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A NEW SUITE OF POST-OROGENIC DIKES IN THE EASTERN NORTH CAROLINA PIEDMONT: PART II. MINERALOGY AND GEOCHEMISTRY

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ABSTRACT

A swarm of post-metamorphic (200 ± 10 Ma) dikes in the northeastern Piedmont of North Carolina and adjacent Virginia comprises predominantly amygdaloidal rhyolite porphyry and ferrobasalt. The felsic samples have phenocrysts of sanidine, quartz, apatite, Fe-Ti oxides, and ferropigeonite, with amphibole microphenocrysts. The more mafic rocks contain phenocrysts of pigeonite, K-Na-labradorite, apatite, Fe-Ti spinel, and ilmenite and carry sanidine in the groundmass. The high-temperature mineralogy and mineral chemistry of these rocks compare well with highly evolved differentiates of intraplate volcanic series and hypabyssal suites. Whole-rock chemical analyses reveal high Fe/Mg, K/Na, and Rb/Sr, as well as extreme enrichment in incompatible elements (Ti, P, K, Zr, Rb, Y). All samples, even the most mafic, are strongly quartz-normative. Three rare-earth element (REE) patterns of mafic members of the suite show steep negative slopes with $La = 110 - 260$ times chondritic abundance. These chemical characteristics serve to distinguish the rocks of this suite, apparently unique in the eastern North America (ENA) region, from nearby olivine diabase, and raise the question of the origin of this distinctive suite.

INTRODUCTION

This is the second of two papers whose aims are to document a suite of previously little-known post-metamorphic dikes that occurs in the northeastern-most Piedmont of North Carolina and adjacent Virginia. The field relationships and petrographic characteristics, as well as the results of paleomagnetic and Rb/Sr studies, were reported earlier (Stoddard and others, 1986). This paper presents the mineralogy, mineral chemistry, and whole-rock geochemistry of rocks of the suite, and addresses the significance and origin of this dike swarm.

ANALYTICAL TECHNIQUES

Mineral analyses were done by the author in the electron microprobe laboratory of the Department of Geological Sciences at Virginia Polytechnic Institute and State University. Natural and synthetic mineral standards were used, as were the correction factors of Albee and Ray (1970). Whole-rock analyses were done by X-ray fluorescence spectrometry at North Carolina State University, using a Philips 1410 spectrometer. Samples were prepared and analyzed using the

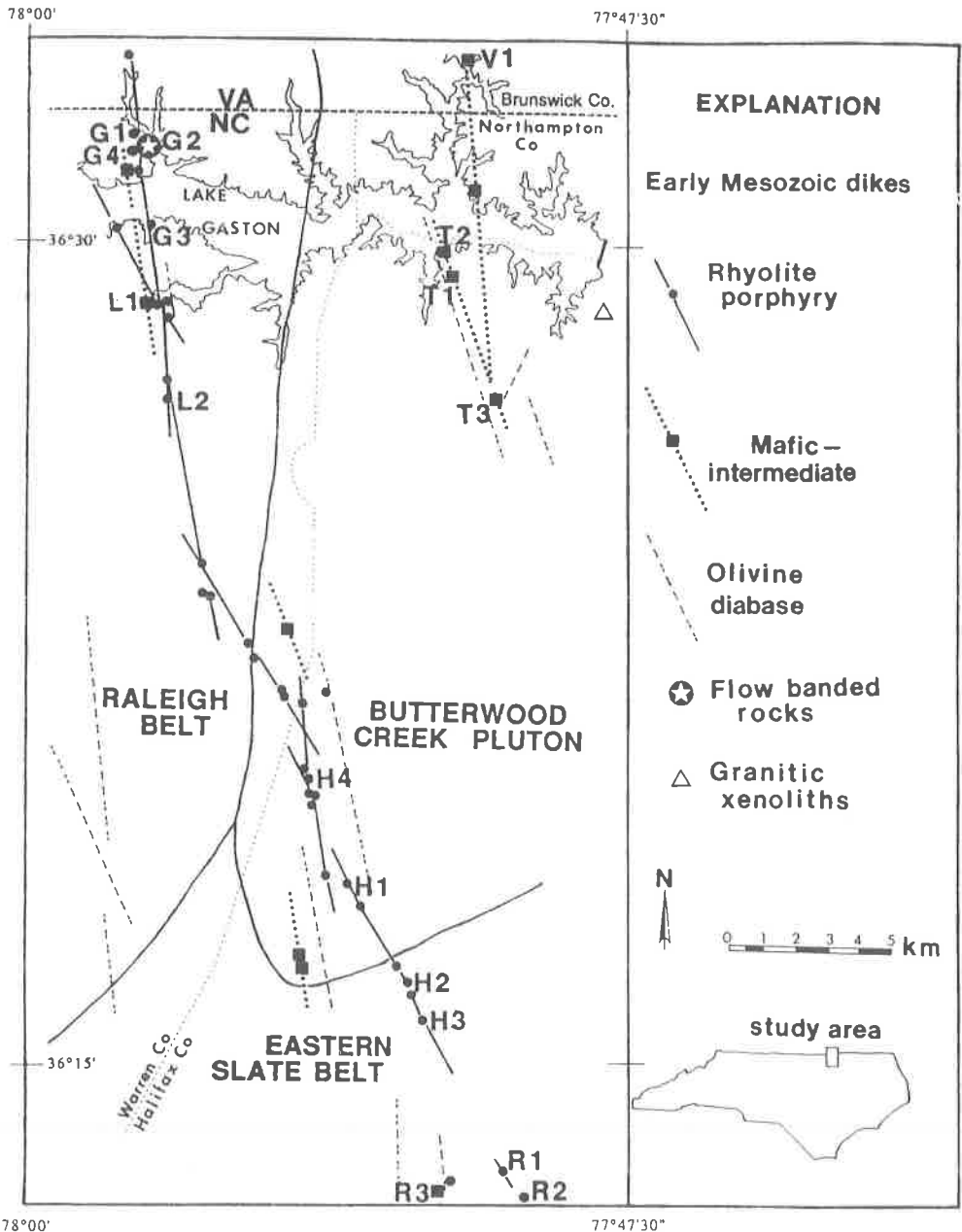


Figure 1. Generalized geologic map of the northeasternmost North Carolina and adjacent Virginia Piedmont, showing locations of sampled early Mesozoic dikes. The Raleigh belt and the Eastern slate belt are late Precambrian to early Paleozoic metamorphic terranes, and the Butterwood Creek pluton is a late Paleozoic granitoid intrusive. Three different varieties of early Mesozoic dikes are shown on the map. Sample localities referred to elsewhere consist of a letter and number combination. The letter indicates the U.S.G.S. 7.5-minute topographic quadrangle in which the samples occur: V, Valentines; G, Gasburg; L, Littleton; T, Thelma; H, Hollister; R, Ringwood.

techniques described by Fabbi (1972). Analytical accuracy and precision are estimated at $\pm 3\%$ of the amount present for most major elements, $\pm 5\%$ for minor elements, SiO_2 and Al_2O_3 , and $\pm 10\%$ for trace elements. Ferric iron was determined using the Pratt titration method described by Jeffery and Hutchison (1981). Rare-earth elements were determined by XRAL Laboratories, Toronto, using instrumental neutron activation analysis techniques. Norms were calculated on a Macintosh SE microcomputer using the BASIC program CIPWNorm (v 2.2) by Douglas R. Mason. Petrologic mixing calculations were done using the program Genmix by R. W. LeMaitre.

ROCK TYPES

The suite has been divided into four petrographic groups (Stoddard and others, 1986): 1. silicic dike rocks; 2. mafic dike rocks; 3. intermediate dike rocks; and 4. flow-banded rocks. Of these, exposures of the silicic rocks are by far the most abundant, outnumbering the mafic occurrences by approximately three to one. The intermediate and flow-banded rocks are known from only one location each. Figure 1 is a map showing the locations of the dikes and the sampled sites. Table 1 summarizes the petrographic features of the samples. Silicic rocks are aphanitic-porphyrific, with prominent phenocrysts of alkali feldspar and quartz. Phenocrysts in the mafic rocks are smaller, and include slender plagioclase laths and pigeonitic pyroxene, as well as abundant Fe-Ti oxide and apatite. Both silicic and mafic dike rocks are amygdaloidal, with amygdules filled with secondary minerals. The intermediate rocks have large blocky plagioclase phenocrysts, but otherwise texturally resemble the mafic rocks. In mineralogy and grain size, the flow-banded rocks resemble the silicic rocks, but are dominated texturally by strongly developed flow-banding, reminiscent of that in some volcanic rocks. For further details of petrography and occurrence, refer to the earlier paper (Stoddard and others, 1986).

MINERALOGY AND MINERAL CHEMISTRY

These dikes contain a distinctive suite of high-temperature phenocrysts. The felsic rocks have sanidine (and locally, anorthoclase), quartz, amphibole, and in some cases, ferropigeonite phenocrysts; the mafic varieties have pigeonite and plagioclase. Both felsic and mafic samples contain Fe-Ti oxides and apatite phenocrysts.

Table 1. Petrographic characteristics of Piedmont dikes*

<u>Group</u>	<u>No.</u> <u>Sections</u>	<u>Vol. % Phenocrysts</u>							<u>Groundmass</u>	<u>% Amygdules</u>
		<u>san</u>	<u>qtz</u>	<u>apa</u>	<u>opq</u>	<u>pla</u>	<u>pyx</u>	<u>olv</u>		
Silicic	54	9	5	1	1	tr	tr		80%	4
groundmass is glassy or felsitic; some with microphenocrysts of Fe-Ti oxides, hastingsite, biotite										
Mafic	12		1	9	30	10			45%	5
groundmass is brownish and alkali feldspar-rich										
Intermediate	2		tr	1	20				75%	5
groundmass of microphenocrysts of alkali feldspar, quartz, hastingsite, and Fe-Ti oxides										
Flow-banded	4	11	1	tr	1	tr	2	tr	83%	2
flow-banded, spherulitic, or glassy										

*from Stoddard and others (1986); san = sanidine-anorthoclase; qtz = quartz; apa = apatite; opq = Fe-Ti oxide minerals; pla = plagioclase; pyx = pigeonite-ferropigeonite; olv = olivine

Feldspars

Phenocrysts of alkali feldspar (sanidine) from the felsic dike rocks and from the flow-banded rocks are most commonly optically homogeneous with a very low 2V (typically 15-30°), optically negative, and monoclinic. They range in composition from Or₆₁Ab₃₂An₇ to Or₄₈Ab₄₁An₁₁ (Table 2). These sanidines commonly occur in glomerocrysts with quartz, apatite, plagioclase, and/or anorthoclase (which displays prominent gridiron twinning). Exsolution is not visible in these feldspars. Powder X-ray diffraction scans suggest they are nearly completely disordered and monoclinic (Wright, 1968), with no evidence of a second phase. A K-feldspar from the groundmass of sample Y, a mafic rock, also falls into the same compositional range as the sanidines from the silicic rocks (Table 3). The alkali feldspars are enriched in Fe (total Fe as FeO = 0.16-0.50 wt%) and Ba (0.41 - 1.30 wt% BaO).

Table 2. Alkali feldspar analyses from silicic rocks

Locality*	R2	G4	G2	H2	H3	H3	R1	H1	G2
Sample	HR2	G4	G2B1	B273	D6A	D6A	HR1	B29A	G2B2
Grain	A	A	A	A	Bcor	Brim	A	A	A
weight percent oxides									
SiO ₂	63.49	63.43	63.16	64.40	64.01	63.21	61.67	61.32	60.03
Al ₂ O ₃	20.51	19.61	20.20	19.96	19.87	19.48	22.35	22.73	22.81
TiO ₂		0.13	0.04						0.12
FeO	0.27	0.20	0.26	0.22	0.25	0.50	0.34	0.34	0.38
MnO		0.02	0.02	0.04	0.03	0.01		0.03	0.02
MgO		0.03	0.03	0.00	0.00	0.00		0.00	0.03
CaO	2.28	1.61	1.97	1.61	1.79	1.36	4.76	5.10	5.36
Na ₂ O	4.53	3.64	3.76	3.67	3.55	3.43	5.36	5.48	4.82
K ₂ O	8.10	9.15	9.21	9.68	9.39	9.64	4.72	3.94	4.79
BaO		0.96	1.30	1.16	0.87	0.73		0.42	0.41
SrO		0.13	0.05	0.00	0.14	0.53		0.33	0.12
total	99.18	98.91	100.00	100.74	99.90	98.89	99.20	99.69	98.89
cations per 8 oxygen atoms									
Si	2.895	2.924	2.893	2.913	2.903	2.916	2.796	2.774	2.749
Al	1.103	1.066	1.091	1.064	1.062	1.059	1.195	1.211	1.231
Ti		0.004	0.001	0.003	0.005	0.001		0.003	0.004
Fe	0.010	0.008	0.010	0.008	0.009	0.019	0.013	0.013	0.015
Mn		0.000	0.001	0.001	0.001	0.000		0.001	0.001
Mg		0.002	0.002	0.000	0.000	0.000		0.000	0.003
Ca	0.111	0.080	0.097	0.078	0.087	0.067	0.231	0.247	0.263
Na	0.400	0.326	0.334	0.322	0.312	0.307	0.471	0.480	0.428
K	0.471	0.538	0.538	0.559	0.543	0.567	0.273	0.227	0.279
Ba		0.018	0.023	0.021	0.015	0.013		0.008	0.007
Sr		0.004	0.001	0.000	0.004	0.014		0.009	0.003
Σ cats	4.990	4.970	4.991	4.969	4.941	4.963	4.979	4.973	4.983
mole percent endmembers									
Or	48.0	55.9	54.2	57.0	56.7	59.4	28.0	23.6	28.6
Ab	40.7	33.9	33.7	32.9	32.6	32.2	48.3	49.9	43.8
An	11.3	8.3	9.8	8.0	9.1	7.0	23.7	25.7	26.9
Cn		1.9	2.3	2.1	1.6	1.4		0.8	0.7

* see Figure 1

Table 3. Other feldspar analyses

Locality* Sample Mineral	G4 G4 pla	R2 RMPD pla	H2 B273 pla	H1 B29A pla	L1 RMWD pla	T1 Y pla	T1 Y ksp	H1 FT10' ksp
weight percent oxides								
SiO ₂	55.82	60.05	57.58	57.25	59.25	54.67	63.15	64.00
Al ₂ O ₃	26.43	24.08	25.86	26.32	25.52	27.45	18.69	18.27
TiO ₂	0.09				0.06	0.10	0.10	
FeO	0.60	0.36	0.56	0.62	0.45	0.71	0.79	0.04
MnO	0.01		0.06	0.01	0.00	0.00	0.00	
MgO	0.07		0.04	0.03	0.06	0.13	0.06	
CaO	9.52	6.84	8.77	9.42	8.15	10.96	2.00	0.03
Na ₂ O	4.78	5.85	5.38	5.30	5.60	4.54	3.67	0.14
K ₂ O	1.22	2.53	1.09	1.03	1.92	0.81	8.93	16.30
BaO	0.07		0.13	0.19				
SrO	0.15		0.22	0.05				
total	98.76	99.72	99.69	100.42	101.01	99.37	97.39	98.78
cations per 8 oxygen atoms								
Si	2.552	2.703	2.598	2.567	2.640	2.494	2.943	2.994
Al	1.424	1.278	1.376	1.401	1.341	1.477	1.027	1.007
Ti	0.003				0.002	0.003	0.004	
Fe	0.024	0.014	0.021	0.023	0.017	0.027	0.031	0.002
Mn	0.000		0.002	0.000	0.000	0.000	0.000	
Mg	0.004		0.002	0.002	0.004	0.004	0.004	
Ca	0.466	0.330	0.424	0.452	0.389	0.536	0.100	0.002
Na	0.423	0.511	0.470	0.460	0.484	0.402	0.332	0.012
K	0.071	0.145	0.062	0.059	0.109	0.047	0.531	0.973
Ba	0.001		0.002	0.003				
Sr	0.004		0.006	0.000				
Σcats	4.972	4.981	4.963	4.967	4.986	4.990	4.972	4.990
mole percent endmembers								
Or	7.4	14.7	6.6	6.1	11.1	4.8	55.2	98.6
Ab	44.1	51.8	49.2	47.4	49.3	40.8	34.5	1.2
An	48.5	33.5	44.3	46.6	39.6	54.4	10.4	0.2

* see Figure 1.

Locally, sanidine phenocrysts have thin, low temperature rims separating them from the groundmass of the rock. In one sample (FT10', see Table 3 and Figure 2), a low-temperature K- feldspar rim adjacent to sanidine is the nearly pure end-member KAlSi_3O_8 .

The plagioclase analyses range from $\text{An}_{33}\text{Ab}_{52}\text{Or}_{15}$ in the most silicic specimen to $\text{An}_{62}\text{Ab}_{36}\text{Or}_{02}$ in the most mafic. Table 3 shows selected plagioclase analyses, with two analyses of alkali feldspar from the groundmass of mafic rocks. Figure 2 is an Ab-An-Or plot of feldspar compositions, also showing isotherms determined by Fuhrman and Lindsley (1988). The alkali feldspars fall into the range of Ca-Na-sanidine, and the plagioclases are K-andesine or K- labradorite, using the nomenclature of Smith (1974). Three analyses of optically homogeneous anorthoclase phenocrysts (G2B2, HR1, and B29A on Figure 2) fall within the two-feldspar field of Smith (1974).

In comparison with other reported natural feldspars, those from the dikes are strongly ternary in composition (*i.e.* K-rich plagioclase; Ca-rich alkali feldspar), and therefore they preserve convincing chemical evidence of a high temperature crystallization history. The distribution of analyses in Figure 2, however, suggests that in most cases where pairs were analyzed, the sanidine may not be in equilibrium with the plagioclase. Application of two- feldspar thermometry, using the algorithm of Fuhrman and Lindsley (1988) with the thermodynamic parameters of Elkins and Grove (1990), converged "with tolerance" only for the feldspar pair

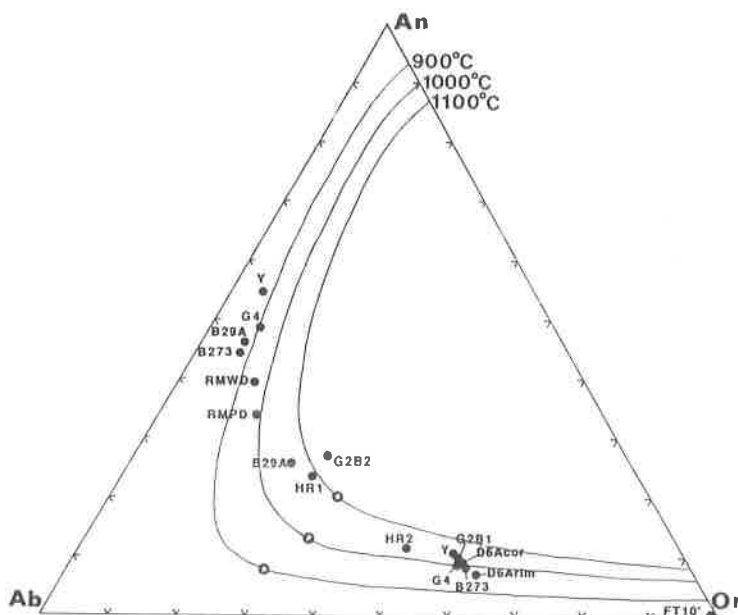


Figure 2. Compositions of analyzed feldspar phenocrysts (Tables 2 and 3) from the Piedmont dike suite. The three isotherms are from 1 kbar data of Fuhrman and Lindsley (1988); the stars indicate the position of the consolute on each isotherm. The analysis labelled FT-10' (at the Or apex) is a thin rim, of apparently low-temperature origin, on a sanidine phenocryst.

from sample B273. For this pair, the average temperature calculated at a pressure of one kbar is 1011°C , with $T_{\text{Ab}}=1021.8^{\circ}$, $T_{\text{Ks}}=1024.6^{\circ}$, and $T_{\text{An}}=985.4^{\circ}$. The non-convergence for other analyses may imply that they do not represent equilibrium pairs.

Quartz and Apatite

The felsic dike rocks contain abundant phenocrysts of quartz. The rounded hexagonal dipyramidal forms, and complete lack of any other SiO_2 polymorphs indicates primary crystallization within the stability field of beta-quartz; however, some quartz phenocrysts are deeply embayed, suggesting partial resorption by the magma. This is especially true of quartz in the flow-banded rocks. Euhedral to subhedral apatite prisms or needles are very common in all rocks of the suite, especially the mafic ones, in which they reach three modal percent. This correlates well with the exceedingly high P_2O_5 in these rocks (Table 7B).

Pyroxene

Sparse phenocrysts of a monoclinic, pleochroic (greenish-tan) pyroxene occur in many of the felsic dike samples and in the flow-banded samples. They appear to be optically homogeneous, except for very narrow reaction rims where they are in contact with the fine groundmass. Table 4 presents four pyroxene analyses from silicic dike rocks and one from a flow-banded rock. These pyroxenes are ferropigeonite (Fs 58-71); they possess high Ca contents compared

Table 4. Pyroxene and olivine analyses

Locality* Sample Grain	G3 CD3 px8	G3 CD3 px1	H2 B273 px5	H2 B273 px3	G2 GS2B px1	T1 Y px	G2 GS2B oliv
weight percent oxides							
SiO ₂	47.36	47.42	47.50	47.62	48.50	52.41	37.96
Al ₂ O ₃	1.64	0.42	1.97	2.03	2.09	0.67	0.00
TiO ₂	0.59	0.38	0.67	0.56	0.62	0.37	0.03
Cr ₂ O ₃	0.03	0.04	0.04	0.00	0.09		0.04
FeO	39.50	40.07	36.24	36.22	32.03	23.14	22.74
MnO	0.85	0.80	0.88	0.77	0.83	0.59	0.31
MgO	4.42	6.09	6.99	7.30	8.48	19.42	39.10
CaO	6.61	4.53	6.18	6.00	6.00	3.92	0.31
Na ₂ O	0.16	0.06	0.21	0.19	0.16	0.03	
total	101.16	99.81	100.66	100.69	98.80	100.55	100.49
site occupancy (per 3 oxygens)							
Si	0.967	0.983	0.961	0.960	0.979	0.985	(4 oxy) 0.985
Al ^{iv}	0.033	0.010	0.039	0.040	0.021	0.015	
Al ^{vi}	0.006		0.008	0.008	0.029	0.000	
Ti	0.024	0.006	0.010	0.009	0.009	0.005	0.001
Cr	0.001	0.001	0.001	0.000	0.002		0.001
Fe	0.675	0.694	0.613	0.611	0.540	0.364	0.493
Mn	0.015	0.014	0.015	0.013	0.014	0.009	0.007
Mg	0.134	0.188	0.211	0.219	0.255	0.544	1.512
Ca	0.145	0.101	0.134	0.130	0.130	0.079	0.009
Na	0.006	0.003	0.008	0.007	0.006	0.001	
Σ vi	1.006	1.007	1.000	0.997	0.985	1.002	2.023
mole percent endmembers							
En	14.1	19.1	22.0	22.8	27.6	55.1	Fo 75.4
Fs	70.8	70.6	64.0	63.7	58.4	36.9	Fa 24.6
Wo	15.2	10.3	14.0	13.5	14.0	8.0	

* see Figure 1.

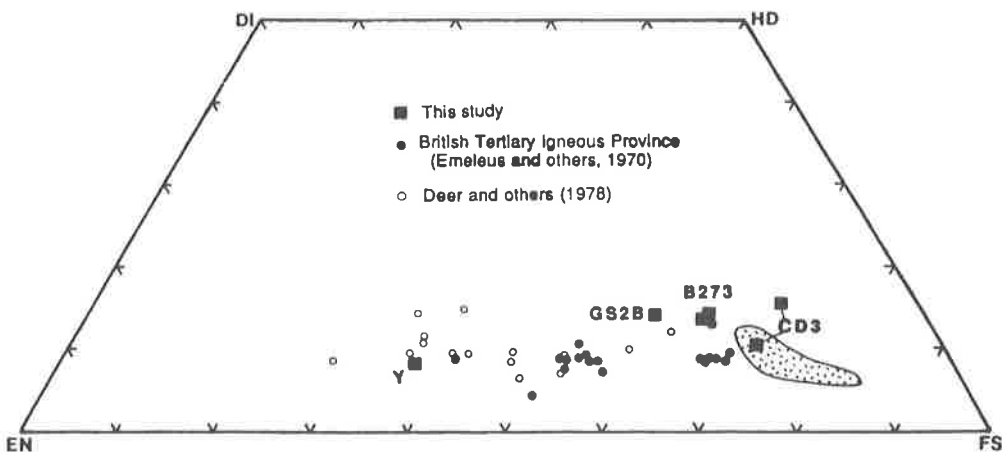


Figure 3. Compositions of pigeonite and ferropigeonite phenocrysts (Table 4) from the Piedmont suite. Some pyroxenes from the literature are shown for comparison. Stippled area represents analyses reported by Deer and others (1978) from the Nain, Skaergaard, and Bjerkrem-Sogndal intrusions.

with other ferropigeonites reported in the literature (Wo 10-15 range). A more Mg-rich pigeonite (En 55) from a sample of mafic dike rock (Y; see Table 4) was also analyzed. Figure 3 is a plot of the compositions of these pyroxenes on the pyroxene quadrilateral. Also shown are analyses reported in Deer and others (1978) and pigeonites from Scotland (Emeleus and others, 1971), which are most similar to the eastern Piedmont suite.

Olivine

A single olivine phenocryst, with a thin reaction rim, was observed and analyzed from a flow-banded sample (GS2B1; see Table 4). Its Mg-rich composition (Fo75, Fa25) compared to the relatively Fe-rich pigeonite (Wo14, En28, Fs 58) from the same rock indicates that the phenocrysts may not represent an equilibrium assemblage in this rock. The sample also contains embayed quartz phenocrysts, further attesting to disequilibrium. The olivine may be a xenocryst; its composition is within the range of olivines formed during early fractionation of the Palisades sill (Walker, 1969).

Amphibole

Microphenocrysts of brown, slightly pleochroic amphibole are very common in the felsic dike rocks. In the coarsest rocks, these range up to only 0.1 mm, in contrast to the phenocryst phases, which may exceed 5 mm. Analyses of the amphiboles (Table 5) show them to be Na, K, Fe-rich solid solutions, with the "A" site about 80% full (with subequal amounts of Na and K). Fe/(Fe + Mg) is very high (.87 - .96; weight percent basis) in these amphiboles, higher than that of coexisting ferropigeonite phenocrysts. Although Fe³⁺ has not been determined, stoichiometric constraints require that considerably less than 1.00 cation of Fe³⁺ per formula unit be present in these amphiboles (per 23 oxygens, on an anhydrous basis); the presence of Ti in the octahedral sites helps balance the charge. Using IMA amphibole nomenclature (Rock and Leake, 1984), the analyzed amphiboles would be classified as any of several different varieties of hornblende (hastingsitic, edenitic, and pargasitic). It is noteworthy that these amphiboles are exceedingly rich in F and Cl: on the average, these halogens occupy greater than 90% of the (OH) sites, suggesting that the magmas which produced these rocks were highly fractionated and may have been nearly anhydrous. In an experimental study of the Na-hornblende hastingsite, Thomas (1982) noted that the mineral is typically found in the "differentiated, Fe-rich members of alkaline and calc-alkaline plutonic suites" and that its occurrence is commonly associated with low oxygen fugacity conditions.

Fe-Ti Oxides

Fe-Ti oxide minerals are common phenocrysts in all samples; a rhombohedral phase (ilmenite) and a spinel phase (ulvospinel-rich solid solution) coexist in many samples (Delorey, 1983). Table 6 shows analyses of the Fe-Ti oxide minerals. The spinels in these rocks are extremely rich in Ti (*cf.* Deer and others, 1962). Small-scale exsolution lamellae of ilmenite were locally observed in reflected light in the spinels, although individual grains were generally homogeneous under the microprobe beam, except for local marginal alteration. For all spinels analyzed,

Table 5. Amphibole analyses

Locality* Sample Grain	G3 CD3 am5	G3 CD3 am7	G3 CD3 am10	H3 CD6A am2	H3 CD6A am4	H3 CD6A am1	H2 B273 am2
weight percent oxides							
SiO ₂	40.68	39.19	40.30	39.40	40.43	39.08	40.64
Al ₂ O ₃	8.72	8.78	7.98	8.69	8.38	8.95	8.10
Cr ₂ O ₃	0.02	0.03	0.17	0.00	0.10	0.02	0.04
TiO ₂	1.55	2.01	1.71	1.85	1.65	2.02	1.87
FeO	30.81	31.78	30.97	29.91	29.55	30.82	28.13
MnO	0.26	0.40	0.27	0.39	0.36	0.38	0.36
MgO	1.52	1.22	1.77	2.37	2.23	1.98	4.07
CaO	10.11	10.45	10.09	10.50	9.67	10.56	10.46
Na ₂ O	2.07	1.74	1.51	1.69	1.64	1.68	1.78
K ₂ O	2.13	1.90	1.70	1.86	1.61	1.90	1.67
Cl	1.09	1.41	1.38	1.16	0.96	1.29	0.93
F	0.79	1.07	0.72	1.37	1.18	1.15	1.54
Subtotal	99.75	99.98	98.57	99.19	97.76	99.83	99.59
O=Cl+F	0.58	0.77	0.70	0.84	0.71	0.77	0.86
total	99.17	99.21	97.87	98.35	97.05	99.06	98.73
site occupancy (per 23 oxygens on anhydrous basis)							
Si	6.571	6.394	6.610	6.427	6.590	6.381	6.541
Al ^{iv}	1.429	1.606	1.390	1.573	1.410	1.619	1.459
Al ^{vi}	0.231	0.082	0.152	0.098	0.200	0.103	0.078
Fe	4.162	4.317	4.161	4.046	3.993	4.114	3.667
Mn	0.035	0.054	0.036	0.054	0.050	0.052	0.048
Mg	0.366	0.296	0.433	0.575	0.542	0.480	0.975
Ti	0.188	0.247	0.211	0.227	0.202	0.248	0.226
Cr	0.003	0.004	0.007	0.000	0.013	0.003	0.006
Ca	1.750	1.827	1.772	1.835	1.689	1.847	1.804
Fe		0.020	0.086	0.034	0.036	0.094	0.118
Mn		0.001					
Na	0.250	0.152	0.142	0.131	0.275	0.059	0.078
Na(A)	0.398	0.397	0.338	0.402	0.243	0.471	0.476
K	0.439	0.396	0.356	0.386	0.335	0.395	0.344
ΣA	0.837	0.793	0.694	0.788	0.578	0.866	0.820
Cl	0.298	0.388	0.415	0.322	0.266	0.356	0.254
F	0.404	0.553	0.442	0.707	0.610	0.592	0.784

* see Figure 1.

ulvospinel and magnetite together account for 90-94% of their compositions; they range from $\text{usp}_{51}\text{mt}_{49}$ to $\text{usp}_{82}\text{mt}_{18}$. The composition of the rhombohedral phase is nearly pure end-member ilmenite, ranging from $\text{ilm}_{91}\text{hem}_{09}$ to $\text{ilm}_{95}\text{hem}_{05}$. If, as discussed by Delorey (1983), the Fe-Ti oxide phases in these rocks have not undergone significant low-temperature modification of their compositions, pairs can be used as $f\text{O}_2$ -T indicators (Buddington and Lindsley, 1964). Using the oxide phases from sample GS-1, approximate conditions would be about 1000°C at $f\text{O}_2$ of about 10^{-11} or 10^{-12} atm, or between the FMQ and WM oxygen buffers (see Lindsley, 1976). This temperature is similar to that calculated from two-feldspar geothermometry, and is consistent with the fact that beta-quartz is the stable SiO_2 phase, if the pressure of crystallization of the phenocryst assemblage was above about 0.5 kbar (using the P-T relations of SiO_2 given by Morse (1980)).

Secondary Minerals

Amygdules filled with secondary minerals are common in these rocks. The

Table 6. Fe-Ti oxide analyses

Locality* Sample Mineral	G4 GS4 ilm	G1 GS1 ilm	H3 CD6A ilm	R1 CD7F ilm	G3 CD3 spl	G1 GS1 spl	H2 B273 spl	H1 B29A spl
weight percent oxides								
SiO ₂	0.05	0.54	0.05	0.05	0.33	0.23	0.26	0.15
Al ₂ O ₃	0.11	0.26	0.11	0.30	2.04	2.71	0.86	1.10
Cr ₂ O ₃	0.13	0.08	0.12	0.07	0.05	0.07	0.09	0.11
TiO ₂	47.29	48.77	48.99	47.14	19.03	22.84	20.54	21.09
FeO	48.12	46.46	48.21	48.92	70.69	64.64	72.15	69.86
MnO	0.53	2.33	0.67	0.58	1.85	3.82	1.64	2.41
MgO	1.22	0.25	0.46	0.54	0.00	0.00	0.02	0.00
CaO	0.11	0.61	0.08	0.09	0.11	0.21	0.26	0.11
total	97.56	99.30	98.69	97.69	94.10	94.52	95.82	94.83
FeO**	39.73	40.91	42.51	40.79	46.54	47.84	48.17	47.64
Fe ₂ O ₃ **	9.32	6.16	6.33	9.04	26.84	18.68	26.65	24.69
total**	98.49	99.92	99.32	98.60	96.79	96.39	98.49	97.30
cations per	3 oxygens				4 oxygens			
Si	0.001	0.014	0.001	0.001	0.013	0.009	0.010	0.006
Ti	0.906	0.923	0.935	0.907	0.551	0.659	0.588	0.610
Al	0.003	0.008	0.003	0.009	0.093	0.123	0.039	0.050
Cr	0.003	0.002	0.002	0.001	0.002	0.002	0.003	0.003
Fe ³⁺	0.179	0.117	0.121	0.174	0.778	0.539	0.763	0.715
Fe ²⁺	0.847	0.861	0.903	0.872	1.499	1.535	1.533	1.533
Mn	0.011	0.050	0.014	0.013	0.060	0.124	0.053	0.079
Mg	0.046	0.009	0.017	0.021	0.000	0.000	0.001	0.000
Ca	0.003	0.016	0.002	0.002	0.005	0.009	0.011	0.005

* see Figure 1.

** recalculated based upon stoichiometry

amygdules constitute considerably less than ten volume percent of the rock, usually less than five percent. Most often containing a dark green or brownish, nearly isotropic, chloritic or clay-like material, they may also contain a carbonate mineral, or less commonly, chalcedonic silica. It should be emphasized that, aside from the amygdules, these rocks are free of any apparent alteration and are exceedingly fresh except for an approximately one cm-thick weathering rind.

WHOLE ROCK GEOCHEMISTRY

Eleven whole-rock analyses for major elements and selected trace elements (Rb, Sr, Y, and Zr) are shown in Table 7.

Silicic Dike Rocks

These rocks (Table 7A) are characterized by SiO₂ values ranging from 61 to 67 weight percent, and by quite high Fe (total Fe as Fe₂O₃ in excess of 8.0 weight percent) and TiO₂ (0.7-1.0 weight percent). K₂O greatly exceeds Na₂O (K₂O/Na₂O > 2.0); and total alkalis are moderately high (6-7 wt. percent). On the basis of modal analyses of phenocrysts, and by most traditional classifications (e.g. Travis, 1955), these rocks would be called rhyolite porphyry. Though they were called trachytes by Spence and McDaniel (1979), they have abundant quartz phenocrysts, and are high in normative quartz (27-35 percent). By the IUGS classification, using either phenocrysts or CIPW norms, these rocks are rhyolites

Table 7. Whole-rock analysis

A. Silicic rocks

Sample Locality*	HR1 R1	HR1A R1	HR1B R1	RW20 L2	FT10' H1	FT10 H1
oxides (weight percent)						
SiO ₂	64.6	61.5	66.4	66.8	64.3	66.2
TiO ₂	0.69	0.89	0.87	0.78	0.93	0.97
Al ₂ O ₃	12.8	13.5	12.4	12.0	12.5	12.6
Fe ₂ O ₃ **	4.81	5.37	4.41	1.55	2.01	3.96
FeO	3.98	3.03	3.86	5.80	6.63	4.42
MnO	0.24	0.15	0.15	0.15	0.29	0.30
MgO	0.40	0.25	0.28	0.30	0.76	0.39
CaO	2.21	2.84	2.72	2.96	2.28	1.45
Na ₂ O	1.8	2.1	2.2	2.1	2.0	1.9
K ₂ O	4.09	4.61	4.31	4.02	4.39	4.72
P ₂ O ₅	0.25	0.37	0.35	0.36	0.34	0.28
L.O.I.	3.5	n.d.	2.7	2.1	2.4	n.d.
Total	99.37	94.61	100.65	98.92	98.83	97.19
Selected trace elements (ppm)						
Rb	137	145	84	116	136	180
Sr	129	132	105	224	120	155
Y	79	89	91	82	142	121
Zr	690	920	920	820	885	963
CIPW Norms						
ap	0.6	0.9	0.9	0.9	0.8	0.7
il	1.4	1.8	1.7	1.5	1.8	1.9
ml	7.3	8.1	6.5	2.3	3.0	5.9
or	25.2	28.8	26.0	24.5	26.9	28.7
ab	15.9	18.8	19.0	18.4	17.6	16.5
an	9.7	12.3	11.4	11.8	9.4	5.5
di	--	--	--	0.8	--	--
hy	3.8	0.7	3.1	9.0	11.8	4.9
q	34.1	27.7	31.5	30.8	27.5	33.4
c	2.1	0.8	--	--	1.2	2.5

B. Intermediate to mafic rocks

Sample Locality*	RWD1 L1	Y T1	Y-N T2	JKS1 T3	V6F V1
oxides (weight percent)					
SiO ₂	57.1	52.0	52.9	53.2	52.8
TiO ₂	1.97	2.64	2.57	2.44	2.87
Al ₂ O ₃	15.5	13.5	13.0	12.9	12.9
Fe ₂ O ₃ **	5.20	4.22	5.17	3.84	4.24
FeO	5.04	8.26	8.15	9.14	9.95
MnO	0.26	0.26	0.25	0.22	0.25
MgO	1.62	1.88	2.49	2.28	2.78
CaO	5.41	6.22	6.09	5.87	6.58
Na ₂ O	2.6	2.4	2.1	2.1	2.3
K ₂ O	3.30	3.08	3.00	3.11	2.61
P ₂ O ₅	1.1	1.7	1.4	1.4	1.5
L.O.I.	2.1	n.d.	1.8	2.6	1.6
Total	101.20	96.16	98.92	99.10	100.38
Selected trace elements (ppm)					
Rb	84	130	45	50	70
Sr	400	440	265	235	300
Y	50	55	54	52	56
Zr	450	365	475	455	415
CIPW Norms					
ap	2.6	4.2	3.4	3.4	3.6
il	3.8	5.2	5.0	4.8	5.5
ml	7.6	6.4	7.7	5.8	6.2
or	19.7	18.9	18.3	19.1	15.6
ab	22.2	21.1	18.3	13.4	19.7
an	19.8	17.7	17.7	17.2	17.4
di	--	2.5	3.3	3.0	4.8
hy	6.3	11.8	11.9	14.7	15.2
q	17.6	12.4	14.5	13.7	12.0
c	0.5	--	--	--	--

* see Figure 1.

** determined by Pratt method (after Jeffery and Hutchison, 1981)

L.O.I.: weight loss on ignition

n.d.: not determined

(Streckeisen, 1979). Compared to most analyzed rhyolites, however, they are silica-poor and iron-rich. Using various other volcanic rock classifications (for example, LeBas and others, 1986), they would be dacites. They show some similarities, at least in terms of major element chemistry, to icelandites from Thingmuli, described by Carmichael (1964), to British ferrorhyolites, referred to by Nockolds and others (1978), and to other rocks of the British Tertiary Igneous Province.

Mafic and Intermediate Dike Rocks

Four analyses of mafic dike rocks and one of an intermediate dike rock are presented in Table 7B. They are quite rich in TiO₂, P₂O₅, and K₂O, low in MgO, and high in Fe/Mg and K/Na. Using the phenocryst assemblage as a basis for classification, these rocks would be called basalt (Travis, 1955), but using the norms as the basis, they would be andesite, K-rich andesite, or possibly basaltic andesite (Streckeisen, 1979; Irvine and Baragar, 1971). They bear chemical and mineralogical similarity to intermediate members of some oceanic island associations (*e.g.* trachyandesite, mugearite), and to the so-called shoshonites of continental interiors (*e.g.* Joplin, 1968), but they are strongly quartz-normative (see Carmichael and others, 1974; Nockolds and others, 1978). The intermediate dike rock (RWDI), is chemically very similar to the four mafic examples, except that its

chemistry suggests it is somewhat more differentiated ($\text{SiO}_2 = 57\%$ vs. 52-53% for the more mafic rocks; total alkalis = 5.9 vs. 4.8-5.5%). Using the rock classification of LeBas and others (1986), the composition of RWD1 falls near the boundary between andesite and latite, while the more mafic samples are basaltic andesite and shoshonite.

Rare-earth and other Trace Elements

Relative to most common rocks of similar degree of apparent differentiation (as suggested by weight percent SiO_2 or MgO), the analyses of Table 7 are higher in Zr, Y, and Rb/Sr, as well as TiO_2 and P_2O_5 . Three analyses of REE in mafic rocks of the dike suite show that the suite is highly enriched in rare-earth elements, especially light rare earths (for example, $\text{La} = 110\text{-}260 \times$ chondritic; Table 8). Chondrite-normalized patterns for REE (Figure 4; chondrite values of Hanson (1980) used) have fairly steep negative slopes, in marked contrast to the low abundances and flat or very gentle positive slopes of typical tholeiitic basalts (*e.g.* Bryan and others, 1977).

Table 8. Rare-earth element analyses*

Sample** Locality***	Y T1	RMWD L1	LK26 R3	LK47 R3
			ppm	
La	86.0	77.0	37.0	1.8
Ce	180	180	90	5
Nd	110	97	52	<5
Sm	23.0	19.0	9.9	1.2
Eu	5.7	5.0	2.8	0.6
Tb	3	3	2	<1
Dy	17.4	14.3	9.9	3.4
Yb	8.0	7.1	4.6	2.4
Lu	1.20	1.10	0.71	0.37

*Analyses by XRAL Laboratories, Don Mills, Ontario.

**The first three are mafic members of the alkali-rich suite;

LK47 is a sample of typical olivine diabase from the area.

*** see Figure 1.

COMPARISON TO EARLY MESOZOIC DIABASE

Studies of the relatively well-known early Mesozoic diabase dikes common in the Appalachian Province (ENA) have distinguished several groups of diabases on the basis of chemical criteria (Weigand and Ragland, 1970; Ragland, 1991; Ragland and others, 1971; see also Manspeizer and others, 1989). Although diabase dikes occur all along the Appalachian trend, quartz-tholeiites are more abundant in the north, while in the south, olivine-tholeiites predominate.

During mapping of the dikes for the present study, tholeiitic diabase dikes were commonly encountered; all are olivine-bearing. Chemical comparisons, using selected major, minor and trace elements, between analyzed samples from the dike suite of this study and the early Mesozoic diabases are given in Figure 5 and Table 9. In addition to the average analyses from the literature (Ragland, 1991), nine samples of olivine diabase from this study area were also analyzed for selected elements for comparison.

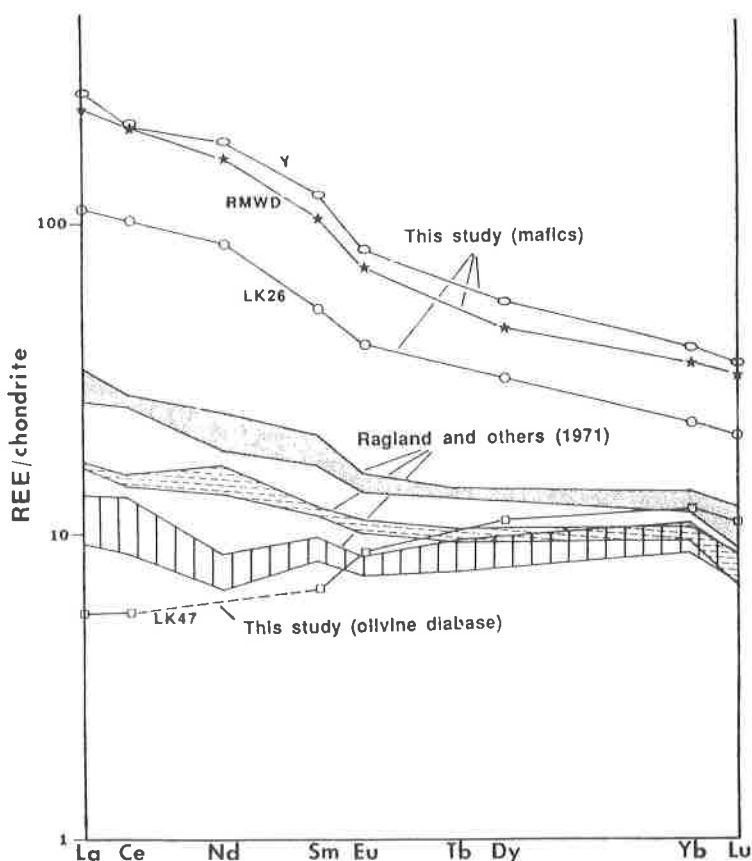


Figure 4. Chondrite-normalized rare earth element patterns for three samples of intermediate to mafic dike rocks from the Piedmont suite, compared with a pattern determined for one sample of olivine diabase (boxes) from the same area. Ranges of patterns determined by Ragland and others (1971) for E.N.A. diabase groups are shown by stippling (high-Ti, quartz normative), dashes (low-Ti, quartz normative), and vertical ruling (olivine normative).

In addition to showing REE patterns for three mafic members of the alkali-rich suite, Figure 4 shows one pattern of an olivine diabase from the study area, as well as patterns determined by Ragland and others (1971) for the major eastern North American (ENA) diabase types. None of the more silicic dikes have been analyzed for REE.

The chemical distinctions between the alkali-rich suite and the tholeiitic diabbases, as exemplified in Table 9 and Figures 4 and 5, are obvious. They raise the strong likelihood that the alkali-rich suite and the diabase suite are not comagmatic. If they were comagmatic, then a diabase magma must have been drastically modified to yield the members of the alkali-rich suite.

A UNIQUE OCCURRENCE?

Southern Appalachian Dikes

Several other occurrences of post-Paleozoic igneous activity have been

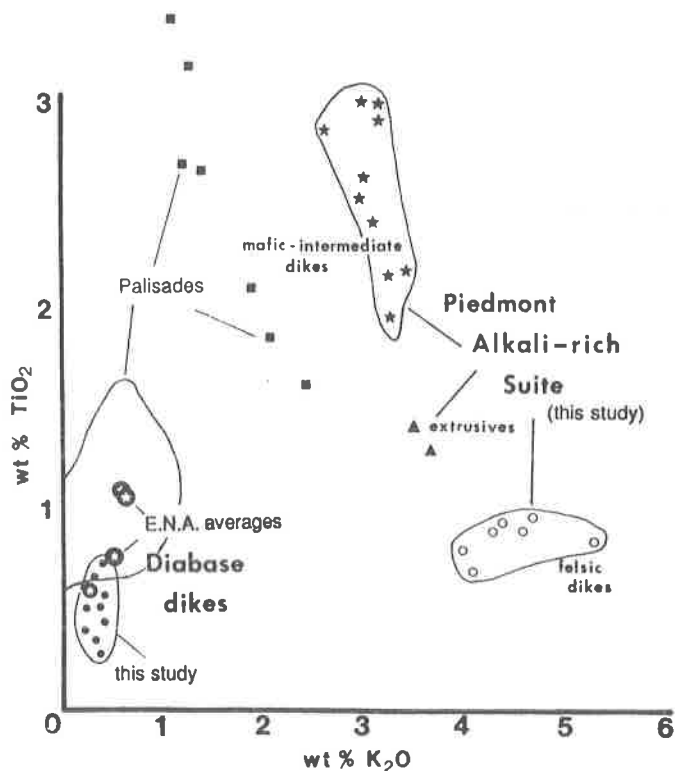


Figure 5. TiO_2 vs. K_2O plot for samples of the Piedmont suite (stars, triangles, and open circles), and for samples of olivine diabase from the same area (dots). Open stars represent averages for groups of E.N.A. diabase (Weigand and Ragland, 1970). Also shown for comparison are analyses of diabase (shown by field centered at 1% TiO_2) and of more evolved samples (individual squares) from the Palisades intrusion (Walker, 1969; Pearce, 1970).

Table 9. Geochemical comparison of Piedmont suite with E.N.A. averages

Group	Alkali-rich Suite		Local olivine diabase	E.N.A. Averages*			
	silicics	mafic		Ol	hi-Fe Q	hi-Ti Q	lo-Ti Q
number of analyses**	12-18	7-10	9	156-222	10-48	17-20	30-37
weight percent oxide							
TiO_2	0.88 ± 0.09	2.59 ± 0.40	0.56 ± 0.16	0.58 ± 0.13	1.13 ± 0.14	1.12 ± 0.05	0.76 ± 0.05
K_2O	4.52 ± 0.04	3.13 ± 0.23	<0.5	0.31 ± 0.17	0.76 ± 0.24	0.66 ± 0.15	0.48 ± 0.15
P_2O_5	0.34 ± 0.04	1.2 ± 0.5	<0.2				
ppm							
Sr	143 ± 32	352 ± 75		112 ± 44	195 ± 57	186 ± 16	127 ± 12
Rb	133 ± 31	93 ± 33		9.9 ± 6.2	18 ± 10	21 ± 6	15 ± 5
Zr	858 ± 93	403 ± 46		48 ± 15	87 ± 30	92 ± 8	60 ± 9

*Averages for eastern North America diabase groups from Ragland (1991) and Weigand and Ragland (1970). Groups are: olivine-diabase with olivine control (column A from Table 10-3 of Ragland (1991)); high-Fe, quartz-diabases (column F from Table 10-3 of Ragland (1991)); high-Ti, quartz-normative and low-Ti, quartz-normative diabases from Weigand and Ragland (1970).

**Some analyses do not include all oxides or trace elements reported here. Lowest figure in each range reflects the minimum number of analyses averaged for each element or oxide reported.

reported from the southeast (see Sutter, 1976). Aside from the common early Jurassic tholeiite dikes, most prominent are the well-documented late Jurassic and Eocene igneous provinces of northwestern Virginia and adjacent West Virginia (Fullagar and Bottino, 1969; Dennison and Johnson, 1971; Johnson and others, 1971). The late Jurassic suite is strongly alkalic and undersaturated, characteristically containing olivine, nepheline, or aegirine in rocks of appropriate compositions (Johnson and others, 1971). Analcite and natrolite are also common. Rocks of this suite belong to a series of alkalic intrusions of the central and eastern U.S. (Zartman, 1977).

The younger (Eocene) suite bears more similarities to the rocks under consideration, being bimodal in nature and consisting largely of basalts and andesites (or felsites) according to Johnson and others (1971); still, the basalts are olivine-rich, and the more felsic rocks contain biotite and hornblende phenocrysts, and not quartz and sanidine. Apparently no chemical analyses of these rocks have been published, although Gray and Gottfried (1986) provide sparse data on U, Th, and Nb.

Amygdaloidal "diabase" dikes and associated "rhyolite dikes" in east-central Georgia have been described by Crawford and others (1966). In addition, porphyritic "post-orogenic" felsic dikes in the central Piedmont of South Carolina, were reported by Ritchie and Fallaw (1978). Chemical analyses are not available, but a cursory petrographic examination (Stoddard, unpublished) of these rocks suggests that they are not related to the suite in question, for the following reasons. All have undergone significant low-temperature alteration and/or metamorphism. The felsic dikes are microgranite or felsite, the mafic dike appears to be a metadiabase, and these dikes are probably pre-Alleghanian.

Lamprophyre dikes also occur in the North Carolina Piedmont (Mauger, 1988). Although the Mesozoic suite under consideration overlaps them in terms of some major-element characteristics, the lamprophyres are considerably older (late Paleozoic), have higher $Mg/(Fe+Mg)$ and $K/(K+Na)$ and are more potassic. The lamprophyres have higher Zr, Sr, and Rb, but lower Y.

Northern Appalachian Dikes

McHone and Butler (1984) have summarized the state of knowledge concerning the Mesozoic intrusions of the New England area by dividing them into four groups. These are, from oldest to youngest: (1) a Coastal New England province, consisting of a very few older (210-240 Ma) alkalic intrusives; (2) the well-known White Mountain Magma Series, which they limit geographically to northern New Hampshire, and which consists of 160-200 Ma plutons; (3) the widespread 160-200 Ma ENA group, consisting predominantly of tholeiitic dikes and flows, but with which they would apparently also include the mildly alkalic basaltic dikes described by Hermes and others (1984), as well as some described by McHone and Trygstad (1982); and (4) a New England - Quebec province, consisting mainly of alkalic plutons of early Cretaceous age, exemplified by the Monteregian Hills.

Rocks of the White Mountain Magma Series and Monteregian Hills bear some similarity, both in age and in overall geochemistry, to the alkali-rich Piedmont suite. However, the White Mountain and Monteregian Hills rocks include a much wider range in levels of emplacement and in rock types, including abundant SiO_2 -undersaturated lithologies, and are geometrically and spatially unrelated to the early

Mesozoic rift basins or to the diabase dikes associated with those basins and the adjacent Piedmont.

Other post-metamorphic dikes from New England have been described by Hermes and others (1984). The authors suggest that the mildly alkalic dikes may have been generated from deeper mantle sources than typical ENA magmas, and point to the fact that the alkalic province lies within crystalline basement well removed from the axis of the nearest Mesozoic basin, whose diabases are tholeiites. This is also true of the North Carolina dikes. However, although the "transitional alkalic" rocks from New England (Hermes and others, 1984) exhibit a great range in degree of silica saturation (from nepheline- and olivine-normative to quartz-normative), they contain olivine phenocrysts and no alkali feldspar, and are still considerably lower in K, P, and Fe/Mg than the North Carolina - Virginia Piedmont suite.

DISCUSSION

Relationship Among Rocks of the Suite

The Piedmont rocks may be truly bimodal, each compositional group having a distinct origin. On the other hand, they may be interpreted as comprising a geochemically coherent suite related by conventional mechanisms of magmatic differentiation. The following observations support such a conclusion. First, Rb and Sr isotopic analyses are consistent with this hypothesis, because they lie on a fairly well-defined isochron with an initial Sr ratio of 0.7067 (Stoddard and others, 1986). Although analyses are few in number, the initial Sr ratios calculated for felsic samples alone (0.707) and for the three more mafic samples (0.706) compare favorably.

Second, the suite of phenocryst minerals supports a comagmatic hypothesis. The pigeonite phenocrysts follow an iron-enrichment trend in going from the mafic to the silicic rocks (Fig. 3). The plagioclase phenocrysts become somewhat more sodic in the silicic rocks. K-feldspar and quartz become important liquidus phases in the silicic rocks.

Major-element geochemistry shows iron and potassium enrichment, but depletion of TiO_2 and P_2O_5 in going from the mafic to the silicic dike rocks. This suggests that the more silicic samples may be related to the more mafic by crystal fractionation involving plagioclase, low-Ca pyroxene, and, importantly, apatite and Fe-Ti oxides minerals. Calculations using such a crystallizing assemblage, with the composition of mafic rock V6F as parent magma, are capable of producing major-element characteristics of the silicic rocks (Table 10). In terms of trace elements, Zr and Rb behave incompatibly, increasing in abundance going from the mafic to the silicic rocks, but Sr decreases, as would be expected with plagioclase fractionation. Sparse phenocrysts of a very high-relief accessory mineral were observed in some thin sections of the mafic dike rocks. They were not positively identified, but are probably monazite, in keeping with the high light rare-earth elements (LREE) and P_2O_5 contents of these rocks. However, the fact that Y and Zr increase in the silicic rocks argues against the unknown mineral being xenotime, zircon, or monazite. REE analyses of the silicic rocks are not available; they would be helpful to evaluate the possible roles of accessory minerals and of plagioclase fractionation.

Table 10. Crystal fractionation model for derivation of silicic magma (=HR1B) from mafic magma (=V6F)

Wt. %	V6F= Parent magma	HR1B= Residual liquid	Calc** diff	Crystallizing assemblage***				
				pyx 13.18%	pla 23.76%	ilm 2.12%	mag 7.68%	apa 3.62%
SiO ₂	52.80	66.4	-0.05	52.40	54.67	0.00	0.00	0.00
TiO ₂	2.87	0.87	-0.03	0.37	0.10	47.65	18.00	0.00
Al ₂ O ₃	12.90	12.4	-0.07	0.67	27.45	0.50	2.50	0.00
Fe(T)*	13.77	7.83	-0.03	23.14	0.71	48.70	73.69	0.00
MnO	0.25	0.15	0.00	0.59	0.00	0.51	1.07	0.00
MgO	2.78	0.28	0.01	19.42	0.13	1.90	0.00	0.00
CaO	6.58	2.72	0.10	3.92	10.96	0.11	0.20	55.00
Na ₂ O	2.30	2.2	0.13	0.03	4.54	0.00	0.00	0.00
K ₂ O	2.61	4.31	0.28	0.00	0.81	0.00	0.00	0.00
P ₂ O ₅	1.50	0.35	-0.20	0.00	0.00	0.00	0.00	42.00

*total Fe as FeO

**For given oxide, this column represents the difference between the composition of V6F and the sum of the oxide in the crystallizing assemblage plus the residual liquid (HR1B). Residual sum of squares = 0.1502.

***pyroxene and plagioclase compositions from sample Y, Tables 3 and 4; ilmenite and magnetite compositions estimated from partial microprobe analyses; apatite composition assumed.

It is thus plausible to infer that these unusual dike rocks are related to one another. The question remains, however, of the origin of the magma(s) which gave rise to the suite. A definitive conclusion is beyond the scope of this paper, but some questions can be addressed with the data at hand.

A Product of Intraplate Magmatism?

The Piedmont rocks are distinguishable from products typical of plate-margin magmatism on the basis of their: (1) enrichment in incompatible elements analyzed (K, Ti, P, Ba, Rb, Sr, Zr, and Y); (2) high K₂O/Na₂O (>1, even at SiO₂ as low as 52%); (3) high Rb/Sr; (4) enrichment in LREE, and steep negative slope of REE patterns; (5) high total Fe; and (6) high Fe/Mg. These properties raise the suggestion that the Piedmont suite may be more akin to intraplate volcanics. If this is the case, then these rocks may represent a separate magma type which was involved in the initial rifting associated with the opening of the Atlantic. They may be the result of fractionation of a partial melt of primitive mantle material enriched in incompatible elements, such as has been associated with hotspots over relatively stationary mantle plumes (*e.g.* Anderson, 1981).

Still, some obvious and possibly significant differences distinguish these rocks from well-documented suites of oceanic islands (*e.g.* LeRoex, 1985), or continental rift or intracratonic magmas (*e.g.* Stolz, 1985). First, although they show alkalic tendencies, none of these rocks is peralkaline, and all (even the most mafic analyzed) contain significant amounts of normative quartz; nearly all thin sections examined contain abundant modal quartz as well. Also, despite the fact that the Piedmont rocks mineralogically resemble those of intracratonic suites (except for complete lack of feldspathoids and of olivine in the more mafic representatives), their Fe/Mg ratios are higher than those from most similar rocks in the literature.

Other objections to this hypothesis can be made. First, no rocks have been found in the area which might be representative of a more "primitive" magma from

which the dikes might have been derived, although they may be (have been) present at another, unexposed level. Second, no deep crustal or mantle xenoliths have been found in the dike rocks, although shallow crustal, granitic xenoliths are present.

A Tholeiitic Derivative?

Because the rocks under consideration here are apparently coeval and occur in dikes having similar N and NW trends as the widespread early Mesozoic tholeiites (Stoddard and others, 1986), the possibility that they represent an extreme late-stage fractionate of a tholeiitic magma should be considered. Such differentiates have been well documented, for example, in the Palisades intrusion (Walker, 1969; Pearce, 1970), the Great Lake and Red Hill intrusions of Tasmania (McDougall, 1964), and the Pageland dike, South Carolina (Steele, 1974; Steele and Ragland, 1976), as well as the Skaergaard intrusion (Wager, 1960). It is evident that, in terms of some chemical components (for example A-F-M, Fig. 6) the Piedmont dikes could be related to a tholeiitic parent by simple crystal fractionation. However, most other chemical characteristics of the suite are difficult or impossible to explain by such a mechanism (*e.g.* the high K/Na and perhaps the REE).

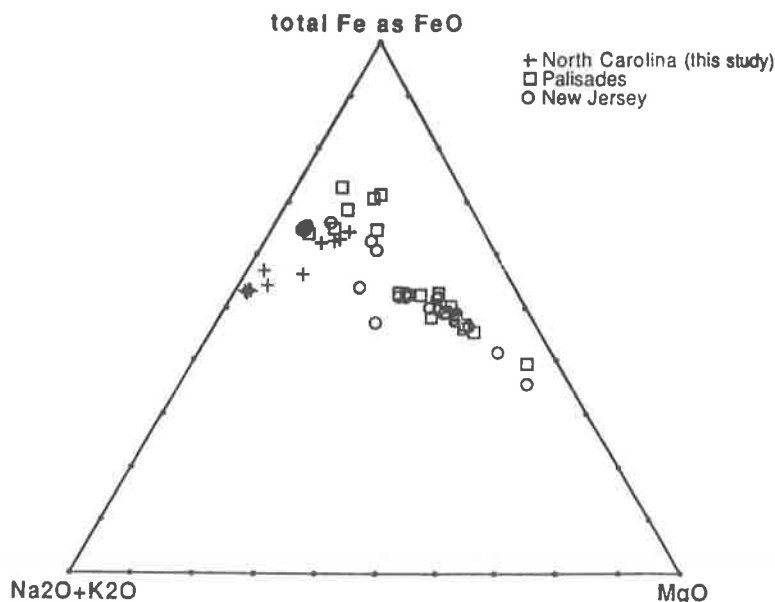


Figure 6. A-F-M diagram combining the analyses from the Piedmont suite (pluses) with analyses from the Palisades intrusion (Walker, 1969), and with a suite of differentiated diabases from New Jersey (Husch and others, 1984).

The more differentiated members of some well-documented ENA sheets bear fairly strong geochemical resemblance to the mafic Piedmont dikes in terms of most elements, including TiO_2 and P_2O_5 (*e.g.* Walker, 1969, p. 77, analyses W-WU5-61 and 16441+; Husch and others, 1984, p. 53, analyses PQ4 and PQ6). Such magmas have been shown to be derivatives of a tholeiitic diabase parent by crystal fractionation of some combination of olivine, pyroxene(s), and plagioclase. However, the K_2O contents of rocks from these studies are much lower than those

of the Piedmont suite, as are the K_2O/Na_2O ratios (always <1 in ENA suites, but considerably >1 in the Piedmont rocks). Attempts to model the Piedmont "parent" magma by crystal fractionation using the ENA average olivine-normative diabase composition (comparable to local olivine diabase) of Weigand and Ragland (1970) encounter this problem.

However, an extensive degree of fractional crystallization of a plagioclase + two-pyroxene assemblage from a high- TiO_2 , quartz-normative diabase magma composition (Weigand and Ragland, 1970, p. 198, analysis #6; Fe_2O_3 set at 1.5 wt%, FeO at 9.0 wt%; P_2O_5 assumed 0.35 wt%) is capable of producing the major element chemistry exhibited by the Piedmont suite (notably their high K_2O , TiO_2 , and P_2O_5). Once sufficiently high levels of these oxides, and SiO_2 , are reached by the evolving magma, then ilmenite, sanidine, quartz, and a phosphate mineral would become important liquidus phases.

Arguing against this model is the fact that no examples of the quartz normative, high- TiO_2 diabase group are known to be present in the vicinity. Furthermore, the extensive fractional crystallization required (almost 80%) is unlikely. Also, there is no significant negative Eu anomaly in the REE patterns of the mafic dikes, as would be expected to result from such major extraction of plagioclase.

Crustal Contamination

It is also possible to model the major oxide chemistry of the suite by an assimilation model. Contamination of a tholeiitic diabase magma with granitic material and then fractional crystallization of a pyroxene-plagioclase assemblage from this magma, provides a fairly good match. In support of this idea is the fact that at one locality abundant granitic xenoliths are present in specimens of fine-grained, dark dike rock. Thin sections of the xenoliths show they are partially digested. However, thermodynamic reasoning (e.g. McBirney, 1979) would argue against the large amount of granitic material required to be assimilated (as much as 20%), as would the fact that extensive assimilation has not been shown to play an important role in ENA magmatism elsewhere. Furthermore, the initial $^{87}Sr/^{86}Sr$ ratio of 0.7067 (Stoddard and others, 1986) of the Piedmont suite is comparable to ratios (0.704 - 0.707) measured for dikes and flows of the ENA province (e.g. Weigand and Ragland, 1970; Gast, 1968).

Immiscibility?

Still other alternative origins are possible. Liquid immiscibility has been correlated with separation of residual tholeiitic magma (Philpotts, 1976; 1982; 1990) or anorthositic magma (Weibe, 1979) into two distinct liquids. High concentrations of certain components, notably P_2O_5 and TiO_2 , expand the two-liquid field (Philpotts, 1990). The contents of these oxides, and the enrichment in other incompatible elements of the Piedmont suite, suggests the possibility that immiscibility may have played a role in the genesis of these rocks, perhaps as the means of producing a less-evolved magma that subsequently underwent fractionation.

CONCLUSION

The Piedmont dike suite constitutes a geochemically coherent group of hypabyssal intrusions with distinctive mineralogy and textural features indicating high-temperature crystallization and shallow emplacement. They are spatially and probably temporally associated with early Jurassic tholeiitic diabase dikes, but are considerably more evolved than diabases and their differentiation products described from ENA. Their high Ti and P, and the fractionation trends within the suite indicate that apatite (or perhaps monazite) and Fe-Ti oxide minerals were important liquidus phases during evolution of the suite. The silicic members of the suite may be derived from the mafic members by simple crystal fractionation of the phenocryst assemblage present in the mafic dikes.

Several hypotheses have been considered for the origin of this suite. Among these are: (1) derivation by small amount of partial melting of an LIL-enriched mantle source with appropriate chemical characteristics (capable of producing apatite and phlogopite perhaps); and (2) derivation by crystal fractionation, liquid separation, or contamination with continental crustal rocks, or any combination of these processes, possibly from a tholeiitic magma parent. Perhaps any of a number of mechanisms is capable of reproducing the geochemical characteristics of the suite. The data available do not permit the rejection of any of the hypotheses, thus permitting the possibility that this dike suite was derived from a mantle source separate from the tholeiitic dikes.

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BANGOR LIMESTONE: DEPOSITIONAL ENVIRONMENTS AND CYCLICITY ON A LATE MISSISSIPPIAN CARBONATE SHELF

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ABSTRACT

The Bangor Limestone accumulated in carbonate shelf and mixed carbonate-clastic basinal settings on the southeastern margin of the North American craton during Late Mississippian time. On the Monteagle Shelf, muddy fossiliferous sediments were deposited in a low-energy, open-marine setting while cross-bedded oolitic grainstones accumulated in high-energy shoals. Muddy, locally dolomitized, fine-grained sediments characterized by tidal channels, flat-pebble breccias, laminae, fenestrae, and evaporite molds were deposited in a facies mosaic of low-energy lagoonal and tidal-flat environments within back-shoal areas.

Shelfal facies-stacking patterns exhibit both small-scale (<1 m) fluctuations within single environmental zones, characterized by interbedding of two or more genetically related facies, and large-scale (3-30 m-thick) shallowing-upward cycles, which exhibit a transition from open-marine to shoal and lagoonal/tidal-flat environments. Facies transitions within a single environmental zone reflect lateral facies shifts and/or short-term variations in energy level due to storms or tides, while shallowing-upward cycles represent long-term shifts in environmental loci owing to seaward progradation of proximal zones over distal ones.

The Bangor Limestone records one major marine transgression, which flooded the interior of the Monteagle Shelf, and several minor ones, which affected only the shelf margin. In the Floyd Synclinorium, a basinal area to the southeast, quiet deep-water sedimentation prevailed throughout the Late Mississippian, broken episodically by thin, grainy oolitic debris flows from the shelf margin and by influx of fine-grained terrigenous clastics from an orogenic collision zone further to the southeast. Correlation of basinal clastics of the middle member of the Bangor Limestone with marine transgressive deposits of the carbonate shelf suggests that clastic influx and shelf flooding shared a common control, probably subsidence of the craton margin owing to tectonic uplift to the southeast and sediment loading of the intervening trough.

INTRODUCTION

The Bangor Limestone is an Upper Mississippian (middle-late Chesterian) formation in the southern part of the Appalachian Valley and Ridge Province and Cumberland Plateau. In the tristate area of northwestern Georgia, northeastern

Alabama, and southeastern Tennessee, the formation caps a 300-450 m-thick Mississippian carbonate sequence which includes the underlying Fort Payne Chert, and Tuscumbia and Monteagle limestones, and is overlain by clastics of the Pennington Formation and Gizzard Group (Figure 1; Thomas and Cramer 1979; Rich 1980, 1982, 1986). Bangor carbonates grade southeastward in northwestern

		SOUTHEAST TENNESSEE	NORTHWEST GEORGIA
PENN.	<div>GIZZARD GROUP</div> <div>PENNINGTON FM.</div> 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Figure 1. Mississippian stratigraphy of southeastern Tennessee and northwestern Georgia. The Bangor Limestone is middle-late Chesterian in age. In northwestern Georgia, the Bangor has been subdivided into carbonate-rich lower and upper members and a clastic-rich middle member (Rich 1986).

Georgia and southwestward in west-central Alabama into coeval 600-800 m-thick mixed carbonate-clastic sequences (Thomas 1972, 1974; Thomas and Cramer 1979). In northwestern Georgia, the Bangor Limestone is subdivided into carbonate-rich lower and upper members and a clastic-rich middle member (Figure 1; Rich 1986), while correlative strata in west-central Alabama are assigned to the predominantly clastic Parkwood Formation.

A Late Mississippian carbonate shelf, herein called the Monteagle Shelf, developed in the tristate area on the northeastern corner of the Alabama Promontory, an oceanward convex portion of the continental margin formed during late Precambrian rifting (Figure 2; Thomas 1977). Demise of the Monteagle carbonate shelf resulted through influx of Pennington-Lee clastics from the central Appalachian Orogen to the northeast and of Parkwood-Pottsville clastics from the Ouachita Orogen to the southwest in late Chesterian time (Thomas 1974). These clastic wedges merged in north-central Alabama, where the youngest portion of the

Bangor Limestone is directly overlain by the Lower Pennsylvanian Pottsville Formation (Thomas 1972).

Upper Mississippian carbonate units of the tristate area have frequently been interpreted as representing sedimentation in shallow-marine environments similar to those of the Bahamas and Florida Keys today (e.g., McLemore 1971; Dinnean

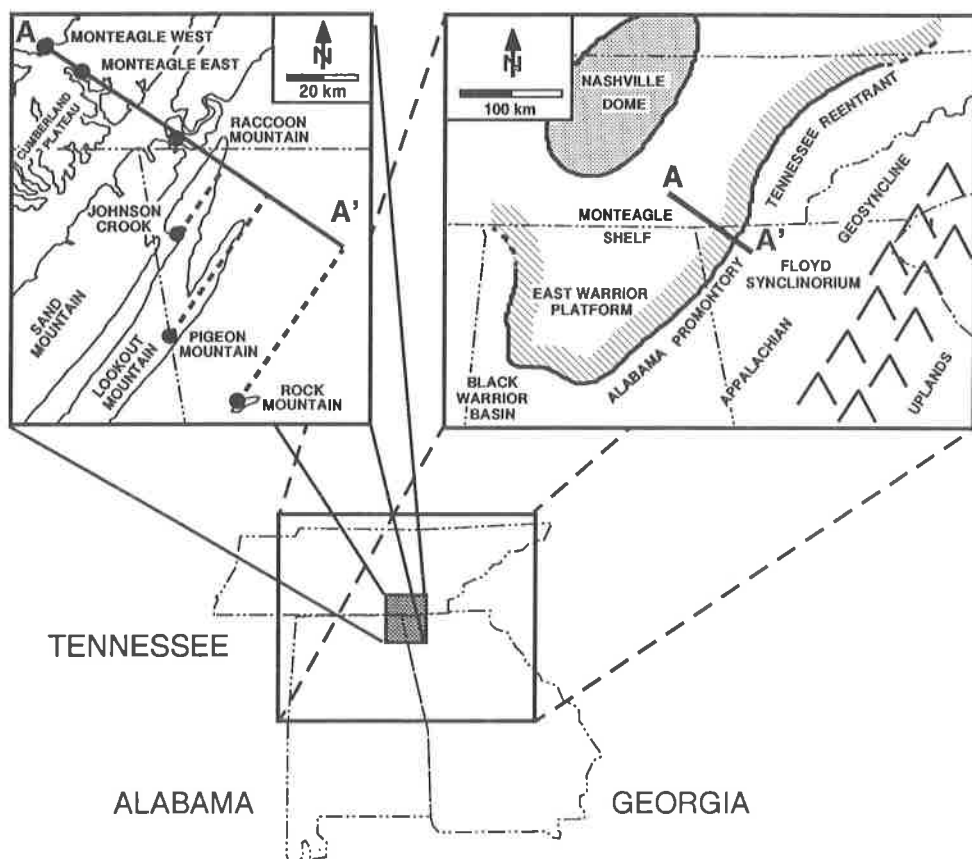


Figure 2. Upper left: Bangor Limestone study locales in relation to present-day topography of southeastern Tennessee, northeastern Alabama, and northwestern Georgia. Monteagle-West, Monteagle-East, and Johnson Crook are outcrop sections; Raccoon Mountain, Pigeon Mountain (near Menlo), and Rock Mountain are cores. Sections are projected onto the line of cross-section A-A' (Figure 3), perpendicular to present-day structural and Mississippian paleogeographic trends (ca. N30E), allowing for reconstruction of Bangor depositional environments in their correct spatial relations. Upper right: Late Mississippian paleogeography of the southeastern United States. Study locales extend from the Monteagle Shelf (southeastern flank of the emergent Nashville Dome) to the Floyd Synclinorium (southwestern extension of the Appalachian Geosyncline).

1974; Handford 1978). However, while individual facies of the Bangor Limestone are analogous to certain modern carbonate facies, significant differences exist between the Bangor Limestone and these modern environments with respect to patterns of sediment accumulation (Algeo 1985). In addition, the Bangor Limestone represents the final 3-5 my of carbonate sedimentation on the

southeastern margin of the North American craton and, as such, is significant as a record of the tectonic and sedimentologic events which brought about the demise of this pericratonic carbonate shelf during the early phases of the Appalachian Orogeny. Therefore, the goals of this study were: 1) to develop an integrated model for Bangor Limestone paleoenvironments and contrast this with modern carbonate settings, and 2) to summarize the depositional history of the Bangor Limestone with respect to eustatic and tectonic events operating on the Late Mississippian shelf of the tristate area.

The Bangor Limestone was studied in three outcrop sections and three cores between the west side of Monteagle Mountain in southeastern Tennessee and Rock Mountain in northwestern Georgia (Figure 2). Sections were measured and sampled at 0.3 m intervals, where possible. A total of 1450 thin-sections from the six study locales were examined and described using transmitted-light microscopy. Dolomite abundance was determined by etching polished core chips and hand samples in a 5% HCl bath for 15-30 sec and visually estimating percent area of high relief. Acetate peels were made of selected thin-sections and core butts to study sedimentary fabric and microstructures.

BANGOR LITHOFACIES

The Bangor Limestone comprises nine lithofacies which formed the basis for environmental interpretation and facies sequence analysis.

Bryozoan-Echinoderm Wackestone/Packstone/Grainstone

Bryozoan-echinoderm wackestone/packstone units are generally thick-bedded (to 10s m) and structureless owing to pervasive bioturbation. They contain a normal marine fauna of abundant bryozoans and echinoderms and subordinant quantities of other stenohaline organisms such as brachiopods, trilobites, and corals. Disarticulation and fragmentation of fossil debris is limited and sorting is poor, indicating deposition within a low-energy environment.

In contrast, bryozoan-echinoderm grainstone units exhibit good sorting and extensive disarticulation and comminution of fossil debris owing to deposition in a moderate- to high-energy environment. Micrite is restricted to interior voids of fossil grains (e.g., gastropod chambers and bryozoan fenestrules) suggesting a source in a muddy, low-energy environment such as that in which bryozoan-echinoderm wackestone/packstone was deposited.

Oolitic Packstone

Oolitic packstone units are generally thin- to medium-bedded (<3 m) and exhibit poor sorting or, occasionally, weak inverse grading. Oolites range from 0.25 to 2.0 mm in diameter and vary from superficial to mature in cortical development. Matrix micrite comprises from 5 to 50% of rock volume. Fossils are more abundant and show less fragmentation than in oolitic grainstone units.

Oolitic Grainstone

Oolitic grainstone units are medium- to thick-bedded (to 10s m) and exhibit planar cross-bedding (sets mostly 1-5 m thick). Oolites range from 0.5 to 1.0 mm

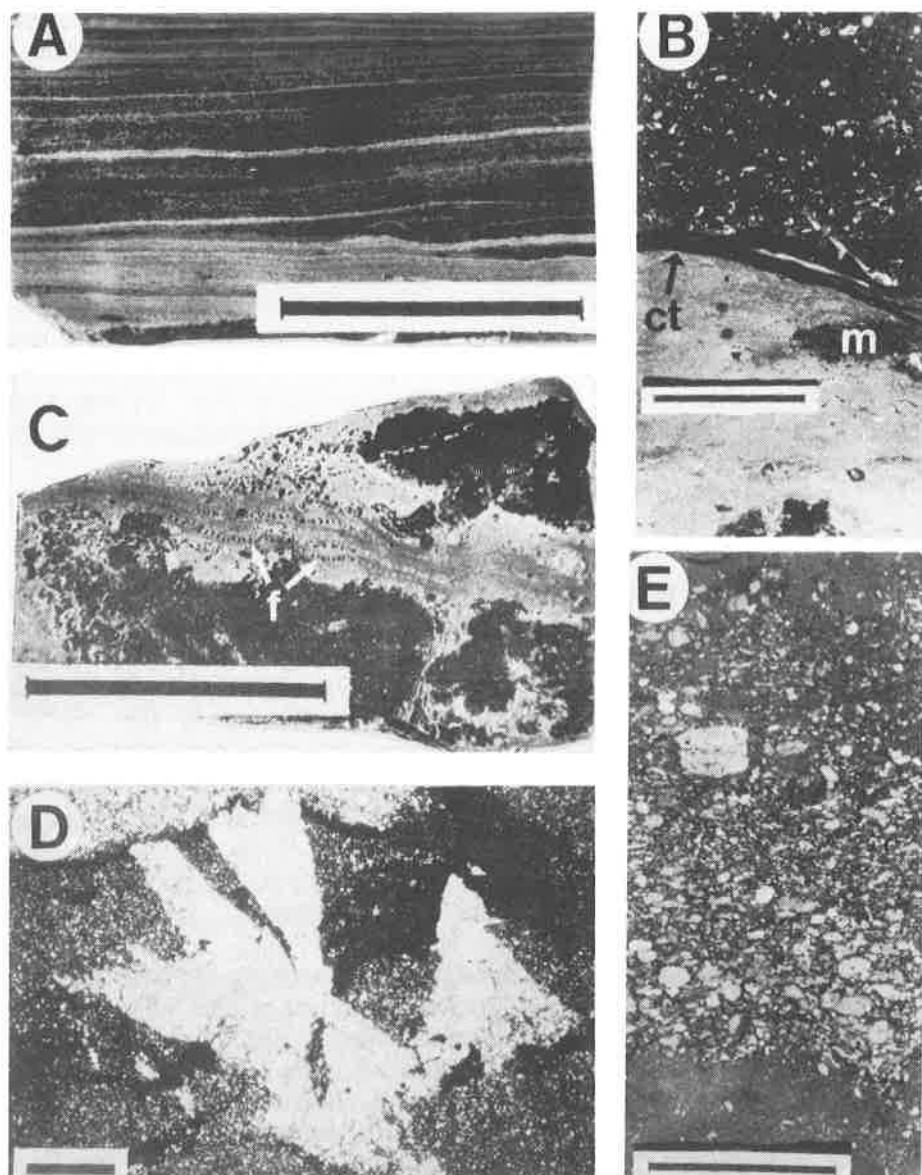


Figure 3. (A) Laminated tidal-flat carbonate/dolomite mudstone from Raccoon Mountain (dolomitic layers are light colored). The scale in this and all subsequent core samples is 1 inch. (B) Contact between two shoaling cycles at Raccoon Mountain. The underlying cycle is capped by a fine-crystalline dolostone containing pods of unreplaced carbonate mudstone (m). Note the absence of dolomite in oolitic packstone basal unit of the overlying cycle. (C) Partially dolomitized bryozoan wackestone from Pigeon Mountain. Note strong dolomitization adjacent to bryozoan fronds; dolomitizing fluids probably moved through open fenestrules (f). (D) Radiating clusters of pseudomorphic dolomite crystals after gypsum from a fine-crystalline tidal-flat dolostone at Raccoon Mountain. Blebs of coarse-crystalline anhydrite are also present in this unit. (E) Tidal-flat intraclast breccia from Raccoon Mountain. Note angularity, coarseness, and lithologic heterogeneity of clasts and fining-upward trend within unit.

in diameter and generally display mature cortical development. Oolite nuclei are mostly normal-marine fossil fragments, although small micritic pellets and intraclasts are an important secondary source. Fossil debris is generally restricted to robust fossil fragments such as echinoderm plates and columnals, and sorting is excellent.

Foram-Ostracod Wackestone

Foram-ostracod wackestone units are generally thin- to medium-bedded (<5 m) and structureless. They are characterized by a restricted-marine fauna composed mainly of ostracods and forams, although sponge spicules are locally abundant. Stenohaline fossil types are found in some samples, but their occurrence as comminuted debris suggests derivation from another source. Matrix micrite is commonly finely pelleted, although compaction locally masks pellet identity.

Carbonate Mudstone

Carbonate mudstone units are generally thin-bedded (<1 m) and exhibit mm-scale lamination (Figure 3A). Fenestrae locally comprise up to 25% of rock volume, ranging from 1 to 10 mm in size and exhibiting infill by geopetal micrite or calcite spar. Fenestrae frequently appear to have originated as voids between soft, partially compacted mud clasts, possibly developing through in-situ brecciation resulting from gas generation (e.g., Mazzullo and Birdwell 1989), although some fenestrae may have formed through dissolution of evaporite crystals, gas entrapment, or dessication-induced sediment shrinkage (e.g., Shinn 1968).

Dolomitic Mudstone/Wackestone

All Bangor dolomite originated through the replacement of matrix micrite and allochems. Evidence for replacement includes remnant patches of lime mud and isolated fossils "floating" in many dolostones (Figure 3B), irregular nonstratified distribution of dolomite in partially dolomitized units (Figure 3C), and cross-cutting relations between dolomite rhombs and allochems. In units exhibiting variable degrees of dolomitization, host-rock components are replaced in a fixed order: 1) matrix micrite, 2) micritic clasts, 3) thin-shelled calcitic fossils (e.g., bryozoans and forams) and 4) massive calcitic fossils (e.g., echinoderms, and sparry calcite cement). Dolomitization commonly ceased after replacement of matrix micrite, leaving fossil debris "floating" in a dolomitic matrix.

Samples were classified as dolomitic mudstone/wackestone where replacement obliterated most of the original texture of fine-grained limestones. This microfacies is characterized by 0.01-0.05 mm dolomite rhombs that commonly exhibit a sugary, intergrown texture and a patchy distribution within host units. Dolomitic units are generally interbedded with carbonate mudstones and foram-ostracod wackestones capping shoaling cycles, and dolomitization tends to decrease in intensity downward from cycle tops (Figure 3B). Association with gypsum (Figure 3D) and dolomitic intraclast breccias (Figure 3E) at or near cycle tops indicates development of hypersaline conditions and pene-contemporaneous dolomitization. In contrast, open-marine bryozoan-echinoderm wackestones and

packstones are marked by low concentrations (10-30%) and relatively even distribution of coarse (0.05-0.10 mm) dolomite rhombs within the rock matrix, and near-total obliteration of the original texture of such units is rare.

(Oo-)Intraclastic Packstone/Grainstone

(Oo-)intraclastic packstone/grainstone units are generally thin- to medium-bedded (<5 m) and exhibit bedding-parallel lamination characterized by mm- or cm-scale variation in clast size. Intraclasts are comprised of lime mud, are generally round and fine-grained (0.2-0.5 mm), and may be either structureless or wispy (algal laminated?) internally. Rare interbeds containing large, angular clasts (e.g., flat-pebble breccias) are generally thin (<10 cm) and fine upward. Local oolitization of intraclastic units is marked by oolites with a high percentage of intraclast nuclei, abundance of superficially-coated grains, and considerable admixture of large lime-mud clasts and algal fragments (Rich 1974). Clast development and ooid formation in a predominantly muddy environment suggest the occurrence of episodic high-energy events within a generally low-energy setting.

Terrigenous Clastics

The most common terrigenous clastic lithofacies are gray to gray-green shale and clayey sandstone composed of angular to subangular, coarse-grained silt and fine-grained sand. Clastic units commonly exhibit mm- or cm-scale interlamination of clay, silt, and sand. Fossils, predominantly weathered echinoderm fragments, are common.

DEPOSITIONAL ENVIRONMENTS

The Bangor Limestone was deposited in three broad environmental zones, each of which is characterized by diagnostic microfacies, sedimentary structures, and facies-stacking patterns. The environments reflect differences in water depth, distance from the paleo-shoreline, and energy level (Figure 4). These environments are: 1) a low-energy open-marine shelf, 2) a high-energy ooid shoal, and 3) a low-energy back-shoal lagoon and tidal-flat complex. Many other shallow-marine carbonate sequences record deposition in similar environments, and numerous ancient and modern examples have now been documented (e.g., Wilson 1975; James 1984).

Distal Low-Energy Open-Marine Shelf

Bangor open-marine shelf facies accumulated below wavebase, where low-energy conditions allowed carbonate mud to settle. Normal marine salinities supported a diverse fauna which contributed abundant skeletal debris to form wackestones and packstones. Intense bioturbation commonly gave rise to structureless units, although storm layers of imbricated coarse fossil debris are present locally. The open-marine shelf facies association of the Bangor Limestone is dominated by bryozoan-echinoderm wackestone/packstone, which is petrographically similar to Mississippian Waulsortian mound facies (Wilson 1975) but which lacks characteristic mound morphology or stromatactid textures.

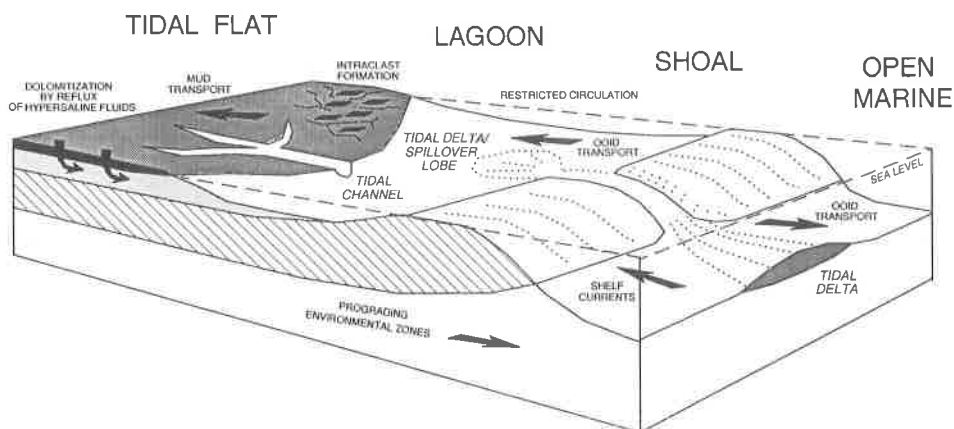


Figure 4. Paleoenvironmental model of Monteagle Shelf facies of the Bangor Limestone. Sedimentation occurred in three distinct energy zones: a low-energy lagoonal/tidal-flat environment, a high-energy oolitic shoal environment, and a low-energy open-marine environment. Environments are distinguished based on characteristic facies associations, facies-stacking patterns, and sedimentary structures. See text for details.

Subtidal lamellibranch mud and muddy sand of the Persian Gulf (e.g., Wagner and van der Togt 1973) are modern examples of such sub-wavebase sediments.

Fossil grainstone, oolitic packstone, sandstone, and shale are also prevalent in Bangor open-marine environments. Grainstones are indicative of episodic higher-energy conditions resulting from storms or strong bottom currents. Thin beds of oolitic packstone are common in foreshoal (<3 m thick) and deep basinal (<1 m thick) environments and probably represent grainy debris flows emanating from the Monteagle Shelf margin. These units may be comparable to debris flow deposits from the lower slope of Little Bahama Bank, which thin (to <1 m) and become grain-supported in distal slope areas (e.g., Mullins and others 1984). Fine-grained sandstone and shale are found in small quantities in shelf areas but accumulated mainly in deep-basinal environments.

Medial High-Energy Shoals

Bangor high-energy shoals developed where wavebase intersected the landward-rising marine shelf, causing wave- and tidal-current energy to dissipate through friction (Figure 4). Such high-energy conditions resulted in shell fragmentation, abrasion, and sorting, and winnowing of carbonate mud. The high-energy shoal facies association is characterized by oolitic and, to a lesser degree, fossil grainstones. Oolitic grainstone represents the highest sustained energy conditions, as ooid formation occurs only through persistent turbulence at shallow depths (Loreau and Purser 1973). Analogy with environments of oolitization in the Persian Gulf (Loreau and Purser 1973), Andros platform in the Bahamas (Purdy 1963b), and Shark Bay, Australia (Hagan and Logan 1974) suggests that maximum depths of formation were no more than 5 m, and commonly less than 2 m.

Bangor oolitic sands form 1-20 m-thick, planar cross-bedded units extending laterally for at least several hundreds of meters. They are generally similar in

thickness and internal structure to Bahamian oolitic tidal bars (e.g., Ball 1967). Oolitic grainstone isolith data from Vail (1959) and Handford (1978) suggest that Mississippian ooid shoals formed northeast-southwest-oriented chains of tidal bars parallel to the paleo-shoreline in northeastern Alabama and eastern Tennessee. Bangor high-energy shoal environments migrated freely across the Monteagle Shelf in response to sea-level fluctuations as evidenced by lateral correlatability of oolitic units over a distance of at least 60 km perpendicular to paleo-depositional strike.

Other facies found in association with oolitic grainstones are not diagnostic of shoal environments. Oolitic packstone occurs not only interbedded with oolitic grainstone, but also in more proximal and distal sequences, indicating that landward and seaward transport of ooids as spillover lobes and tidal deltas was common (Figure 4; e.g., Ettensohn and Chesnut 1985). Similar muddy oolite is forming in modern environments at Joulters Cay in the Bahamas, where backshoal sandflats are several times wider than the active shoal (Harris 1977), and along the Trucial Coast of the Persian Gulf, where seaward transport of ooids occurs in tidal-bar channels (Loreau and Purser 1973).

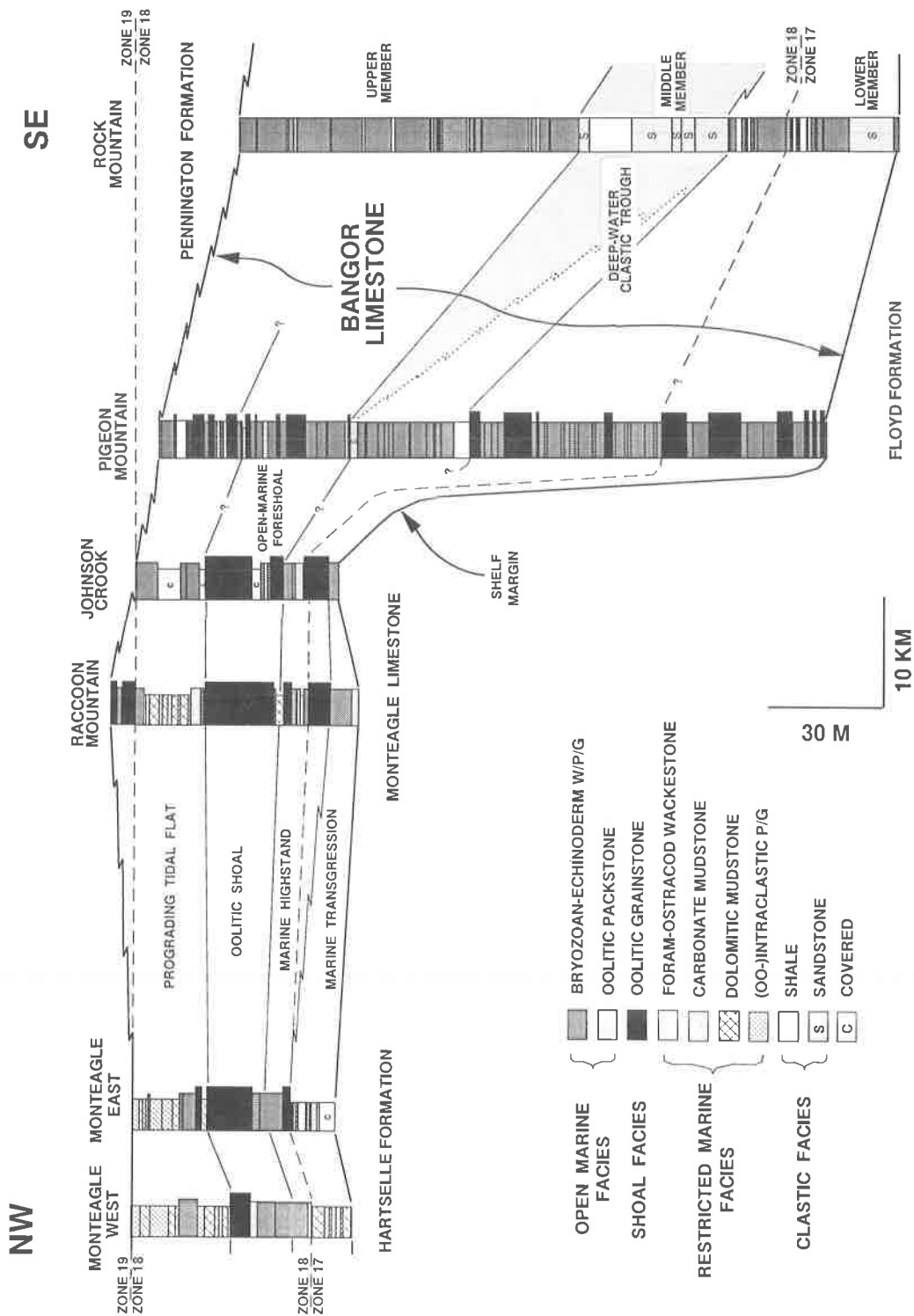
Bryozoan-echinoderm grainstone is found most commonly underlying oolitic grainstone units, indicating deposition in foreshoal areas just seaward of prograding shoal environments. Foreshoal energy conditions were thus intermediate: sufficient to winnow fines but insufficient to initiate oolitization. Foreshoal skeletal sand beds may have served as the major source of ooid nuclei as shelf currents swept sand into the shoal environment. Similar skeletal grainstone facies are found in modern foreshoal environments of the Persian Gulf (Wagner and van der Togt 1973) and in the Bahamas (Purdy 1963b; Multer 1977).

Proximal Low-Energy Lagoons and Tidal Flats

The most proximal environments in which the Bangor Limestone accumulated were lagoons and tidal flats located shoreward of the high-energy shoal zone. Bangor backshoal sequences are characterized by a diversity of lithologies deposited in thinly-bedded units, suggesting a complex and rapidly shifting facies mosaic and/or variable energy conditions. This is typical of backshoal environments, which are broad, shallow low-energy zones subject to episodic high-energy events such as storms and tides (e.g., Ginsburg and Hardie 1975), and in which slight changes in depth result in large changes in sediment type (e.g., Hagan and Logan 1974).

Lagoonal Association: Bangor lagoonal units accumulated in shallow subtidal areas bounded basinward by high-energy shoals and shoreward by tidal flats (Figure 4). The lagoonal facies association includes carbonate mudstone, bryozoan-echinoderm wackestone/packstone, and oolitic packstone, but the most diagnostic facies is ostracod-foram wackestone. While a stenohaline fauna of this type indicates restricted-marine conditions, burrows, mottled fabrics, and absence of primary laminae are evidence of bioturbation and oxygenated bottom-waters, suggesting elevated and/or fluctuating salinity or temperature rather than anoxia as the cause of restriction. Similar restricted-marine sediments are found in modern lagoonal settings such as the shallow subtidal platform west of Andros Island (Purdy 1963b; Shinn and others 1969).

Hydrographic and bathymetric conditions within the Bangor lagoon were



probably similar to those of modern tropical lagoons. Water depths in analogous settings on the Andros Island platform and in the Persian Gulf are generally less than 5 m and only rarely exceed 10 m. In addition, restricted circulation and increased evaporation tend to elevate salinity. In the Bahamas, salinities increase from normal marine (35 o/oo) at the platform edge to 38-46 o/oo just west of Andros Island (Purdy 1963a), while salinities in restricted lagoons of the Persian Gulf commonly range between 50 and 70 o/oo (Loreau and Purser 1973; Wagner and van der Togt 1973). Similar conditions probably existed in the Bangor lagoon, in which variations in faunal abundance and diversity reflect variations in water circulation, temperature, or salinity.

Tidal Flat Association: Tidal flats are the most proximal environments preserved in the Bangor Limestone (Figure 4). The tidal-flat facies association comprises carbonate and dolomitic mudstone and (oo-)intraclastic packstone/grainstone. Dolomitic mudstone of the Bangor Limestone is similar petrographically and stratigraphically to fine-grained, tan to medium-gray tidal-flat dolomites of the Persian Gulf, which form thin, laterally continuous beds extending from the supratidal zone into adjacent subtidal lagoonal units (Illing and others 1965).

Bangor tidal-flat sedimentary structures include mm-scale laminae, fenestrae, intraclast breccias, dolomite pseudomorphs after gypsum, and tidal channels (Figure 3). Laminae and fenestrae are ubiquitous on modern tidal flats such as on Andros Island, where they are found in beach ridge, levee, and marsh sediments, but are absent in nearby lagoonal and intertidal units (Shinn and others 1969). Bangor intraclast breccias occur as thin, fining-upward layers of angular to subangular dolomitic clasts in a carbonate mud matrix (Figure 3E). In modern tidal flats, they develop when semilithified supratidal crusts are broken by plants, dessication, or storms (Ginsburg and Hardie 1975; Logan 1974; Woods and Brown 1975). While anhydrite blebs and dolomitic pseudomorphs after gypsum crystals

Figure 5. Stratigraphic cross-section of the Bangor Limestone in southeastern Tennessee and northwestern Georgia. Sections at Monteagle West and Monteagle East are dominated by tidal-flat sedimentation punctuated by a single marine transgression. Sections at Raccoon Mountain and Johnson Crook are characterized by oolitic-shoal sedimentation and exhibit one major and several minor marine transgressions. The Pigeon Mountain section was located in an upper slope fore-shoal environment marked by thinly interbedded open-marine fossiliferous and oolitic units. The Rock Mountain section records quiet deep-basinal fine-grained carbonate sedimentation interrupted episodically by thin, grainy oolitic debris flows and influx of terrigenous clastics. Cross-section datum: zone 18/19 contact of Mamet's global foram zonation. Although the lower Bangor Limestone correlates with the upper part of Mamet's zone 17, most of the Bangor is coeval with zone 18 (Rich 1980, 1986). The upper contact of the Bangor Limestone is approximately coincident with the zone 18/zone 19 contact across the Monteagle Shelf but is found within the upper part of zone 18 at Rock Mountain, reflecting a time-transgressive relationship between the Bangor Limestone and the overlying Pennington Shale (Rich 1986). The section at Raccoon Mountain has not been biostratigraphically zoned, and stratigraphic relations presented here are based on lithologic correlation.

occur sporadically in the Bangor Limestone, the relative paucity of evaporites suggests that Bangor tidal flats may have been of the humid-zone variety, such as those found in the Bahamas (Hardie and Garrett 1977), rather than of the strongly evaporitic Persian Gulf type. Small tidal channels (<2 m deep and 2-20 m wide) are common in Bangor tidal-flat sequences as on modern tidal flats.

Energy conditions were variable on Bangor tidal flats as evidenced by cm-scale interbedding of low-energy mudstone with higher-energy (oo-)intraclastic packstone/grainstone. Similar facies associations are found in modern carbonate tidal flats such as Shark Bay, Australia, where intraclasts form in high intertidal/low supratidal areas through storm-induced fragmentation of indurated micritic crusts and are distributed across the intertidal and supratidal zones (Hagan and Logan 1975; Woods and Brown 1975). Oolitic coating of intraclastic sediments required frequent moderate turbulence, possibly through wave and tidal action within the intertidal zone (Figure 4). Similar oolitic tidal-flat sands have been described from the Persian Gulf (Loreau and Purser 1973).

SEQUENCE ANALYSIS AND TECTONIC CONTROLS ON SEDIMENTATION

Paleogeography

Late Mississippian paleogeography of the tristate area was controlled by structural elements that had persisted since the early Paleozoic (Thomas 1977; DeWitt and McGrew 1979). Inherited tectonic elements included: 1) the Nashville Dome, a low-lying landmass which was part of the structurally-high Cincinnati Arch, 2) the Monteagle Shelf, a shallow-marine area on the southeastern flank of the Nashville Dome, and 3) the Floyd Synclinorium, a deep-water trough to the southeast of the Monteagle Shelf (Figure 2). The Monteagle Shelf was bounded by a sharply-defined shelf margin approximately coincident with the northern side of Pigeon Mountain in northwestern Georgia.

The regional distribution of Bangor facies and environmental zones reflects this paleogeography (Figure 5). Sections from Monteagle West in the northwest to the northern side of Pigeon Mountain in the southeast were located on the Monteagle Shelf, while Pigeon Mountain and Rock Mountain, the southeasternmost sections, were located on the northwestern margin of the Floyd Synclinorium. Low-energy lagoonal and tidal-flat units are most abundant in the Monteagle West and Monteagle East sections, which represent inner-shelf areas flanking the Nashville Dome (Figure 6). High-energy shoal sediments occur most frequently in the Raccoon Mountain and Johnson Crook sections on the midshelf, while low-energy open-marine sediments dominate the Pigeon Mountain section on the outer shelf. Basinal sediments, comprised of open-marine carbonate and terrigenous clastic units, make up the entire section at Rock Mountain.

Facies-Stacking Patterns

Variability in facies-stacking patterns in the Bangor Limestone reflects differences in proximity to paleo-shorelines and shelf margins. While alternation between two or more facies within the same environmental zone is common at all locales, shallowing-upward cyclicity is restricted to shelf areas (Figure 5).

Inner Shelf: In inner-shelf areas at Monteagle West and Monteagle East,

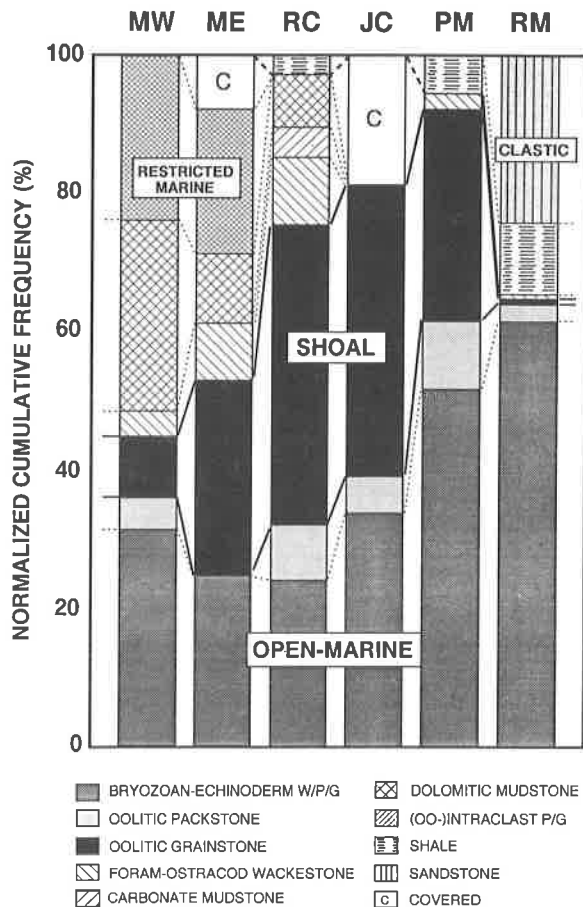


Figure 6. Facies distribution by study locale (MW = Monteagle West; ME = Monteagle East; RC = Raccoon Mountain; JC = Johnson Crook; PM = Pigeon Mountain; and RM = Rock Mountain). Frequencies are based on sampling at 0.3-m intervals and are normalized to total sample count. Facies are grouped according to dominant environment of occurrence: bryozoan-echinoderm wackestone/packstone/grainstones and oolitic packstones are predominantly open-marine facies; oolitic grainstone is predominantly a shoal facies; and foram-ostracod wackestone, carbonate and dolomitic mudstone, and (oo-)intraclastic packstone/grainstone are predominantly restricted-marine facies.

intrazonal transitions between dolomitic mudstone, (oo-)intraclastic packstone/grainstone, and foram-ostracod wackestone of the lagoonal/tidal-flat association are dominant (Figure 7). The high frequency of occurrence of dolomitic mudstone/intraclastic packstone transitions at Monteagle West may reflect its proximal location and abundance of tidal-flat facies (Figure 4). In contrast, at Monteagle East, foram-ostracod wackestone/intraclast packstone transitions predominate, reflecting dominance of lagoonal depositional environments.

Middle-Outer Shelf: Sedimentation at Raccoon Mountain, a mid-shelf locale, was dominated by shallowing-upward cyclicity, although intrazonal facies shifts

to lagoonal/tidal-flat units) are found in all shelf sequences of the Bangor Limestone (Figure 8). In inner-shelf areas, one major shoaling cycle (25-30 m thick) is present, while in mid- to outer-shelf areas several additional minor cycles (3-10 m thick) are recorded (Figure 5). Cycle boundaries are marked by sharp, disconformable lithologic contacts, separating proximal restricted-marine sediments of the underlying cycle from distal open-marine sediments of the overlying cycle (Figure 3B). Cycle tops commonly exhibit strong dolomitization decreasing rapidly downward over a few tens of centimeters, while the bases of overlying cycles show no alteration.

Each cycle commenced with a marine transgression marked by a 0.3 to 2.0

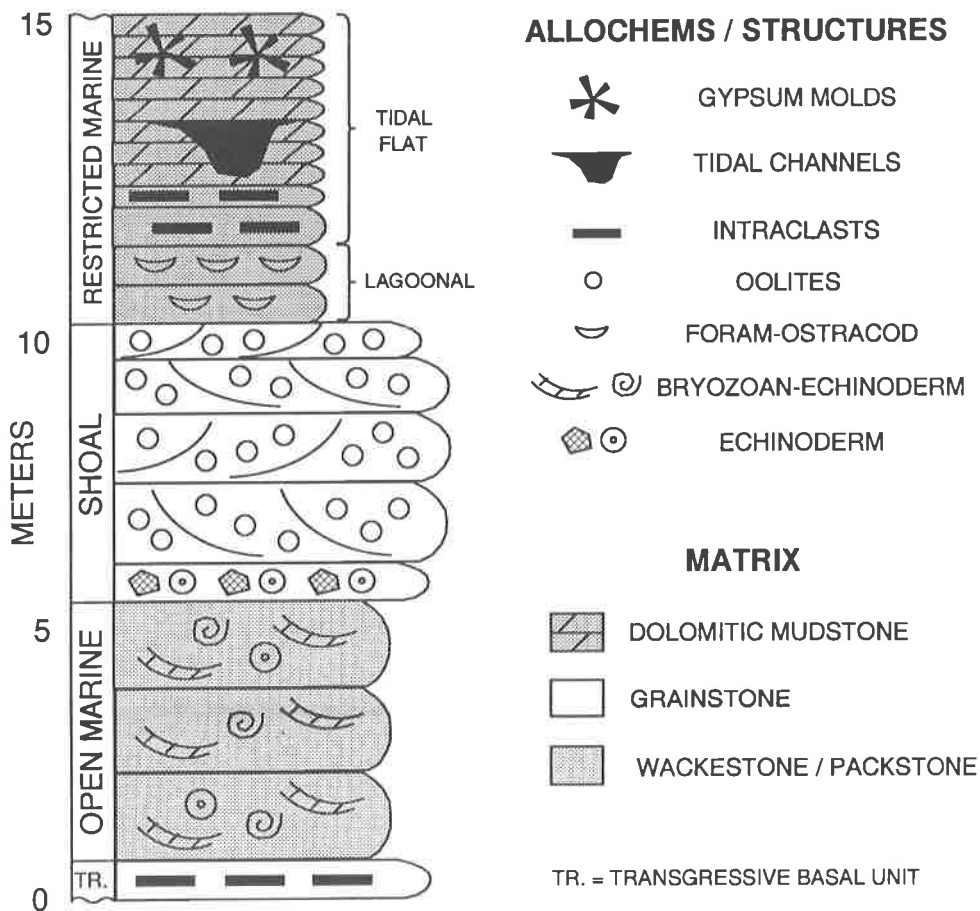


Figure 8. Bangor Limestone shallowing-upward cycle. This model emphasizes the vertical relations of energy zones, variations in bedding thickness, and facies-specificity of allochems and sedimentary structures.

m-thick basal unit composed of intraclastic-oolitic packstone/grainstone exhibiting poor to moderate sorting and considerable admixture of quartz sand and glauconite (Figure 8). Basal-unit intraclasts are of diverse lithologies, indicating penecontemporaneous erosion and mixing of semi-lithified sediment from several facies within the lagoonal/tidal-flat association. Sparry matrices, ooid and

intraclast formation, and coarse-grained allochems are evidence of high-energy conditions. These observations are consistent with deposition of basal units during transgression of partially consolidated and dolomitized shelf sediments. The thinness of the transgressive deposits may reflect either rapidity of the transgressive events or their inherently erosional nature.

Intrazonal facies shifts and interzonal shoaling patterns probably had fundamentally different origins. Frequent alternation between two or more facies within an environmental zone suggests that such sequences developed through short-term fluctuations in energy conditions and/or lateral facies shifts which may have been unrelated to larger-scale cyclic processes. The latter are marked by an orderly shoaling pattern reflecting seaward progradation of energy zones during episodes of stable or falling relative sea level. Subsequent marine transgressions re-established open-marine conditions and reinitiated the shoaling process.

Tectonic Control of Shelf-Basin Subsidence

A single major transgression affected the entire Monteagle Shelf, flooding areas at least as far landward as the Monteagle West section (Figure 5). While this flooding event could have resulted from either eustatic or regional tectonic movements, correlation of the clastic middle member of the Bangor Limestone at Rock Mountain with highstand units on the Monteagle Shelf suggests that the latter is more probable. Progradation of a coarse clastic wedge derived from an active orogen to the southeast into the Floyd Synclinorium flysch basin may have resulted in subsidence of both the basin and the adjoining shelf, and transgression of the latter (Rich 1986). If the transgressive event had been due to eustatic rise without tectonic movement, then coarse-grained clastics should have been trapped closer to the receding shoreline on the southeastern margin of the Floyd Synclinorium, rather than flooding into the basin.

While the Bangor thickens into the Floyd Synclinorium by a factor of 2.5-3.0 in relation to Monteagle Shelf sections (Figure 5), the locus of thickening, i.e., whether distributed uniformly through the section or concentrated at a particular horizon, is unknown. If the middle member of the Bangor Limestone is equivalent to the Upper Floyd Shale/"Hartselle interval," then thinning of the shelf section is concentrated in units correlative to the thick clastic middle member of the basal sequence (Rich 1982, 1986). This would be consistent with drowning-induced low sedimentation rates on the shelf, a widely-observed phenomenon in carbonate environments (e.g., Read 1985). In this event, the carbonate-rich lower and upper members of the Bangor Limestone at Rock Mountain correlate with the thick, regressive, predominantly tidal-flat sequences of inner-shelf locales.

CONCLUSIONS

1) The Bangor Limestone was deposited in shallow carbonate shelf and basal mixed carbonate-clastic settings on the southeastern margin of the North American craton during Late Mississippian (middle and late Chesterian) time.

2) Deposition occurred in three broad environmental zones: a distal low-energy open-marine shelf, medial high-energy shoals, and a proximal low-energy lagoonal and tidal-flat complex.

3) Each environmental zone is characterized by diagnostic microfacies, sedimentary structures, and facies-stacking patterns which reflect differences in

water depth, distance from the paleo-shoreline, and energy level.

4) Fluctuations within an environmental zone, marked by cm-scale interbedding of two or more genetically-related facies, are the most common form of facies transition. Such transitions reflect lateral facies shifts and/or short-term variations in energy level due to storms or tides.

5) Shelf sequences are characterized by 3-30 m-thick cycles that shallow upward from open-marine through shoal to lagoonal/tidal-flat sediments, reflecting long-term shifts in environmental loci owing to seaward progradation of proximal environmental zones over distal ones.

6) Inner areas of the Monteagle Shelf are dominated by tidal-flat sedimentation, punctuated by a single major marine transgression, while outer-shelf areas are dominated by ooid-shoal and foreshoal sedimentation and record several minor flooding events in addition to the major one.

7) In the Floyd Synclinorium, a basinal area to the southeast of the Monteagle Shelf, quiet deep-water sedimentation prevailed throughout the Late Mississippian, broken episodically by thin, grainy oolitic debris flows from the shelf margin and by influx of fine-grained terrigenous clastics from an active orogen to the southeast.

8) Correlation of the clastic middle member of the Bangor Limestone with marine-highstand units of the Monteagle Shelf suggests that clastic influx and shelf flooding share a common control, probably subsidence of the craton margin owing to uplift of tectonic highlands to the southeast and sediment loading of the intervening flysch trough.

9) Bangor Limestone deposition terminated when the Monteagle Shelf was overwhelmed by southwestward- (Pennington Formation) and northeastward-prograding (Parkwood Formation) clastic wedges in late Chesterian time.

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STRATIGRAPHY AND STRUCTURE OF MEADOW BRANCH SYNCLINORIUM, WEST VIRGINIA

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ABSTRACT

Detailed geological mapping has led to a revision of the stratigraphy and structure of Meadow Branch synclinorium, Berkeley County, West Virginia. The synclinorium is composed of five formations (Rockwell, Purslane, Hedges, Myers, and Pinkerton) long considered to comprise the Mississippian Pocono Group. The three basal formations (Rockwell, Purslane, and Hedges formations) are retained as Lower Mississippian (Kinderhookian and Osagean). Only the Purslane (equivalent to the Burgoon Sandstone in Pennsylvania) is correlative with the original Pocono Formation of northeastern Pennsylvania. The Myers Shale is now correlated with the Mississippian Maccrady Formation (Meramecian). The Pinkerton Sandstone is possibly equivalent to the missing Mississippian (Meramecian) strata at the Maccrady-Greenbrier unconformity. This correlation is supported by lithology, thickness, regional structure, and paleobotanical identification of *Rhodeopteridium* cf. *vespertina* (Read) and *Triphyllopteris* cf. *lescuriana* (Meek).

The newly-mapped Woods fault truncates the four lower formations on the overturned eastern limb of the synclinorium. This 35 km (22 miles) long thrust fault places the Devonian Hampshire Formation on the Pinkerton Sandstone. At this location the Pinkerton reaches a thickness of 350 m (1,150 feet). Sleepy Creek fault, a back thrust, cuts out the Purslane, Hedges, and Myers formations and places older Rockwell on younger Pinkerton rocks near the southern end of the synclinorium.

Mississippian clastics, perhaps 2,400 to 3,050 m (8,000-10,000 feet) thick in the central Appalachians, were transported a minimum of 48 km (30 miles) westward and eroded with only a few remnants remaining as isolated synclines. Extensive folding associated with the development of Meadow Branch synclinorium has produced a structural complex that partly accommodates the westward displacement of the allochthonous Martinsburg sheet.

INTRODUCTION

Meadow Branch synclinorium, located in Berkeley County, West Virginia, has traditionally been mapped as five formations, all comprising the Mississippian Pocono Group (Kinderhookian and Osagean). The entire sequence was considered Pocono based on floral assemblages identified by David White (Campbell, 1903) while the five formations were mapped and named by Stose and Swartz (1912). Recent geological mapping by the authors has shown numerous difficulties with this stratigraphic correlation and structural interpretation. However, examination of the literature reveals why the traditional stratigraphic and structural interpretations have persisted. The interested reader is referred to Berg and Edmunds (1979) and Brezinski (1989) for excellent discussions of the checked history of the term "Pocono."

The original mapping was done by William Barton Rogers in 1838 and on his map of 1874 he placed the rocks of this basin in his "Lowest Coal Group" (originally called Formation X or Vespertine, and finally Pocono). Campbell (1903) notes that David White collected fossil plants here in 1902, which he determined were Pocono floras. Griffith (1902, in White, I.C., 1903) also refers to the entire sequence as Pocono, but I.C. White notes (p. 709):

"Mr. David White has examined the fossil plants occurring with these coals, and unhesitatingly pronounces them of Pocono age, in entire agreement with the conclusions of Prof. William B. Rodgers [sic], who first examined and described the coal beds in his annual reports of the Geological Survey of Virginia, sixty odd years ago. The writer had hoped that some of these basins might contain coal of Pottsville age instead of Pocono, but David White's conclusions from the flora effectually disposes of any such possibility."

Stose and Swartz (1912) mapped the geology and introduced the term "Pocono Group" for the entire sequence above the Devonian Catskill Formation (now Hampshire Formation in West Virginia) based on David White's paleobotany identification. They divided the Pocono Group into five formations, in ascending order, Rockwell, Purslane, Hedges, Myers, and Pinkerton (Figure 1). The major coal-bearing horizon is the Hedges Shale from which David White collected most of his fossil plant specimens in 1902. However, Stose and Swartz (p. 14) noted that:

"The red color and general character of the shale of the Myers formation, together with its stratigraphic position, suggest its correlation with the Mauch Chunk shale of the Mississippian. No fossils have been obtained from this formation. Scant collections of fossil plants from coal beds in the overlying Pinkerton sandstone have been pronounced by David White to be older Mississippian, and the Myers shale is therefore also regarded as of Pocono age."

Grimsley (1916) followed the stratigraphy and structure established by Stose and Swartz and retained a Pocono age for all five formations of Meadow Branch synclinorium with additional fossil plants again identified by David White. Grimsley (p. 138), however, noted:

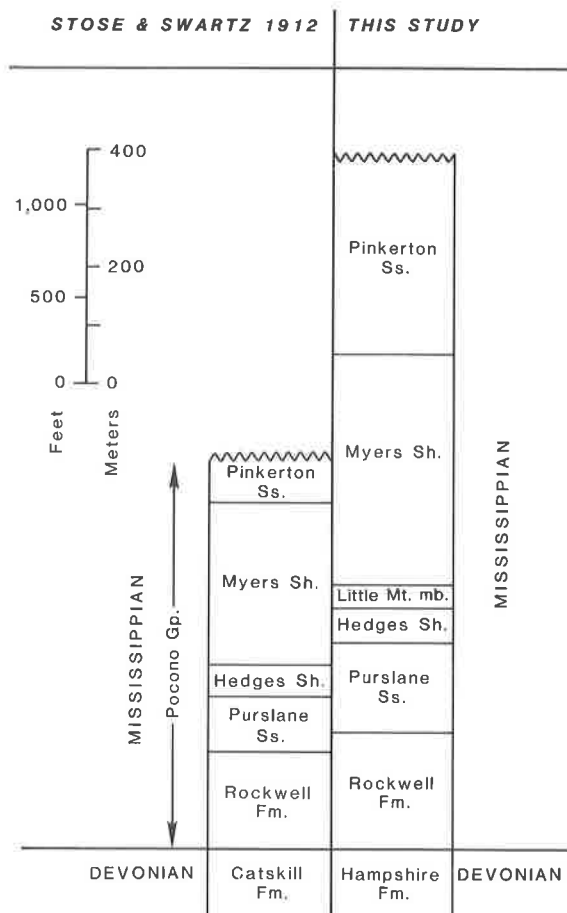


Figure 1. Comparison of stratigraphy by Stose and Swartz (1912) and the present interpretation in Meadow Branch synclinorium, West Virginia.

“The bright red shales of this section [Myers] suggest the Mauch Chunk, but fossil plants found in the upper sandstone formation [Pinkerton] were determined to be Pocono in age.”

In 1955, Read extensively discussed the Pocono flora and made several interesting comments concerning the area in question (p. 10):

“In the Meadow Branch syncline, West Virginia and Maryland, it has already been indicated that there is a question regarding the correlation of the units assigned in the past to the Pocono formation [sic] (Stose and Swartz, 1912). The coal-bearing Hedges shale contains a flora, which was reported in the Paw-paw-Hancock folio (Stose and Swartz, 1912) and which the writer has examined more recently. It is as follows:

Rhodea vespertina Read

Triphyllopteris lescuriana (Meek)

Carpolithes sp

Lepidodendropsis scobiniformis (Meek)

This flora is clearly that of the *Triphyllopteris* zone, widespread in the

upper part of the Pocono. Its position, according to past classifications of the strata, is about in the middle of the so-called Pocono group.

Above the Hedges is the red Myers shale, and above the Myers is the light-gray Pinkerton sandstone containing rounded quartz pebbles. Coaly beds at the top of the Pinkerton are reported [by David White] to yield material referable to *Lepidodendron* [*Lepidodendropsis*] cf. *L. corrugatum*. If this identification were positive, there could be no reasonable doubt regarding the Pocono age of the Pinkerton sandstone. However, the writer has examined the material and does not concur with the previously expressed opinion. Rather, he feels that the material is not identifiable, except in a general way as *Lepidodendron*. It therefore seems only fair to re-open the question of the age of the post-Hedges formations. Lithologic evidence favors the correlation of Myers with Mauch Chunk, and Pinkerton with basal Pennsylvanian [Pottsville], and the flora of the Hedges indicates an upper Pocono age for that unit."

Read and Mamay (1964) use *Triphyllopteris* to define floral zone 2 as upper Pocono (Osagean), which is the floral zone found in the Hedges. Floral zone 3 (*Fryopsis* and *Spheropteridium*) is Mississippian Mauch Chunk (Chesterian). Floral zone 4 (*Neuropteris pocahontas* and *Mariopteris eremopteroides*) and floral zone 5 (*Mariopteris pottsvillea* and *Aneimites*) are Lower Pennsylvanian Pottsville (Morrowan). Read and Mamay do not define a floral zone between 2 and 3, because stratigraphically it represents the Maccrady (early Meramecian) and the marine Greenbrier Limestone (late Meramecian and early Chesterian), neither of which contain plant fossils (compare with Wagner, 1984).

STRATIGRAPHY

Recent geological mapping (Lessing and others, 1989, 1991) of the five "Pocono" formations revealed several perplexing problems concerning the stratigraphy and structure (Figures 1 and 2). The Rockwell Formation is 200 m (650 feet) thick and composed of fine- to coarse-grained sandstone (sublitharenite, Bjerstedt, 1986) with minor conglomerate. Near the top, siltstone and shale are common. At the northern end of the synclinorium, approximately 12 m (40 feet) above the base, is a 2 m (6 feet) thick polyolithic diamictite containing clast of quartzite and sandstone up to 20 cm (8 inches) in diameter set in a sand and mud matrix. This diamictite has been noted elsewhere in the Rockwell (Berg and Edmunds, 1979; Bjerstedt, 1986; Brezinski, 1989). The marine Riddlesburg shale has not been found. The Purslane Sandstone is a 150 m (500-feet) thick sequence of medium to coarse sandstone with minor conglomerate beds. This unit lacks the diamictite and shale beds of the Rockwell. Several coal prospects have been noted, particularly near the top. This sandstone formation has been correlated with the Burgoon Sandstone in Pennsylvania (Berg and Edmunds, 1979). The Hedges Shale is poorly exposed, but estimated to be 58 m (190 feet) thick. It is primarily a gray shale with perhaps 25 percent sandstone. The Hedges contains thin, semi-anthracite coal beds that have been sheared and crushed, but none of the numerous coal prospects ever proved economical. The Rockwell, Purslane, and Hedges are slightly thicker than originally mapped in 1912. The Rockwell and Purslane sandstone formations exist in the northern half and western limb of Meadow Branch synclinorium, but the southern portion of the east limb is now

considered to be entirely Pinkerton sandstone. Only the Rockwell and Purslane formations can be correlated into Maryland (Brezinski, 1989).

Above the Hedges, which Read and Mamay (1964) place in upper Pocono (Osagean), is 396 m (1,300 feet) of Myers red shale including a basal yellow-buff sandstone approximately 30 m (100 feet) thick. The authors have named this sandstone the Little Mountain member (informal) for excellent exposures at Little Mountain. The State geologic map (Cardwell and others, 1968) erroneously replaced Myers with Hampshire in Meadow Branch synclinorium based on the unlikely correlation proposed by Burford and others (1969).

Overlying the Meyers is an additional 350 m (1,150 feet) of Pinkerton sandstone with extensive conglomeratic zones. The areal extent of the Pinkerton is considerably greater than indicated by Stose and Swartz (1912) and Grimsley (1916). What they mapped as Rockwell and Purslane in the southern part of the synclinorium is now shown to be Pinkerton with a thickness of 350 m (1,150 feet) as traced from the type locality on Pinkerton Knob.

The stratigraphic correlation of the five formations that constitute Meadow Branch synclinorium creates interesting problems primarily because the entire sequence does not exist elsewhere (Figure 3). Most authors agree that the Rockwell can be correlated with the lower Price Formation, Huntley Mountain Formation, or the Spechty Kopf Formation. Further, the Purslane is equivalent to the Burgoon Sandstone, which in turn, is equivalent to the Pocono Formation (Berg and Edmunds, 1979). The Hedges and Little Mountain member are not described outside of Meadow Branch to our knowledge. The thick 396 m (1,300 feet) red shales of the Myers has its closest affinity to the Maccrady Formation that reaches 500 m (1,640 feet) thick (Warne, 1990). Finally, the 350 m (1,150 feet) of Pinkerton Sandstone can not be correlated beyond Meadow Branch synclinorium. Its stratigraphic position indicates it belongs between the Maccrady and Greenbrier at the regional unconformity.

PALEOBOTANY

There are 29 coal prospect pits reported in the literature for Meadow Branch synclinorium. The authors have confirmed 15 locations and found 16 pits not previously reported. Nine old coal-prospect pits are known to occur in the Pinkerton Sandstone, but none of these sites yielded identifiable floras. All of the coal is crushed and sheared, semi-anthracite and consequently, spores have also been obliterated. However, a coal location in the upper Pinkerton not previously mined was excavated by the authors and *Rhodeopteridium* cf. *vespertina* (Read) and *Triphyllopteris* cf. *lesciana* (Meek) have been identified (Figure 4) from associated shale and siltstone. These floras are identical to those in the Hedges. These Pinkerton shale samples were also analyzed for spores, but they have been obliterated as in the coal.

The occurrence of *R. vespertina* and *T. lesciana* tentatively restricts our correlation to the Lower Mississippian. These floras of zone 2 are traditionally limited to Osagean time by Read and Mamay (1964)(Figure 3) because overlying strata do not carry plant fossils (ie, Maccrady, Greenbrier, and Mauch Chunk). However, these floras have now been identified 760 m (2,500 feet) above this traditional Osagean horizon in what appears to be the only place in the Appalachians where such plant-bearing rocks have been preserved. Wagner (1984) places the top of the *Triphyllopteris* zone in the middle of the Meramecian,

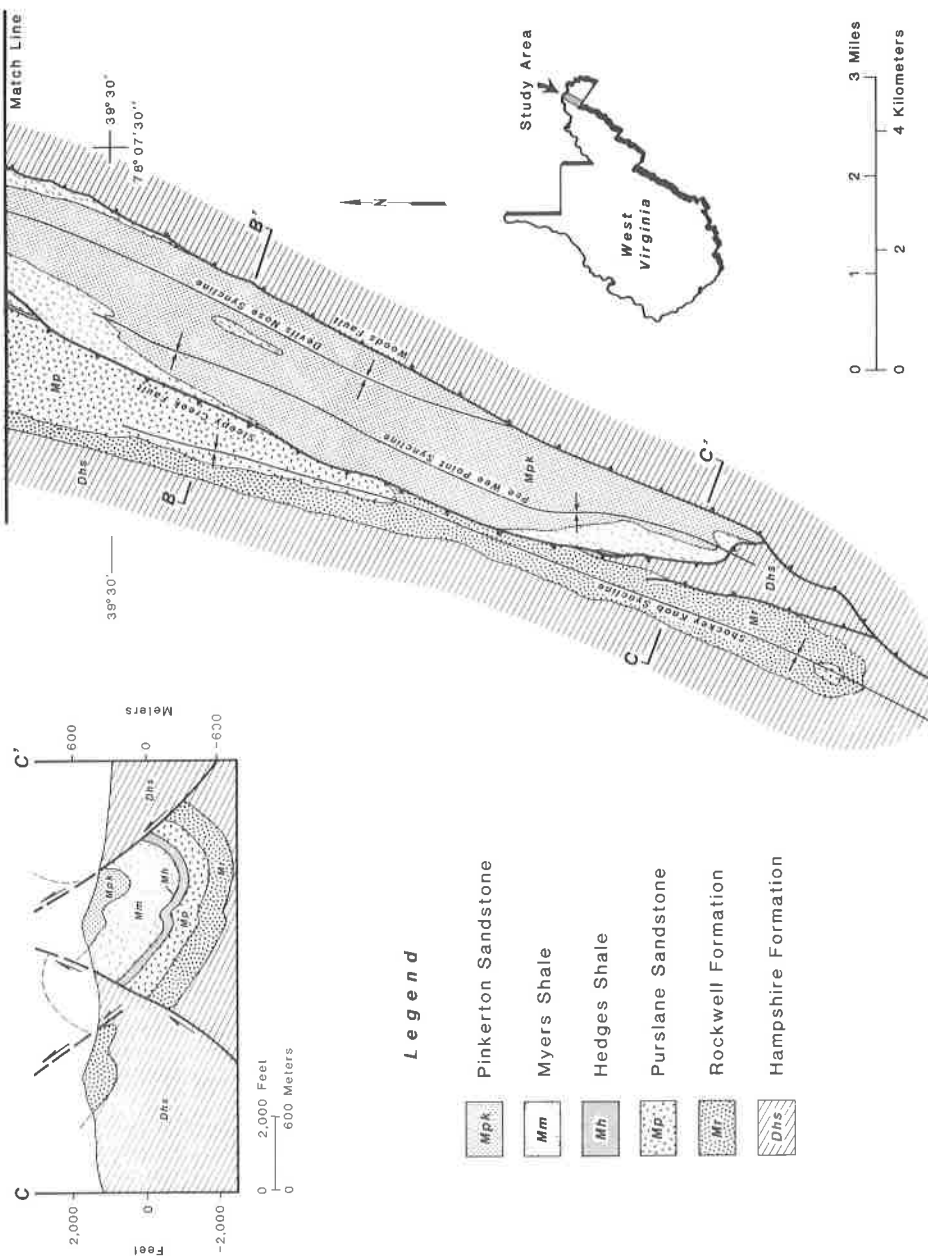


Figure 2. Generalized geological map and cross sections of Meadow Branch synclinorium, West Virginia based on detailed 1:24,000-scale mapping by the authors.

SYSTEM	SERIES	STAGES		Ma	FLORAL ZONE	Southwestern Virginia	Meadow Branch West Virginia	Broad Top Pennsylvania	Southern Anthracite Pennsylvania
		Europe	North America			1.		2.	3.
PENNSYLVANIAN		Bashkirian	Morrowan	330	4	Pottsville Gp.		Pottsville Gp.	Pottsville Gp.
MISSISSIPPIAN	Upper	Serpukhovian	Chesterian	340	3	Mauch Chunk Fm.		Mauch Chunk Fm.	Mauch Chunk Fm.
					2	Greenbrier Ls.			
		Visean	Meramecian	350	1		Pinkerton Ss.		
							Myers Sh.		
							Little Mt. mb.		
	Lower	Tournaisian	Osagean		2	Maccrady Fm.	Hedges Sh.		
							Purslane Ss.	Burgoon Ss.	Pocono Fm.
		Knoxian		365	1	Price Fm.		Rockwell Fm.	Spechtly Kopf Fm.
							Rockwell Fm.		
DEVONIAN		Famennian	Conewangoan			Chemung Fm.	Hampshire Fm.	Catskill Fm.	Catskill Fm.

Figure 3. Proposed correlation of Mississippian stratigraphy in Meadow Branch synclinorium, West Virginia (from COSUNA Charts and 1. Warne, 1990; 2. Sevon, 1986; 3. Wood and others, 1969; Edmunds and others, 1979).

which supports our present correlation (Figure 3).

Several authors have suggested that the Meyers may be equivalent to the Mauch Chunk Formation (Chesterian), while the Pinkerton may be correlated with the Pennsylvanian Pottsville (Morrowan) (White, 1903; Stose and Swartz, 1912; Grimsley, 1916; Read, 1955; and Lessing and others, 1991). However, with the authors recent discovery of *R. vespertina* and *T. lescuriana* near the top of the Pinkerton, we must correct ourselves and partially concur with David White's original determination of Lower Mississippian. Our proposed stratigraphic correlation however, necessitates the range of *Triphyllopteris* be extended upwards at least through early Meramecian, in agreement with Wagner (1984).

STRUCTURE AND REGIONAL DEPOSITION

The Rockwell, Purslane, Hedges, and Myers are clearly truncated on the eastern side of the synclinorium by the Woods fault and do not continue to the ridge top and south along Third Hill Mountain to Pee Wee Point as mapped 75 years ago. The authors do not believe that the fold mapped by Stose and Swartz (1912) south of Pinkerton Knob exists, as there is no structural evidence for it in the Pinkerton or underlying Hampshire Formation. The newly defined Woods fault, extending at least 35 km (22 miles) in West Virginia from Virginia to Maryland, has thrust the Devonian Hampshire Formation onto the Pinkerton at the southern end. At the northern end, this fault has thrust Short Mountain syncline

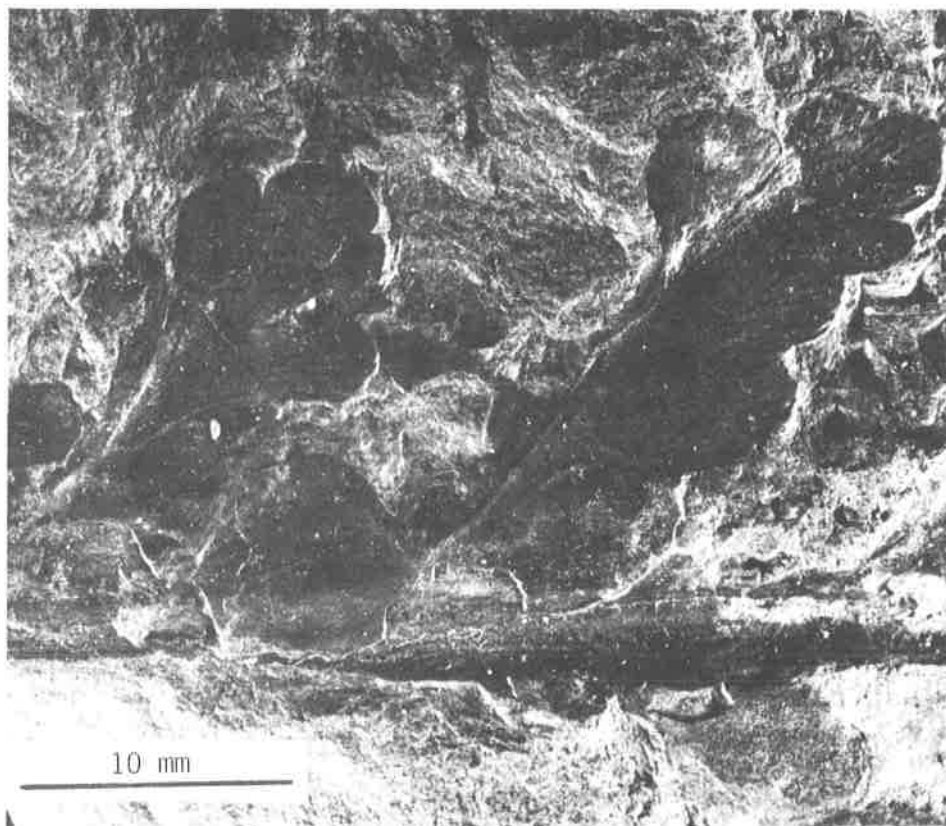


Figure 4. Photograph of *Triphyllopteris* cf. *lescuriana* (Meek) from the Pinkerton Sandstone, Meadow Branch synclinorium, West Virginia.

against Devils Nose syncline and truncated Short Mountain syncline.

Prior to displacement along the Woods fault, the back-thrusted Sleepy Creek fault cut out Purslane, Hedges, and Myers formations near the southern end of the synclinorium. Cross sections accompanying Figure 2 illustrate the complexity associated with these two new major faults.

Restoration of folds, faults, and layer-parallel shortening places the depositor of rocks within Meadow Branch synclinorium a minimum of 48 km (30 miles) east (Kulander and Dean, 1986). Dennison (1961) estimated that Meadow Branch was 32 km (20 miles) east on his palinspastic map, while Gwinn (1970) suggested 80 km (50 miles) of westward displacement for the Paleozoic sequence in the Great Valley and Valley and Ridge of southern Pennsylvania.

Based on isopach maps for the Mississippian by Colton (1970) and DeWitt and McGrew (1979, plate 7), it would appear reasonable that a thick wedge was deposited in the Appalachians during this time period. Mississippian thicknesses of 1,220-2,440 m (4,000 to 8,000 feet) are not unrealistic along the western margin of the Blue Ridge. Wood and others (1969) and DeWitt and McGrew (1979) report nearly 2,440 m (8,000 feet) in the southern anthracite field of Pennsylvania, while Edmunds and others (1979) suggest upwards of 3,050 m (10,000 feet) at this location. Averitt (1941) reported 2,073 m (6,800 feet) of Mississippian in southwestern Virginia and Warne (1990) notes that the Price and Maccrady alone

exceeds 1,006 m (3,300 feet) in the same area. DeWitt and McGrew (1979) further note that Hurricane Ridge in Mercer County, West Virginia has 1,830 m (6,000 feet) of Mississippian. The Mississippian sequence in Meadow Branch synclinorium is nearly 1,220 m (4,000 feet) thick and could easily have been 1,830-2,440 m (6,000 to 8,000 feet) had the upper Mississippian not been eroded.

The isopach data and lithologic characteristics indicate that during Mississippian time a large source area east of the Appalachian basin supplied clastic material to the basin in the order of 1,830- 3,050 m (6,000 to 10,000 feet) thick. The major exception represents the marine transgression that deposited a maximum of 915 m (3,000 feet) of Greenbrier Limestone in southern West Virginia and southwestern Virginia. Periodic uplift and erosion certainly took place generating coarse conglomerates and unconformities. The Maccrady, Myers, and Mauch Chunk redbeds also indicate deposition in a low-relief area, such as a deltaic plain. Minor deposits of coal, diamictite, dolomite, gypsum, and the marine Riddlesburg unit provide evidence for a more complex depositional history which is beyond the scope of this paper.

We envision that thick Mississippian deposition took place in the Meadow Branch area when it was located near the present-day Blue Ridge Mountains. Subsequent deformation caused by the Alleghenian orogeny moved Meadow Branch westward a minimum of 48 km (30 miles) to its present location, riding on the Martinsburg décollement. Later erosion removed large portions of this thick Mississippian clastic wedge leaving only five scattered synclines with exposed Mississippian rocks in West Virginia's Valley and Ridge province (Meadow Branch, Sideling Hill, Spring Gap, Whip Cove, and Town Hill) (Lessing and others, 1990). Only Meadow Branch contains Myers Shale and Pinkerton Sandstone; the other synclines are too shallow.

CONCLUSION

The Meadow Branch synclinorium is composed of five formations as originally reported by Stose and Swartz (1912). In ascending order, they are the Rockwell Formation, Purslane Sandstone, Hedges Shale, Myers Shale, and Pinkerton Sandstone, totaling 1,154 m (3,786 feet). All five formations are considered Mississippian based on the occurrence of *Triphylopteris* cf. *lescuriana* and *Rhodeopteridium* cf. *vespertina* floras found in the Pinkerton Sandstone.

This unique stratigraphic sequence presents difficulties with correlation, particularly with the Myers and Pinkerton. We believe the Myers is equivalent to the Maccrady Formation (Meramecian) while the Pinkerton (also Meramecian) is missing entirely from all other parts of the Appalachians.

The synclinorium is intensely folded and faulted reflecting minimum westward transport of 48 km (30 miles) of the allochthonous Martinsburg sheet. The newly-mapped Woods fault along the eastern portion of the synclinorium extends at least 35 km (22 miles) in West Virginia from Maryland to Virginia.

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