma pulse of the composite Palisades intrusive, the New City Park Dike, and the Orange Mountain Basalt may each be classified as belonging to the high-Ti type of eastern North American tholeiitic magma (Weigand and Ragland, 1970).

Walker's (1969) chemical analyses of drill core samples from Haverstraw, New York indicate that both of the two magma phases are present within the northern Palisades and that the contact is somewhere within a transition zone located 33 to 50 m above the base of the intrusive. The basal 33 m of the dike at Haverstraw that represents the first magma phase is quite homogeneous and closely resembles the composition of the chill zone of the Edgewater, New Jersey section of the sill. The composition of the diabase through the 17-m-thick transition zone and up to the 144 m level above the base of the dike undergoes an increase in Fe, Ti,

Table 6

Igneous rocks of the northern Newark Basin

	N. Pal,	N.C.P.D.	Laden.	Union H.	Pal,CZ.*	Pal.Int.*	Hi.TiO_2*	Orange M.*	Preak.*
No. of Analyses:	19	4	5	5	7	6	20	8	11
SiO_2	52.39	52.51	51.69	51.43	51.98	51.70	52.10	50.66	52.24
TiO_2	1.24	1.02	1.26	1.25	1.22	1.58	1.12	1.07	1.20
Al_2O_3	14.55	14.48	14.24	14.26	14.48	14.08	14.22	14.08	13.70
Fe_2O_3	1.68	1.62	2.35	2.22	1.37	2.51	11.65	2.79	2.15
FeO	9.86	9.15	9.28	9.30	8.92	9.18	-	7.62	9.43
MgO	5.30	7.14	6.23	6.44	7.59	6.63	7.41	8.17	6.62
CaO	9.65	9.80	9.72	9.89	10.33	9.86	10.66	11.20	9.83
Na_2O	2.22	2.04	2.63	2.47	2.04	2.49	2.12	2.11	2.64
K_2O	0.62	0.63	0.61	0.54	0.84	0.82	0.66	0.58	0.65
H_2O^+	1.70	1.96	0.75	0.77	1.04	-	-	0.75	0.80
H_2O^-	-	-	0.29	0.25	-	-	-	0.35	0.21
Total	99.21	100.35	99.37	98.95	99.81	98.85	100.13	98.38	99.09
					Trace Elements (PPM)				
Ba	-	-	149	141	195	220	-	182	160
Cu	-	-	150	73	110	143	111	127	81
Sr	-	-	222	156	175	188	186	183	139
V	-	-	282	334	235	350	-	257	333
Y	-	-	32	36	29	30	-	28	34

* Palisades chill zone (Pal.CZ) and the interior or second magma phase of the Palisades Sill 30 to 268 m above the base (Pal.Int.) after Walker (1969), High TiO_2 magma type after Weigand and Ragland (1970), Orange Mt. and Preakness Basalts after Puffer and Lechler (1980).

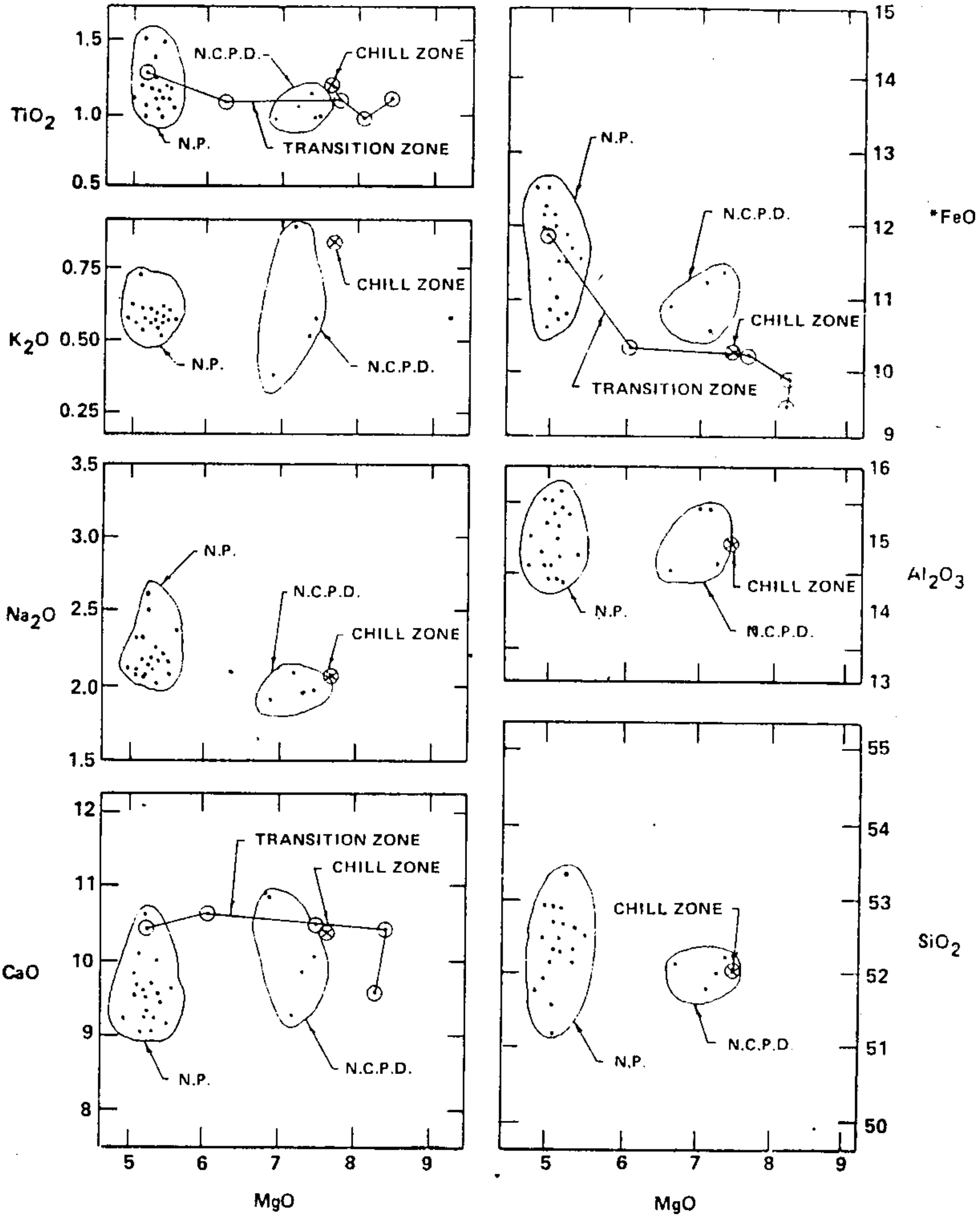

Figure 18 MgO Variation Diagrams: Northern Palisades Diabase (N.P.), New City Park Dike (N.C.P.D.) Edgewater Section Chill Zone (Walker, 1969), and Haverstrow Section Transition Zone (33m to 144m Above the Base of the Intrusion) Between the First Magma Phase and the Second Magma Phase

*FeO Includes Fe$_2$O$_3$ Recalculated as FeO

Ba, Cu, Sr, Zn and Zr and a decrease in Mg, Ca, Cr, and Ni corresponding to the Palisades trend fractionation displayed by the Edgewater section through the sill. The TiO_2, Fe_2O_3 (total) and CaO content of the transition zone (Walker, 1969) is plotted on figure 18 and connects the field representing the New City Park Dike and the first magma phase with the field representing the exposed portion of the northern Palisades. The composition of the diabase at the 144-m level closely matches the composition of each of the samples from the exposed portion of the northern Palisades (table 6). Apparently, fractionation processes did not continue to affect the magma above the transition zone.

Analyses of samples from the exposed portion of the northern Palisades consistently plot within a single rather restricted field (fig.18) regardless of the fact that the samples were collected from several widely scattered locations. There is not any clearly discernible fractionation pattern within the restricted field represented by our samples of northern Palisades diabase. We interpreted the chemical data of Table 6 as indicating a high degree of compositional homogeniety throughout the portion of the northern Palisades exposed west of Long Clove (fig.17). The kind of layering that includes the olivine zone of the southern Palisades (Walker, 1969) is absent from the western sickle-shaped portion of the northern Palisades for reasons that may be related in part to the discordant structural attitude of the western portion.

Gravity-controlled crystal settling was probably not involved in the formation of the cryptic layering displayed throughout the southern Palisades. McBirney and Noyes (1979) have pointed out some of the difficulties with crystal settling as a differentiation mechanism except for the separation of phases that are very dense and/or heavy relative to the silicate liquid phase. Even the olivine layer at the base of the second magma phase of the southern Palisades probably did not involve much crystal settling. Walker (1969) has found that the olivine crystals are too small, too Fe-rich, and too uncorroded to have settled very far through a diabase magma as an initial cumulate phase.

Nevertheless, the presence of a gravitational field directed perpendicular to the floor of the southern Palisades was probably partially responsible for the asymmetric development of the Palisades fractionation trend from the base to near the top of the second magma phase because of its control on the upward migration of volatiles. The homogeneous nature of the northern Palisades probably resulted from 1) its discordant implacement resulting in the absence of a gravitational field and a thermal gradient directed perpendicular to a horizontal intrusive floor; 2) the relatively shallow and therefore more rapid crystallization of the magma resulting in the advancement of a crystallizing solid front at a faster rate than diffusion and convection rates. The differentiation mechanism, therefore, simply broke down above the transition zone of the northern Palisades.

INTERPRETATION AND STRATIGRAPHIC IMPLICATIONS

Our interpretation of the structural and geochemical data indicates a sequence of events that begins with the intrusion of the first pulse of Palisades magma as a sill throughout most of its length. The composite Palisades intrusive, however, becomes increasingly discordant at its northern end west of Verdrietge Hook (Lowe, 1959; Koutsomitis, 1980). The first pulse of magma was emplaced in Rockland County as the New City Park Dike (fig.17) and the basal 33 m of the northern Palisades (Walker, 1969). The second pulse of magma passively penetrated into the still liquid interior of the first magma phase before it had completely crystallized. Since the base of the second magma phase, as represented by the bronzitic dolerite located 34 m above the base of the northern Palisades at Haverstraw, is slightly more fractionated than the lower chill zone of the first phase (Walker, 1969), the second magma phase was a more highly fractionated magma. Fractionation of the second pulse magma continued throughout the transition zone (fig.18) until a rather homogeneous zone was developed as represented by the rock above the transition zone and all the northern Palisades exposures (fig.17) that we analyzed (fig.18).

Some of this rather homogeneous and fractionated second pulse magma probably penetrated through to the surface at some point near Mt. Ivy where it flowed out onto the floor of the Newark Basin. Our geochemical (table 6) and petrographic data (table 5) agrees with field evidence and geophysical evidence (Kummel, 1900; Savage, 1968; Geiger and others, 1980; Ratcliffe, 1980; and Kodama and others, 1981) indicating a genetic connection between the Ladentown Basalts and the northern Palisades. Fractional crystallization following the emplacement of the northern Palisades magma has resulted in rock that is somewhat more fractionated than the extrusive Ladentown Basalts that did not undergo similar post-emplacement fractionation. A high degree of similarity in the chemical composition of the northern Palisades and Ladentown Basalt is, however, retained (table 6). Field evidence that indicates a genetic connection includes the southwesterly paleoflow direction of the Ladentown Basalt based on the orientation of pahoehoe tongues and lobate bulges (Ratcliffe, 1980; Geiger, 1982 ms). The basalt flowed along an apparent channel that cuts across the northeast-dipping strata of the longitudinal syncline occupied by the Ladentown Basalt (Kummel, 1900; Ratcliffe, 1980). As the lava flowed through the southwesterly incised drainage system, it became ponded against the border fault (Ratcliffe, 1980). The contact between the basalt flows and the sediments at the western edge of the Ladentown Basalts may indicate that the lavas flowed into locally developed sag ponds along the active border fault. Ratcliffe (1980) has found boulders and cobbles of rhyodacite in fanglomerates exposed between the Ladentown and Union Hill Basalts that were probably carried south from the Proterzoic Rosetown, New York dike swarm, thus indicating a southward-flowing drainage system.

Gravity and ground magnetic survey data (Kodama and others, 1981) indicate two near-vertical feeder dikes that extend westward from the apparent termination of the northern Palisades and join a third feeder dike still closer to the Ladentown Basalts. The genetic connection is therefore well established.

The evidence that indicates that the source of the Ladentown Basalts was from the north does not contradict Manspeizer's (1980) interpretation, based largely on pipe vesicle orientations, that the Orange Mountain and Preakness Basalts flowed from the south. Manspeizer's data pertains to the central portion of the Newark Basin whereas the Ladentown and Union Hill Basalts occupy restricted basins along the Ramapo Fault.

At least two interpretations are consistent with both a northern Palisades source of lava and a high degree of chemical resemblance among the various fractionated rocks of Rockland Co.; one involves two episodes of magmatic activity whereas the second alternative involves three episodes.

Two-Magma Interpretation

The most straightforward treatment of the geochemical data would be to incorporate only two pulses of magma. The first pulse of the composite Palisades diabase intrusive would have emplaced the New City Park Dike, the exterior layers of the Palisades, and then extruded to form the Orange Mountain Basalt. The second pulse would have then intruded into the still-liquid interior of the previous intrusive, then extruded to form the Ladentown, Union Hill, and Preakness Mountain Basalts, all of which are chemically similar in most respects (table 6 and fig.18). The difficulty with this proposal is that it does not leave very much time for the Feltville Formation to have been deposited. If it is assumed that the first pulse of magma intruded under the approximately 3000 m of Passaic Formation sediment separating it from the Orange Mountain Basalt, it can be estimated on the basis of the crystallization rate model of Irvine (1970) that the first pulse would have crystallized within approximately 1000 to 2000 years after the first pulse to have intruded into a still-liquid interior. Van Houten (1969), however, estimated that the Passaic Formation underlying the Orange Mountain Basalt was deposited at the rate of approximately 0.3 millimeters per year. If the 170-m thick Feltville formation was deposited at the same rate as the Passaic Formation, a 0.6 m.y. time span would separate the Orange Mountain Basalt from the Preakness Mountain Basalt.

The depositional rate of sediment in the Newark Basin probably accelerated by as much as an order of magnitude within the zone of extrusives (Cornet, 1977; Olsen and others, 1981) because of increased tectonic activity accompanying magmatic activity. This would decrease the time span between the first magma pulse (the Orange Mountain Basalt) and the Preakness Bountain Basalt to as low as 0.06 m.y. (60,000 years) but still too much time to be compatible with the 1000-2000 year time span based on cooling-rate estimates. It is, therefore, unlikely that the second pulse of the Palisades can be time-stratigraphically correlated with the Preakness Mountain Basalt.

Three-Magma Interpretation

The second alternative also begins with the intrusion of the first (high-TiO_2) magma phase of the Palisades and the accompanying extrusion of the Orange Mountain Basalt. The second magma pulse then quickly intruded into the still liquid interior of the first magma phase. Only minor fractionation occurred within the magmatic source during the short time separating the first from the second magma pulses. The base of the second pulse as represented by Walker's (1969) bronzitic dolerite layer at Haverstraw is vertually equivalent to the composition of the lower chill zone at Edgewater (Walker, 1969). Once the second magma pulse had been emplaced, however, shallow, in-situ fractionation quickly proceeded (within 1000 to 2000 years) to develop what is commonly referred to as the Palisades fractionation trend. In-situ fractional crystallization resulted in the cryptic layering of the southern Palisades at Edgewater. Within the discordant northern Palisades, very shallow fractionation proceeded to the point where a homogeneous magma was developed above a thin transition zone then apparently stopped (fig.18). This fractionated second-pulse magma represents the major exposed portion of the northern Palisades. Some of this fractionated second magma phase presumably then extruded onto the surface at some point near Mount Ivy and flowed south to form the Ladentown Basalt (and perhaps the Union Hill Basalt). The flows were deposited on Passaic Formation sediments and were followed by the deposition of the Feltville Formation in New Jersey.

While Feltville sediments were being deposited a fractionation of high-TiO_2 magma was slowly occurring somewhere deep within the crust or perhaps at the base of the crust. Similar deep crustal ponding and fractionation of high-TiO_2 type eastern North American magma has been proposed by Weigand and Ragland (1970), Smith and others (1975), Ressitar (1979), Puffer and Lechler (1980), and Puffer and others (1981). If a large supply of high-TiO_2 magma had been ponded at depths approaching the base of the crust then the rate of fractionation may have been slow enough to allow the 170 m of Feltville sediments to have been deposited. This magma may have been tapped as a third magma thus forming the Preakness Mountain (Second Watchung) Basalt at approximately the same stage of fractionation experienced by the Ladentown Basalt (the second magma pulse).

The principal advantages of this interpretation are that it allows enough time for the Feltville Formation to have been deposited and it accounts for some of the compositional and textural differences between the Preakness Mountain Basalt and the Ladentown Basalt. The Ladentown Basalt contains about the same amount of Cu and Sr (150 ppm and 222 ppm) as the second pulse of the southern Palisades (143 ppm and 188 ppm) but much more than the Preakness Basalt (81 ppm and 139 ppm) or the Union Hill Basalt (73 ppm and 156 ppm). Low Cu and Sr values are also characteristics of the Holyoke Basalt of Connecticut that has been interpreted as a deep seated fractionation product of high-TiO_2 type magma (Puffer and others, 1981). The higher Cu and Sr values of the Preakness and Union Hill Basalts compared to the second magma phase of the Palisades and the Ladentown Basalt may, therefore, have been caused by contrasting fraction trends that were occurring at different depths.

The most notable difference in the composition of the northern Palisades and Ladentown Basalt compared to the Preakness Basalt and Union Hill Basalt is the copper content (table 6). The fact that the copper content of the Preakness Basalt is lower than the copper content of the underlying Orange Mountain Basalt indicates that considerable copper was separated from the magma source some time during the fractionation process that developed the Preakness Basalt. The separation of copper probably occurred: 1) as a venting of copper-bearing volatiles during or shortly following the extrusion of the Orange Mountain Basalt; or 2) as a dense immiscible sulfide phase that may have separated from the fractionating magma source perhaps in a way resembling the development of the immiscible sulfide layers at the base of the Sudbury Norite and within the Merensky Reef of the Bushveld layered intrusion. The upward increase in the copper content of the Palisades is, in contrast, probably caused by the upward migration of copper-bearing deuteric volatiles. The chalcopyrite of the Palisades is typically disseminated within the rock but also fills thin fractures and does not appear to have precipitated in equilibrium with the silicate phases.

The enrichment of other elements such as titanium and strontium in the second magma phase of the Palisades and in the Ladentown Basalt compared to the Preakness Basalt and Union Hill Basalt may be due to the contrasting confining pressure and vapor pressure environments and contrasting crystallization rates of the two (shallow versus deep) fractionating magmas.

The composition of the Union Hill Basalt, therefore, more closely resembles the composition of the Preakness Mountain Basalt than the Ladentown Basalt and may be an erosional remnant or tectonically displaced portion of Preakness Mountain Basalt. The glass content and plagioclase and opaque oxide phenocryst content of the Union Hill Basalt also more closely resemble the Preakness Mountain Basalt than the Ladentown Basalt (table 6).

We, therefore, favor the three-magma interpretation but recognize that the geochemical and petrographic evidence derived from the Union Hill Basalt remains somewhat ambiguous. We also recognize that the Hook Mountain (Third Watchung) Basalt of New Jersey and perhaps still more recent igneous events followed the extrusion of the Preakness Mountain Basalt and Union Hill Basalt (Puffer and Lechler, 1980).

In summary, the three magma sequence is: 1) A high-TiO_2 type magma intruded as the outer layers of the northern and southern Palisades, the New City Park Dike, and extruded as the Orange Mountain (First Watchung) Basalt onto the Passaic Formation. 2) A second magma pulse, quickly following the first, passively intruded into the interior of the Palisades. Shallow, in-situ, fractionation within the northern Palisades proceeded to develop a rather thick layer of homogeneous copper enriched magma at which point the process was interrupted by the extrusion of the Ladentown Basalt onto the Passaic Formation. 3) A third magma slowly fractionated out of high-TiO_2 magma ponded at a deep level that extruded onto the Feltville Formation as the Preakness Mountain Basalt (the Second Watchung) and perhaps the Union Hill Basalt. The copper depletion of the third magma may have been due to gravitational separation of an immiscible sulfide liquid.

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CHAPTER FIVE

MESOZOIC BASALTIC ROCKS FROM WEST-CENTRAL NEW JERSEY AND PENNSYLVANIA: MAJOR AND TRACE ELEMENT GEOCHEMISTRY OF WHOLE-ROCK SAMPLES

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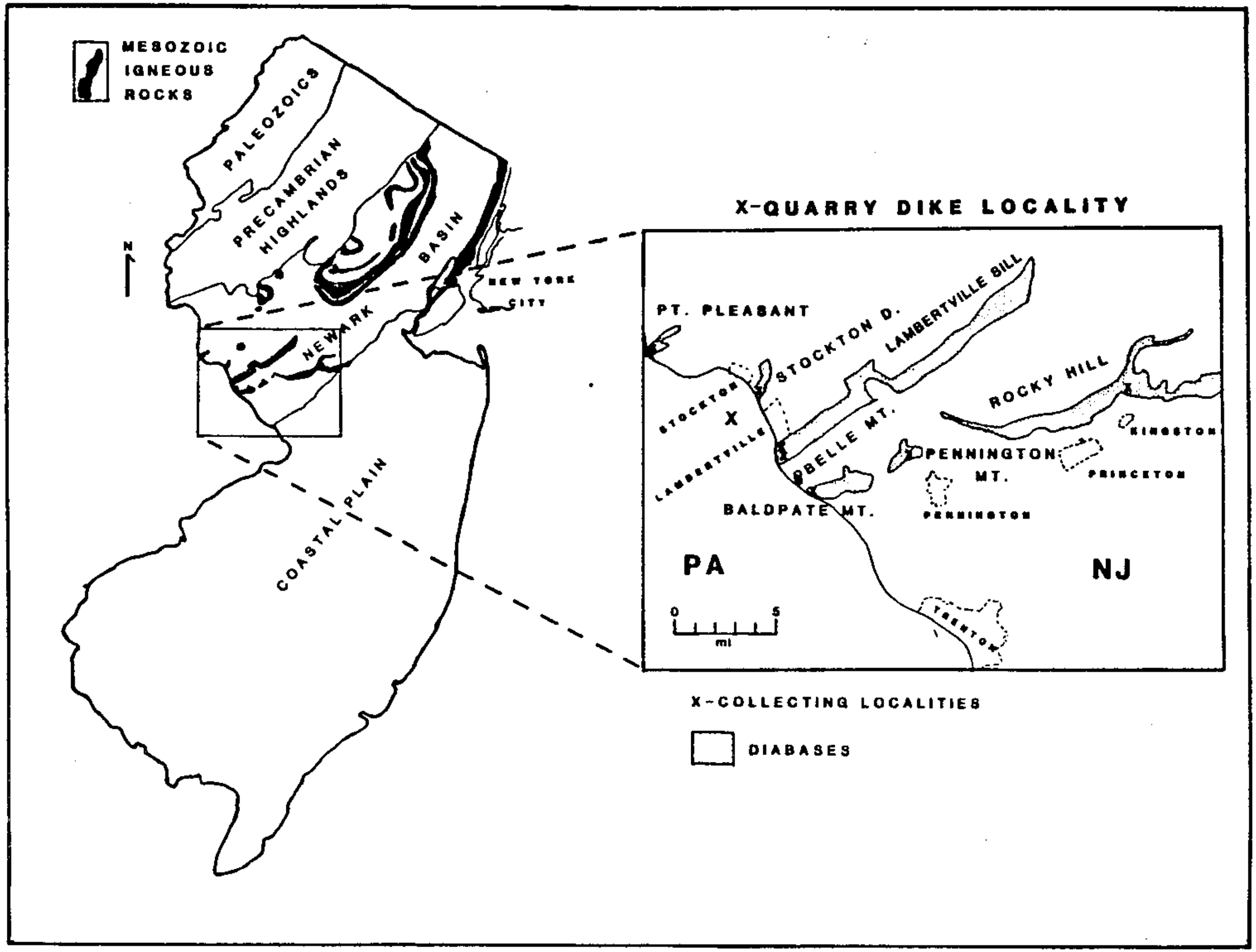
The basaltic rocks of west-central New Jersey and eastern Pennsylvania are located in the Mesozoic Newark/Gettysburg basin and include at least seven individual diabases with sill-like structures associated with numerous smaller fine grained dikes. All of these intrusives are part of the late Triassic/early Jurassic Eastern North America (ENA) Mesozoic Basalt Province. They are of interest because they form a magmatic "link" between the intensively studied ENA basaltic rocks of southern Pennsylvania (Hotz, 1953; Smith et al., 1975) and northern New Jersey (Walker, 1969; Puffer and Lechler, 1980). Because of this important regional position, forty-five whole-rock ENA samples from west-central New Jersey and eastern Pennsylvania have been analysed for major and trace element concentrations (Husch et al., 1984; Husch and Schwimmer, 1984). All analyses utilized direct current plasma emission spectrometry (Carr and Walker, 1982), an analytical method that has proven to be fast, inexpensive, and accurate.

The seven ENA diabases from west-central New Jersey (fig. 1) can be divided geographically into northern and southern groups. The northern group consists of the Point Pleasant (Byram) Diabase, Stockton Diabase, and Lambertville Sill (Solebury Mountain Diabase). The southern group consists of the Belle Mountain (Bowman Hill) Diabase, Baldpate Mountain Diabase, Pennington Mountain Diabase, and Rocky Hill Diabase. Analyses for northern and southern group samples (Husch et al., 1984) are presented in tables 1 and 2, respectively.

Compositions of southern group samples are more fractionated, in general, than compositions of northern group samples (figs. 2 and 3). Figure 4 shows that southern group compositions are more closely comparable to the quartz normative high-iron type of Weigand and Ragland (1970), whereas northern group compositions are more closely related to the quartz normative high-titanium (HTQ) type of Ragland and Whittington (1983). Although measured amounts of compositional variability within any one diabase are limited, the combined differentiation trend of all seven diabases bodies is almost identical to the Palisades trend (figs. 2 and 3) as determined by Walker (1969). This would suggest similar crystal fractionation schemes. However, for the diabase bodies of west-central New Jersey, a variety of parental magma compositions are indicated. Fractionation of a crustally ponded HTQ type magma, coupled with the periodic injection of the evolving magma to higher crustal levels is a viable mechanism for having produced this


MESOZOIC IGNEOUS ROCKS
N
PALEOZOICS
PRECAMBRIAN HIGHLANDS
NEWARK BASIN
NEW YORK CITY
COASTAL PLAIN
X-QUARRY DIKE LOCALITY
PT. PLEASANT
STOCKTON D.
LAMBERTVILLE SILL
ROCKY HILL
STOCKTON
X
LAMBERTVILLE
BELLE MT.
PENNINGTON MT.
KINGSTON
PRINCETON
BALDPATE MT.
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X-COLLECTING LOCALITIES
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Figure 1.

Table 1

Northern Group Analyses

	PP1	PP2	PP5	PP6	LS2	LS3	LS8	LS9	LS11	LSGRAB	LQ1	LQ4
SiO_2	52.58	52.42	52.51	52.57	52.18	52.63	53.07	52.78	52.75	52.55	52.82	52.81
TiO_2	1.13	1.11	0.75	1.10	0.90	1.11	1.03	0.97	0.89	1.06	1.00	1.00
Al_2O_3	14.63	14.19	10.29	14.26	11.50	14.23	18.09	15.47	14.99	13.44	14.50	14.27
FeO	8.26	8.22	7.95	8.30	8.23	8.26	7.20	7.99	7.70	8.20	8.77	8.98
Fe_2O_3	1.94	1.93	1.87	1.95	1.93	1.94	1.69	1.88	1.81	1.93	2.06	2.11
MnO	0.16	0.16	0.18	0.16	0.17	0.16	0.15	0.17	0.16	0.17	0.19	0.20
MgO	7.51	8.14	15.39	8.03	11.97	8.04	4.65	6.92	7.75	9.05	6.98	7.15
CaO	10.69	11.00	9.03	10.95	10.94	10.93	10.55	10.83	11.22	11.17	10.64	10.52
Na_2O	2.34	2.08	1.56	1.98	1.67	1.99	2.84	2.35	2.25	1.86	2.35	2.23
K_2O	0.61	0.61	0.35	0.55	0.47	0.59	0.58	0.55	0.35	0.54	0.58	0.60
P_2O_5	0.14	0.14	0.11	0.14	0.03	0.12	0.14	0.09	0.11	0.03	0.11	0.13
Ba	145	147	90	140	225	506	152	201	105	157	139	217
Cr	399	496	791	533	810	437	42	152	301	757	42	45
Ni	270	396	660	341	495	252	186	236	297	338	269	200
Rb	21	16	10	18	13	19	17	19	13	18	18	15
Sc	35	36	37	36	40	36	30	35	36	37	40	40
Sr	165	169	109	161	121	162	218	173	102	149	183	171
V	256	264	229	258	241	260	232	261	245	235	283	300
Zr	96	94	65	96	71	93	96	84	80	87	82	82

Major elements normalized to 100 percent; Fe_2O_3/FeO set at .235

Table 2

Southern Group Analyses

	KQ6	KQ7	KQ8	PQ2	PQ3	PQ4	PQ6	WH1	WH2	WH3	MS1	MS2	MS3
SiO_2	53.08	52.90	53.43	53.89	53.75	61.13	54.86	52.98	53.64	52.90	61.30	61.64	61.59
TiO_2	1.06	0.97	1.08	1.84	1.65	1.52	2.40	1.37	1.12	1.30	1.63	1.59	1.51
Al_2O_3	17.01	15.82	16.70	14.55	15.73	12.11	12.10	16.40	16.85	16.32	12.05	11.85	12.11
FeO	7.86	8.01	7.75	10.64	9.59	10.53	13.33	9.08	7.73	8.43	10.46	10.24	10.38
Fe_2O_3	1.85	1.88	1.82	2.50	2.25	2.47	3.13	2.13	1.82	1.98	2.46	2.41	2.44
MnO	0.16	0.17	0.15	0.18	0.18	0.26	0.27	0.17	0.16	0.19	0.18	0.22	0.21
MgO	5.12	6.54	5.29	3.56	3.67	1.05	2.04	4.07	5.19	5.63	1.09	0.96	1.00
CaO	10.37	10.43	10.37	8.51	9.13	4.74	5.68	8.37	9.92	7.33	4.53	4.80	4.50
Na_2O	2.65	2.33	2.58	2.84	2.88	3.23	3.77	3.57	2.86	4.46	3.57	3.37	3.34
K_2O	0.72	0.83	0.70	1.28	0.96	2.49	1.87	1.74	0.60	1.33	2.29	2.41	2.42
P_2O_5	0.12	0.12	0.12	0.20	0.20	0.47	0.55	0.11	0.11	0.13	0.45	0.50	0.49
Ba	169	164	175	280	236	520	396	281	174	303	451	478	508
Cr	49	119	68	17	19	17	12	23	85	144	9	9	12
Ni	150	206	168	88	102	---	12	115	203	227	---	---	---
Rb	20	27	20	41	25	79	54	53	18	36	66	69	73
Sc	31	34	31	34	33	21	30	31	35	34	21	19	20
Sr	192	187	191	195	198	147	173	225	221	204	142	142	150
V	233	224	231	331	295	29	68	271	257	266	23	22	25
Zr	89	83	104	161	141	323	208	109	115	108	306	279	296

Major elements normalized to 100 percent; Fe_2O_3/FeO set at .235

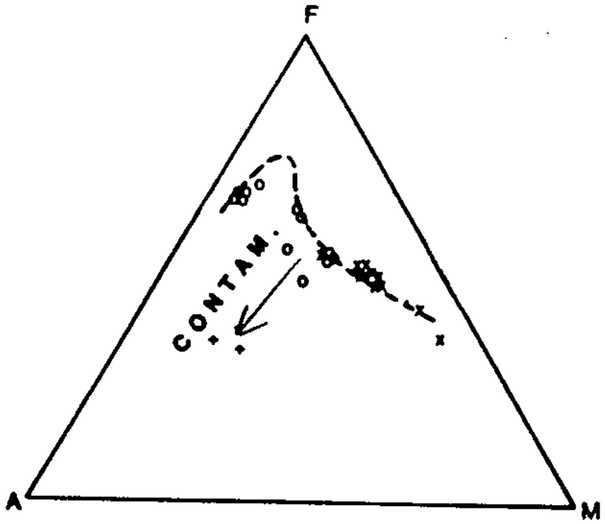

Figure 2. AFM (Na_2O+K_2O, Fe_2O_3* (total), MgO) ternary diagram. Dashed curve shows Palisades trend of Walker (1969). Northern group compositions (X). Southern group compositions (O). Sediments (+).

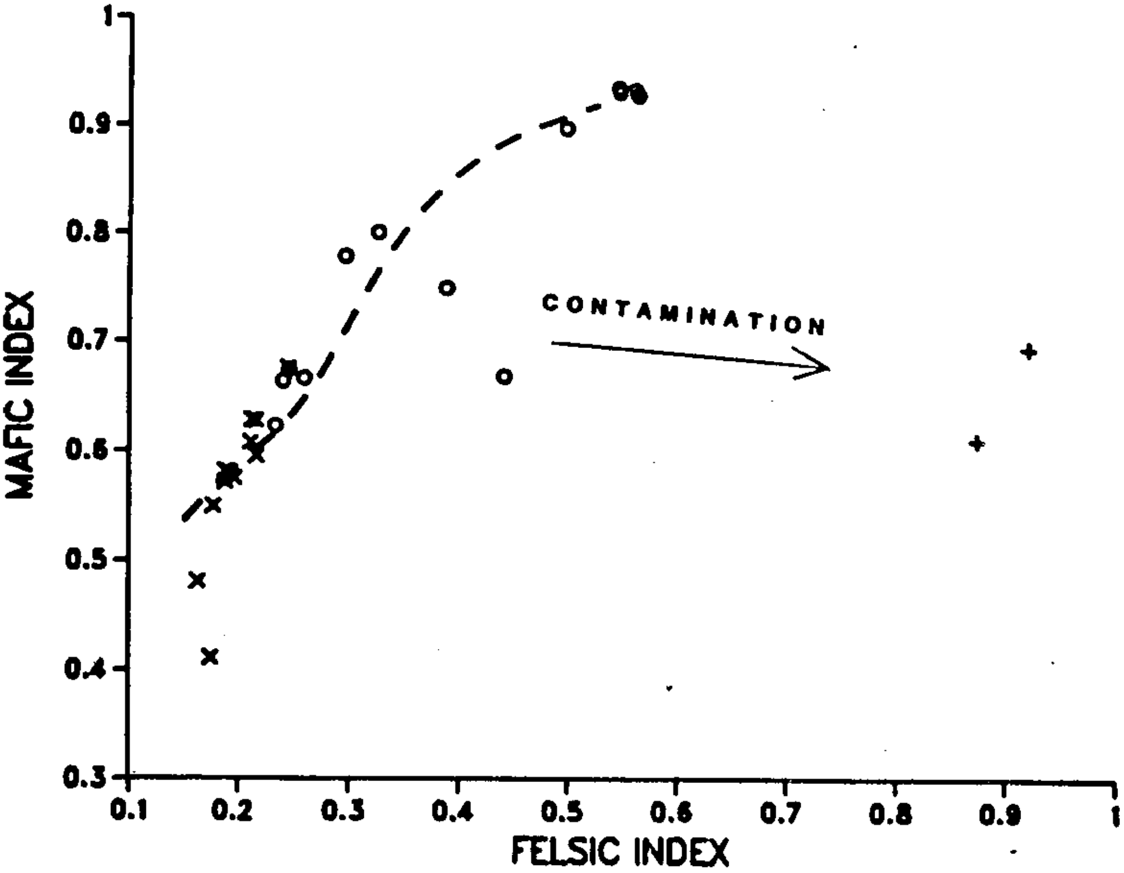

Figure 3. Mafic index (Fe_2O_3*/Fe_2O_3*+MgO) versus felsic index (K_2O+Na_2O/K_2O+Na_2O+CaO). Dashed curve shows Palisades trend of Walker (1969). Symbols as in figure 2.

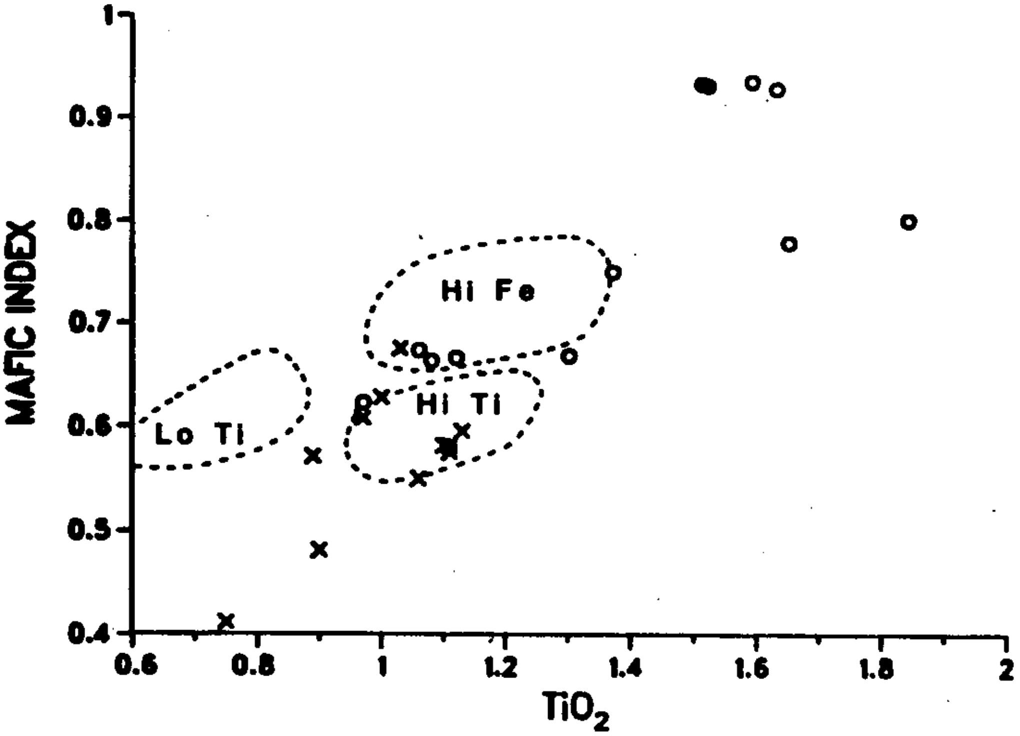

Figure 4. Mafic index versus TiO_2. Fields after Weigand and Ragland (1970) and Ragland and Whittington (1983). Symbols as in figure 2.

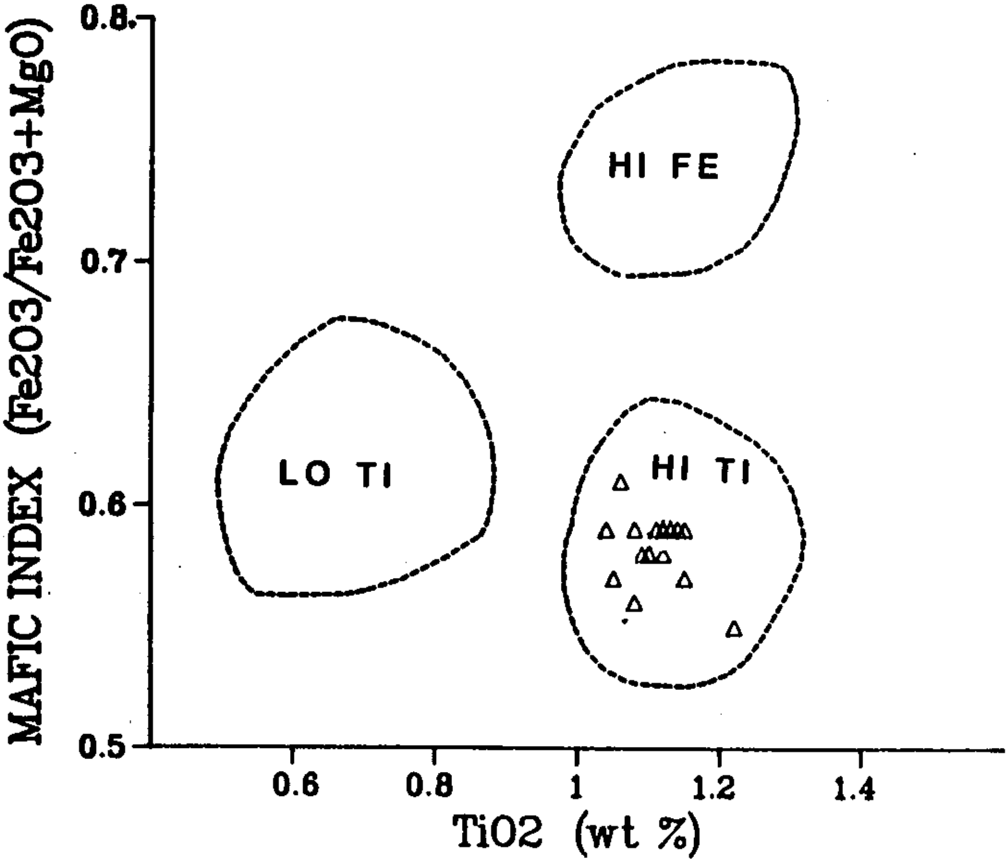

Figure 5. Same diagram as figure 4 with plotted positions of Quarry Dike compositions (triangles).

compositional variabilty (Weigand and Ragland, 1970; Smith et al., 1975; Puffer and Lechler, 1980).

The Point Pleasant Diabase and Lambertville Sill contain compositions significantly enriched in Mg, Cr, and Ni and depleted in Al, Rb, Sr, and Zr (samples PP5 and LS2). Mixing calculations indicate that orthopyroxene accumulation in a HTQ magma could have produced these characteristics (Husch et al., 1984). Finally, the effects of local crustal contamination are seen in only two samples, both from the Belle Mountain Diabase (samples WH1 and WH3). Figures 2 and 3 show that their compositions plot off the general fractionation trend and towards the plotted positions of two analysed Lockatong Argillite samples. This strongly suggests that Lockatong Argillite (or a rock of similar compositions) was the source of contamination.

Because many of the west-central New Jersey diabase samples are course grained, there are varying uncertainties with regard to how closely analysed compositions approximate actual magma compositions. The analysis of twenty samples collected across the fine grained Quarry Dike (Husch and Schwimmer, 1984), located near New Hope, Pennsylvania (fig. 1), provides results that are less ambiguous. These analyses are presented in table 3.

The Quarry Dike represents a HTQ type basalt (fig. 5) which has undergone an extensive amount of local contamination; highly localized reaction with dolomitic country rock also is indicated (Husch and Schwimmer, 1984). The calculated Quarry Dike parental

TABLE 3

Sample	1	2	3	4	5	6	7
SiO_2	54.46	52.84	52.67	52.77	52.97	52.75	52.91
TiO_2	1.27	1.15	1.11	1.12	1.13	1.13	1.09
Al_2O_3	16.37	14.01	14.10	14.03	14.41	14.45	14.34
Fe_2O_3	1.62	1.97	1.95	1.98	1.98	1.98	1.97
FeO	6.88	8.40	8.28	8.43	8.43	8.39	8.41
MnO	0.02	0.17	0.16	0.17	0.16	0.17	0.17
MgO	12.16	7.97	7.94	7.99	7.88	7.90	7.97
CaO	3.07	10.50	10.54	10.28	9.57	8.30	8.85
Na_2O	2.58	1.90	2.02	1.99	1.96	2.45	2.10
K_2O	1.41	0.96	1.11	1.11	1.38	2.34	2.05
P_2O_5	0.16	0.13	0.12	0.13	0.13	0.14	0.14
Ba	133	145	149	153	176	197	187
Cr	311	282	291	275	279	270	304
Cu	72	90	78	69	71	78	74
Ni	96	86	88	83	83	84	89
Rb	35	28	36	34	40	63	55
Sc	32	35	34	36	36	35	36
Sr	155	190	170	186	218	258	200
V	274	241	254	229	241	238	257
Zr	96	101	98	100	102	96	98

Sample	8	9	10	11	12	13	20
SiO_2	52.96	52.63	52.85	52.79	52.71	52.00	52.80
TiO_2	1.07	1.08	1.06	1.10	1.13	1.23	1.08
Al_2O_3	14.23	14.30	14.24	14.46	14.39	14.96	14.41
Fe_2O_3	1.97	1.98	1.94	1.95	1.97	1.97	1.86
FeO	8.37	8.42	8.31	8.33	8.40	8.39	7.91
MnO	0.17	0.18	0.17	0.17	0.18	0.12	0.17
MgO	8.08	8.36	8.27	8.22	8.03	9.30	8.25
CaO	8.93	7.84	7.89	8.06	8.50	6.82	8.34
Na_2O	2.05	2.61	2.73	2.31	2.37	3.09	2.81
K_2O	2.05	2.47	2.41	2.46	2.18	1.98	2.24
P_2O_5	0.12	0.13	0.13	0.14	0.15	0.15	0.13
Ba	174	201	227	207	184	152	195
Cr	264	334	374	314	293	360	333
Cu	56	64	82	106	102	142	105
Ni	77	89	105	97	93	101	99
Rb	48	68	76	70	63	51	67
Sc	34	34	40	35	37	40	36
Sr	202	201	229	211	165	266	160
V	209	244	281	270	266	307	270
Zr	88	89	106	90	97	106	93

TABLE 3 (cont.)

Sample	19	18	17	16	15	14
SiO_2	52.93	52.53	52.50	52.80	52.90	52.79
TiO_2	1.05	1.15	1.13	1.13	1.16	1.13
Al_2O_3	14.61	14.51	13.94	14.40	15.00	14.58
Fe_2O_3	1.96	1.96	2.00	1.94	1.89	1.96
FeO	8.36	8.34	8.51	8.27	8.05	8.38
MnO	0.17	0.17	0.18	0.18	0.15	0.16
MgO	7.94	7.92	8.10	7.88	8.32	7.76
CaO	7.88	8.00	8.26	8.28	8.27	9.68
Na_2O	2.20	2.62	2.76	2.44	2.40	2.10
K_2O	2.75	2.65	2.47	2.55	1.71	1.30
P_2O_5	0.14	0.15	0.15	0.14	0.15	0.15
Ba	221	221	233	225	175	173
Cr	307	301	296	303	311	321
Cu	112	118	95	116	117	128
Ni	92	93	92	95	99	101
Rb	80	82	74	80	52	47
Sc	37	36	37	37	37	40
Sr	182	171	189	181	188	208
V	273	286	271	276	288	299
Zr	99	94	99	101	103	103

Major elements normalized to 100 percent anhydrous.
Fe_2O_3/FeO set at .235
Samples are presented in order from west to east across the dike.

magma composition (table 4) is extremely similar to the chilled margin compositions of the Point Pleasant Diabase (sample PP1) and Lambertville Sill (sample LS3). This suggests that all three intrusions were tapping the same magma reservoir. However, the Point Pleasant Diabase and Lambertville Sill show no compositional evidence for having undergone any significant amounts of local contamination.

Quarry Dike variation trends (figs. 6 and 7) are very different from those for the Palisades Sill and west-central New Jersey diabases (figs. 2 and 3). Quarry Dike trends suggest that the source of contamination was a Mesozoic sediment similar in composition to Lockatong Argillite, the same potential contaminant as for the Belle Mountain Diabase. Contamination is marked by an increase in alkali content at essentially constant FeO*/MgO (figs. 6 and 7) and silica (fig. 8) values. Other mechanisms of magmatic differentiation, such as crystal fractionation, differences in percentage of partial melting, and liquid immiscibility appear not to have played an important role in the geochemical variability found across the Quarry Dike (Husch and Schwimmer, 1984).

In summary, analytical results show the most dominant ENA magma type in west-central New Jersey and eastern Pennsylvania to be a HTQ type associated with its more iron-rich derivatives. The HTQ compositions found in this region are very similar to those known previously from northern New Jersey and southern Pennsylvania. Thus, there appears to be no significant

TABLE 4

	QD*	PP1	LS3
SiO_2	52.77	52.58	52.63
TiO_2	1.13	1.13	1.11
Al_2O_3	14.23	14.63	14.23
Fe_2O_3	1.96	1.94	1.94
FeO	8.35	8.26	8.26
MnO	0.16	0.16	0.16
MgO	7.89	7.51	8.04
CaO	10.24	10.69	10.93
Na_2O	2.01	2.34	1.99
K_2O	1.12	0.61	0.59
P_2O_5	0.13	0.14	0.12
Ba	156	145	506#
Cr	298	399#	437#
Cu	99	---	---
Ni	92	270#	252#
Rb	37	21	19
Sc	36	35	36
Sr	189	165	162
V	265	256	260
Zr	101	96	93

* Quarry Dike parental magma compositions calculated by averaging samples 2, 3, and 4. These three samples are from the chilled margins and they do not exhibit any geochemical characteristics associated with contamination or wall rock reaction.

\# Analytical uncertainty is high

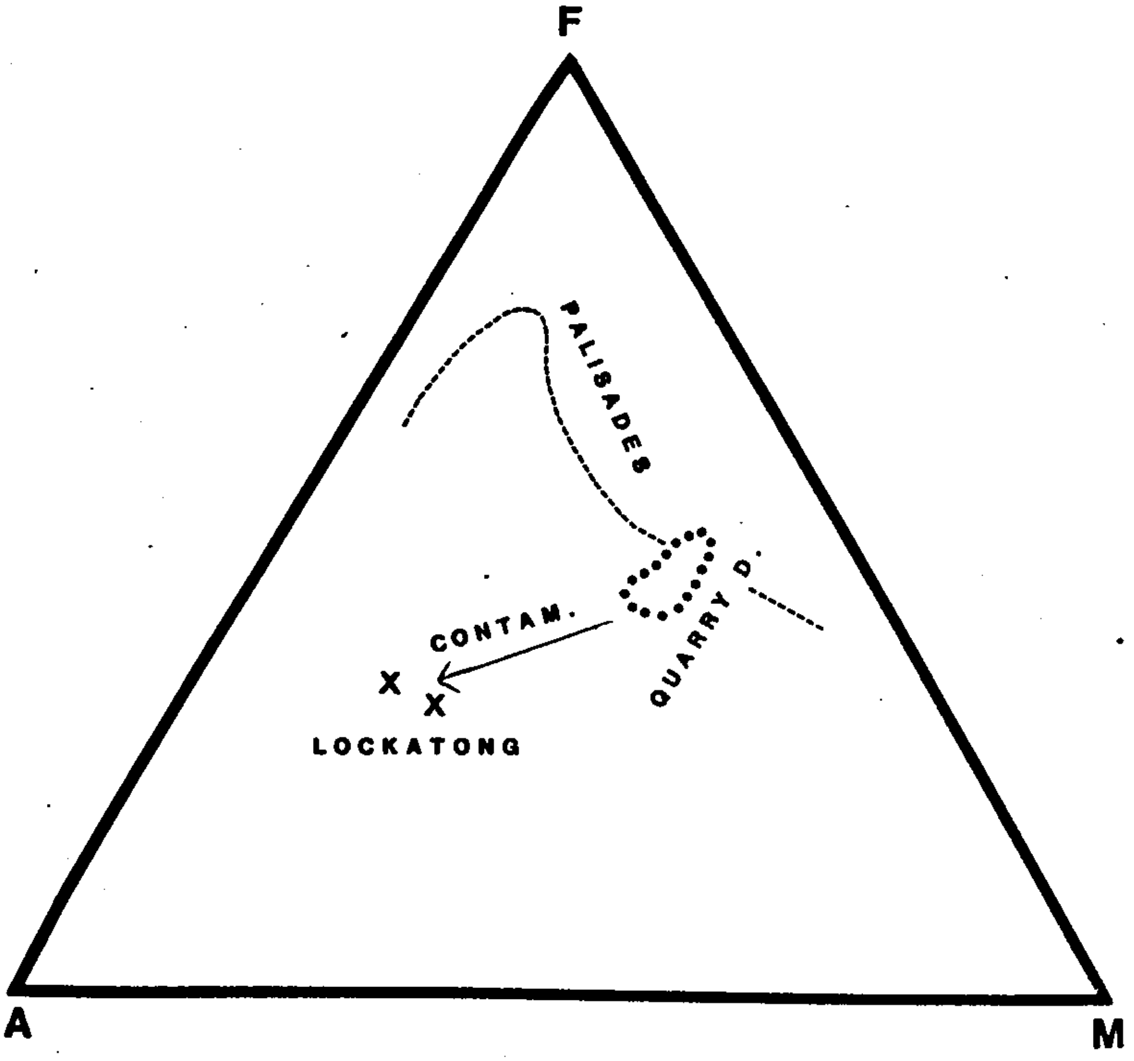

Figure 6. AFM diagram showing distribution field of Quarry Dike compositions (dotted field). Palisades trend (dashed line) and Lockatong Argillite compositions (X) also shown.

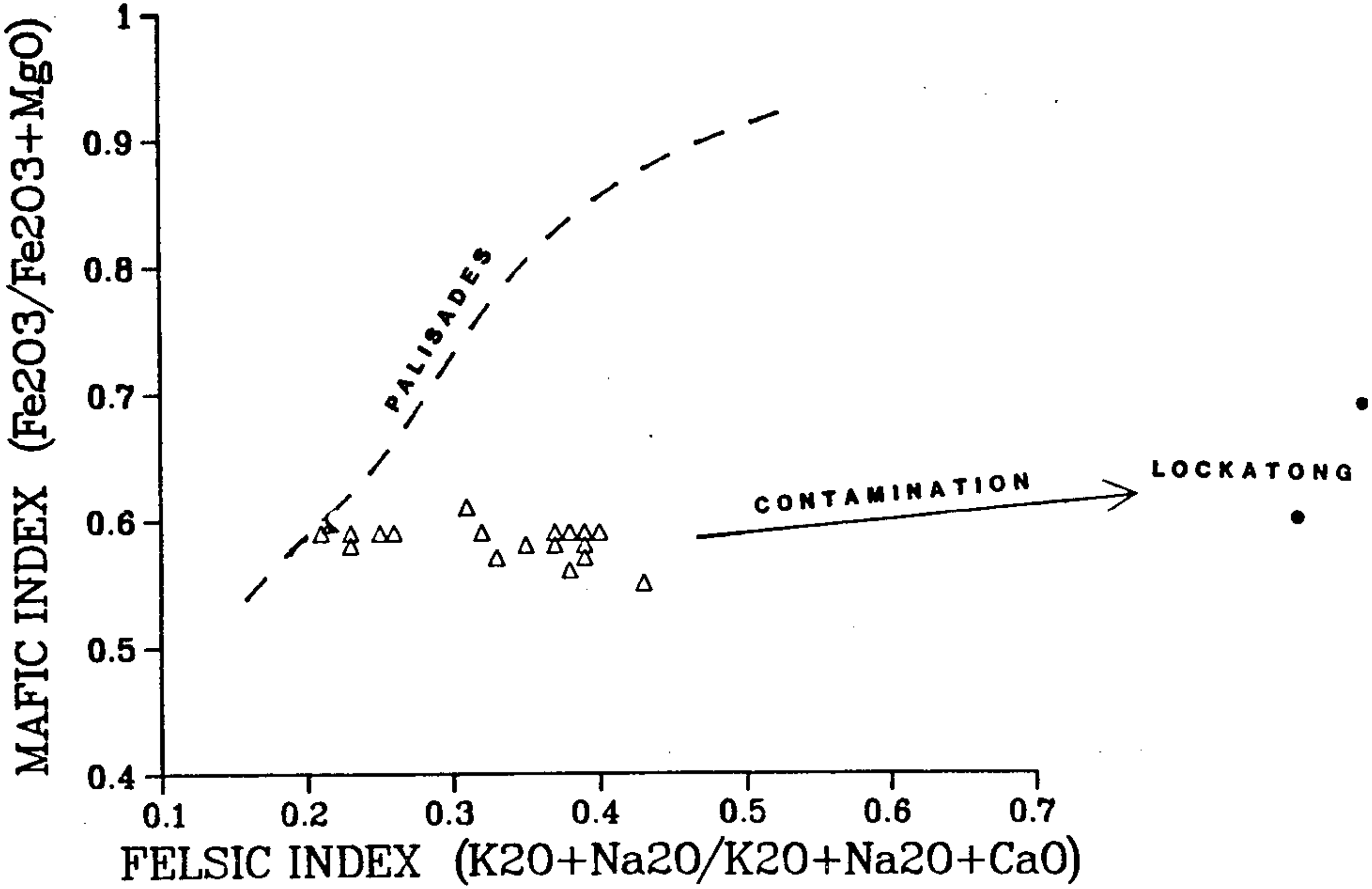

Figure 7. Mafic index versus felsic index diagram with Quarry Dike compositions shown by triangles. Palisades trend (dashed line) and Lockatong Argillite compositions (filled circles) also shown.

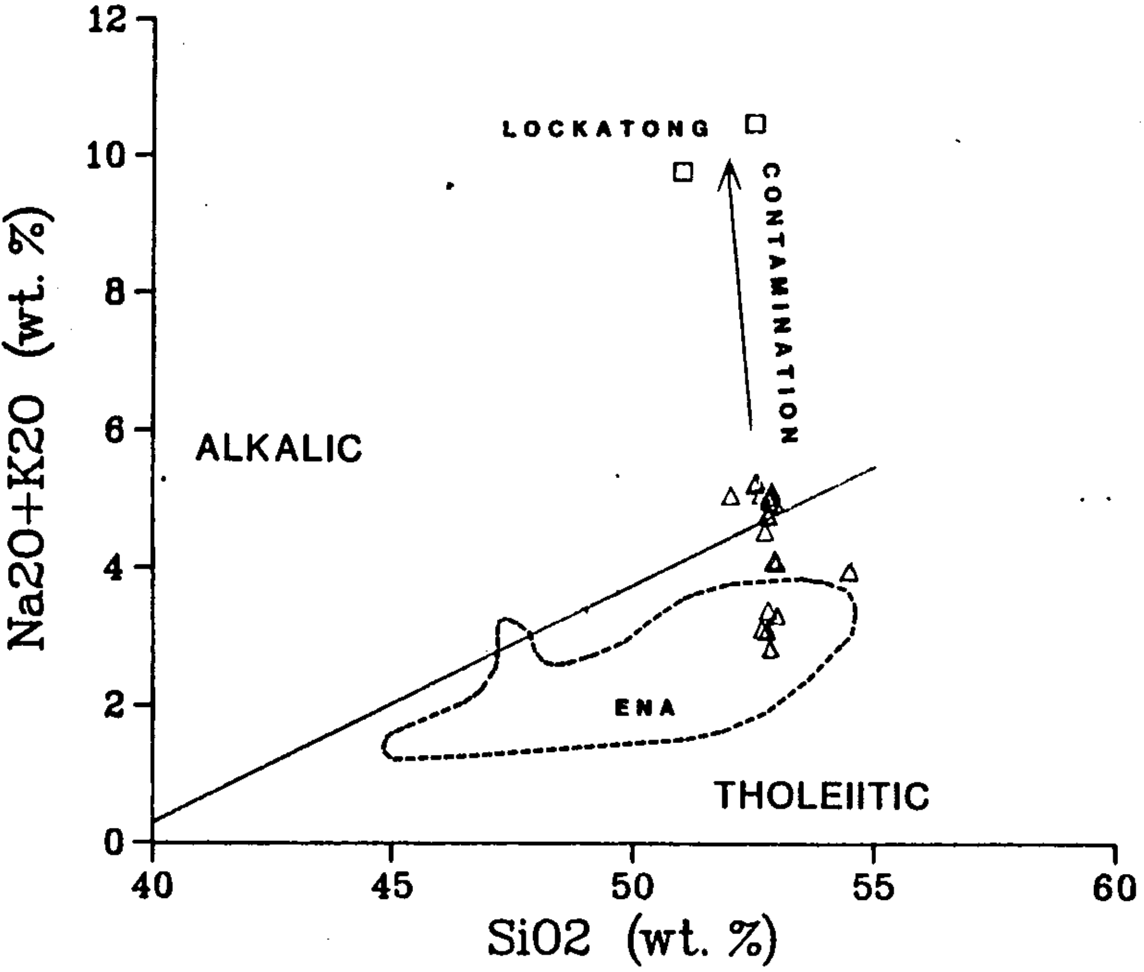

Figure 8. Alkali versus silica diagram with Quarry Dike compositions shown by triangles. Typical ENA compositions plot within dashed field. Lockatong Argillite compositions (squares) also shown.

variations in HTQ basalt compositions throughout the entire Newark/Gettysburg Basin.

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CHAPTER SIX

THE MINERALS OF BERGEN HILL, NEW JERSEY*

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INTRODUCTION

The Triassic traprock quarries of New Jersey have been a prolific source of beautifully crystallized zeolites and associated species since the early 1800s. The purpose of this article is to review, summarize and supplement previous publications dealing with New Jersey traprock minerals, especially Manchester (1931) and Mason (1960). This article will focus on the Bergen Hill occurrence in the Palisades diabase. The geology, mineralogy and history of Bergen Hill localities is summarized and brought up to date.

HISTORICAL BACKGROUND

The area known as Bergen Hill consists of several railroad cuts and borings through a 19-kilometer-long ridge of diabase stretching from Bergen Point, New Jersey at the south to Piermont, New York to the north (Manchester, 1931). Six tunnels or cuts produced thousands of superb mineral specimens (Fig. 1). In 1830, a group of Newark businessmen sought to charter a railroad to transport manufactured goods to New York City markets via Jersey City. Goods destined for markets at that time were transported by barge via the Morris Canal. However, this waterway was often frozen solid in winter making passage difficult or impossible. Their charter was granted by the New Jersey legislature on March 7, 1832. In order to reach Jersey City by rail, Cunningham (1951) stated that a cut through the diabase ridge at Bergen Hill was necessary as "Bergen Hill was astride any route between Paterson or Newark and Jersey City, impossible for the locomotive power of the day to surmount." The first railroad cut through Bergen Hill to produce minerals was jointly excavated by the New Jersey Railroad and the Camden and Amboy Railroad between 1832 and 1838. According to Bourne (1841), well-crystallized specimens were not found in any quantity in the New Jersey Railroad cut prior to 1837. Bourne (1841) lists 14 different calcite-lined veins in the diabase which produced a number of fine specimens of stilbite, calcite, apophyllite, and other associated species (Table 1). Beck (1843) obtained mineral specimens from Bergen Hill as early as 1838. He enumerated 12 species, of which pectolite, apophyllite, mesolite, analcime, datolite and stilbite were most important (Table 1). Specimens from this early occurrence at Bergen Hill are not represented in the mineral collections of the American Museum of Natural History or the Paterson Museum, and it is doubtful that other institutional mineral collections contain documented specimens from this Bergen Hill locality.

After producing a fair number of specimens between 1837 and 1343, Bergen

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*Information for this article originally appeared in Peters, J.J. (1984). Triassic traprock minerals of New Jersey. Rocks & Minerals, 59:157-183.

TABLES 1-6

All species or varieties reported in literature as occurring at the localities covered in this article are listed in Tables 1-6. Valid species names are capitalized; invalid, out-moded and group names are in lower case. Several minerals reported in the literature are dubious; these are indicated with asterisks. Specimens represented in the mineral collections of the American Museum of Natural History and the Paterson Museum are indicated by the abbreviations AMNH and PM.

TABLE 1. MINERALS REPORTED FROM BERGEN HILL, HUDSON COUNTY, NEW JERSEY

New Jersey Railroad Cut, Jersey City (1832-1838)

ANALCIME	Bourne (1841), Beck (1843)
APOPHYLLITE	Bourne (1841), Beck (1843)
CALCITE	Bourne (1841), Beck (1843)
CHABAZITE	Bourne (1841), Beck (1843)
DATOLITE	Bourne (1841), Beck (1843)
EPISTILBITE*	Bourne (1841)
GALENA	Bourne (1841), Beck (1843)
HEULANDITE	Bourne (1841), Beck (1843)
LAUMONTITE	Bourne (1841)
MESOLITE	Bourne (1841), Beck (1843)
NATROLITE	Bourne (1841)
PREHNITE	Bourne (1841), Beck (1843)
PYRITE	Bourne (1841), Beck (1843)
SCOLECITE*	Bourne (1841)
SPHALERITE	Bourne (1841)
STELLERITE*	Beck (1843)
STILBITE	Bourne (1841), Beck (1843)

Erie Railroad Tunnel, Jersey City (1857-1860)

ANALCIME	Dana (1872), AMNH
APOPHYLLITE	Dana (1872), AMNH
CALCITE	Manchester (1931)
DATOLITE	Dana (1872), AMNH
diabantite	Manchester (1931)
NATROLITE	Dana (1872), AMNH
PECTOLITE	Dana (1872), AMNH
QUARTZ (milky)	Manchester (1931)
SPHALERITE	Manchester (1931)
STILBITE	Dana (1872), PM

Delaware, Lackawanna & Western Tunnels, Jersey City (1874-1876)

NATROLITE	AMNH
PECTOLITE	AMNH
PREHNITE	Manchester (1931)
THOMSONITE	Manchester (1931)

Pennsylvania Railroad Cut, Mt. Pleasant (1880s?)

apatite	Canfield (1889)
BARITE	Canfield (1889), Manchester (1931)
COPPER	Manchester (1931)
GALENA	Manchester (1931)
MAGNETITE	Manchester (1931)
OPAL (fire)	Canfield (1889), Manchester (1931)
OPAL (hyalite)	Manchester (1931)
OPAL (hydrophane)	Canfield (1889)
NATROLITE	AMNH
QUARTZ (chalc.)	Manchester (1931)

New York, Susquehanna & Western Railroad, Edgewater (1880s?)

apatite	Manchester (1931)
APOPHYLLITE	Manchester (1931)
CALCITE	Manchester (1931)
chlorite	Manchester (1931)
DATOLITE	Manchester (1931)
diabantite	Manchester (1931)
HEULANDITE	Manchester (1931)
NATROLITE	Manchester (1931), AMNH
PECTOLITE	Manchester (1931)
PYRITE	Manchester (1931)
QUARTZ (amethyst)	Manchester (1931)
QUARTZ (rock crystal)	Manchester (1931)
SCOLECITE*	Manchester (1931)
STILBITE	Manchester (1931)

New York, West Shore & Buffalo Tunnel, Weehawken (1881)

ALBITE	Manchester (1931)
ALLOPHANE*	Manchester (1931), Canfield (1889), AMNH
ANALCIME	Darton (1882) & (1883), Manchester (1931), AMNH
apatite	Manchester (1931)
APOPHYLLITE	Darton (1882) & (1883), Manchester (1931) Canfield (1889), AMNH
ARAGONITE*	Canfield (1889), Manchester (1931)
CALCITE	Darton (1882) & (1883), Manchester (1931), AMNH
CHABAZITE	Darton (1882), AMNH
CHALCOPYRITE	Manchester (1931)
chlorite	Darton (1882), Manchester (1931)
DATOLITE	Darton (1882), Manchester (1931), AMNH
GMELINITE	AMNH
hayesine*	Darton (1882), Canfield (1889)
HEULANDITE	Canfield (1889), Darton (1889), Manchester (1931)
ILMENITE	Manchester (1931)
LAUMONTITE	Darton (1882) & (1883), Canfield (1889), Manchester (1931), AMNH
leonhardite	Canfield (1889), Manchester (1931)
MESOLITE	Manchester (1931)
NATROLITE	Darton (1882) & (1883), Manchester (1931), AMNH
OPAL (hyalite)	Canfield (1889), Manchester (1931)
ORTHOCLASE	Manchester (1931)
paperspar	Manchester (1931)
PECTOLITE	Darton (1882) & (1883), Manchester (1931), AMNH
PREHNITE	Darton (1882) & (1883), AMNH
PYRITE	Darton (1882) & (1883), Manchester (1931)
QUARTZ (amethyst)	Manchester (1931)
QUARTZ (rock crystal)	Canfield (1889), Manchester (1931)
SIDERITE	AMNH
STILBITE	Darton (1882) & (1883), Manchester (1931)
TITANITE	Manchester (1931)
ULEXITE	Refer to Hayesine listed above and to text.
vermiculite	Manchester (1931)
wad	Manchester (1931)

Bergen Archways (Erie Cut), Jersey City (1906-1910)

ANALCIME	Manchester (1919)
APOPHYLLITE	Manchester (1919), AMNH, PM
CALCITE	Manchester (1919), AMNH, PM
CHALCOPYRITE	Manchester (1919), Wherry (1919), AMNH
chlorite	Manchester (1919)
DATOLITE	Manchester (1919), PM
GMELINITE	Manchester (1919)
LAUMONTITE	Manchester (1919)
NATROLITE	Manchester (1919), AMNH
PECTOLITE	Manchester (1919), PM
PREHNITE	PM
PYRITE	Manchester (1919), PM
QUARTZ (milky)	Manchester (1919)
SPHALERITE	Manchester (1919)
STILBITE	Manchester (1919), PM

*Dubious

Hill remained relatively quiescent until the 1870s when mineralized veins were uncovered in the Erie tunnel, Jersey City (Fig. 1). The Erie Railroad Company (with financial aid from the Morris & Essex Railroad) blasted a tunnel through Bergen Hill between 1857 and 1860 (Cunningham, 1951). While specimens may have been uncovered in the Erie tunnel during or shortly after its construction, record of specimen collecting during the 1860s is lacking. It seems likely that the onset of the Civil War in 1860 discourged mineral collecting at Bergen Hill for several years. Preserved in the Clarence S. Bement collection at the American Museum of Natural History are many fine specimens of Bergen Hill datolite, natrolite, pectolite, analcime and apophyllite acquired by Mr. Bement between 1871 and 1874. The Erie tunnel presumably was the source of this material. The late 1860s to early 1870s encompassed the great "Railroad Wars" waged between the New York Central's Cornelius Vanderbilt and the Erie Railroad's Jay Gould, Jim Fisk and Daniel Drew. In 1878, the Lackawanna Railroad leased the Morris & Essex Railroad thereby gaining access to the Erie tunnel through Bergen Hill. Two years later, the Lackwanna built its Boonton Branch via Paterson and sought to link up with the Morris & Essex via the Erie tunnel. The Erie felt this constituted an infringment of its Paterson territory, and in December of 1870, Jim Fisk temporarily withdrew his "troops" engaged in "battle" with the minions of the New York Central in order that he might engage another "enemy" - the Lackawanna Railroad (Cunningham, 1951). Fisk had an Erie locomotive placed across the mouth of the tunnel and stationed one thousand "troops" nearby to insure it would not be moved. Governor Randolf of New Jersey arrived on the scene with a company of state militia and was successful in removing the locomotive. Wishing to avoid a lengthly battle with the Erie, the Lackawanna Railroad cut its own tunnel through Bergen Hill between 1874 and 1876. Manchester (1831) states that two tunnels were eventually cut by the Lackawanna (the Delaware, Lackawanna & Western tunnels (Manchester, 1919). The Bement collection contains two specimens specifically cited from the Lackawanna tunnels - a natrolite obtained in 1885 via George F. Kunz (AMNH 13177) and a pectolite obtained via E.P. Hancock in 1884 (AMNH 9833).

The next tunnel cut through Bergen Hill was the New York, West Shore & Buffalo tunnel at Weehawken (Fig. 1). Cunningham (1951) relates that..."through the flinty rock of Bergen Hill at Weehawken the West Shore sliced its 4,255-foot tunnel in 1881. It cut over the north portion of the Meadows and sped through pleasant Bergen County terrain to Orangeburgh, New York." The West Shore tunnel was one of the more prolific sources of crystallized mineral specimens. Manchester (1931) listed 27 varieties or species from this tunnel. Darton (1882) reported pectolite, datolite, analcime, prehnite, calcite, natrolite, apophyllite, quartz (amethyst), stilbite, laumontite, titanite, heulandite, chabazite and pyrite. Pectolite was especially abundant as evidenced by a number of superb specimens acquired by Clarence Bement from George Kunz in 1885. The author was unable to gather much information concerning the New York, Susquehanna & Western Railroad tunnel at Edgewater or the Pennsylvania Railroad cut at Mount Pleasant (Fig. 1). Three specimens of natrolite are preserved in the Bement collection from the Susquehanna and Pennsylvania tunnels. Bement obtained a natrolite specimen (AMNH 13178) from the Susquehanna tunnel via George L. English in 1903. Similarly, two natrolite specimens from the Pennsylvania tunnel (AMNH 13163, 13166) were obtained via George Kunz in 1885.

The Erie Railroad develped a second cut through Bergen Hill between 1906 and 1910. Manchester (1919) named this combination of open cut and tunnel the Bergen Archways (Fig. 1). A crew of over 1100 men labored round-the-clock cutting a 1300-meter path through Bergen Hill. The average depth of the cut was 25 meters, and the width was 17 meters. Most of the cut was open, but

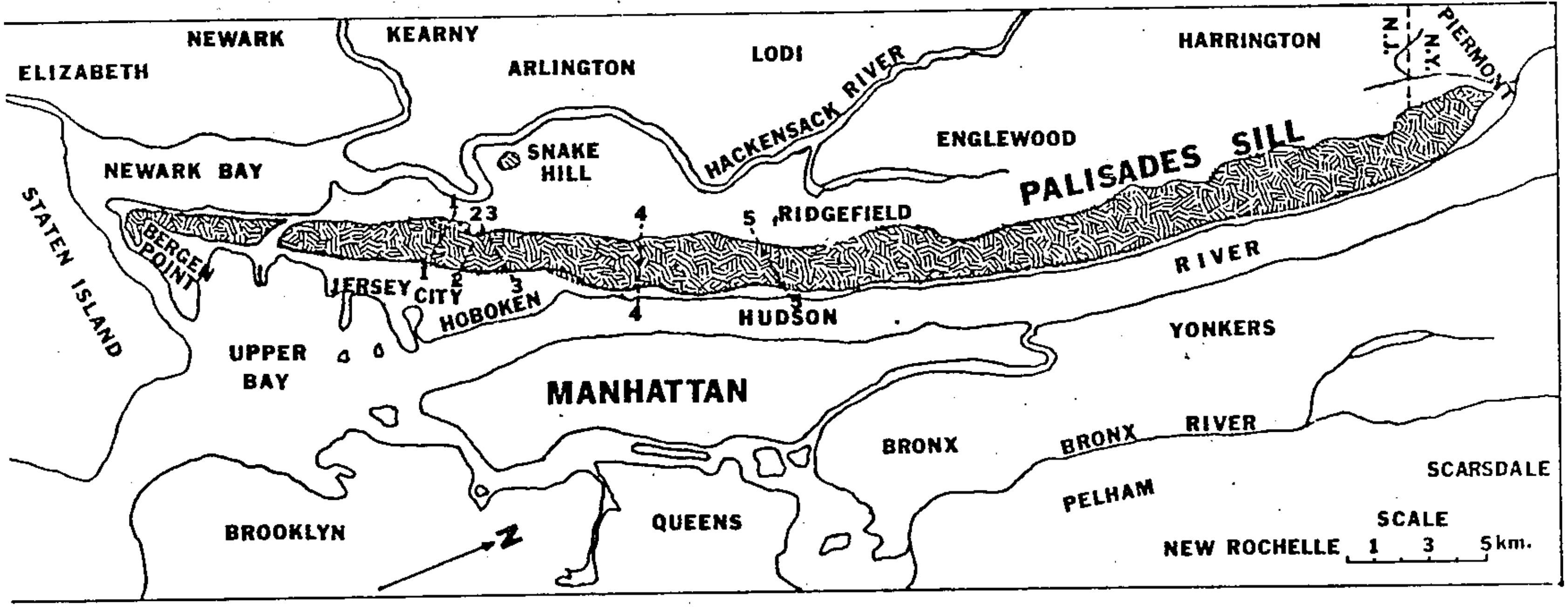

Figure 1. Location map for Bergen Hill modified from Manchester (1931). 1) Pennsylvania Railroad cut (1880s), Mt. Pleasant; 2) Bergen Archways (1906-1910) and the Erie Railroad tunnel (1857-1860), Jersey City; 3) Delaware, Lackawanna & Western tunnels (1874-1876 & ?), Jersey City; 4) New York, West Shore & Buffalo tunnel (1881), Weehawken; 5) New York, Susquehanna & Western Railroad tunnel (1880s?), Edgewater. The precise location of the New Jersey Railroad cut (1832-1838), Jersey City, is unknown.

intersecting streets forced over 400 meters of tunnel to be bored (Manchester, 1919). The following minerals were found at the Bergen Archways: stilbite, laumontite, gmelinite, analcime, natrolite, apophyllite, pectolite, datolite, quartz, calcite, pyrite, chalcopyrite, sphalerite, prehnite and "diabantite" (Manchester, 1919). Unfortunately, collecting at any of the Bergen Hill localities has not been possible since the 1920s. A list of species reported from the various Bergen Hill localities is given in Table 1.

GEOLOGY

Bergen Hill and the Watchung Mountains have been studied by geologists since the early 1800s. Bourne (1841) and Beck (1843) both refer to the rock of Bergen Hill as being "greenstone." The **American Geological Institute's Glossary of Geology (1966)** define greenstone as "an old field term applied to altered basic igneous rocks which owes their color to the presence of chlorite, hornblende, and epidote." While early investigators recognized the volcanic nature of the traprock, it wasn't until the 1850s and 1860s that it was termed basalt. Opinions differed as to whether New Jersey traprock was intrusive or extrusive. Davis (1883) summarized earlier views on the origin of the New Jersey traprock and cited evidence that the Watchung Mountains are predominately extrusive, whereas the Palisades Sill is intrusive.

Modern studies, especially Walker (1969) and Faust (1975), support Davis' conclusions regarding the origin of New Jersey traprock. Faust (1975) provides a detailed account of the geologic history of this region. He states that during the Late Triassic and continuing into the Early Jurassic period, the supercontinent of Pangaea (then containing all of the world's landmasses) began to split apart due to tensional stresses. Huge rifts or cracks developed in the crust of Pangaea and concomitant faulting created a series of grabens or basins. Basaltic lava extruded along these newly-created sedimentary basins pouring out and covering the accumulating sediments that eventually lithified into arkosic sandstones and siltstones, (i.e., the "brownstones"). Basaltic magma that was unable to reach the surface was injected between sedimentary layers, forming sills. The basins were later faulted (down-dropped) and their contents arched (folded). Subsequent erosion removed most of the sediments overlying the more resistant basaltic layers creating a series of three basalt ridges, the Watchung Mountains and the related Palisades Sill.

MINERALOGY

The occurrence of zeolites and associated secondary minerals at Bergen Hill and at the First Watchung Mountain localities are similar, but different. Both occurences are the product of low-temperature hydrothermal solutions derived from an almost-crystallized basaltic magma. After the primary minerals of the magma - plagioclase feldspar, pyroxenes, and amphiboles - had crystallized at relatively high temperatures (1000 to 800 degrees centigrade), the residual became enriched in volatile elements not utilized as essential components in the rock-forming silicates. In the case of Bergen Hill (and most other diabase intrusives), during the late stages of crystallization, contraction cracks or fractures developed in the cooling rock body. These fractures acted as conduits for the pent-up residuals enriched in such volatiles as boron, fluorine, and water. The residual fluids also reacted with the wall-rock of the veins, deriving the calcium, aluminum, and silica needed to crystallize zeolites. Datolite (with boron) and apophyllite (with fluorine and hydroxyl) were more abundant at Bergen Hill localities that at First Watchung quarries. This may be due to higher crystallization temperatures or to a greater abundance of dissolved volatiles in the intrusive Palisades Sill.

The zeolites and related secondary minerals do not occur in veins at the First Watchung Mountain localities, but rather they fill the interstices between basalt pillows. The basalt pillows developed when molten lava poured into lakes or water-saturated ground. Mineralization also occurred on a much smaller scale with small (usually 1 cm or less) vesicles which represent gas bubbles trapped in the congealing basaltic lava (Peters and Peters, 1978). As was the case with the Bergen Hill occurrence, the secondary zeolites crystallized from hydrothermal fluids containing residual volatiles plus cationic elements derived from wall-rock reactions. However, unlike Bergen Hill, these hydrothermal fluids mixed with saline pond waters rich in anionic sulfates and carbonates. The influx of saline pond waters into the hydrothermal system would explain why sulfates like anhydrite and glauberite crystallized in the First Watchung flows, but not at Bergen Hill where such an influx did not occur. Fenner (1910) and Schaller (1932) discuss the crystallization sequence at the First Watchung localities. Readers interested in further information concerning the paragenesis of the secondary minerals of the First Watchung quarries are directed to these references.

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CHAPTER SEVEN

SECONDARY MINERALIZATION OF PATERSON AREA TRAP-ROCK QUARRIES

by

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Secondary Mineralization of Paterson Area Trap Rock Quarries

Introduction

The basalts of the Watchung Mountains are mineralized with some spectacular occurrences of zeolites and other secondary minerals. Museums from around the world have chosen the prehnite, heulandite, datolite, and chabazite of the Paterson, New Jersey area as some of the most presentable examples for their displays. Crystal clusters of amethyst and calcite from the Paterson area are also quite presentable and are found throughout the area. Less showy but more pervasive secondary mineralization or alteration products (chlorite, albite, etc.) are the result of spilitization processes and a sequence of secondary events experienced by the basalt flows. A list of the primary and secondary minerals of the Paterson area is presented as table 1.

Locations

Secondary mineralization is not evenly distributed throughout the Watchung Basalts and instead tends to be highly localized within specific structures and influenced by the chemistry, particularly the volatiles, of the depositional environment. Variations in the susceptability of the basalt to low grade metamorphism also control the distribution of secondary mineralization.

In general the most prevasively altered portions of the Watchung Basalts are: 1) the second or upper flow unit of the Orange Mountain (First Watchung) Basalt, and 2) the amygdaloidal zones of the Hook Mountain (Third Watchung) Basalt. Pillowed subaqueous basalt occurrences such as much of the second flow of the Orange Mountain Basalt within the Paterson area, and amygdaloidal flow tops and bottoms are deeply altered and veined with secondary minerals. But occasional gas pockets lined with prehnite or other secondary minerals occur at unpredictable locations within the Watchung Basalts.

Access to sampling locations within the Paterson area is difficult. Virtually all of the basalt exposures are located where permission to collect

TABLE 1

Minerals Reported from Paterson Area Trap Rock Quarries
(and confirmed by Peters, 1984)

Actinolite
Albite
Analcime
Anhydrite
Apophyllite
Babingtonite
Barite
Bornite
Calcite
Chabazite
Chalcocite
Chalcopyrite
Chlorite
Chrysocolla
Covellite
Datolite
Epidote
Galena
Gmelinite
Goethite
Greenockite
Gypsum
Hematite
Heulandite
Hornblende
Kaolinite
Laumontite
Limonite
Malachite
Mesolite
Native Copper
Natrolite
Opal
Orpiment
Pectolite
Prehnite
Psilomelane
Pumpellyite
Pyrite
Pyrolusite
Quartz
Realgar
Silver
Sphalerite
Stevenlite
Stilbite
Stipnomelane
Talc
Thaumasite
Thomsonite

samples is legally required. The current operators or owners of each of the Paterson area quarries have adopted a strict policy of denying access largely as protection against legal responsibility for personal injury. Somehow, however, serious mineral collectors seem to overcome access problems and continue to find beautiful clusters of several secondary minerals. The most currently productive collecting sites are the operating quarries that have exposed the upper flow of the Orange Mountain Basalt, but good crystal clusters are also being recovered from each of the operating quarries and some of the abandoned quarries and road cuts.

Paterson area trap rock quarries:

1) Prospect Park (also known as Vandermade and Sowerbutt quarry)

The Prospect Park quarry has been in operation since 1901 (Peters, 1984). The upper part of the lower Orange Mountain flow is currently exposed. Large gas pockets lined with prehnite are common, as well as good amethyst vugs and veins. Until recently the pillow basalt of the upper flow was exposed on active quarry workings but the upper flow has been largely mined away. The pillow basalts of the Paterson area have been described in detail by Bello (1980) and Bello and Manspeizer (1983).

2) New Street Quarries (Upper and Lower on both sides of New Street in Paterson)

The Upper New Street Quarry was opened in 1893 and the Lower New Street was opened in 1900 (Mason (1960). Papke (1908) has reported spectacular calcite, apophyllite, heulandite, stilbite and chabazite mineralization at the Upper New Street Quarry and magnificent prehnite and natrolite at the Lower New Street Quarry.

Pillow basalts of the upper flow of the Orange Mountain Basalt are well displayed at the New Street Quarries. Secondary mineralization associated with the pillow structures is currently very abundant, and from a historical perspective the New Street Quarries and Prospect Quarry have probably yielded more museum quality prehnite and associated minerals than any of the Paterson area quarries or perhaps any quarry in the world. As of September 1984 the New Street Quarries are in the early stages of being landscaped for a garden apartment complex which will effectively terminate mineral collecting upon completion of the project.

The volcanic structures displayed at the New Street Quarries have been described by Manspeizer (1980).

3) Little Ferry Asphalt Quarry (formerly known as UBC, Union Building and Construction Quarry) located on the Clifton and West Paterson border north of Montclair State College Quarry.

The Little Ferry Quarry is one of the larger operating Paterson area quarries. The lower flow of the Orange Mountain Basalt is well displayed along quarry walls. Beautiful amethyst and prehnite are concentrated within some specific volcanic structures of the quarry, particularly the diapir-like structures described in the volcanic structure section of this chapter.

The most spectacular prehnite I have ever seen was collected August 1984 by Christopher Laskowich of Paterson in a huge gas pocket (approximately 1.5 meters long and 0.6 meters wide and 0.5 meters high located within the lower colonnade near the entablature (see figure 1-isolated pockets). Samples of this prehnite will be on display during the Friday evening (October 19) symposium session.

4) Great Notch Crushed Stone Quarry now abandoned and partly occupied by the Ward's Trucking Company Warehouse, in Little Falls, just northwest of Montclair State College quarry.

The lower flow of the Orange Mountain Basalt is well exposed at the abandoned Wards Trucking Quarry. The quarry is described in Chapter 10 of this guidebook.

5) Montclair State College Quarry (also known as Houdaille Upper Montclair Quarry) is wholly located within the easternmost portion of Little Falls, New Jersey.

The lower flow of the Orange Mountain Basalt is also well exposed at an abandoned quarry on the campus of the Montclair State College University.

The lower contact is particularly well exposed revealing a very amygdaloidal zone that is mineralized largely with calcite.

Structural Control of Mineralizing Fluids

Secondary mineralization is highly localized within some specific structures. Most of these structures are characteristics of specific subdivisions of the Orange Mountain Basalt. These subdivisions or rock units are as follows:

1. LOWER FLOW UNIT
    1. Lower Colonnade (lower chill zone)
    2. Entablature (central columnar basalt)
    3. Upper Colonnade (upper chill zone)

11. UPPER FLOW UNIT
    1. Massive and Columnar Basalt
    2. Subaqueous Flow Lobe
    3. Bedded Pillow Lava

Description of Rock Units

1. LOWER FLOW UNIT

1) Lower Colonnade (Fig. 1-a)

The lower colonnade commonly contains a white vertically jointed zone at the top of the layer. The white coating on the joint faces are due to thin coatings of prehnite, pectolite and datolite. Below this bleached zone is one of the following: (a) a dark zone characterized by interfingering near-horizontal joints which are usually wet with ground water, with no white coating and with few amygdules (Fig. 1-b); (b) pahoehoe or pillow-pahoehoe basalt (Fig. 1-c) as found at Montclair State College and Little Falls, N. J. This lava probably occupied shallow river channels and contains stilbite on pahoehoe ropy-stringy lava surfaces or among very amygdaloidal pillows; (c) an amygdaloidal zone with no pahoehoe or pillow lava (Fig. 1-e).

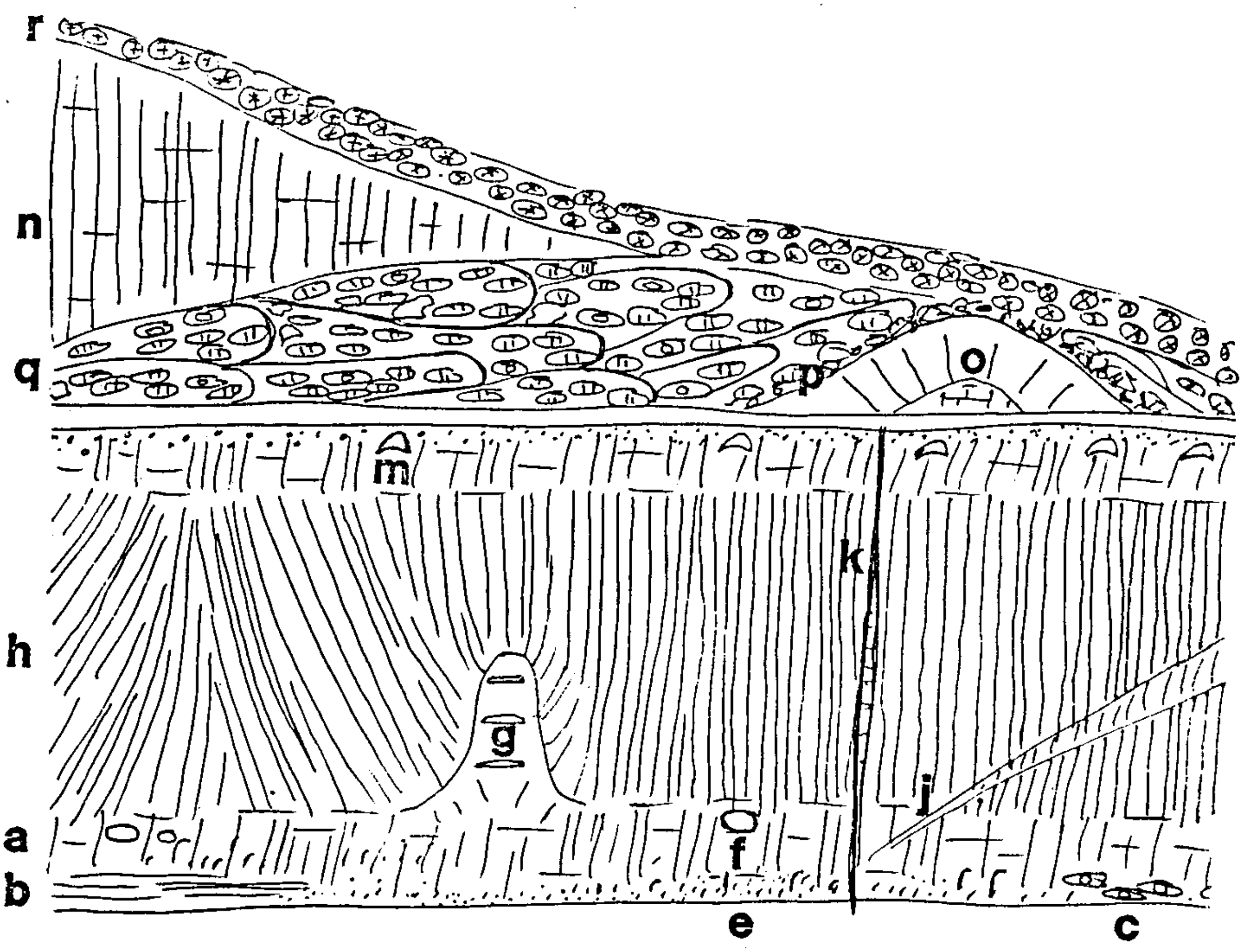

Fig. 1 Composite sketch of structures characteristic of Orange Mountain Basalt: a. white jointed lower colonnade, b. horizontal jointed lower colonnade, c. pahoehoe zone, e. amygdalar zone of lower colonnade, f. isolated gas pockets, g. diapir-like structure, h. entablature, j. low-angle faults, k. strike-slip faults, m. gas pockets of upper colonnade, n. massive and columnar basalt of upper flow unit, o. tumulus, p. vesicular and brecciated basalt layer, q. subaqueous flow lobe, r. bedded pillow lava.

The amygdaloidal zone is characterized by pockets of quartz, calcite, gypsum, anhydrite casts and agate. Many of the secondary minerals occur as pseudomorphs after large anhydrite, glauberite and calcite crystals as described in detail by Schaller (1932). Large anhydrite crystals have commonly disolved away leaving hollow casts lined most commonly with quartz or prehnite. Also included in the white jointed portion of the lower colonnade, just below the entablature, are isolated pockets (Fig. 1-f), commonly located within a small rise in the lower colonnade boundary.

Over 30 of these isolated pockets, typically lined with prehnite or datolite, but in combination with pectolite, chalcopyrite and rarely natrolite, have been found at the Little Ferry Asphalt and the Wards Trucking Quarry. They are also common throughout the Prospect Park and New Street Quarries. All occurred within the top meter of the lower colonnade or within noticeable rises of the colonnade into the entablature. Three hollow pockets were found approaching 2 meters in length, and more than one half meter in height and width. These pockets do not seem to be associated with any other structures.

Another structure associated with the lower colonnade but intruding into the entablature is a diapir-like structure (Fig. 1-g). There have been about a dozen of these structures, each resembling salt diapirs, observed within the Paterson area. Within them, there are layers of prehnite with altered basalt, perhaps a meter thick, separated by layers of unaltered basalt one to three meters thick. At least three of these horizontal layers occur within each diapir-like structure. Thin vertical veins of prehnite and chalcopyrite are also typically present, about one half cm thick. It is estimated that some of the diapirs attained a height of 15 meters or more, and were up to 10 meters in diameter.

Two diapir-like structures observed on the working face of quarry walls in 1981 and 1983 by Chris Laskovich contained zeolites in the lower layers, and quartz or agate in the lowest portion of the diapir. Massive fine grained basalt typical of the lower colonnade actually sweeps up alongside the diapir surface, and may also contain isolated pockets of prehnite and datolite. All diapirs found thus far have been in the Little Ferry Asphalt Quarry, with the exception of two recent diapirs, now destroyed, found in a quarry a few killometers to the south.

Isolated pockets are thought by Laskovich to be failed diapirs, that contained insufficient quantities of volatiles to allow them to rise any higher into the overlying entablature.

2) Entablature

The Webster's Ninth New Collegiate Dictionary (1984) definition of an entablature "a horizontal part in classical architecture that rests on the columns and consists of architrave, frieze, and cornice" is not a good description of the interior zone of the lower flow unit, but the term as used by Tomkeieff (1940) to describe the closely spaced curved columnar jointing of the Giants Causway does apply to some exposures of the interior zone. The interior zone or entablature (Fig. 1-h) is consistently coarser grained than the upper and lower colonnades and the joints are typically more closely spaced, but the columns of the entablature are not consistently curved.

Low-angle faults (Fig. 1-j) and strike-slip faults (Fig. 1-k) are common throughout the lower flow unit. The majority of the strike-slip faults trend N-S, the same trend as the large vertical joints. The low angle faulting took place before columnar jointing, and is commonly mineralized with prehnite. The strike-slip faulting took place after

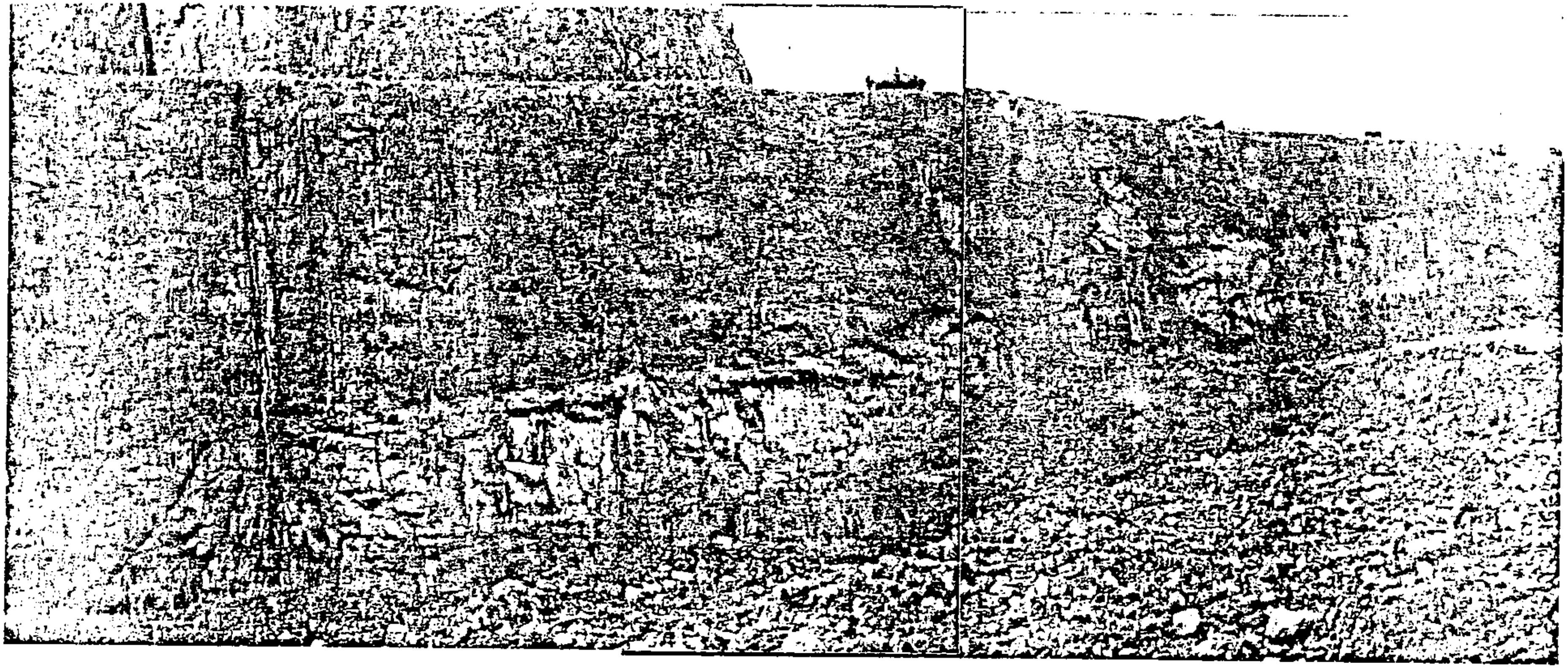

Fig. 2 Lower flow unit at Little Ferry Asphalt Quarry displaying a white jointed lower colonnade over a black horizontal jointed layer, with a strike slip fault on the left and a diapir-like structure on the right.

columnar jointing, and is mineralized with calcite, quartz, and minor datolite, hematite and sulphides deep within the entablature and lower colonnade. However, within the upper colonnade and the upper part of the entablature, strike slip faults may contain stilbite and heulandite, such as those of the route 80 road cut.

Columnar joint surfaces are typically barren of any secondary mineralization but are occasionally lined with manganese oxides, limonite, or encrustations of fine grained clay. The joint surfaces, however, facilitate ground water circulation that has locally chemically weathered the basalt.

There is one major normal fault in the Patterson area, the Clove Road Fault, which is easily traced from Clove Road, Little Falls to route 80, Patterson. The down block is to the east, with a throw of about 15 meters.

3). Upper Colonnade

This uppermost layer of the lower flow unit contains a frothy amygdalar crust on top, which was weathered before the upper flow unit resumed lava outpouring. Just below the crust are gas pockets, lined with prehnite, calcite, and chalcopyrite (Fig. 1-m) that are about 12 cm wide, and rarely up to one meter. These pockets have a flat bottom with a convex upward surface.

11. UPPER FLOW UNIT

1) Massive and columnar basalt

Some of the upper flow unit formed as massive and columnar basalt (Fig. 1-n) particularly as thick lenses close to the source of volcanic activity and as the central protion of flow channels or as the proximal end of flow lobes. These lenses display each of the three Tomkeieff structures more typical of the lower flow unit.

Occasionally where an advancing flow channel becomes clogged under hardened but still plastic basalt, the overlying crust will arch upward creating a structure known as a tumulus (Fig. 1-o) also known as a pressure dome or schollendome. If basalt splatters through cracks in the tumulus it may form a layer of vesicular and brecciated basalt (Fi. 1-p) such as that exposed over the tumulus of field trip Stop 2 (Chapter 10). Most tumuli in the Newark Basin are about 7 m high and 15 m wide (Manspeizer, 1980) and exhibit a well-defined lower layer characterized by vertical joints overlain by an upper layer characterized by radial joints.

2) Subaqueous Flow Lobes

Pillow basalts of the upper flow of the Orange Mountain Basalt exhibit two distinctly different modes: (1) subaqueous flow lobes and (2) bedded pillow lava.

The subaqueous flow lobes (Fig. 1-q) of the Orange Mountain Basalt overly the vesicular upper surface of the lower flow unit and are typically composed of three zones (Manspeizer, 1980): (1) a proximal zone of massive and columnar basalt, (2) a transitional zone of columnar basalt with isolated pillows, and (3) a distal steeply dipping zone containing ellipsoidal flow lobe pillows and pahoehoe toes. The pahoehoe toes as described by MacDonald (1953) are characterized by an ellipsoidal cross section with an aspect ratio of 3 or 4, concentric structures, and common lava tubes resulting in central cavities that in the Paterson area are typically mineralized with quartz and zeolites.

3) Bedded Pillow Lava

Layers consisting of bedded pillow lava (Fig. 1-r) commonly overlie subaqueous flow lobes in the Paterson area. Unlike the pahoehoe toes, the pillows of the bedded layers are not as vesicular or ellipsoidal, with an aspect ratio of less than 3, and exhibit radial structures. Secondary mineralization is typically abundant in the interstices between pillows.

Paragenesis

The first major examination of the secondary mineralization of the Paterson area trap rocks was written by Fenner (1910) who proposed three periods of alteration:

Period 1 (Boric Acid Period)

- stage 1 - albite, quartz, garnet, amphibole, hematite, sulfides.
- stage 11 - datolite, prehnite, pectolite, amphibole, hematite, sulfides.

Period 11 (Zeolite period)

- zeolites, amphibole, chlorite, hematite, sulfides.

Period 111 (Calcite period)

- thausmanite, calcite, gypsum, amphibole, chlorite, hematite, sulfides.

Schaller (1932) agreed with most of Fenners interpretation but organized his interpretation of the paragenesis into six overlaping periods:

Trap rock and basaltic glass period

- olivine, labradorite, diopside, magnetite, glass.

Saline period (from lake water)

- anhydrite, glaubernite, (probably some calcite)

Quartz period

- quartz, albite, babingtonite , amphibole, garnet, hematite, sulfides

Prehnite period

prehnite, pectolite, datolite, amphibole, hematite, sulfide.

Zeolite period

zeolites, chlorite, hematite, sulfides.

Calcite period

calcite, babingtonite, gypsum, thausmanite, amphibole, chlorite, hematite, sulfides

Both Fenner (1910) and Schaller (1932) meticulously describe paragenetic details that were based on solid criteria, particularly pseudomorphism. Their paragenetic sequence remains unchallanged and is endorced by several subsequent investigators (Mason, 1960; Peters and Peters, 1978; Sassen, 1978 Peters, 1984, and this study). Assuming that the Fenner - Schaller paragenesis is approximately accurate the next logical step seems to be the positioning of the paragenesis into the sequence of geologic events that may have effected the mineralogy of the Watchung Basalts. A sequence of eight such events are probably responsible for most of the Watchung mineralogical development:

1) Primary igneous crystallization

Samples of Watchung Basalt taken from massive, structureless portions of the basalt flows that have escaped most secondary alteration effects are composed of: plagioclase (labradorite - bytownite), pyroxene (augite, pigeonite, and minor clinopyroxene some of which may be mantle xenocrysts (Philpotts and Reinchenbach, 1983), a glass phase (or perhaps two partially devitrofied immiscible glass phases, Philpotts and Doyle, 1983); minor olivine (although unaltered olivine has not been found), ilmeno-magnetite, ilmenite, apatite, pyrite, chalcopyrite; and traces of biotite, amphibole, K-feldspar, and quartz found in the Preakness and Hook Mountain Basalts that may or may not be primary igneous phases.

2) Deuteric activity

Late stage igneous build-up of volatiles including water, sulphur, carbon, nitrogen, chlorine, and boron, led to the separation from the igneous magma of a very corrosive fluid phase (probably supercritical). Alteration of olivine to chlorite, some sericitization of plagioclase, precipitation of minor disseminated biotite, amphibole, K-Spar, quartz, and precipitation of chalcedony, and other minerals in gas vesicles was probably deuteric. Deuteric fluids precipitated veins of magnetite in the Jurassic diabase of Laurel Hill, New Jersey (Puffer and Peters, 1974) but similar magnetite veins have not been found in any of the Watchung Basalts. Copper, barium, and boron mineralization in joints, vugs, and adjacent sediments, however, particularly chalcocite, native copper, barite, and either datolite or its precursor may have been deuteric to hydrothermal precipitates. Fenner (1910) describes the Boric Acid Period as the first alteration period of his paragenetic sequence that includes datolite and copper sulfide precipitation. Recrystallization of any deuteric copper minerals to chalcocite and native copper, however, may not have taken place until diagenetic, metamorphic, or hydrothermal processes effected the rock.

3) Spilitization and contact metasomatism

Spilite as defined in the AGI Glossary (1977) is "an altered basalt characteristically amygdaloidal or vesicular, in which the feldspar has been albatized and is usually accompanied by chlorite, calcite, epidote, chalcedony, prehnite or other low-temperature hydrous crystallization products characteristic of a greenstone. Spilite often occurs as submarine lava flows and exhibits pillow structure."

It is clear that varying degrees of spilitization have effected the Watchung Basalts particularly wherever lava encountered saline lake waters. The upper or second flow of the Orange Mountain Basalt, for example, as exposed at the New Street Quarries and along Rt. 46 through West Paterson is pervasively spilitized but elsewhere, throughout most of the Watchungs spilitization effects are not apparent, probably because of extrusion onto a dry land surface.

Most of the albitization and chloritization of the effected Watchung

Basalts probably occured in response to reaction with saline waters. These alteration processes resulted in major geochemical changes that are typical of spilitization such as the effects described by Thompson (1973) for ocean floor basalts. Comparison of an analyzed sample of pillow basalt from the upper flow of the Orange Mountain Basalt at Prospect Park with an approximately unaltered sample from the Rt. 280 Orange Mountain roadcut (Table 2) indicates the degree of such alteration and is very similar to the ocean floor alteration described by Thompson (1973). The major changes are a decrease in SiO_2, MgO, CaO, and an increase in Fe_2O_3/FeO, Na_2O, K_2O, and H_2O.

The influx of sulfate into the lava from the saline lake waters was a major factor in the mineralization of amygdules and other open spaces in the basalts. Schaller (1932) suggests that most of the mineralization of Paterson area trap rock is pseudomorphic after anhydrite and glauberite mineralization and possibly calcite mineralization that precipitated during his "saline period". Anhydrite was apparently a major metasomatic entry into the basalts. Although most of the original anhydrite has leached away, evidence of its former presence is abundant. Schaller (1932) suggests that most of the mineralization in rectangular cavities (up to 5 by 3½") and in lamellar cavaties (up to 3 by 1/15") is pseudomorphic after anhydrite. He also describes common rhombic cavities (up to 3¼" across) that were originally filled with glauberite.

4) Diagenetic reactions

Burial of volcanic flows under layers of overlying sediment and subsequent extrusions have provided the temperature and pressure conditions condusive to development of diagenetic or sub-metamorphic reactions. Diagenetic reactions and subsequent metamorphic reactions, however, are not pervasive and are only locally developed wherever appropriate structural and chemical conditions were available. There is little evidence that any of the Paterson area trap rock assemblages equilibrated to prevailing diagenetic or metamorphic environments

Table 2

Alteration of Orange Mountain Basalt

	Fresh (1st-53b)	Altered (1st-11alt)
SiO_2	51.44	49.21
TiO_2	1.04	1.09
Al_2O_3	14.49	14.95
Fe_2O_3	1.32	5.05
FeO	9.20	7.02
MnO	0.16	0.17
MgO	8.17	6.12
CaO	10.40	6.22
Na_2O	2.09	3.45
K_2O	0.52	0.78
P_2O_5	0.13	0.10
H_2O^+	0.43	3.41
H_2O^-	0.24	0.89
Total	99.63	98.46

beyond very limited dimensions represented by individual veins or vugs.

Hay (1966) has described the diagenesis of some volcanic rich detritus that was subsequently subjected to low grade metamorphism similar to the Jurassic rocks of the Paterson area. Hay (1966) describes the devitrification of the glass phase and the precipitation of a mordinite, clinophilolite, heulandite, stilbite, analcite assemblage and an analcite, mordernite, clinophilonite, heulandite, albite assemblage as diagenetic mineralization. Although mordinite and clinophilolite has not been reported in the Paterson area analcite, heulandite, stilbite, albite, and quartz mineralization is widespread and probably represents diagenetic to early zeolite facies metamorphism. The stability field of heulandite (Fig. 3) based on Coombs and others (1959), and Turner (1968) plots within a lower temperature and pressure range than most other zeolites. Analcite is described by Schaller (1932) and Sassen (1978) as the first mineral to precipitate during the zeolite period of alteration and is followed by natrolite and heulandite.

5) Zeolite Facies Metamorphism

Burial of Orange Mountain lava under the combined thickness of the overlying Feltville Formation, Preakness Mountain Basalt, Towaco Formation, Hook Mountain Formation, Boonton Formation, and unknown thicknesses of Cretaceous sediments probably generated environmental conditions within the stability field of zeolite facies mineral assemblages.

Armstrong and Besancon (1970) suggest that zeolite facies metamorphism is widespread throughout the Newark Supergroup from Nova Scotia through New Jersey. The abundant laumontite observed by Heald (1956) in the New Haven Arkose of the Hartford Basin (underlying the Talcott Basalt) is cited as evidence that zeolite facies metamorphism has effected K-Ar dating of the basalts (Armstrong and Besancon, 1970).

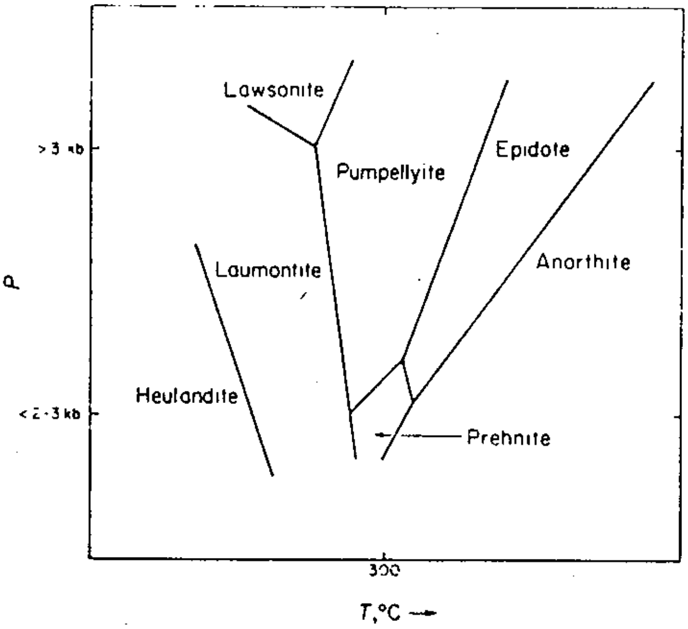

Fig. 3 Possible stability pattern of assemblages containing calcium-aluminum silicates in metamorphic rocks.
(after Coombs and others, 1959; and Turner, 1963)

Van Houten (1969) however, has shown that at least portions of the Passaic Formation, Lockatong Formation, and Stockton Formation stratigraphically below the Orange Mountain Basalt, have not re-equilibrated to zeolite facies conditions although he does describe abundant evidence of widespread diagenetic activity and groundwater induced alteration. The generally very low permeability of these sediments and contact metamorphosed hornfels may have retarded the development of zeolite facies metamorphism except locally where water and other volatiles were able to circulate. Porous and permeable zones within the Orange Mountain Basalt, particularly the vesicular and pillow basalt zones, were probably the most sensitive to zeolite facies metamorphism. Coombs and others (1959) have observed that zeolite facies assemblages are not found in Recent volcanic rocks and suggest that wherever such minerals are found in volcanic rocks they are probably the result of metamorphism.

Hay (1966) suggests a laumontite - albite assemblage as representative of zeolite facies metamorphism. Coombs (1959) suggests that the conversion of heulandite to laumontite is a characteristic of zeolite facies metamorphism according to the reaction: heulandite ↔ laumontite + 3 SiO_2 + H_2O. The upper boundary of laumontite stability is the reaction:

$$\text{laumontite} \leftrightarrow \text{anorthite} + 2\ \text{quartz} + 11\ H_2O$$

as described by Thompson (1970). Laumontite, albite, and quartz mineralization is also common throughout the Paterson area (Peters, 1984) and probably precipitated under zeolite facies conditions. Fiske and others (1963) however, suggest that laumonite precipitation is not particularly pressure sensitive and may also form at low pressures as a hydrothermal alteration product.

6) Transition to prehnite-pumpellyite facies metamorphism

Seki (1961) and Coombs (1960) each proposed a prehnite-pumpellyite facies to fill the gap between the zeolite facies and the greenschist facies. Seki (1961) describes the reaction:

30 anorthite + 18 diopside + 46 H_2O ↔ 12 pumpellyite + chlorite + 20 quartz as transitional to prehnite-pumpellyite facies metamorphism. Hyndman (1972) lists the following reactions as transitional to the prehnite-pumpellyite facies:

$$\text{laumontite} + \text{calcite} \leftrightarrow \text{prehnite} + SiO_2 + H_2O + CO_2$$

$$2\ \text{laumontite} \leftrightarrow \text{prehnite} + \text{pyrophyllite} + SiO_2 + 6\ H_2O$$

$$\text{laumontite} + \text{calcite} + \text{chlorite} \leftrightarrow \text{pumpellyite} + SiO_2 + H_2O + CO_2$$

$$\text{laumontite} + \text{prehnite} + \text{chlorite} \leftrightarrow \text{pumpellyite} + SiO_2 + H_2O$$

$$\text{chlorite} + 4\ \text{prehnite} + H_2O \leftrightarrow 2\ \text{pumpellyite} + SiO_2$$

Turner (1968) proposes the environmental boundary between the zeolite facies and the prehnite-pumpellyite facies as 6 - 10 km ($P_t = 3$ kb, $T = 100$ °C) and lists six diagnostic mineral assemblages as stable within the prehnite-pumpellyite facies:

1. Quartz, albite, prehnite, pumpellyite, chlorite, sphene in graywackies
2. Quartz, prehnite in veins.
3. Albite, prehnite, chlorite, sphene, epidote, quartz in spilitic lava
4. Prehnite, calcite in limestone
5. Quartz, albite, muscovite, pumpellyite, epidote, stilpnomoline in graywacke
6. Quartz, albite, muscovite, chlorite in schist.

Hay (1966) suggests the assemblage pumpellyite, prehnite, albite as representative of prehnite pumpellyite facies conditions.

Prehnite, pumpellyite, albite, epidote and chlorite are widespread constituents of altered Paterson area trap-rock (Peters, 1984) suggesting that at least some early stages of prehnite-pumpellyite facies metamorphism has effected the Paterson area basalts. Lausonite, however, has not been reported within the Paterson area and on the basis of Fig. 3 suggests pressures less than about 3 kb.

7) Hydrothermal alteration

There is little, if any, clear evidence that the Orange Mountain Basalt has been effected by hydrothermal solutions emanating from sources that post-date Orange Mountain vulcanism. In addition to any ground water driven out of underlying sediments because of compaction due to the weight of the thick layer of volcanic rock some hydrothermal water may have been driven out of these sediments by the thermal effects of the underlying Palisades Sill whose intrusion accompanied the extrusion of the Orange Mountain Basalt. Additional hydrothermal solutions may have included those that were generated during Preakness Mountain or Hook Mountain igneous activity or may have been squeezed out of underlying sediments during zeolite facies metamorphism but any clear hydrothermal effects have been masked by the effects of several kinds of fluids associated with Orange Mountain volcanism. As deuteric supercritical fluids condensed and cooled through the classic hydrothermal temperature range (500 - 100 °C) quartz, calcite, and presumably some other minerals were probably precipitated. In addition some of the chlorite - epidote - calcite alteration common throughout the Paterson area may be due to hydrothermal induced propylitic alteration.

8) Weathering

Although glacial activity has removed vertually all of the pre-Pleistocene weathered zone, some post-pleistocene weathering has effected the basalts of the Paterson area. A brown limonite rich encrustation characterizes the exposed basalt surfaces. Sassen (1978) lists gypsum, agate, opal, chrysocolla, malachite, brochantite, goethite, and pyrolusite as products of weathering within the Chimney Rock quarry. Presumably these and others have also developed as weathering products throughout the Paterson area. Chlorite after pyroxene, various clay minerals after feldspar, hematite after magnetite, and fiberous rutile after ilmenite have been identified in thin section and scanning electron microscopic examination of a weathered sample of Orange Mountain basalt from the Prospect Park quarry.

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CHAPTER EIGHT

COPPER MINERALIZATION OF THE NEWARK BASIN

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A useful review of the historic copper mining industry of New Jersey was written by Woodward (1944). Other important contributions were made by Kummel (1901), Lewis (1907) and Weed (1911). As indicated in Fig. 1 most of the mines are located near the base of the Orange Mountain Basalt and in the sediments of the Passaic Formation. A list of the 11 most important of these mines appears in Table 1.

The mines located at the base of the Orange Mountain Basalt were clearly the largest and most productive copper mines in New Jersey (Fig. 1). The Schuyler Mine was the only important mine not associated with Orange Mountain Basalt. The American Mine of Somerville, New Jersey was probably the most productive copper mine in New Jersey and was clearly the largest of those located at the base of the Orange Mountain Basalt. Morse (1812) reported that before 1754 as much as 1900 pounds of native copper had been found in the Somerville area and that before the Revolution many small tunnels had been made along the base of the First Watchung. During the Civil war production was near its peak but then gradually decreased and then came to an end by 1908. Kummel's (1901) State Geologist evaluation of the New Jersey copper industry was quite optimistic but by 1906 the entire copper industry was almost finished (Lewis (1907).

The ore minerals at the American Mine were chiefly chalcocite and native copper but common chysocolla, and malachite and minor bornite, cuprite, and tenorite, also occurs there in both siltstone and basalt.

The richest mineralization is in the siltstones within one meter of the contact. The siltstone of the contact zone has been altered into a purple porous hornfels-like rock that contains common bleached roughly spherical structures mineralized with native copper and chalcocite. Copper mineralization has also commonly filled cracks in the sediment and basalt

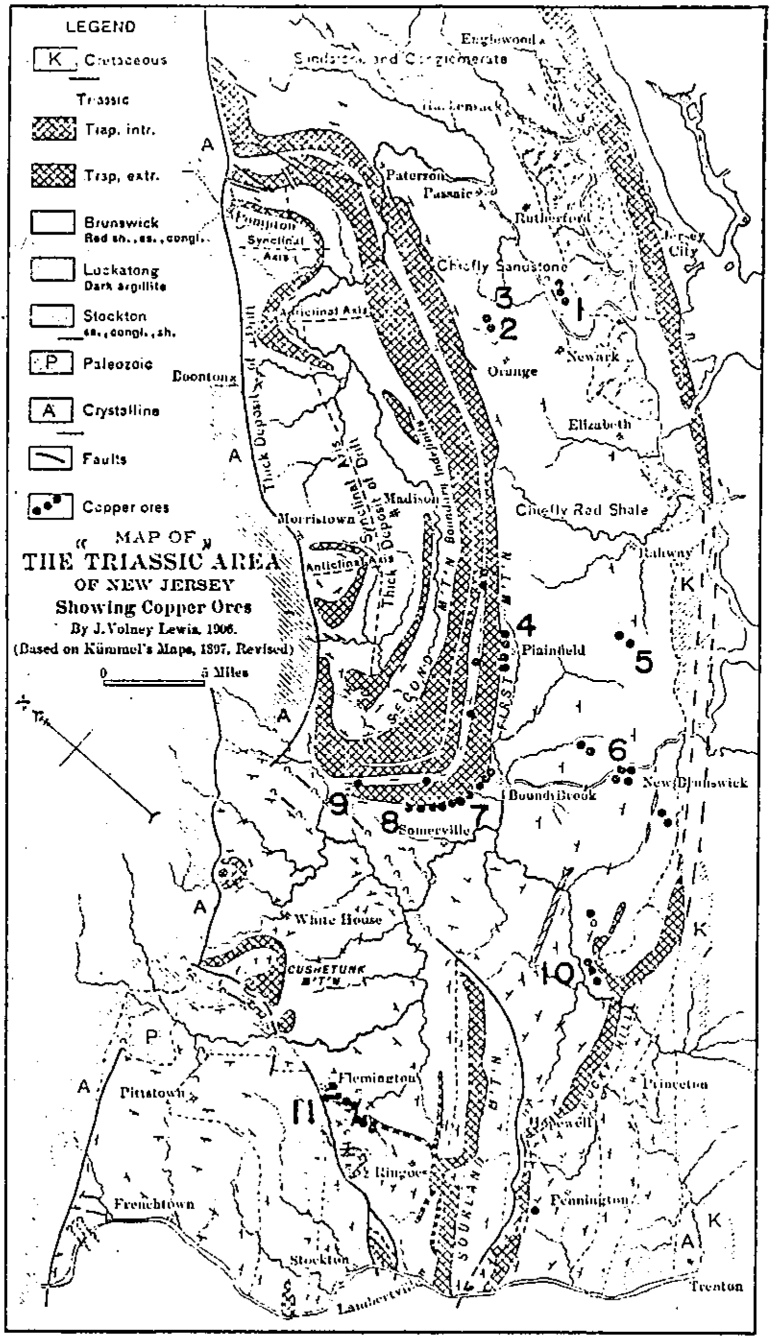

Fig. 1 Copper deposits of the Newark Basin (after Lewis, 1907), Deposits 1-12 are listed in Table

Table 1

Major Copper Deposits of the Newark Basin

1. The Schuyler Mine (also called the Belleville, Arlington, or Victoria mine). The ore occurs in sandstone lenses adjacent to thin branching diabase dikes and sills.
2. The Dod Mine was a small mine in beds of Passaic Formation in East Orange, New Jersey, first mined in the 1720s.
3. The Glen Ridge Mine was a small mine in Passaic Formation stratigraphically equilivent to the Dod Mine.
4. The Stoney Brook Mine (also called the Green Valley Prospect) was located slightly north-west of Plainfield, New Jersey where Stoney Brook cuts through the Orange Mountain Basalt.
5. The Menlo Park Mine (also called the Edison Mine) was located in Raritan Twp., New Jersey, first mined in 1784.
6. The New Brunswick Mine (also called the French Mine) was located in New Brunswick, New Jersey. It is an open pit mine located near a farm where Morse (1812) reports that prospectors found nuggets of native copper from five to thirty pounds in the soil.
7. The Chimney Rock Mine was located in Bound Brook, New Jersey. The ore occurs in Passaic Formation sediments and in fractures and amygdules in Orange Mountain Basalt. One native copper nugget weighing 76 pounds was found in 1927 when it jammed the crusher. The nugget is in the Rutgers university mineral collection.
8. The American Mine (also called the Bridgewater Mine) located north of Somerville, New Jersey was the most productive copper mine in New Jersey. The ore occurs chiefly as thin veins in altered Passaic Formation at the base of the Orange Mountain Basalt and as veins and amygdules in the basalt.
9. The Hoffman Mine was an important mine located near Pluckemin, New Jersey at the base of the Orange Mountain Basalt.
10. The Griggstown Mine (also called the Franklin Mine) was located in Franklin Twp., New Jersey and is probably the oldest copper mine in New Jersey. Production probably began some time before 1699 (Weiss, 1963).
11. The Flemington Mine was a small mine located in Flemington, New Jersey.

and commonly occurs in amygdules in the basalt. The major gangue minerals found with the ore are quartz, calcite, prehnite, and zeolites.

ORIGIN

Published speculation on the origin of the New Jersey copper deposits includes:

1) Weed (1902) who suggested that the ore was leached from the First Watchung and precipitated out of descending meteoric water that passed through the basalt.

2) Kemp (1907) suggested that chalcocite and native copper was derived from the breakdown of chalcopyrite disseminated through the trap rock or from copper released from the alteration of the pyroxenes of the basalt.

3) Lewis (1907) proposed the Palisades Sill as the source of copper for all the New Jersey deposits and suggested that ascending solutions precipitated copper under the overlying Orange Mountain Basalt.

Before evaluating these or other suggestions at least four observations seem pertenant:

1) All significant copper mineralization within the Newark Basin is closely associated with the Orange Mountain Basalt or to a lesser extent with its magmatic equilivent, the Palisades diabase. The copper mineralization associated with Orange Mountain and the absence of significant copper mineralization associated with the overlying flows, however, is inconsistent with the distribution of copper in the various flows. The copper content of relatively unaltered basalt samples from Orange Mountain, Preakness, and Hook Mountain Basalt averages 127, 81 and 188 ppm respectively.

The drop-off in copper content during Orange Mountain-Preakness fractionation probably took place deep within the magmatic source of both basalts. Gravitational separation of copper from Orange Mountain magma as an immiscible sulfide phase was suggested by Puffer and others (1982), and by Gotfried (1983).

The enriched copper content of the Hook Mountain basalt, however, is inconsistent with its lack of copper ore deposition, if the basalts were the source of copper mineralizing solutions. The absence of copper deposits associated with Hook Mountain may instead be related to its relatively shallow stratigraphic position and its tendency to be less effected by depth dependent regional metamorphism or diagenesis (more on this point to follow).

2) The copper content of the Palisades Sill increased during fractionation. After the Palisades magma was emplaced as a shallow sill, fractionation proceeded rapidly under a high thermal gradient, unlike the deep and slow Orange Mountain-Preakness Basalt fractionation. Walker (1969) has shown that the copper content of the Palisades Sill increased during fractionation from 110 ppm in the lower chill zone to 500 ppm in the upper layers.

Some copper escaped into a volatile phase and precipitated as chalcopyrite veins in some shallow diabase offshoots such as Laurel Hill, New Jersey and as chalcocite in overlying sediments such as those of the Schuyler Mine. Copper mineralization is not widespread throughout the sill but is associated with some very shallow (low pressure) branching dikes and sills. Palisades copper, therefore was released almost exclusively into either shallow intrusive or extrusive (Orange Mountain Basalt) occurrences.

3) The copper mineralization of the Newark Basin is vertually identical to its Hartford Basin counterpart. Most of the several copper deposits of the Hartford Basin, Connecticutt, are associated with the lower contact of the Talcott Basalt, the time stratigraphic and lithostratigraphic correlative of the Orange Mountain Basalt. Several of the copper mines of the Hartford Basin are described in a recent NEIGC Guidebook (Gray,

1982) including the well known Newgate Prison deposit. Orange Mountain type copper mineralization is, therefore, widespread and probably not due to any local set of unique events.

4) The copper mineralization of the Newark Basin is similar in many respects to the copper mineralization of the Keeweenawan Basalts of the Lake Superior area:

(a) Orange Mountain - Talcott Basalt and the Keeweenawan Basalt are each continental quartz tholeiites of very similar chemical composition and vertually identical copper content. The copper content of the Orange Mountain Basalt averages 127 ppm, the Talcott Basalt averages 125 ppm, and the Keeweenawan Basalt averages 125 ppm (18 samples, Prinz, 1967).

(b) The copper minerals and associated amygdule assemblages of the Orange Mountain-Talcott and Keeweenawan occurrences are approximently the same. The principal copper minerals of the Keeweenawan ores are native copper and chalcocite, and the amygdaloidal basalts are mineralized with quartz, calcite, prehnite, heulandite, analcime, laumontite, chlorite, pumpellyite, chabazite, and even datolite (Broderick, 1929; and Stoiber and Davidson, 1959). A distinct zonation of secondary minerals within the Keeweenawan flows has been mapped by Broderick (1929). The zones are arranged according to depth with the deep zone characterized by sericite, ankerite, chalcocite, and copper arsenides. The intermediate zone is characterized by increasing adularia, prehnite, analcime, and datolite The shallow zone is characterized by abundant datolite, prehnite, apophyllite, analcime, natrolite, stilbite, and laumontite. The secondary minerals of the Watchung Basalt most closely approximates the minerals of the shallow zone.

(c) The copper mineralization of each province is concentrated in amygdaloidal zones and adjacent sediment layers. Keeweenawan and Watchung Basalt copper occurs in finely disseminated form and in fractures. Fine grained copper is commonly intergrown with prehnite, and pumpellyite. Keeweenawan copper is rare where laumontite is prominent (Stoiber and Davidson, 1959).

(d) Both provinces have been effected by low grade regional metamorphism or at least diagenesis as implied by the zeolite facies mineral assemblages. Stoiber and Davidson (1959) suggest that copper was mobilized during metamorphism of the Keeweenawan Basalt. Pumpellyite alteration was suggested as particularly effective. Their suggestion agrees with DeVore (1955) who suggested that ore fluids commonly result from the mobility of metals under metamorphic conditions.

Coombs and others (1959) suggest that the mineralization zones described by Broderick (1929) are metamorphic zones analogous to those of the graywackes and semischists of southern New Zealand. Coombs and others (1959) also point out that zeolite mineral assemblages are rare or absent in the vesicles of newly cooled lavas. This is particularly true of continental lavas and probably applies to the Keeweenawan and Watchung Basalts. Some copper may have been mobilized during the deuteric stage of basalt cooling in a low pressure environment and precipitated as sulfides in the sediments adjacent to hypabyssal branching offshoots of the Palisades Sill and as sulfides in vesicules or late joints in disseminated Orange Mountain and Hook Mountain Basalts and to a lesser extent in the Preakness Basalt. Subsequent burial to levels consistent with zeolite facies mineral stability may have resulted in remobilization of primary, deuteric or hydrothermal copper and recrystallization of copper enriched glass, early alteration products, or copper enriched primary silicate phases into zeolite facies minerals and native copper and chalcocite. Massive diabase units and very impermeable portions of the Lackatong argilite, particularly portions that had been thermally metamorphosed would be unlikely to respond to the zeolite facies metamorphic processes. The porous and permeable and mineralogically complex amygdaloidal layers of the Orange Mountain Basalt, however were probably particularly susceptable to zeolite facies processes. Zeolite facies metamorphism, therefore may have been capable of mobilizing any loosely held copper in vugs or veins or altered basalt and induced reprecipitation as native copper or chalcocite. The copper of the overlying and relatively shallow Hook Mountain Basalt was apparently less effected by zeolite facies metamorphism. The copper of the Hook Mountain Basalt, therefore, remains locked into the silicate and sulfide phases disseminated throughout the basalt in less than ore grade concentrations.

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CHAPTER NINE

PALEOMAGNETISM IN THE CENTRAL NEWARK BASIN

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ABSTRACT

Three Newark Basin intrusives (Byram, Sourland Mountain, and Rocky Hill) and one lava flow (Sand Brook) were studied to determine the orientation of their magnetic vectors. This study yielded a reliable virtual geomagnetic pole for the Byram Diabase of 65.7 N and 104.1 E, and preliminary poles for the Sourland Mountain (63.4 N and 91.6 E), Rocky Hill (68.0 N and 91.6 E), and Sand Brook (60.4 N and 103.1 E) bodies. These results correlate well with the polar wandering curve of Harrison and Lindh (1982) for the period 180 to 200 million years, suggesting that these intrusions are about that age, and probably were emplaced at the same time as the Palisades Sill, Orange Mountain Basalt, and, possibly, the Preakness Basalt. A comparison of our data prior to the application of a structural correction which removed the dip of bedding, and after applying the correction demonstrate that these bodies were emplaced prior to the bulk of the tectonism in the region.

INTRODUCTION

Our study of the paleomagnetism of small intrusions in the Newark Basin began in 1982 when Hozik, Baxter, and Lorsbach collected some samples from the Byram Diabase in an attempt to document the rotation of small fault blocks within that intrusion

(Hozik and others, 1983). The present study extends that work to a continuation of the Byram Diabase in Pennsylvania, two other intrusions, and one lava flow in the basin.

We decided to extend the original study for the following reasons:

1. We are interested in the structure of the basin, and hoped that we could use the paleomagnetic results in conjunction with structural studies to document the rotation of fault blocks around steeply dipping axes.
2. There has been very little study of the paleomagnetism of the smaller intrusions in the basin.
3. We hoped to be able to place some constraints both on the timing of intrusive events relative to deformation, and on the correlation among intrusives.

So far we have collected and analyzed samples from three intrusions and one lava flow (See Figure 1 for locations). They are:

1. Byram Diabase and its extension into Pennsylvania
2. Rocky Hill Diabase
3. Sourland Mountain Diabase
4. Sand Brook Lava Flow

PREVIOUS WORK

The diabase intrusives in the Mésozoic basins are an obvious target for paleomagneticists, and there have been several studies of them. The earliest investigation we have come across was a study of Connecticut Valley igneous rocks by DuBois and others (1957). The first study of such rocks in the Newark Basin was by

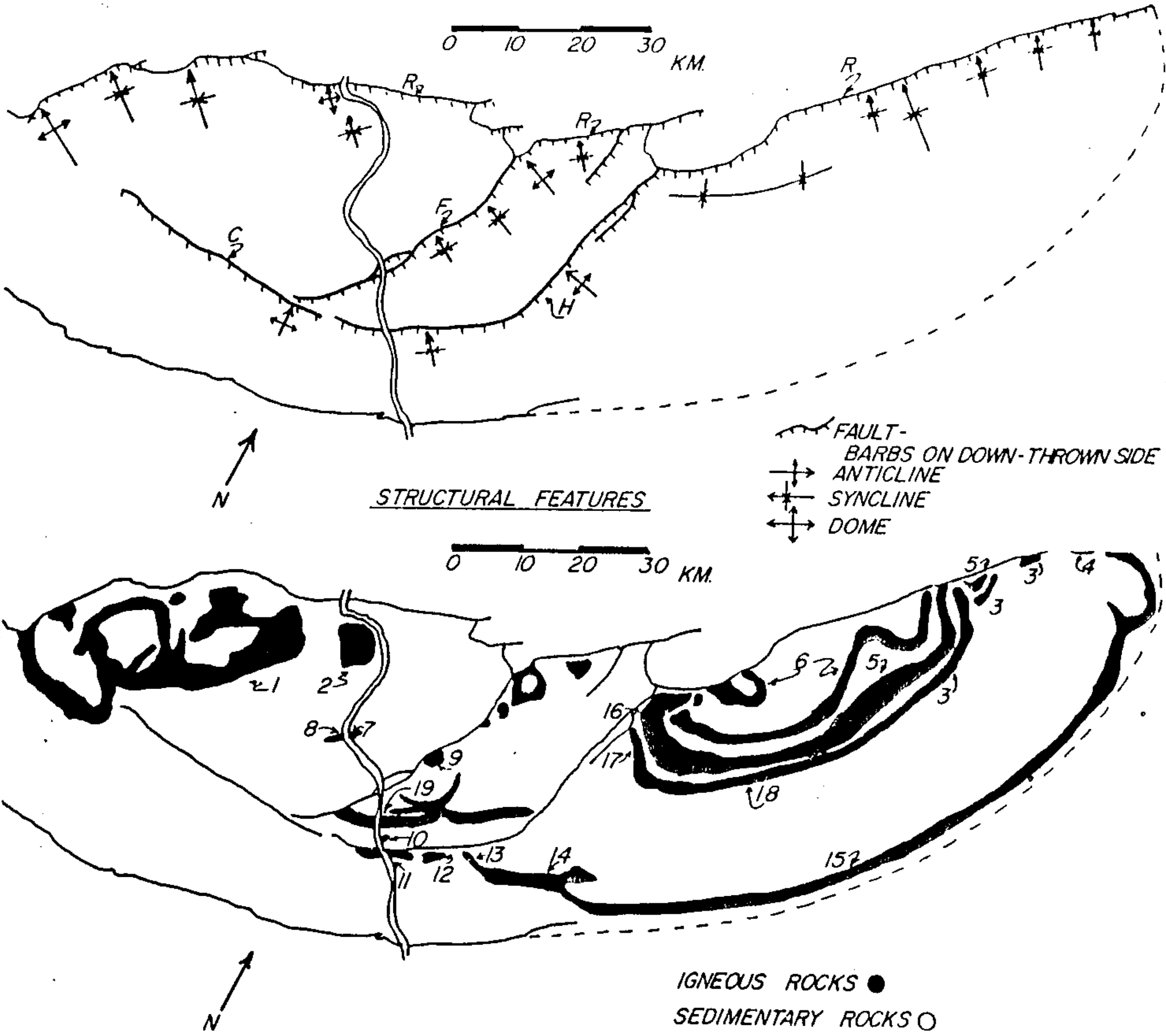


Figure 1: Locations of intrusions, extrusions, and major structural features in the Newark Basin. Igneous bodies: 1-Quakertown Diabase; 2-Coffman Hill Diabase; 3-Orange Mt. Basalt and some equivalents; 4-Ladentown Flow; 5-Preakness Basalt; 6-Hook Mt. Basalt; 7-Byram Diabase (NJ localities); 8-Byram Diabase (PA localities); 9-Sand Brook Flow; 10-Belle Mt. Diabase; 11-Baldpate Diabase; 12-Pennington Mt. Diabase; 13-Rocky Hill Diabase (this study); 14-Rocky Hill Diabase (Opdyke, 1961); 15-Palisades Sill; 16-Preakness Basalt (Opdyke, 1961); 17 & 18--Orange Mt. Basalt (Opdyke, 1961); 19-Sourland Mt. Diabase (this study and Opdyke, 1961). Structural features shown are: R-Ramapo Fault and its extensions; H-Hopewell Fault; F-Flemington Fault; and C-Chalfont Fault (modified from Olsen, 1980).

Opdyke (1961). Although there were many subsequent paleomagnetic studies of Mesozoic igneous rocks on the east coast, most of them did not examine Newark Basin rocks (Bowker, 1960; Irving and Banks, 1961; DeBoer, 1967, 1968; Beck, 1965, 1972; Smith, 1976; DeBoer and Snider, 1979; Smith and Noltimier, 1979; and Dooley and Smith, 1982).

With the exception of Beck (1972) who worked only as far east as the Delaware River, most of the work in the Newark Basin has been directed at the lava flows of the Watchung Mountains. Opdyke (1961) reported on 13 samples from 5 sites. McIntosh, Hargraves, and West (in press) analyzed 158 specimens from 120 samples collected at 18 sites. The largest and most famous intrusion in New Jersey, the Palisades Sill, is represented by 3 samples (12 specimens) from one site reported by Opdyke (1961)! In fact, Opdyke's (1961) study of the magnetization in the sedimentary rocks and some of the intrusions represents most of the published paleomagnetic work in the New Jersey portion of the Newark Basin. Hargraves and Young (1969) conducted an interesting study on the source of remanent magnetization in the Lambertville (Sourland Mountain) Diabase, but never published data on the orientation of the magnetic vector. McIntosh, Hargraves, and West have a major study of the magnetization in the sediments in press which includes data from the Orange Mountain, Preakness, and Hook Mountain Basalts and the Palisades Sill. Data from all of these studies are presented in Table 2.

Table 2: Virtual Magnetic Pole Positions From Triassic Basins

UNIT	N	K	A	D	I	LAT	LONG	SOURCE
Orange Mt.	12	85		3	34	67.9	97.8	Opdyke (1961)*
	8	50		3	23	61.2	99.4	Opdyke (1961)*
	31	251	4.8	359.8	29.9	65.2	106.1	H, M, W **
Preakness	10	157		18	25	58.3	70.5	Opdyke (1961)*
	8	350		19	14	52.5	73.4	Opdyke (1961)*
	55	21	11.5	13.9	27.9	61.4	76.4	H, M, W **
Hook Mt.	12	52		10	31	64.3	83.1	Opdyke (1961)*
	18	55	12.5	4.0	27.1	63.3	97.0	H, M, W **
Haycock	13	21		360	11	55.1	104.5	Opdyke (1961)*
Baldpate	12	85		4	36	69.4	94.4	Opdyke (1961)*
Rocky Hill	12	27		354	25	66.3	109.8	Opdyke (1961)*
	8	**27**	**10.9**	**5.3**	**34.4**	**68.0**	**91.6**	**This Study**
Belle Mt.	12	31		339	31	60.2	148.8	Opdyke (1961)*
Sourland Mt.	8	88		358	31	66.3	109.8	Opdyke (1961)*
	3	**109**	**11.9**	**11.5**	**29.5**	**63.4**	**91.6**	**This Study**
Palisades	12	73		4	30	65.0	96.9	Opdyke (1961)*
Byram	**32**	**94**	**2.6**	**0.2**	**29.8**	**65.7**	**104.1**	**This Study**
Sand Brook	**9**	**54**	**7.0**	**4.0**	**47.0**	**60.4**	**103.1**	**This Study**
Pa Intrusives	78	118	3.2			62.0	104.5	Beck (1972)
NE N Am (190)	72	56	2.3			63.0	83.2	Smith & Noltimier (1979
NE N Am (175)	156	92	1.4			65.3	103.2	Smith & Noltimier (1979)
Newark Group						63	108	Opdyke (1961)
Gettysburg	7	54	8.3			65.0	103.0	Beck (1972)
Yorkhaven	37	237	1.5			61.0	105.5	Beck (1972)
Birdsboro	26	85	3.2			62.0	105.5	Beck (1972)
Quakertown	9	143	4.3			66.0	105.0	Beck (1972)
Mass Lavas						55	88	Irving & Banks (1961)

* Lat and long of pole calculated for this study from data presented in Opdyke (1961)

** Lat and long of pole calculated for this study from data presented in McIntosh, Hargraves, and West (in press)

Table 1: Paleomagnetic data from this study.

		UNCORRECTED FOR STRUCTURE						CORRECTED FOR STRUCTURE					
						VGP						VGP	
Sample	N	K	A	D	I	LAT	LONG	K	A	D	I	LAT	LONG
By 1	6	121	6.1	12.6	40.0	69.4	70.1	110	6.4	3.0	33.0	67.5	97.5
By 2	4	586	3.8	13.8	42.2	70.2	65.0	392	4.6	3.5	35.0	68.8	95.8
By 4	7	257	3.7	-1.4	37.9	70.8	108.9	245	3.8	-8.7	28.0	63.5	124.1
By 5	5	1188	2.2	6.5	37.8	70.0	87.0	432	3.7	-1.0	29.0	65.2	107.3
By 8	3	181	9.0	10.0	41.0	70.9	75.1	235	9.0	2.8	28.0	65.7	99.1
By 1-8	25	136	2.5	7.2	39.6	71.1	83.5	121	2.6	-1.0	30.8	66.3	107.1
By 23	3	390	6.0	5.1	33.7	67.6	92.1	446	6.0	0.7	20.4	60.3	103.6
By 24	1	-	-	4.0	35.0	68.8	115.4	-	-	-8.0	23.0	60.9	121.2
By 25	3	428	6.0	21.7	42.1	65.7	49.8	621	5.0	13.3	32.3	64.5	74.2
By-Pa	7	67	7.4	10.5	37.8	69.0	76.8	61	7.6	4.5	26.0	63.2	95.2
All By	**32**	**113**	**2.4**	**8.0**	**39.2**	**70.7**	**81.8**	**94**	**2.6**	**0.2**	**29.8**	**65.7**	**104.1**
RH 14	3	68	15.0	-2.6	43.9	75.1	114.4	78	14.0	4.4	23.4	61.6	96.1
RH 16	5	26	15.2	-7.4	61.4	84.1	-140.6	27	15.0	6.0	41.0	72.4	86.7
All RH	**8**	**26**	**11.1**	**-5.1**	**54.8**	**83.6**	**145.3**	**27**	**10.9**	**5.3**	**34.4**	**68.0**	**91.6**
S 7	**3**	**121**	**11.0**	**17.5**	**37.8**	**65.7**	**62.0**	**108**	**11.9**	**11.5**	**29.5**	**63.4**	**79.6**
SB 17	3	104	12.0	-6.1	45.7	75.8	127.8	142	10.4	-3.7	47.9	78.1	122.3
SB 18	3	156	10.0	6.9	50.8	79.4	71.2	220	17.0	7.2	53.1	81.1	62.4
SB 19	3	25	25.0	1.8	40.4	72.5	99.6	25	25.0	3.9	42.0	73.5	92.4
ALL SB	**9**	**139**	**11.0**	**0.7**	**46.0**	**76.7**	**102.4**	**54**	**7.0**	**4.0**	**47.0**	**60.4**	**103.1**

By = Byram Diabase--Samples 1, 2, 4, 5, & 8 collected in New Jersey; Samples 23, 24, & 25 collected in Pennsylvania; Their mean is designated as By-Pa.

RH = Rocky Hill Diabase

S = Sourland Mountain Diabase (Lambertville Diabase)

SB = Sand Brook Lava Flow

N = Number of specimens
K = Fisherian precision parameter
A = Radius of cone of 95% confidence
D = Declination
I = Inclination
VGP = Virtual geomagnetic pole

REGIONAL GEOLOGIC SETTING

The Newark Basin is a structural basin approximately 200 km long and 60 km wide, beginning in Rockland County, New York and trending southwesterly across the state of New Jersey to near Lancaster, Pennsylvania. The basin is interpreted to be a rift valley which formed in response to the stresses responsible for the opening of the modern Atlantic Ocean, and into which 5 to 6 km of continental sediments were deposited (Van Houten, 1969, 1980; Manspeizer, 1980). Lava flows, interpreted to be of Lower Jurassic age (Hettangian) are interbedded with the sediments in the upper part of the sequence (Olsen, 1980). Seidemann and others (1984) report ages of 191 +/- 8 my for the Orange Mountain Basalt, 194 +/- 4 my for the Preakness Basalt, and 185 +/- 4 my for the Hook Mountain Basalt.

The sediments interbedded with the lava flows have also been intruded by a series of dikes and sills, the largest of which is the Palisades Sill. Conventional wisdom dictates that all of the intrusions are of roughly the same age, which is approximately 190-200 million years (Erickson and Kulp, 1961; Armstrong and Besancon, 1970; Dallmeyer, 1975). If correct, this means that the intrusions are also approximately the same age as the Orange Mountain Basalt and possibly the Preakness Basalt.

Since many of the lava flows and intrusions are cut by faults, at least some of the deformation in the basin occurred after the igneous activity. Faill (1973) has suggested that the bulk of the tectonic activity in the basin occurred after the deposition of most of the sediments. In marked contrast with this idea are observations of Ratcliffe (1980, p. 292) that raise the possibility that "faulting and folding of Triassic strata preceded eruption of the lava flows at about 193 m.y. or earliest Jurassic time." Resolution of the timing and magnitude of tectonic activity is crucial to the development of an accurate model of the formation of the basin.

METHODOLOGY

The typical habit of Mesozoic diabases is as topographic highs covered by highly weathered, well-rounded boulders. Large natural outcrops are rare, and they usually are severely altered by chemical weathering. As a consequence, most of our samples came from artificial exposures such as road cuts and quarries. Our few experiments with large, natural exposures have been disappointing. In the field, we collected oriented blocks from locations as widely spaced as the site would allow. At many sites it was possible to collect only one block. Orientation of the blocks was accomplished by measuring the strike and dip of two, non-parallel faces, marking a strike line and dip direction on each face with an indelible marker, and recording all the information in a field notebook. In addition, structural information was recorded from nearby outcrops showing the attitude of bedding.

In the lab, a minimum of three cores were drilled from each block by using a diamond drill bit. Cores were marked with the sample number and orientation information, and cut into 2.5 mm segments for spinning. All magnetic measurements were carried out at the University of Massachusetts in Laurie L. Brown's

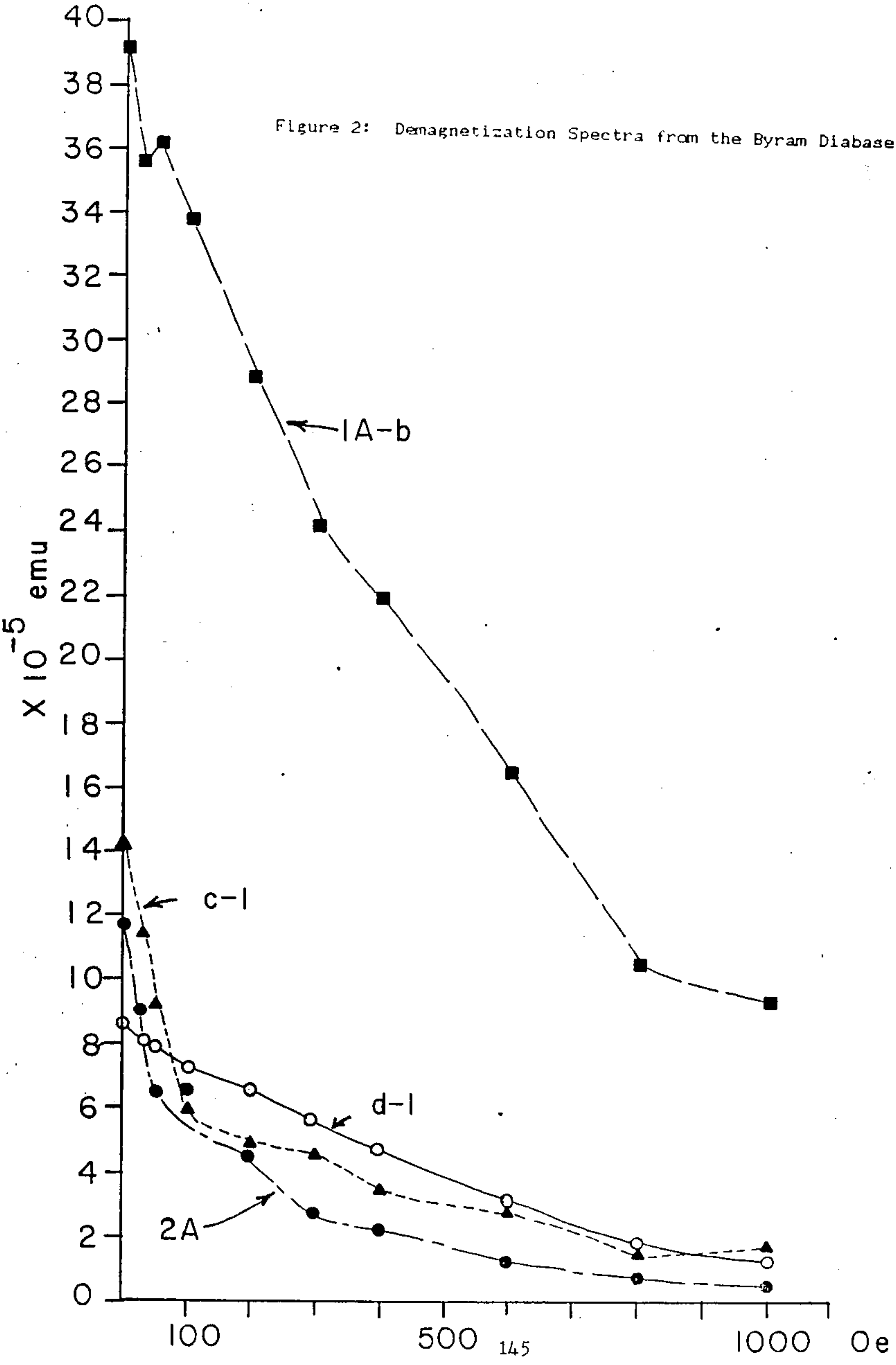


Figure 2: Demagnetization Spectra from the Byram Diabase

laboratory. Spinning was done on a Molespin Minispin 1 spinner magnetometer. Demagnetization was carried out on a Schoenstedt Alternating Field Demagnetizer.

All samples were spun to determine natural remanent magnetization. Pilot samples from each site were selected and demagnetized in steps in fields of 25, 50, 100, 200, 400, 600, 800, and 1000 Oe. The magnitude and orientation of the magnetic vector was measured after each step in the demagnetization sequence. Typical demagnetization curves are shown in figure 2.

Zijderveld Diagrams (Zijderveld, 1967) are presented for 4 representative samples in Figure 3. The plot presented for the Byram sample is typical of that intrusive in that the direction of magnetization does not become stable until demagnetization at the 400 Oe level. This is a significantly higher level than is used. is most studies (150 Oe is typical). The simplest explanation for this difference is that these rocks have acquired a much more stable secondary remanent magnetization than those studied by others, and hence a higher alternating field is necessary to remove it. DeBoer and Snider (1978) raise the possibility that samples which have unstable remanent magnetizations which are not removed until demagnetization at levels in excess of 350 Oe. have acquired an unusually stable viscous remanent magnetization. They suggest that the thermal remanent magnetization is removed at lower demagnetization levels leaving. the viscous component to be removed at higher levels. We are currently exploring this question, although our observations to date do not appear to support this interpretation. At present it is clear that some of the samples studied by us did not exhibit stable remanence at

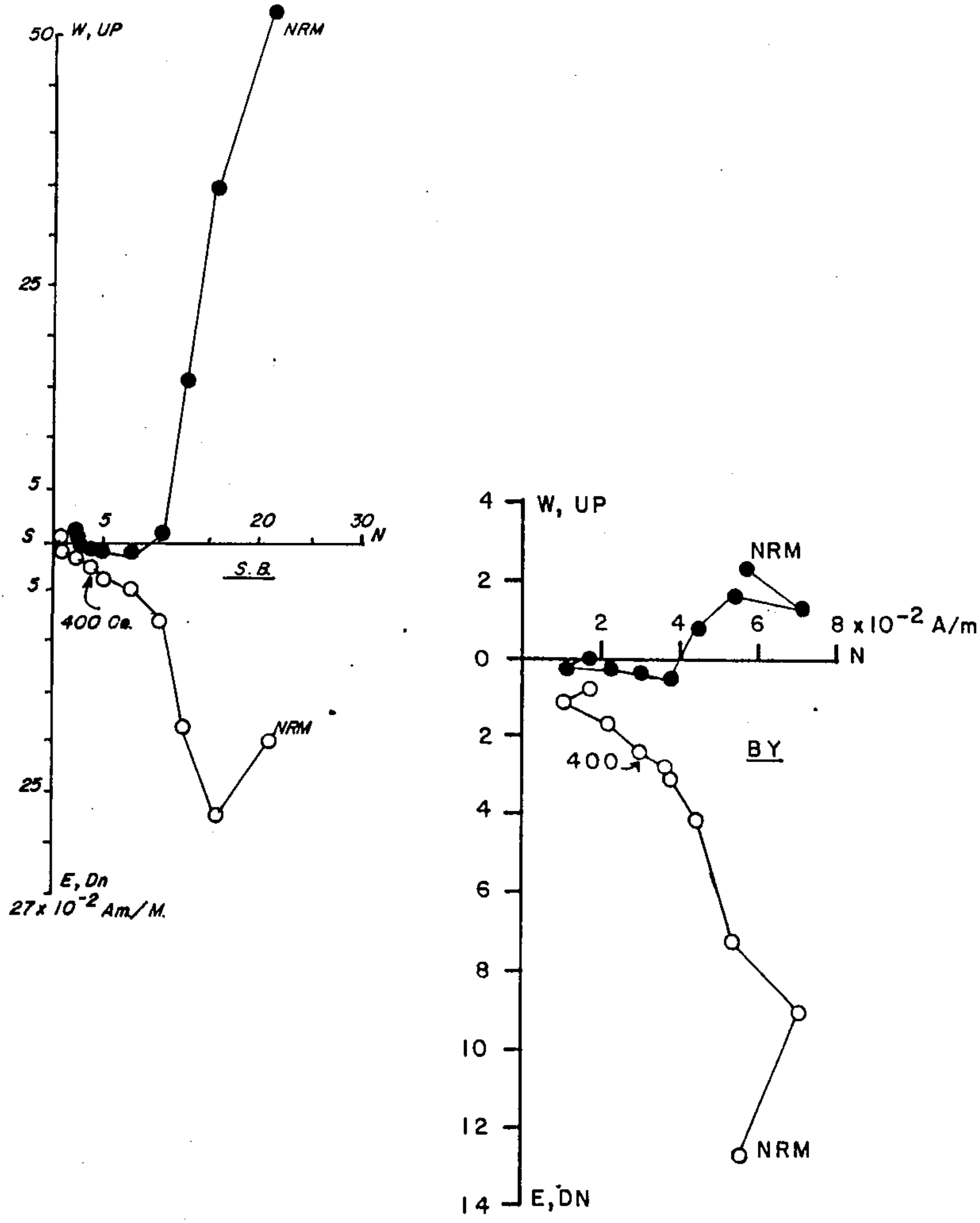


Figure 3a: Zijderveld diagrams from the Byram Diabase Sill and the Sand Brook Lava Flow.

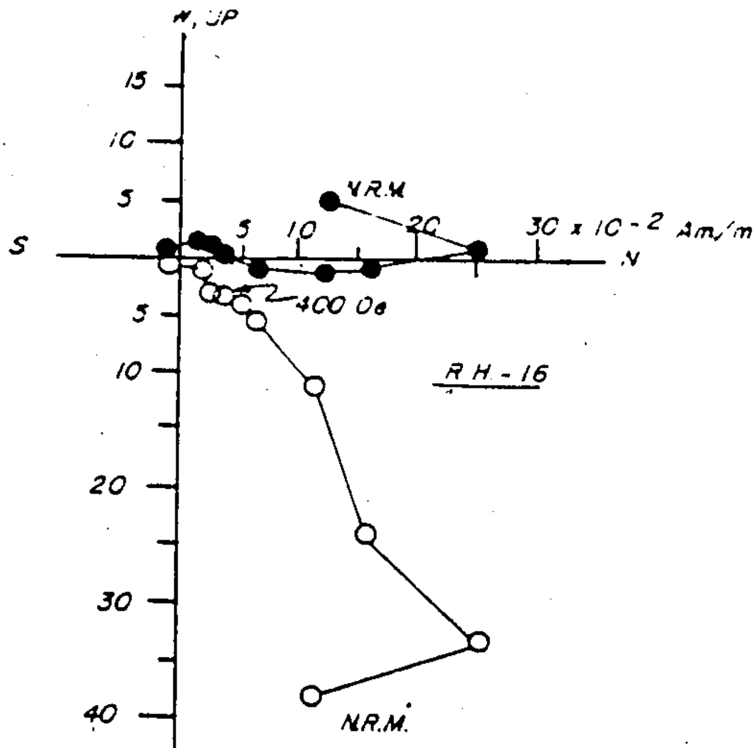
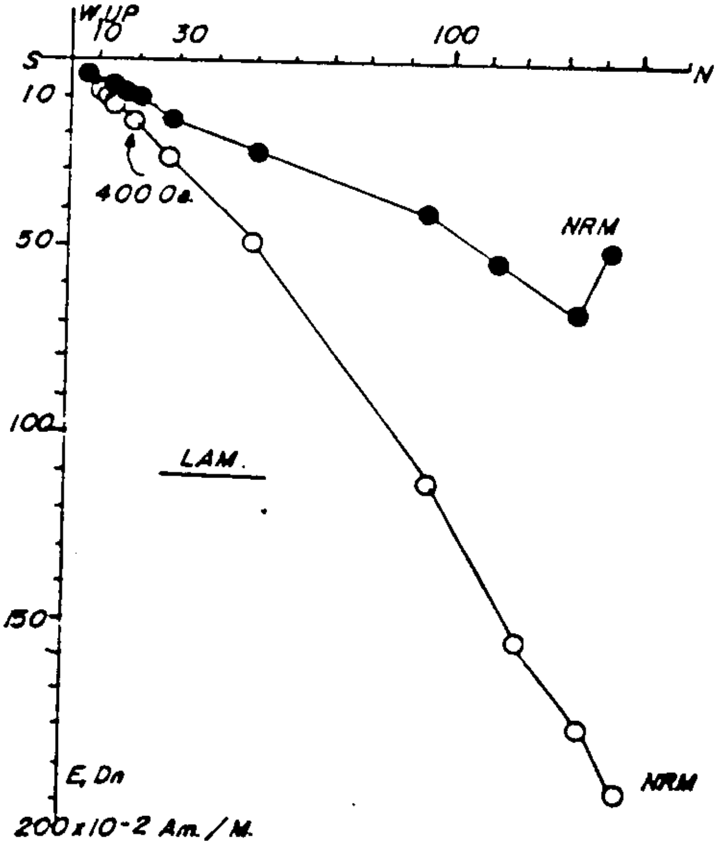


Figure 3b: Zijderveld diagrams from the Rocky Hill and Sourland Mt. Diabases.



demagnetization levels much below 400 Oe. As a result of the analysis of the Zijderveld Diagrams, all samples were demagnetized to 400 Oe. The improvement in the consistency of orientations resulting from demagnetization to 400 Oe is shown in figure 4 for the rocks we studied.

Two samples of the Byram Diabase were rejected. One was rejected because the calculated magnetic vector was nearly 180 degrees away from the mean of the rest of the samples, and we concluded that we had sampled a boulder. A second was rejected because it yielded internally inconsistent results suggesting orientation problems with the cores. One sample from Rocky Hill was rejected because the results were statistically extremely poor and the mean from the block was markedly at variance with the means from the other blocks.

Measurements from individual cores were combined to compute means for each block. Measurements from individual cores (not means from each block) were combined to compute mean directions for each intrusion. These data are presented in Table 1. Structural correction was accomplished by rotating the magnetic vector around the strike of bedding in nearby rocks until the dip of the beds was zero. Means from each block, and means from each intrusion were used separately to compute virtual geomagnetic pole positions.

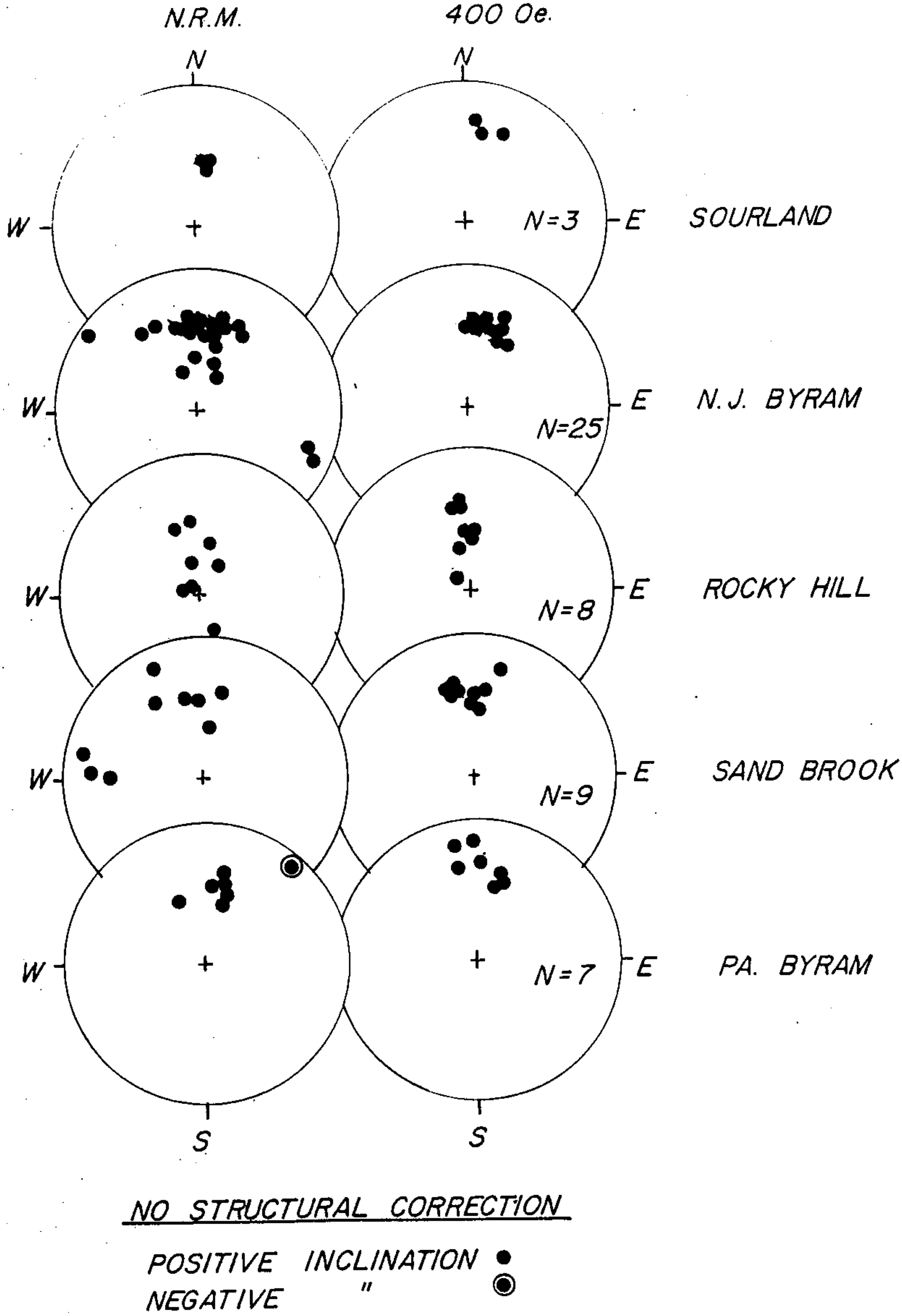


Figure 4: The orientations of the magnetic vectors for each sample prior to cleaning by Alternating Field Demagnetization (NRM) and after cleaning (400 Oe).

RESULTS

MAGNETIC VECTORS

The results of the samples analyzed to date are presented in Table 1. Essentially, we have good results on one intrusion and preliminary results on two other intrusions and one lava flow.

Virtual geomagnetic poles represent the position of the pole required to produce the measured declination and inclination of the magnetic vector at the sampling site. Ideally, virtual geomagnetic poles from all rocks magnetized at the same time would coincide. This is in marked contrast to the magnetic vectors for individual intrusions of the same age which will vary as a function of geographic position of the body. To compensate for this variation, virtual geomagnetic poles were calculated for all of our bodies, and used in all subsequent comparisons. In addition to the declination of the magnetic vector for each sample, Table 1 lists the latitude and longitude of the virtual geomagnetic pole for each block, rotated and unrotated, and the mean pole for each intrusion. Table 2 shows the latitude and longitude of all the virtual geomagnetic poles from our work, important poles from previous workers, as well as the source of the information. Table 3 lists some frequently cited Mesozoic pole positions.

BYRAM

Our best, and most complete data come from the Byram Diabase, a small intrusive along the Delaware River in New Jersey and Pennsylvania (See Figure 1). Data on 32 samples from 8 sites yield very consistent results and correspond to a virtual geomagnetic pole at 65.7 N and 104.1 E. This pole very nearly coincides with the 175 m.y. pole of Smith and Noltimier (1979) at 65.3 N and 103.2 E, and led us to speculate (Hozik and others, 1983) that the Byram Diabase, and perhaps other diabase intrusions in the region might be of that age (See Figure 5).

ROCKY HILL

Rocky Hill is the southernmost intrusion sampled, and crops out just north of Princeton (See Figure 1). Three samples were collected at each of three sites in one quarry. Results from one of the sites were discarded because they were not statistically consistent. The agreement of our pole with the results of Opdyke (1961), obtained on samples collected at another locality, is not impressive. The mean virtual magnetic pole determined from our analysis as 68.0 N and 91.6 E plots equidistant from Smith and Noltimier's (1979) 190 and 175 my poles (See Figure 5). It does plot very close to the polar wander path proposed by Harrison and Lindh (1982), approximately half way between their 180 and 190 my poles. The pole calculated from Opdyke's (1961) data plots away from the bulk of the Newark Basin magnetic data (See Figure 7).

SOURLAND MOUNTAIN

The Sourland Mountain Sill is a large, apparently continuous sill which crops out at the town of Lambertville along the Delaware River, and extends northeastward until it is truncated by the Hopewell Fault (See Figure 1). Three samples were collected at one site and yielded acceptable results, which are not in particularly good agreement with previous work of Opdyke (1961).

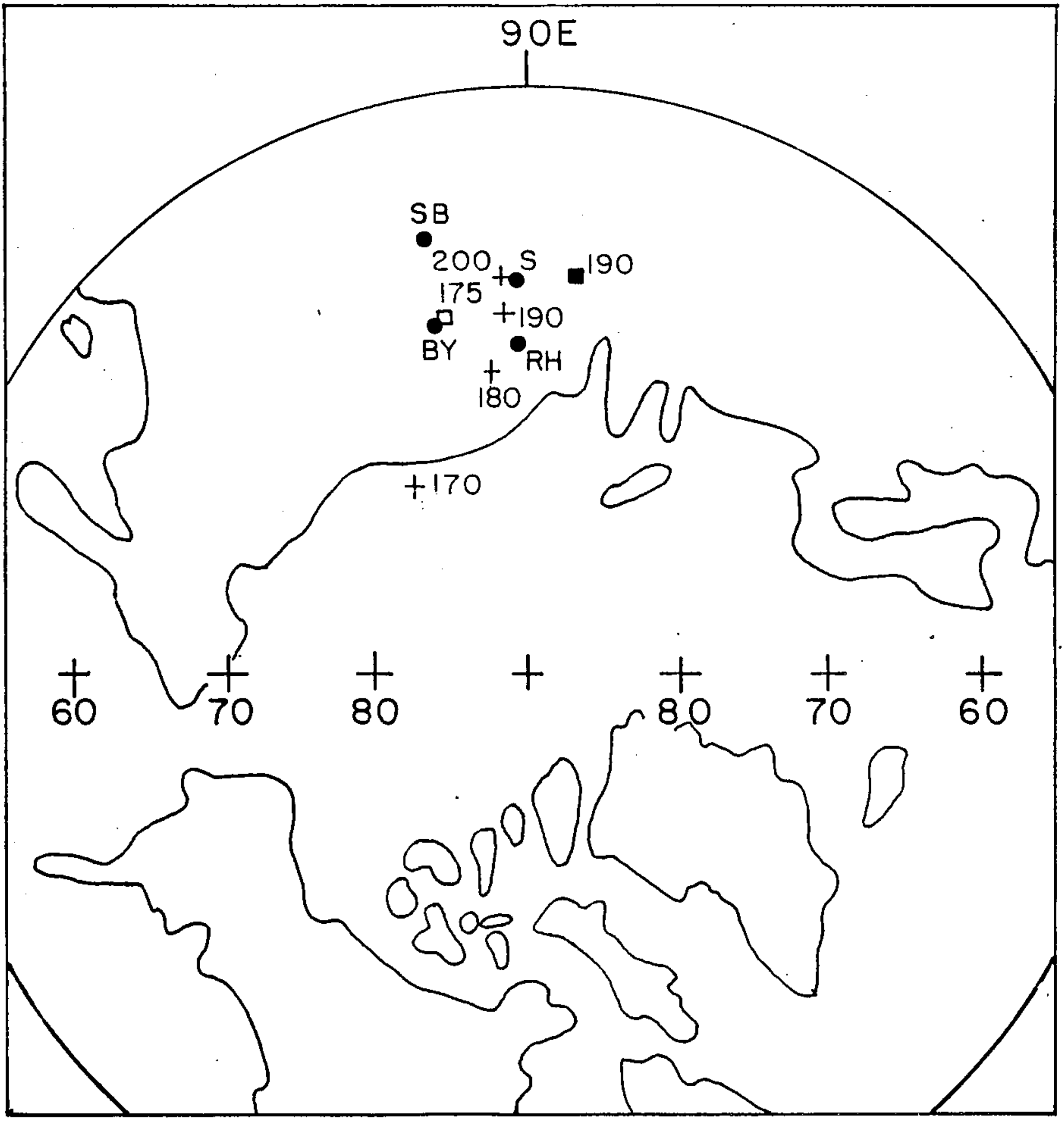


Figure 5: Virtual geomagnetic poles determined in this study (solid circles) with points on the polar wander curve (crosses) of Harrison and Lindh (1982) and the 175 (open square) and 190 my (solid square) poles of Smith and Noltimier (1979). SB-Sand Brook Flow; BY-Byram Diabase Sill; RH-Rocky Hill Diabase; S-Sourland Mt. Diabase.

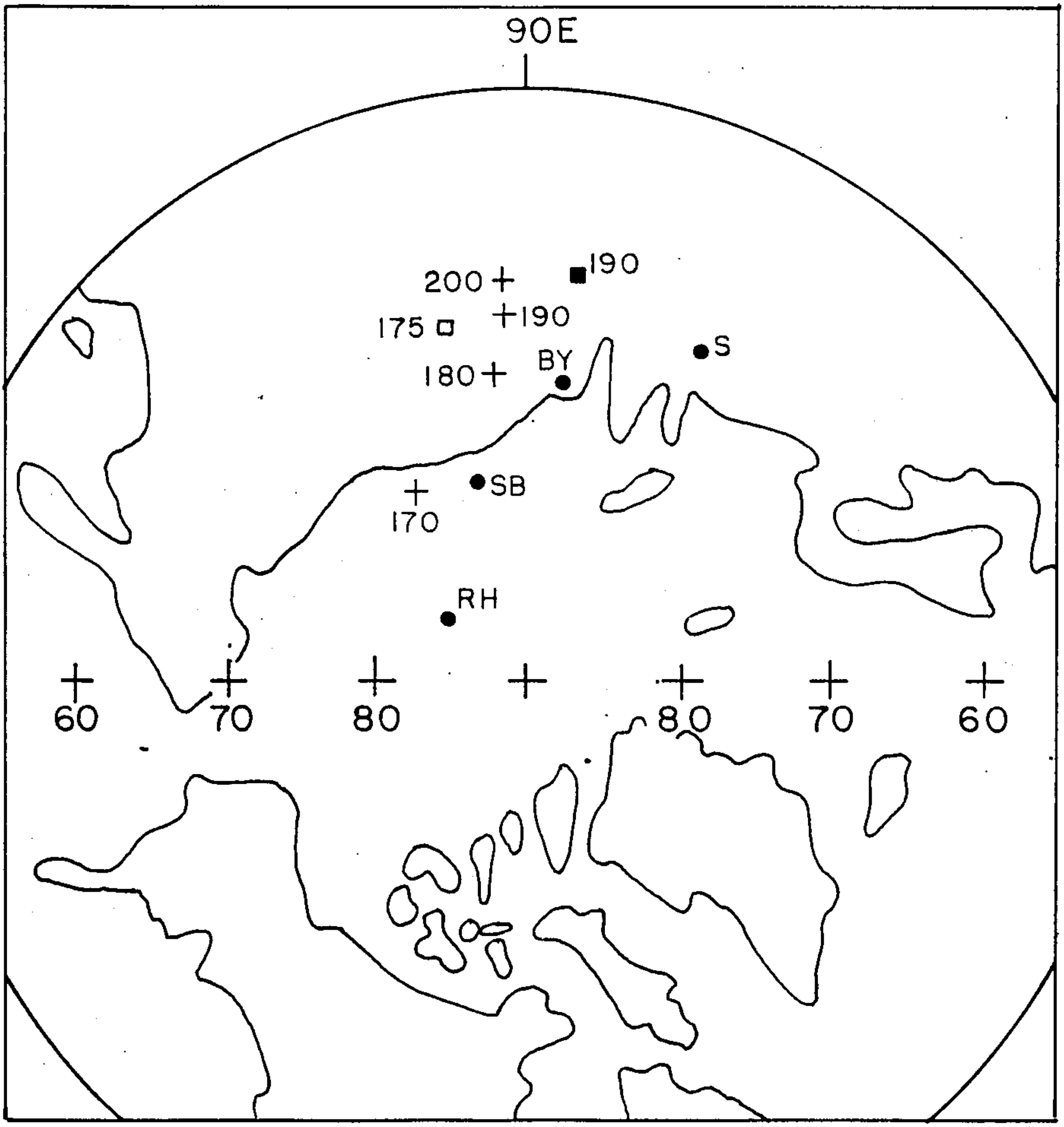


Figure 6: Virtual geomagnetic poles calculated in this study prior to applying corrections for the inclination of bedding (solid circles). Other points as in Figure 5 are shown for reference.

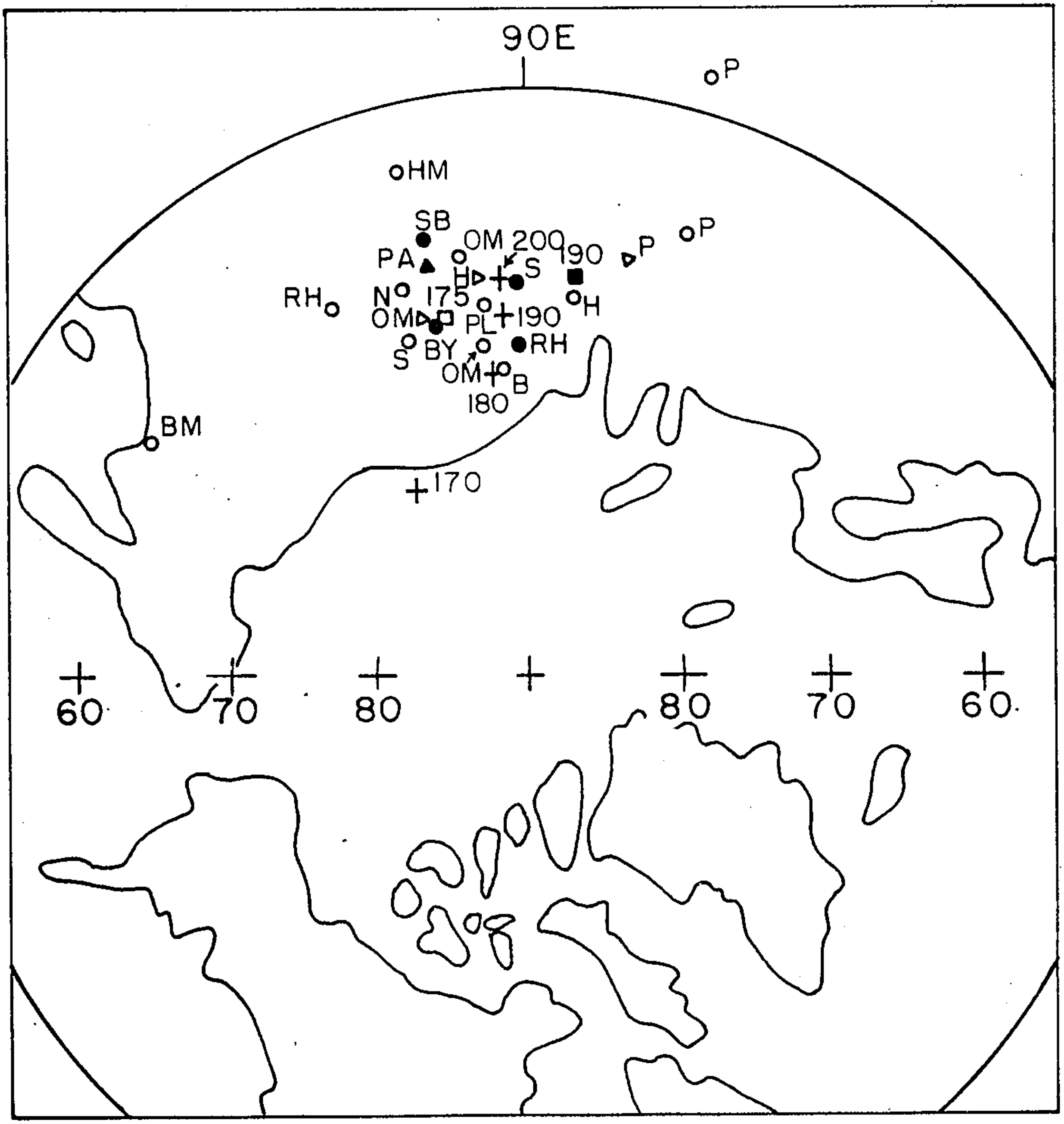


Figure 7: Virtual geomagnetic poles pertinent to this study. Solid circles indicate data from this study. Open circles are from Opdyke (1961). Crosses are from the polar wandering curve of Harrison and Lindh (1982). Open triangles are from McIntosh, Hargraves, and West (in press). Solid triangle is from Beck (1972). The 190 my (solid square) and 175 my (open square) poles of Smith and Noltimier (1979) are also shown. Key to letters: SB-Sand Brook; RH-Rocky Hill; BY-Byram; S-Sourland Mt.; HM-Haycock Mt.; BM-Belle Mt.; PA-Pennsylvania intrusives; PL-Palisades Sill; OM-Orange Mt. Basalt; P-Preakness Basalt; H-Hook Mt. Basalt, N-Newark Basin sediments.

The virtual magnetic pole calculated from our data plots at 63.4 N and 91.6 E, which is half way between Smith and Noltimier's (1979) 175 and 190 my poles. It is quite close to the 200 my point on the polar wander curve of Harrison and Lindh (1982). The pole calculated from Opdyke's data plots closer to the 175 my pole of Smith and Noltimier (1979) as shown in Figure 7.

SAND BROOK

The Sand Brook Lava Flow is a small lava flow, interpreted by Sanders (1962) to lie in a small syncline along the Flemington Fault (See Figure 1). It is generally thought to be correlative with the Orange Mountain Basalt. Results from three closely spaced sites yielded consistent results, and a mean geomagnetic pole of 60.4 N and 103.1 E (See Figure 5). This is not in particularly good agreement with our other results, and we are re-evaluating the structural correction we used.

DISCUSSION OF RESULTS

Figure 6 is a plot of the mean poles which have not been structurally corrected for each of our bodies. Also plotted on that diagram are the polar wander path of Harrison and Lindh (1982) and the 175 and 190 my poles of Smith and Noltimier (1979). There is no correlation. A comparison of this figure with Figure 5, which shows the same results after application of structural corrections based on the attitude of adjacent sediments, shows considerably more agreement. This leads to the conclusion that the magnetization was acquired prior to structural tilting. In this part of the basin then, tectonism post-dated the intrusions.

This is a result inconsistent with models of the evolution of the basin in which the bulk of the tectonic activity occurred prior to the igneous activity.

In Figure 5, the scatter of our data make it impossible to distinguish different 175 and 190 my events as suggested by Smith and Noltimier (1979). Our data do, with the possible exception of the Sand Brook Lava Flow, plot near the polar wandering path of Harrison and Lindh (1972) for the time period 180 to 200 my. If this correlation is valid, our bodies should have been emplaced during that time period, as were the Palisades Sill and Orange Mountain Basalt.

Figure 7, which adds data from other sources to that presented in Figure 5, strengthens the interpretation of the ages of these bodies. Virtually all of the data tend to form a wide cluster around the polar wander curve of Harrison and Lindh (1982), suggesting ages in the range of 180 to 200 my. The exceptions are the Belle Mountain, Haycock Mountain, and Rocky Hill data of Opdyke (1961), and all data on the Preakness Basalt. We can offer an explanation only for the Belle Mountain site. That locality is in a fault zone, and the rocks are extremely highly sheared and serpentinized. We suspect that alteration has partially destroyed the original thermal remanent magnetism.

Finally, the data we have collected to date do not appear to suggest major rotation of fault blocks about sub-vertical axes. Given that there are two major faults cutting across the Newark Basin which have been interpreted to have major strike-slip components of motion on them, it seemed possible that some rotation of structural blocks would have occurred. We would

Table 3: Mesozoic Pole Positions

UNIT	N	K	A	LAT	LONG	SOURCE
Kayenta Fm (Utah)	105	32	2.5	61.7	72.7	Steiner & Helsley (1974)*
Kayenta (Utah)	7	102	6.0	61.2	82.5	Johnson (1976)*
Summerville Fm (Utah)	15	71	4.6	67.5	110.6	Steiner (1978)*
Topley Intrusives (British Columbia)	13	12	9.1	72.0	128.6	Symons (1973)*
Lwr Morrison Fm (Utah)	32	23	5.4	61.4	143.0	Steiner & Helsley (1975)*
Upr Morrison Fm (Utah)	68	35	3.0	67.8	163.8	Steiner & Helsley (1975)*
Pole (170 my)	7		9.9	75.9	121.3	Harrison & Lindh (1982)
Pole (180 my)	13		5.4	69.9	96.6	Harrison & Lindh (1982)
Pole (190 my)	12		3.5	66.2	94.0	Harrison & Lindh (1982)
Pole (200 my)	16		2.8	63.6	94.1	Harrison & Lindh (1982)

* Data from Smith and Noltimier (1979) who cited the original references.

expect to find our data plotting as discrete clusters correlating with the fault block in which the intrusive or flow was located. This is not obvious in our data. If rotations exist at all, they are of a magnitude too small to resolve with the level of accuracy we have been able to obtain.

CONCLUSIONS

In the rocks we have examined, there is no evidence to suggest that tilting of the sediments occurred prior to the intrusion of the magma. Smith and Noltimier (1979) have evidence that one of their intrusions was emplaced after the rocks were tilted. Ratcliffe (1980) has suggested the possibility that some folding and faulting occurred prior to the igneous activity. This leads one to suspect that there should be lava flows or intrusions which post-date the tilting of the rocks. We are still looking for examples of this relationship.

When we began working on the Byram intrusive and determined a virtual geomagnetic pole that correlated so well with the 175 my pole of Smith and Noltimier (1979), we hoped that we would be able to document two ages of intrusive and extrusive activity in the Newark Basin. Thus far, our data suggest that this is not the case. It now appears more likely that there was one major period of igneous activity. It is hoped that further work will clarify this issue.

There is no evidence to support major rotation of blocks about steeply plunging axes. This probably lends support to Manspeizer's (1980) suggestion of complementary motion on normal faults and strike-slip faults to produce rhomb-graben with very little rotation of blocks about sub-vertical axes.

FURTHER WORK

There is ample opportunity for further paleomagnetic work in the intrusives and extrusives in the Newark Basin. We are currently collecting data on three small lava flows which lie in small synclines along the Flemington Fault. Our hope is to determine the timing of the lava flows relative to the formation of the folds. The same type of study should be done for the Ladentown Flow, on which there are already gravity and magnetic data linking it to the Palisades Sill (Kodama, 1983), and structural data suggesting it cross-cuts folds (Ratcliffe, 1980).

ACKNOWLEDGEMENTS

The authors would like to thank Dr. Laurie L. Brown for the use of her lab, without which this study would have been impossible. We would also like to thank Ms. Joyce Castro and Ms. Patricia Weisse for their assistance in helping us find our way around the lab. Hozik received a Research and Professional Development Grant from Stockton State College which supported some of the field work for this project.

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CHAPTER TEN

ROAD LOG

IGNEOUS ROCKS OF THE NEWARK BASIN: PETROLOGY, MINERALOGY, AND ORE DEPOSITS

by

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Note - Most rock exposures viewed on this trip will be at quarries and roadcuts. Please keep away from vertical cuts and loose rock; you are advised to wear a hard-hat and safety glasses.

Mileage from Kean College, Union, New Jersey.

0.0 From parking lot turn left onto Morris Ave. (north)

2.0 Turn right onto Parkway-Rt. 22 entrance, bear left onto Garden State Parkway.

7.0 Turn right onto Route 280 at exit 145, follow signs to 280 west.

11.1 Type section of Orange Mountain Basalt, displaying the westward dipping lower flow unit and a structural sequence comparable to that of the Giants Causway of Northern Ireland as described by Tomkeieff (1940). The lower colonnade is exposed as a fine-grained columnar jointed but otherwise massive layer about 6 m thick. The overlying entablature is a medium-fine grained layer, about 35 m thick, characterized by radiating and curved joint patterns. The upper colonnade is exposed as a fine-grained layer, about 9 m thick, characterized by poorly defined columnar and horizontal joints.

About 40 percent of the total thickness of the Orange Mountain Basalt is exposed along Route 280. Its thickness along Route 280 is about 125 m but it varies from about 100 m to 200 m throughout the Watchung Syncline (Olsen, 1980). The lower flow unit is separated from the upper flow or flow sequence by a red volcanoclastic bed typically less than one meter thick. Chemical analyses of samples from along this roadcut indicate no discernible differences in the composition of the three structural portions of the lower flow unit (Puffer and Lechler, 1980).

11.1 East rim of valley cut into Feltville Formation.

12.6 Type section of Preakness Basalt, exposing almost the entire and unusually thick lower flow unit, and an almost complete Tomkeieff structural sequence. The lower colonnade is exposed as a medium to fine grained and massive layer, about 5 m thick, with columnar jointing. The entablature is an unusually coarse-grained central layer, about 80 m thick, characterized by closely spaced columnar jointing. The upper colonnade is a medium-grained layer, about 10 m thick, with poorly defined columnar and horizontal jointing. The overlying one or two flows of the Preakness Basalt are separated from the first by a thin red siltstone layer that is not exposed along Route 280. The combined thickness of the Preakness flows averages 215 m. Samples of Preakness Basalt from the Route 280 exposure have been chemically analyzed by Puffer and Lechler (1980) and are described in Chapter Three.

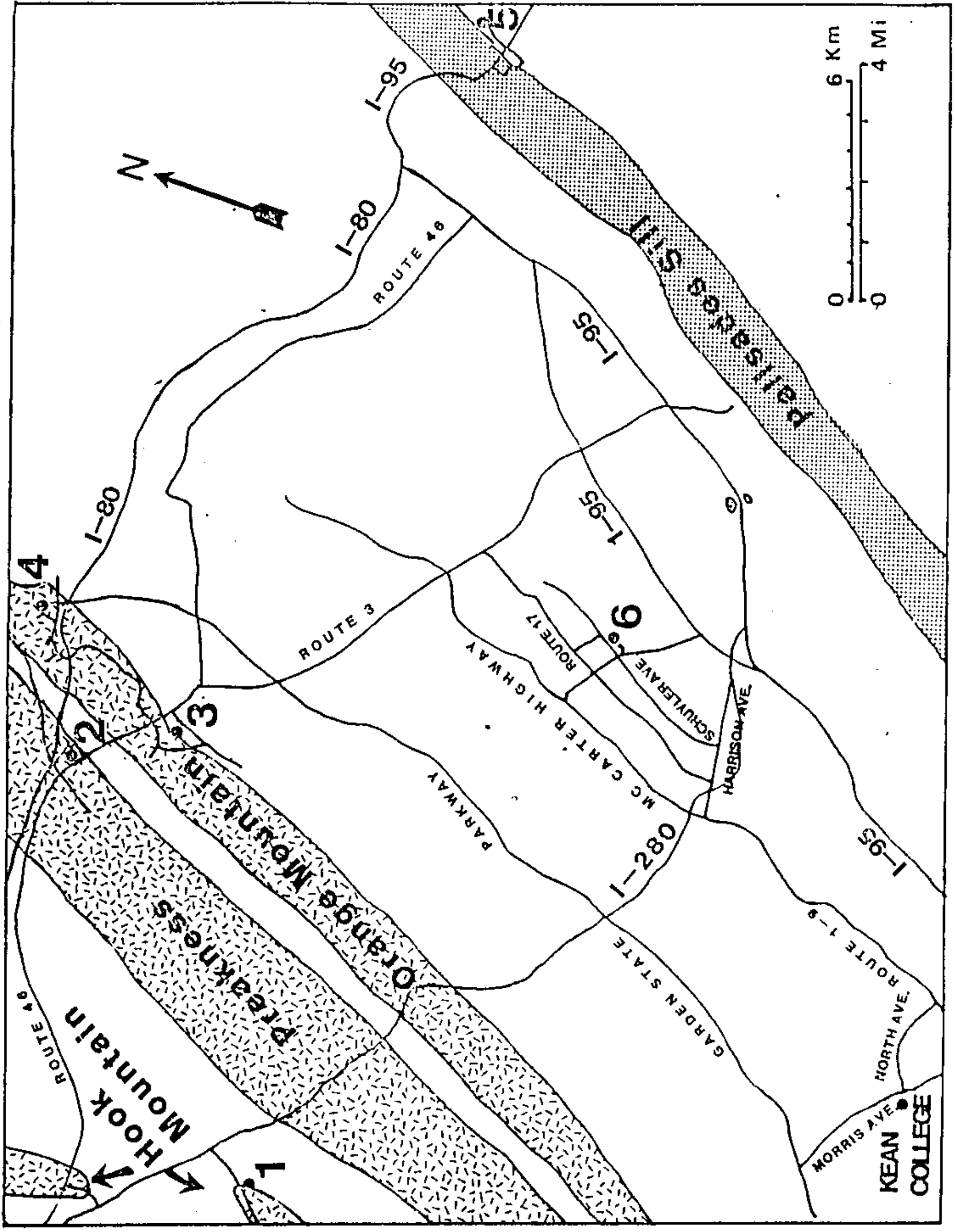
16.1 Take Exit 4A onto Eisenhower Parkway, proceed south.

17.2 Park in back of Merrigan's Bar.

STOP 1 Hook Mountain Basalt.

Walk along Eisenhower Parkway exposure toward lower contact (east) with the Towaco Formation. Note poorly developed columnar structures and abundance of vesicules and amygdules in contrast to the Preakness and Orange Mountain Basalts. Most of the larger vesicules, or gas pockets up to one-half meter across, are lined with prehnite. Prehnite

Fig. 1


N
0 6 Km
0 4 Mi
Palisades Sill
5
I-95
I-80
ROUTE 46
I-95
I-95
I-95
ROUTE 3
ROUTE 17
6
SCHUYLER AVE.
HARRISON AVE.
MC CARTER HIGHWAY
PARKWAY
I-280
GARDEN STATE
ROUTE 1-9
NORTH AVE.
MORRIS AVE.
KEAN COLLEGE
4
3
2
Orange Mountain
Preakness
Hook Mountain
ROUTE 46
1

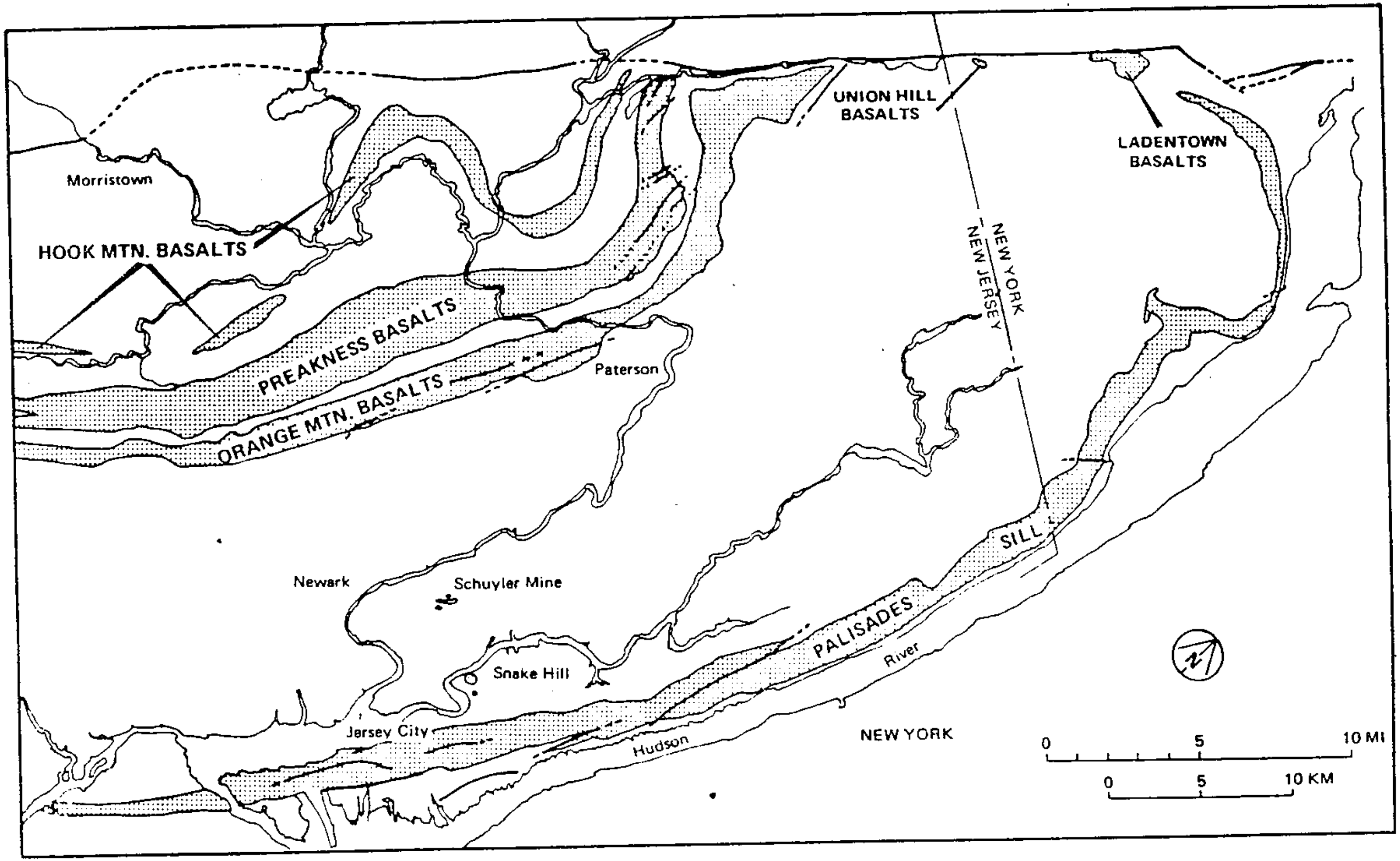


Figure 2 Igneous Rocks of the Northern Newark Basin (after Mason, 1960)

is abundant along the road-cut and provides a good collecting opportunity. A highly altered zone of basalt, about one meter thick, at the base of the flow, rests upon a gray, bleached layer of Towaco Formation that marks the contact. The highly altered zone is loaded with xenoliths of Towaco sediment and appears to have strongly reacted with the sediments. Pipe vesicules are common and have been interpreted by Manspeizer (1980) as indicating southerly flow in contrast to the underlying basalt units.

17.2 Turn north onto Eisenhower Parkway.

18.1 Turn right onto Route 280 west.

20.5 Turn right onto Edwards Road (Exit 1).

20.6 Turn right, follow signs to Route 46 east.

21.4 Turn right onto Route 46.

21.8 Exposure of Hook Mountain Basalt on left.

28.8 Turn right at Union Blvd. exit (across from Totowa Cinema), cross over Route 46.

29.1 At second light turn right onto Lackawana Ave.

29.4 Turn right onto Riverview Drive, the Passaic River is on the left, proceed under bridge, pass 8 telephone poles and park along road.

29.8 STOP 2 Preakness Mountain Basalt.

The tumulus exposed here is described by Manspeizer (1980). The tumulus (also known as pressure domes or schollendomes) is about 7 m high and 15 m wide consisting of a central core of massive basalt with columnar joints, an overlying zone of radiating joints (entablature), and an upper zone of vesicular basalt (upper colonnade with poorly defined joints (Manspeizer, 1980). The tumulus is overlain by an irregular thickness of a pillow-pahoehoe layer that was presumably fed by the tumulus. The pillow-pahoehoe layer is very vesicular and glassey with pillow buds, pahoehoe toes, flattened lava tubes, and a variety of volatile generated structures. A second layer of basalt overlies both structures and consists of massive basalt with bent pipe amygdules.

Tumuli are described by Swanson (1973) as the result of the upward arching of lava over active lava tubes or distributary tubes where they have become clogged. The upward arching is due to the pressure from advancing lava through tunnels or tubes. Occassionally the lava breaks through to the surface and extrudes, here as the pillow-pahoehoe layer. Tumuli, or pressure domes are eliptical in plain view and grade into elongate pressure ridges.

29.8 U-turn on Riverview Drive; be very careful of blind curves.

30.3 Turn left onto Lackawana Ave.

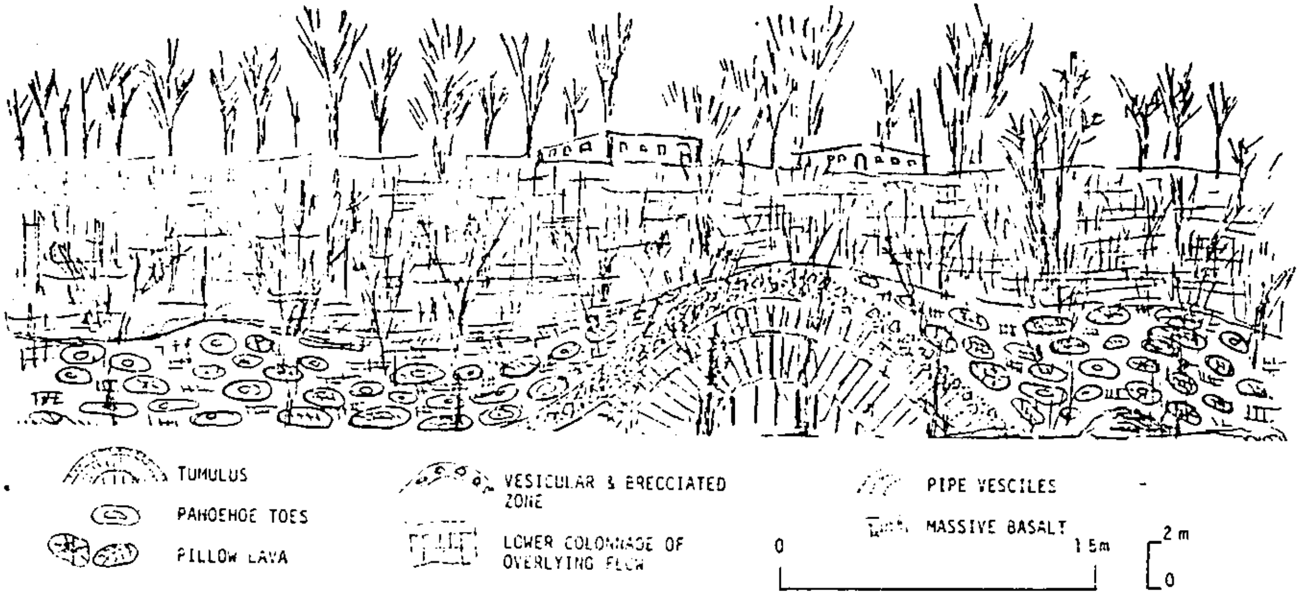


Fig. 3 Field sketch of tumulus with overlying pillow-pahoehoe complex that thins along the crest of the tumulus and is overlain by columnar basalt of the second flow unit; Little Falls.

(from Manspeizer, 1980)

30.6 Turn left onto Union Blvd., follow signs to Route 46 east.

30.7 Turn onto Route 46 east.

32.6 Pillow basalts of upper flow of Orange Mountain Basalt exposed on right.

33.1 Turn right onto Clove Road (at Shell Station) proceed past Wards Trucking Company, bear to the left.

33.4 Park at Montclair State College parking lot.

STOP 3 Orange Mountain Basalt.

Carefully walk across Clove Road; watch out for dangerous traffic Walk along path through fence while crossing Clove Fault. The path is on an exposure of Passaic Formation, with common rills, ripple marks and tracks. Large pits on the left, while facing the quarry wall, were dug by quarry operators to accomodate stilbite collectors. The lower chill zone or colonnade and central columnar zone, or entablature, is exposed on the quarry wall and the amygdaloidal zone of the lower colonnade is exposed on the quarry floor. Walk along the upper quarry terrace toward the notch eroded into a strike-slip fault. Note the slickenside and fault gouge along the fault. Carefully climb down the fault plane to the lower terrace. Several large gas pockets lined with prehnite are exposed on the quarry walls with some brown heulendite and reddish chabazite. Spheroidal weathering that superficially resembles pillow structure is common throughout the quarry. The upper exposed quarry surfaces have been glacially polished and striated.

33.4 Continue south on Clove Road.

34.0 Turn right onto Long Hill Road at stop sign.

34.7 Turn right at stop sign (stay on Long Hill Road).

34.9 Turn right.

35.3 Turn left at "46 west" sign and cross over Route 46.

35.4 Turn left onto Rifle Camp Road but be careful of dangerous cross traffic. Rifle Camp Road, almost throughout its entire length, rests on the top of the lower flow unit of the Orange Mountain Basalt. The only section which rests on pillow lava is the southern part. Exposures of basalt along the west side (left) of Rifle Camp Road are the pillowed second flow unit of the Orange Mountain Basalt. Exposures east of the road are the lower flow unit. Clove Road Fault is parallel to Rifle Camp Road on the east (right) side.

37.2 Bear to the left.

37.6 Continue through intersection onto New Street

40.2 Pass the Upper New Street Quarry on the right and the Lower New Street

Quarry on the left. The quarries are in the process of being relandscaped and developed into apartment units but as of September 1984 the large quarry wall of the Upper Quarry exposes the pillows of the upper flow unit of the Orange Mountain Basalt together with abundant secondary mineralization. The volcanic structures, some of which are still exposed, have been described in detail by Manspeizer (1980).

40.3 Continue north across Route 80.

40.5 Turn right onto Grand Street.

40.7 Turn left onto Spruce Street at light.

41.0 Turn right onto Market Street.

41.1 Turn right into parking lot in back of the Paterson Museum.

STOP 4 Orange Mountain Basalt (lunch stop)

During the lunch break, about one hour, try to divide your time between:
1) Great Falls viewed at statue of Alexander Hamelton, with park benches. Note several vertical strike-slip fault plains through the lower flow unit of Orange Mountain Basalt and the well developed columnar jointing. The Tomkeieff structural sequence is again quite apparent. Glacial activity has removed the overlying friable pillowed flow unit. The S.U.M. over the door of the building near the river at the base of the view area stands for "Society of Useful Manufactures" an organization founded to promote local trade. The hydroelectric plant operated from 1914 through 1968. It was closed in 1969 because of flood damage but is currently being restored.
2) The path through Upper Raceway Park provides an excellent view of the lower flow unit of the Orange Mountain Basalt. The upper raceway and middle raceway systems were built during 1792 through 1838 to bring water power to Paterson's early industries, particularly the silk industry.

Southwest plunging pipe amygdules and vesicles are exposed along the lower southwest dipping contact with the Passaic Formation. Manspeizer (1980) plotted these and other pipe amygdule orientations onto steriographic projections and together with paleocurrent data determined that the Orange Mountain lava flowed to the northeast transgressing a regional paleoslope inclined to the southwest.
3) The paterson Museum exhibits some of the best examples of the secondary mineralogy found in the Paterson area trap rock quarries particularly prehnite (the official State Mineral of New Jersey), heulandite, datolite, chabazite, stilbite, other zeolites, and amethest.

41.1 Turn left onto Market Street.

41.2 Turn left onto Spruce Street.

41.3 Turn left onto Grand Street.

41.4 Turn right onto Route 80 entrance ramps bear to the right, follow signs to Route 80 east.

46.2 Bear to the right; take Route 80-local.

51.2. Continue east toward the George Washington Bridge on Route 95

54.0 Pass upper contact of Palisades Sill with the Lockatong Formation. The upper contact is completely concordant here and is rarely exposed anywhere. Olsen (1980) correlates the hornfels exposed here with that exposed at the Granton Quarry to the south.

54.5 Bear to the right, use local traffic lanes.

55.1 Turn off Route 95 at Exit 67-Fort Lee (last exit in N. J.), continue east, toward the Hudson River, on Bridge Plaza.

55.6 Turn right onto River Road.

56.0 Turn left into entrance to Palisades Interstate Park, follow park road east.

56.1 Observe olivine zone of Palisades Sill on left exposed as a weathered brown layer. Follow park road north along the lower Palisades Sill.

56.5. About 0.1 miles south of the George Washington Bridge is a good exposure of an irregular contact of the Palisades with the Lockatong Formation as sketched by Olsen (1980), Fig. 6 . No parking along the park road is permitted.

56.6 Pass under the George Washington Bridge.

56.7 Observe approximately conformable contact on left.

57.0 Pass through traffic circle in park road, turn off to the right down hill.

57.3 Park in lot at Ross Dock.

STOP 5 Lower contact of Palisades Sill at Ross Dock. Walk south along road to boat launch then walk up stone steps, through tunnel under road and out onto road level. Walk north along the contact with the Lockatong hornfels. The contact at this exposure has been sketched by Olsen (1980) Fig. 7 .

Most of the Lockatong Formation as described by VanHouten (1964) is arranged in short "detrital" and "chemical" cycles. Detrital cycles, about 4-6 m thick, occur as black pyritic shales in the lower part of the formation overlain by dark gray carbonate rich mudstone and by massive gray calcareous argillite in the upper part. Some detrital cycles contain lenses of siltstone and feldspathic sandstone. The portions of the detrital cycles that have escaped contact metamorphism are typically composed of abundant Na-feldspar, illite-muscovite, some K-feldspar, chlorite,

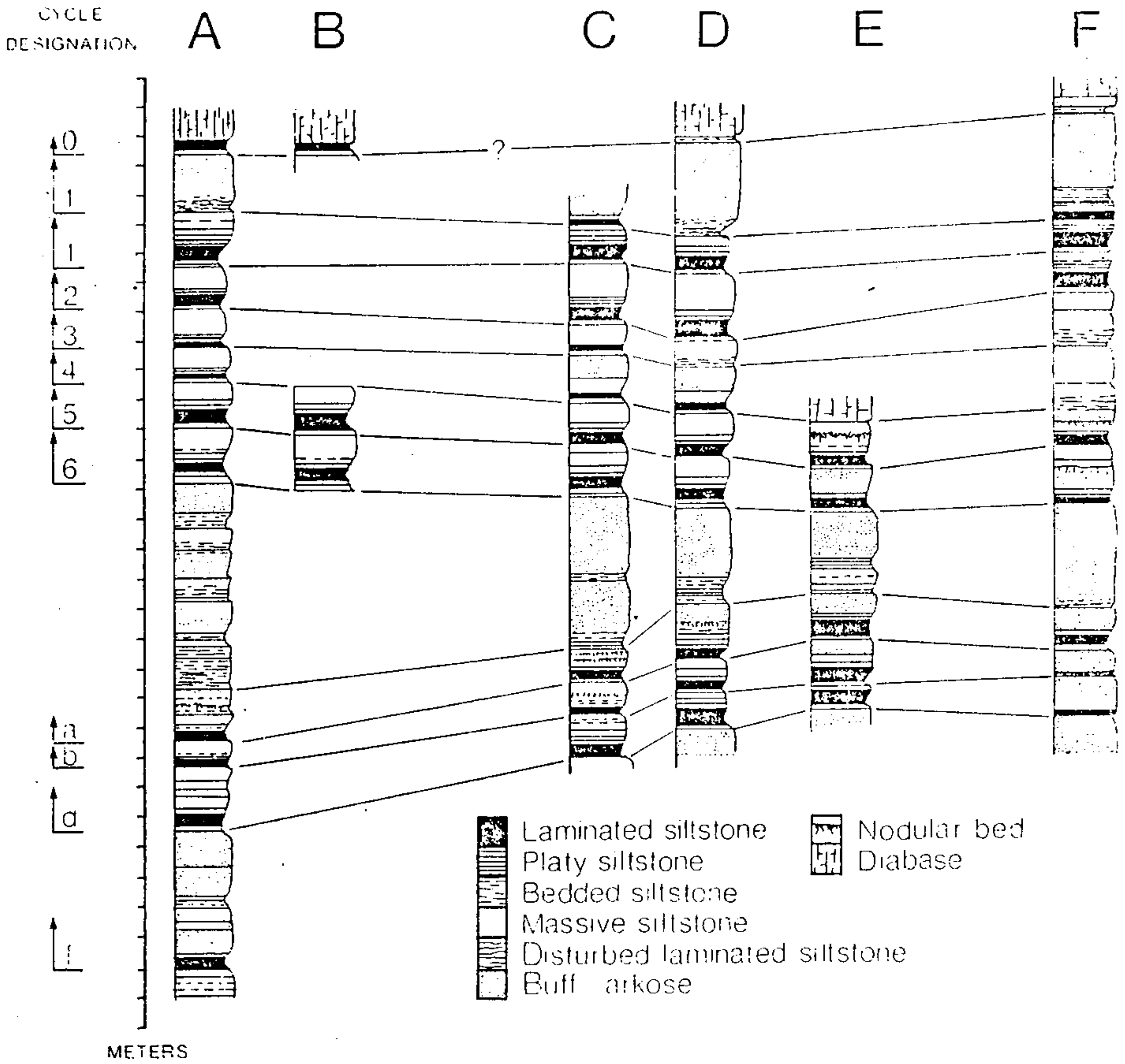


Fig. 4 Correlation of cycles from Kings Bluff, Weehawken to Ross Dock, Fort Lee; A, Kings Bluff exposure; B, Gratacap's (1886) Weehawken locality and exposure of cycle 0 to the north; C, Gorge and River Roads exposure, Edgewater; D, exposure at east portal northern tunnel for old New York, Susquehanna Southwestern Railroad; E, "old trolley route" below old Palisades Amusement Park, Fort Lee; F, exposures west of Ross Dock, Palisades Interstate Park, Fort Lee. Exposures A and F are 12 km apart; the other sections are positioned to scale.

(after Olsen, 1980).

Metamorphic minerals

(adapted from Van Houten 1969)

generalized cycle

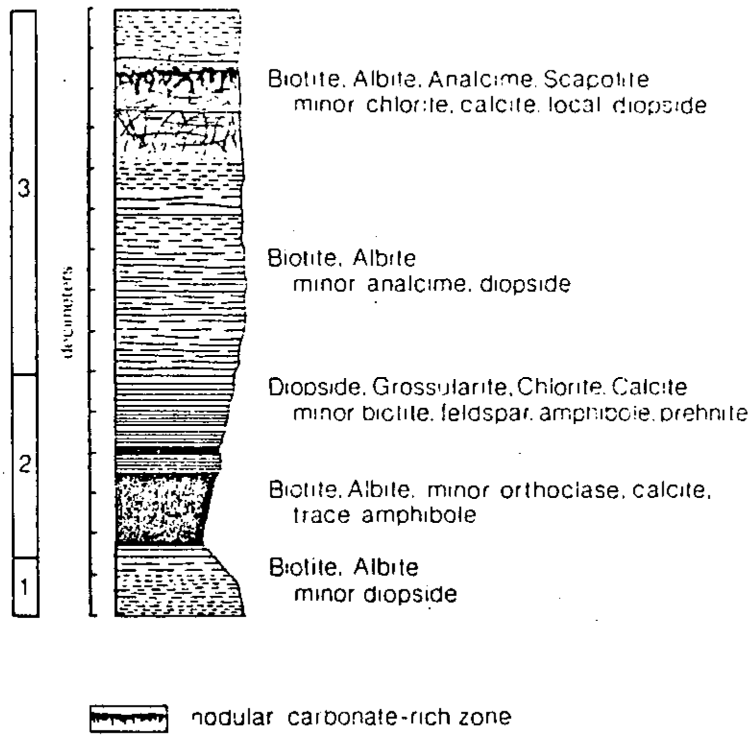


Fig. 5 (after Olsen, 1980)

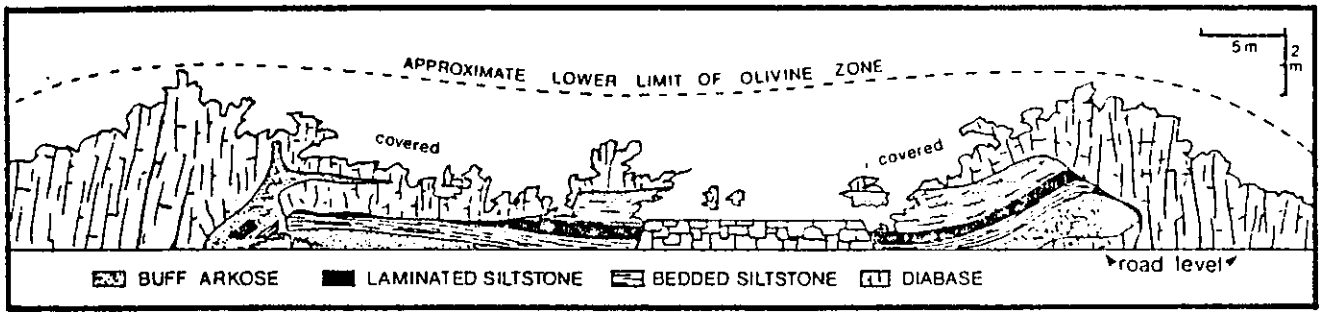


Fig. 6 Exposures of discordant contact of Palisade Diabase and Lockatong Formation, south of George Washington Bridge on road from River Road to Ross Dock in Palisades Interstate Park, Fort Lee. (after Olsen, 1980)

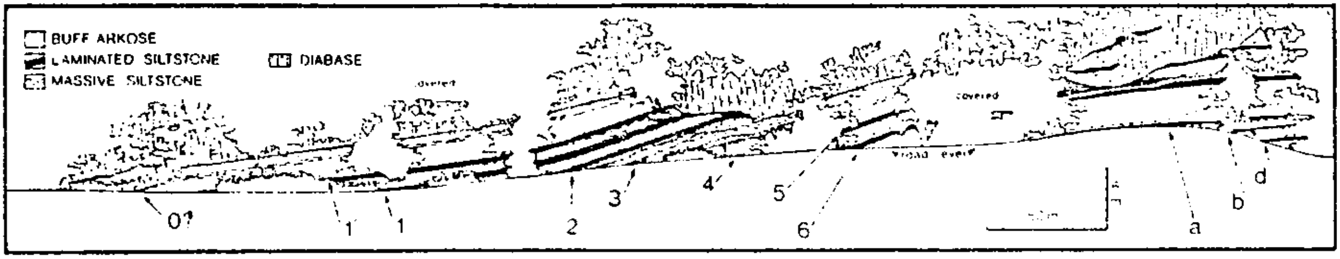


Fig. 7 Exposures of Palisade Diabase and cycles 0-d west of Ross Dock, Palisades Interstate Park, Fort Lee.

(after Olsen, 1980)

calcite and a little quartz. Short cycles dominated by dolomitic mudstone and analcime - dolomite rich argillite, about 2-4 m thick, are limited to the central portion of the Newark Basin VanHouten (1964).

The cycles exposed at the lower contact (Fig. 7) are detrital cycles that have been metamorphosed into hornfels and have been correlated by Olsen (1980) to five other exposures along the lower contact up to 12 km south of Ross Dock (Fig. 4). Metamorphic mineral assemblages found in hornfels exposed along the lower contact as described by VanHouten (1969) include:

1) biotite, albite, analcime, scapolite, with minor calcite and local diopside.
2) biotite, albite, with minor analcime and diopside.
3) diopside, grossularite, chlorite, calcite with minor biotite, feldspar, amphibole, and prehnite.
4) biotite, albite, minor orthoclase, calcite, and trace amphibole.
5) biotite, albite, mihor diopside.

While proceeding north along the road from the stone path observe:

1) The decreasing grain size of the diabase as the lower contact is approached.
2) The buff arkose cycle exposed 0 to 75 m north of the contact (cycle 1').
3) The platy and laminated siltstone poorly exposed 75 to 100 m north of the contact (cycle 1).
4) The platy and laminated siltstone well exposed 100 to 180 m north of the contact (cycle 2).
5) The buff arkose exposed 180 to 250 m north of the contact.
6) The laminated and massive siltstone well exposed 250 to 300 m north of the contact (cycles 5 and 6).
7) The buff arkose exposed 300 to 440 m north of the contact (cycle a).

Also note the occurrence of porphyroblasts that are particularly common in the black biotite-albite hornfels developed out of the pelitic layers. Most of the common green spots found in these hornfels are pinite pseudomorphs after cordierite (Miller and Puffer, 1972). Large tourmaline porphyroblasts are also common in Lockatong hornfels. Even larger green spherical structures, up to 4 cm across, composed largely of clinozoisite, are less common in the hornfels but have been found at several locations throughout the contact metamorphosed portions of the Newark and Culpeper Basins. I am currently studying these green structures and tentatively suggest progressive stages of thermal metamorphic development capable of developing the large, up to one meter, roughly spherical, phaneritic structures found in the meta-graywackies of the Ducktown Tennessee ore deposit and the phaneritic syenites that Barker and Long (1969) interprete as having been generated out of Lockatong Formation.

58.6 Return to park entrance, leave park and turn right.

59.0 Turn left at Rt. 95 south sign onto Bridge Plaza, follow Rt. 95 signs.

59.6 Turn right at 95-south Detour sign, cross over Rt. 95, turn left and follow all 95-south signs. <u>Avoid</u> 95-north signs which will take you across the George Washington Bridge.

62.6 Bear to the left, follow Rt. 95-south; avoid Rt. 80-west.

64.6 Bear to the right where 95-south divides.

68.3 Take exit 16W to Route 3-west, drive through exit roads, follow all signs to Rt. 3-west.

69.2 Turn right onto Route 3 west.

70.7 Exit onto Route 17 south-Lyndhurst.

72.3 Turn left at Ridge Road (Rt. 17)

73.9 Turn left at Noel Drive.

74.4 Turn right at Schuyler Ave.

75.6 Turn left onto Belleville Ave.

75.7 Turn left onto Porete Ave.

75.9 Bear to the left around sharp curve.

76.2 Park in parking lot to the left at the base of the cliff.

STOP 6 Schuyler Copper Mine, North Arlington, New Jersey. The Schuyler Mine is also called the Belleville Mine, the Arlington Mine and the Victoria Mine. The ore occurs in lenses of gray sandstone to siltstone within the red sandstone and siltstone of the Passaic Formation. The Arlington diabase sill and some shallow branching dikes and sills (offshoots of the Palisades Sill) are exposed within and stratigraphically under the ore deposit.

The Schuyler deposit was mined intermittently throughout the longest historic interval of any of the New Jersey copper mines. It was first mined shortly after 1700 and was finally abandoned in 1923 (Woodward, 1944). It was the second most productive copper mine in New Jersey, second only to the American Mine, and had the most extensive network of underground workings of any of them (Fig. 8).

The chief ore minerals at the Schuyler deposit was chalcocite. Considerable chrysocolla is also present with some azurite and malachite and minor native copper, cuprite, covellite and brochantite. A few rich pockets of chalcocite were extracted with some disseminated copper ore between pockets. The ore was presumably precipitated out of hypogene solutions emanated out of the underlying Arlington sill and branching dikes.

76.2 Turn right onto Porete Ave., return to Belleville Ave.

76.6 Turn right onto Belleville Ave.

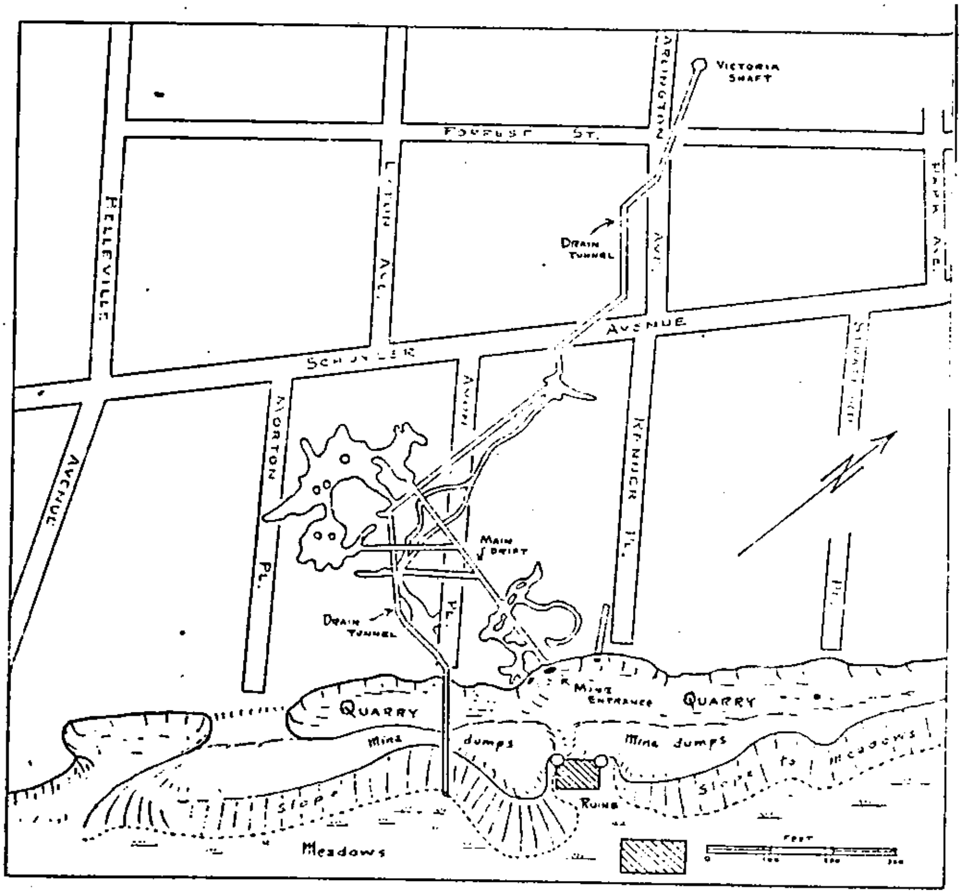


Fig. 8 Generalized sketch map of the vicinity of the Schuyler or Arlington copper mine. There is no present surface indication of the old Victoria shaft at street level, and considerable areas over the large chambers near Morton Place are partially caved. The reduction plant of the Arlington Copper Company is mostly in ruins, save the building on the meadow floor, which is currently occupied. Location of the underground workings is based upon a published map by O. Ivan Lee.

76.8	Turn left onto Schuyler Ave.
79.0	Turn right onto Harrison Ave.
80.1	Turn left onto McCarter Highway (Rt. 21)
82.4	Follow all signs to Rt. 21 south, climb ramp, follow signs to Rt. 1-9 south (stay on local lanes).
84.9	Bear to the right, follow all signs to North Ave.
85.2	Bear to the left.
85.3	Exit to the right, follow all signs to North Ave. west.
87.6	Turn right at Morris Ave.
87.8	Turn into Kean College Parking lot (the end).

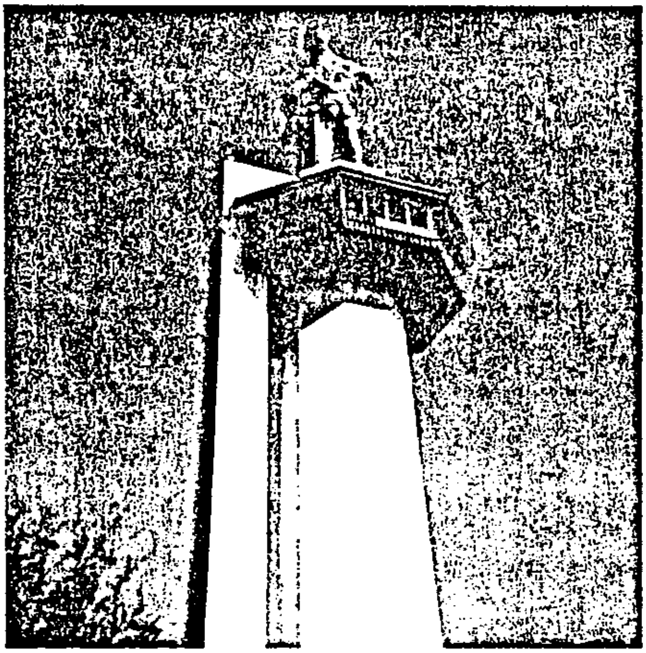


Fig. 9 Back-side of Vulcan, presumably the God of volcanology, Birmingham, Alabama.

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STRIKE-SLIP NEWARK-TYPE BASINS (TRIASSIC-JURASSIC) ALONG THE ATLANTIC PASSIVE MARGIN OF EASTERN NORTH AMERICA AND NORTHWEST AFRICA

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STRIKE-SLIP NEWARK-TYPE BASINS (TRIASSIC-JURASSIC) ALONG THE ATLANTIC PASSIVE MARGIN OF EASTERN NORTH AMERICA AND NORTHWEST AFRICA

For over one-hundred years, Newark-type Triassic-Jurassic rift basins were considered grabens produced by extension at right angles to the rift axis. The same strain field, however, may have formed by a left-lateral, east-west oriented shear couple yielding detrital and evaporite pull-apart basins. Oriented primarily by the fabric of the underlying Hercynian-Variscan Orogene, these basins may have originated initially from a heated and stretched crust that ultimately failed along conjugate, en-echelon strike-slip (transform) faults as sea-floor spreading occurred along the axis of the proto-Atlantic Ocean.

Where the basement had been pulled apart creating deep structural troughs, as in the Newark-Gettysburg Basin or the Argana Basin of Morocco, plutons intruded the axis of the basin in the form of dikes, lava flows and subaqueous fissure flows. Differential horizontal shear along strike-slip faults had created asymetric basins with an upthrown leading plate and a subsiding trailing plate. Strata within the basins record a history of recurrent, but alternating, transtentional and transpressional episodes in an overall wrench-tectonic regime. While the borderfault facies is marked by complex unconformities, young basin sediment, volcanics, en-echelon folds, fanglomerates, turbidites and deep-water lacustrine deposits with organic-rich black shale, sediments on the trailing plate are marked by an older suite of gently inclined fluvial-deltaic sands that rest with profound unconformity on the Hercynian-Variscan basement.

Where shallow marine waters of the Tethys Ocean transgressed sagged pull-apart basins (as in the Khemisset and Berrichid Basins of Morocco) or where the basement was faulted by straight, non-branching transforms (as in Grand Banks), vast salt flats occurred forming thick deposits of halite and potash salt. The extent of Tethyan transgression and concomitant subsidence of these basins is marked by salt diapirs in the Baltimore Canyon Trough and in the Aaiun Basin of Africa.