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Abstract

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PETROLOGY OF THE MAFIC IGNEOUS COMPLEXES IN THE
SOUTHEASTERN U. S. PIEDMONT: II. THE BUFFALO MAFIC
IGNEOUS COMPLEX, UNION COUNTY, SOUTH CAROLINA

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ABSTRACT

Detailed petrologic studies on the Buffalo mafic igneous complex, located three miles southwest of Buffalo, Union County, South Carolina, show it to be composed of mostly gabbro, some norite, and spatially related biotite-quartz monzonite and pyroxene hornfels. It is two by four miles in size and intrudes the Kings Mountain metamorphic belt.

The complex consists of olivine (0-25%), orthopyroxene (1-42%), clinopyroxene (7-51%), brown amphibole (0.3-41%), plagioclase (10-67%), and lesser amounts of biotite and opaque and corona minerals occur in the mafic complex. Electron-probe analyses indicate that the olivine (F₀₆₈ to F₀₈₀), orthopyroxene (En₇₅ to En₇₈), and clinopyroxene (Wo₄₂ En₄₆ to Wo₄₄En₄₁) display little or no discernable trend within the complex. The plagioclase compositions range from An₆₀ to An₆₆ for the gabbros and An₆₇ to An₈₀ for the norites.

Spectrochemical and modal analyses on 32 rock samples suggest that the mafic complex is layered. Trend surface and isochemical diagrams show modal plagioclase and Al₂O₃ decrease from the margin inward, whereas orthopyroxene, clinopyroxene, brown amphibole, MgO and solidification index increase in that same direction. CaO, total iron, olivine and mafic and felsic indices remain about constant.

The mineralogy and petrochemistry of the complex indicate both tholeiitic and calc-alkaline affinities. The gross outline, textures and mineralogies suggest that the mafic complex has undergone little or no regional metamorphism.

INTRODUCTION

Until recently, research on the mafic intrusions in the southeastern U. S. Piedmont was largely neglected or at most reconnaissance in scope. However, beginning in the early 1960's, several workers began concentrated studies on these intrusions in an attempt to elucidate their distribution, shape, petrology, petrochemistry, age, and role in the geologic history of the southeastern Piedmont, especially in the Carolinas and Georgia.

Results of these studies have been published by Butler (1964, 1966, 1969), Butler and Ragland (1966, 1969), Medlin (1968, 1969), Hermes (1966, 1968, 1969, 1970), Price (1969), Matthews (1967, 1969), Myers (1968), Overstreet and Bell (1965a, 1965b) and Larrabee (1966). The approach and scope of these studies have been diverse. They range from reconnaissance mapping to detailed petrologic and petrochemical characterization of individual intrusions. These diverse approaches have revealed that the mafic rocks can be divided into the following categories:

1. Those exhibiting regional metamorphic textures and structures.
2. Those showing little or no regional metamorphic textures and structures; and containing no regional metamorphic minerals.
3. Those exhibiting combinations of 1 and 2, above.

Detailed work on individual mafic plutons has revealed definite differences in rock associations, mineralogies, and petrochemistries, which indicate at least two different types of mafic intrusions are recognizable - gabbro-diorite-syenite and gabbro-norite. The gabbro-diorite-syenite complexes are represented by the Mt. Carmel complex in South Carolina (Medlin, 1968; Medlin *et al.* in preparation) and Concord complex in North Carolina (Legrand and Murdorff, 1954; Bell, 1960; Butler and Ragland, 1969). Examples of the gabbro/gabbro-norite associations are the Mecklenburg complex in North Carolina (Hermes, 1966, 1968, 1969, 1970) and in Georgia, the Gladesville norite (Matthews, 1967, 1969) and Presley Mill gabbro (Myers, 1968). According to Hermes (1970, p. 137) the Mecklenburg complex "... consists of a regionally metamorphosed metagabbro of early Paleozoic age intruded by pre-metamorphic adamellite, diorite-tonalite and a post-metamorphic gabbro of late Paleozoic age." Butler and Ragland (1969) concluded that the more than 30 separate gabbro-diorite-syenite intrusions (they grouped the gabbro-diorite-syenite and gabbro/gabbro-norite categories) in the Carolinas and Georgia form an alkali-calcic magma series showing moderate iron enrichment. In the same study Butler and Ragland (1969, p. 183) indicate that the "parental magma" for the Mecklenburg and Concord complexes was a high-alumina basalt. Earlier studies by Butler and Ragland (1966) suggested that these mafic intrusions were derived from a common magma of tholeiitic composition. Studies by Medlin (1968) suggest that the Mt. Carmel rocks were derived from an

alkali-basalt magma.

The results from the above studies indicate that to fully understand these mafic intrusions more detailed work is required. This study is directed toward the characterization of one of these complexes, the Buffalo mafic complex, in terms of the distribution of: 1) petrology, 2) mineralogy, 3) petrochemistry, and 4) the comparison of it with other mafic complexes.

Acknowledgments

The authors are indebted to a number of individuals and organizations. Thanks are extended to the College of Earth and Mineral Sciences Experiment Station, The Pennsylvania State University, under the directorship of M. E. Bell, for financial assistance so that the rock and mineral analyses could be undertaken. Deep appreciation is acknowledged for the unlimited co-operation and advice given by Norman Suhr, acting director, and J. B. Bodkin and D. H. Anderson, of the Mineral Constitution Laboratory, during the rock and mineral analyses.

Thanks are due to the Society of Sigma Xi for a grant-in-aid defraying most of the field expenses and to Lauren Wright and the Department of Geology and Geophysics, The Pennsylvania State University, for additional financial aid. Publication expenses were supported by a West Georgia College Faculty Research Grant.

METHODS

Initial studies on the Buffalo complex involved detailed field mapping and grid-sampling. These were followed by detailed thin-section studies on some 50 samples, spectrochemical analyses on some 40 samples and electron microprobe analyses on three minerals from six samples. Thin-section studies were designed to determine mineralogy, rock fabric, paragenetic trends, and modal variations not evident in hand specimens. Thirteen hundred points were counted for each modal analysis (Table 1); plagioclase compositions were estimated from extinction angles on sections perpendicular to a. The equipment, excitation conditions, accuracy, and precision of the spectrochemical method used are discussed by Suhr and Ingamells (1966) and Suhr (1967); results are given in Table 1.

Electron microprobe procedures and trend-surface analyses are discussed under the appropriate headings.

GENERAL GEOLOGY

Regional Geology

The distribution of some of the mafic plutons in the Carolinas and Georgia are shown on Figure 1, along with the geologic belts in South Carolina (Overstreet and Bell, 1965b). A comprehensive discussion of the "belt" concept as used in the Piedmont can be found in King (1955), Overstreet and Bell (1965a) and Butler and Ragland (1969). The Buffalo complex (Figure 1) was intruded into rocks of the northeasterly trending Kings Mountain metamorphic belt (Overstreet and Bell, 1965b). According to Overstreet and Bell (1965a) the metamorphic rocks of this belt have assemblages indicative of the upper greenschist facies.

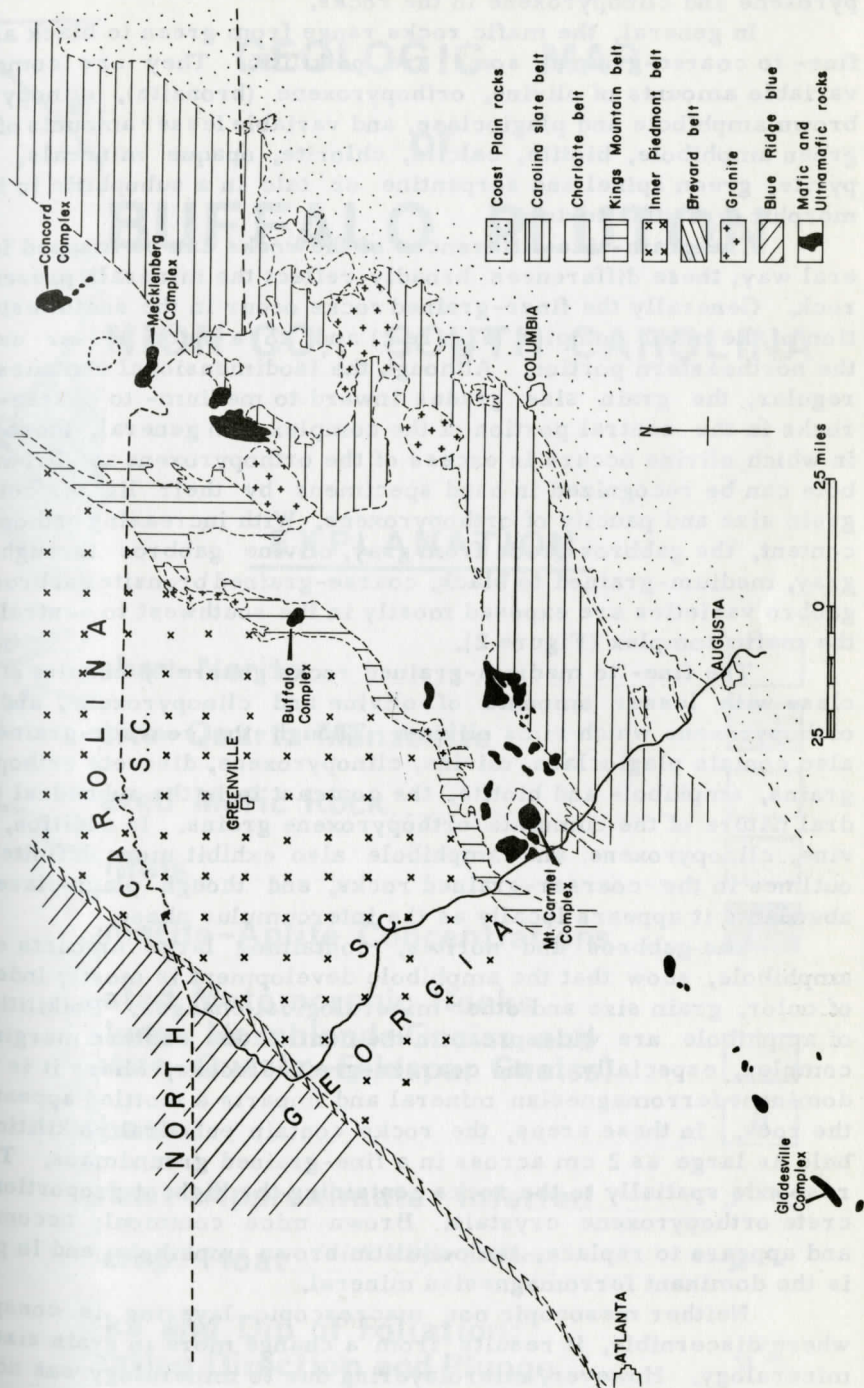
Areal Geology

The Buffalo complex is approximately two by four miles in size, with its longer dimension trending northwest. The mafic complex is composed mostly of several varieties of gabbro and norite, altered mafic rock, hornfels, biotite-quartz monzonite, and minor dikes and veins. This mafic character was recognized by Lieber (1858) and Lounsbury et al. (1914). Subsequently, the complex was grossly outlined by Overstreet and Bell (1965b). Butler and Ragland (1966) reported a pyroxenite consisting of 76 percent clinopyroxene and 11 percent olivine in this mafic intrusion.

Generally the mafic rocks crop out through the soil cover in the form of knobs or as spheroidally weathered float boulders. Narrow ridges preferentially occur on outcrops of those rocks cut by veinlets 1 to 3 mm in width. Black, fine-grained amphibole-plagioclase dikes are found locally.

Gabbro-Norite - The Buffalo mafic rocks display a variety of colors, textures and mineral assemblages. This variable physical appearance, along with a rarity of continuous exposure and with random outcrop patterns, make the rocks difficult to classify or subdivide in the field. However, detailed petrographic studies suggest that the gabbro and norite may grade into each other and that a division into several norite varieties and seven gabbro varieties is possible (Medlin, 1968).

Figure 1. Index map showing location of the Buffalo and other mafic complexes and the metamorphic belts in South Carolina, North Carolina and Georgia (map modified from following sources: Geologic map of North Carolina, 1958; Overstreet and Bell, 1965b; Matthews, 1967, Geologic map of Georgia, 1939; Medlin, J. H., unpub. maps.).



This gabbro-norite division is based on the relative amounts of orthopyroxene and clinopyroxene in the rocks.

In general, the mafic rocks range from green to black and from fine- to coarse-grained; some are poikilitic. They are composed of variable amounts of olivine, orthopyroxene (bronzite), clinopyroxene, brown amphibole and plagioclase, and variable lesser amounts of fibrous green amphibole, biotite, calcite, chlorite, opaque minerals, apatite, pyrite, green spinel and serpentine or talc in a subophitic to hypidiomorphic-granular texture.

The grain-size differences of the rocks can be mapped in a general way; these differences broadly reflect the minerals present in the rock. Generally the finer-grained rocks occur in the southwestern portion of the mafic complex (Figure 2) and, to a much lesser extent, in the northeastern portion. Although the isodimensional contours are irregular, the grain size grades inward to medium- to coarse-grained rocks in the central portion of the complex. In general, those gabbros in which olivine occurs in excess of the orthopyroxene and brown amphibole can be recognized in hand specimens by their lighter color, fine grain size and paucity of orthopyroxene. With increasing orthopyroxene content, the gabbros grade from gray, olivine gabbros through dark-gray, medium-grained to black, coarse-grained bronzite gabbros. These gabbro varieties are exposed mostly in the southwest to central part of the mafic complex (Figure 2).

The fine- to medium-grained rocks generally consist of plagioclase with lesser amounts of olivine and clinopyroxene, and a little orthopyroxene which rims olivine. Though the coarser-grained rocks also contain plagioclase, olivine, clinopyroxene, discrete orthopyroxene grains, amphibole and biotite, the contrast is in the subhedral to euhedral nature of the cumulate orthopyroxene grains. In addition, the olivine, clinopyroxene and amphibole also exhibit more definite crystal outlines in the coarser-grained rocks, and though plagioclase is less abundant, it appears locally as the intercumulus phase.

The gabbros and norites, containing large amounts of brown amphibole, show that the amphibole development is mostly independent of color, grain size and other mineralogical changes. Poikilitic grains of amphibole are widespread in the central and western margins of the complex, especially in the coarser-grained rocks, where it is the predominant ferromagnesian mineral and imparts a mottled appearance to the rock. In these areas, the rocks contain euhedral poikilitic amphibole as large as 2 cm across in a fine-grained groundmass. This corresponds spatially to the rocks containing the highest proportion of discrete orthopyroxene crystals. Brown mica commonly accompanies, and appears to replace, the poikilitic brown amphibole; and in places it is the dominant ferromagnesian mineral.

Neither mesoscopic nor macroscopic layering is conspicuous; where discernible, it results from a change more in grain size than in mineralogy. However, microlayering due to mineralogy was noted in a

few thin-sections and a megalayering is apparent from a trend surface analysis of modal and chemical data (see pages 12-21; 25-26).

A typical example of microlayering is found in U127-23 where layers, 1 to 2 mm wide, of plagioclase laths plus subordinate clinopyroxene and olivine are separated by layers, 1 to 2 mm wide, of clinopyroxene and olivine grains with subordinate plagioclase. The layers are not constant in thickness, and may pinch out in a short distance. Locally (U127-27), granular patches composed of 10 to 20 grains of orthopyroxene, clinopyroxene and plagioclase, all less than 0.2 mm in diameter, occur in a coarser-grained rock. These may represent hornfels inclusions or segregated mineral clusters.

In several places, coarse-grained gabbro is enclosed by a finer-grained variety. These include a round "inclusion" or pod of coarser-grained rock completely surrounded by the finer-grained gabbro; and a coarse-grained stringer-like inclusion with color and mineralogy the same as the finer-grained enclosing rock. In both cases there is no apparent reaction across a sharp contact.

Where muscovite-quartz-feldspar aplites and pegmatites intrude the mafic rocks, the latter are altered to a dark-green rock composed of amphibole, chlorite, epidote and sericitic plagioclase. In extreme cases the rock is composed predominantly of green to brown mica resembling vermiculite, which may occur as pod-like or string-like masses within the pegmatites or aplites as well as along their contacts.

The attitude of these pegmatite and aplite intrusions appears to be variable, changing from horizontal sinuous masses to almost vertical dikes within a few feet. Their spatial distribution does not appear to be controlled by the mafic rock-quartz monzonite contact, because they occur near the contact as well as within the mafic rocks; they are later than either of these rock types.

Anastomosing veinlets, 0.5 mm or less in width, cross the mafic rocks and produce an alteration zone visible microscopically, about 1 mm wide along the veinlet edges. Within and adjacent to this zone the ferromagnesian minerals are changed to colorless to green amphibole rosettes, serpentine, and opaque grains; the plagioclase, to sericite and calcite. From the presence of many opaque grains and rare green spinel (hercynite?) in or near the green amphibole, the following breakdown relationship is deduced: ferromagnesian minerals -- green amphibole + opaque grains + green spinel + serpentine.

Altered Mafic Rock - Altered mafic rock crops out in the northernmost sector of the pluton (Figure 2). A slight rise in elevation, five to ten feet, commonly marks the areal extent of the rock. Outcrops are rare; float is common in some forest and field areas.

The rock is dark-green and coarse-grained; it is cut by numerous anastomosing veinlets, 2 mm or less in width, which impart a "net-veined" appearance to the hand specimens. Very small shear fractures, less than 1 mm in width are also present; these impart a faint foliation to the rock. The veinlets, composed mostly of epidote and quartz,

Table 1. Chemical and Modal Analyses of Buffalo Complex Rocks.

Sample No.	Norites										Gabbros	
	U185-7	U127-9	U125-29	U125-31	U125-6	U127-15	U127-1	U127-3	U127-4	U127-5	U127-7	
SiO ₂	44.00	44.50		45.00	46.10	51.00	46.00	45.00	46.00	45.00	44.50	
TiO ₂	1.32	.50		.40	1.24	1.00	.49	.60	1.01	1.00	.47	
Al ₂ O ₃	11.80	13.20		7.60	18.00	8.80	13.30	19.40	20.20	17.80	8.70	
FeO		8.08 ^a		9.63 ^a		8.26 ^a	5.81 ^a	5.55 ^a	4.64 ^a	6.09 ^a	8.69	
Fe ₂ O ₃	11.80	3.00 ^a		2.10 ^a	12.30	2.13 ^a	1.55 ^a	3.04 ^a	4.12 ^a	3.29 ^a	2.25	
MnO	.16	.155		.18	.135	.20	.115	.11	.12	.115	.17	
MgO	15.70	15.70		23.20	7.90	15.10	13.50	9.70	6.72	9.80	21.00	
CaO	10.20	12.60		9.50	11.30	12.00	13.80	14.50	12.40	13.20	11.30	
Na ₂ O	1.78	.74		.72	2.25	1.06	1.62	1.34	2.45	1.55	.72	
K ₂ O	.56	.14		.17	.43	.17	.10	.11	.37	.14	.16	
SrO	.08	.07		.025	.145	.04	.085	.125	.16	.106	.065	
BaO	.03	.005		.005	.015	.005	.005			.015	.002	
Cr ₂ O ₃	.03	.10		.37	.08	.11	.045			.06	.19	
NiO	.03	.10		.08	.03	.07	.005			.02	.065	
Total	97.49	98.88		98.93	99.92	99.94	96.42	99.30	98.19	98.19	98.28	
Felsic Index	18.7	6.5		8.5	19.2	9.3	11.0	9.1	18.5	10.9	7.2	
Mafic Index	42.9	43.3		35.5	60.9	42.8	37.2	48.7	56.5	50.3	36.0	
Solid. Index	52.6	54.9		62.9	34.5	54.7	58.1	47.6	36.7	45.8	62.4	
Olivine	5.4	7.6	trace	20.9			10.8		4.2	6.1	21.5	
Plagioclase	14.4An ₇₉	31.2An ₈₀	30.0An ₇₄	10.4An ₇₅	52.6An ₇₆	13.6An ₇₅	39.2An ₈₀		63.7An ₇₁	47.8An ₇₉	19.7An ₈₆	
Clinopyroxene	14.5	19.3	9.9	16.1	7.2	36.3	39.6		7.8	27.2	32.8	
Orthopyroxene	23.1	28.1	12.8	18.0	12.0	42.2	3.5		1.9	3.1	8.0	
Opaque Mineral	1.1	2.1	.6	2.6	4.5	.6	.8		3.8	4.1	1.0	
Brn. Amphibole	41.5	7.6	27.9	29.7	22.9	6.3	3.7		15.7	.8	8.4	
Biotite			.4		.8						.4	
Grn. Amphibole		3.7	17.8	.9			1.6		2.3	10.3	7.9	
Other		.4	.6	1.4		1.0	.8		.6	.8	.3	
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00		100.00	100.00	100.00	
Spec. Gravity	3.17	3.12		3.24	2.99	3.21	3.13	3.01		2.99	3.27	

Sample No.	U127-11	U127-13	U127-16	U127-17	U127-18	U127-22	U127-23	U127-25	U127-27	U127-29	U127-30
SiO ₂	44.80	46.20	46.20	47.50		44.00	45.30	44.50	49.60		46.00
TiO ₂	.40	.45	.36	.31		1.00	.28	.44	.87		.46
Al ₂ O ₃	19.30	8.70	14.10	17.70		20.90	22.00	14.50	12.30		19.20
FeO	4.72	7.63 ^a	5.82 ^a	3.95 ^a			3.23 ^a	7.58 ^a			4.23 ^a
Fe ₂ O ₃	2.96	2.53 ^a	1.94 ^a	1.10 ^a		8.70	1.92 ^a	1.48 ^a	8.80		2.30 ^a
MnO	.11	.16	.13	.11		.076	.08	.14	.15		.10
MgO	10.20	18.50	14.90	9.95		8.10	8.70	16.00	12.10		9.00
CaO	14.20	13.00	13.90	15.80		14.00	15.90	11.50	13.20		15.30
Na ₂ O	1.72	.68	1.20	1.20		2.24	1.22	1.12	1.14		2.07
K ₂ O	.18	.12	.16	.09		.12	.13	.06	.68		.11
SrO	.136	.05	.085	.12		.145	.135	.095	.09		.14
BaO	.005	.005	.007			.005	.002	.02	.02		.01
Cr ₂ O ₃	.015	.15	.09			.065	.085	.06	.075		.02
NiO	.04	.04	.01			.005	.005	.005	.02		.005
Total	98.79	98.21	98.90	97.83		99.35	98.98	97.49	99.04		98.94
Felsic Index	11.8	5.8	8.9	7.5		14.1	7.9	9.5	12.1		12.4
Mafic Index	44.6	37.3	36.0	33.6		51.8	38.7	38.2	42.1		43.7
Solid. Index	50.2	61.0	60.4	61.1		42.3	55.9	59.1	53.3		49.5
Olivine	2.8	14.7	7.1	6.3	9.5	5.4	4.1	23.1	3.7	18.2	2.2
Plagioclase	47.2An ₇₂	21.8An ₇₆	32.1An ₇₇	47.2An ₈₀	17.7An ₇₉	58.3An ₇₉	56.8An ₇₃	39.8An ₈₃	47.6An ₇₃	9.7An ₇₄	62.6An ₈₂
Clinopyroxene	23.4	38.0	41.9	34.3	32.0	25.2	28.5	25.7	35.6	49.3	22.2
Orthopyroxene	6.8	19.5	8.7	6.6	11.8	.6	3.4	2.1	5.7	10.0	4.0
Opaque Mineral	3.4	.6	1.0	.2	.6	2.5	.3	.2	.2	.4	1.2
Brn. Amphibole	7.0	5.4	9.2	5.0	24.4	6.9	5.8	6.9	6.3	10.1	5.0
Biotite				.2						1.5	.5
Grn. Amphibole	9.4			1.9	1.1		.7	1.7	.2	.4	2.3
Other				.4	1.9		.4	.5	.7	.4	
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
Spec. Gravity	3.08	3.17	3.11				3.00	3.05	2.96		3.10

^aFeO and Fe₂O₃ determined by J. B. Bodkin, Mineral Constitution Laboratory, The Pennsylvania State University

Table 1. Continued

Sample No.	Gabbros cont.										
	U127-31	U125-1	U125-6	U125-8	U125-14	U125-32p	U185-1	U185-9	U185-13	U185-30	U185-32
SiO ₂	44.40	45.20	46.10	48.40	51.00		46.00	46.00	47.30	48.00	
TiO ₂	.68	.50	1.24	.72	.78		.40	.28	.28	.42	
Al ₂ O ₃	19.30	8.40	18.00	5.90	9.80		16.70	21.90	21.00	18.80	
FeO	5.62	8.61 ^a	6.84	7.08	5.65		6.42 ^a	3.50	3.81	4.44	
Fe ₂ O ₃	4.06	1.94 ^a	4.70	1.84	2.53		2.47 ^a	2.02	1.67	1.77	
MnO	.115	.165	.135	.154	.145		.13	.09	.08	.11	
MgO	9.70	19.70	7.90	14.70	12.40		12.80	8.80	8.30	9.40	
CaO	14.70	11.20	11.30	15.10	15.10		13.50	14.80	15.60	15.30	
Na ₂ O	1.29	.16	2.25	1.70	1.28		1.50	1.36	1.25	2.12	
K ₂ O	.08	.17	.43	.67	.43		.11	.14	.12	.09	
SrO	.125	.045	.145	.018	.055		.09	.15	.11	.13	
BaO	.002	.005	.015	.02	.02		.01	.01	.005	.015	
Cr ₂ O ₃	.005	.12	.08	.15	.22		.07	.035	.05	.035	
NiO	.015	.04	.03	.015	.005		.035	.005	.005	.005	
Total	100.09	97.15	99.16	96.47	99.41		100.23	99.09	99.57	100.63	
Felsic Index	8.6	9.9	19.2	13.6			10.6	9.2	8.0	12.6	
Mafic Index	51.5	37.1	60.9	39.8			42.8	40.1	41.5	39.2	
Solid. Index	45.4	60.6	34.5	54.9			53.3	57.9	53.3	48.7	
Olivine	9.7	9.3			14.3		10.9	8.3		1.9	4.3
Plagioclase	51.6An ₇₆	17.8An ₈₂	52.6An ₇₆	10.5An ₆₄	47.4An ₇₉	47.0An ₇₆	35.3An ₈₂	49.7An ₆₉	An ₈₃	67.2An ₈₀	56.3An ₈₀
Clinopyroxene	31.2	31.2	7.2	58.0	24.0	21.0	40.5	25.7		18.8	26.1
Orthopyroxene	1.2	18.2	12.0	16.3	5.3	16.1	4.5	7.4		5.0	4.8
Opaque Mineral	2.8	.3	4.5	.1	3.9	.7	.1	.4		.3	.4
Brn. Amphibole	1.1	21.8	22.9	8.7	3.6	13.6	3.9	6.0		3.4	5.4
Biotite		.1	.8	6.3		1.1					
Grn. Amphibole	2.1	.5		.1	.15	.5	3.4	2.2		1.8	2.7
Other		.8				.5	.1	.3		1.6	
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00		100.00	100.00
Spec. Gravity	3.19	3.18	2.99	3.16		3.09		3.03	2.88	2.91	2.97

Sample No.	Hornfels					Altered Mafic Rock			
	U185-34	U185-36	U185-37	U185-41	U185-29	U125-11	U125-21	U125-4	U125-33
SiO ₂	47.20	47.50	46.80	47.10	50.20	42.00	49.00	48.50	49.50
TiO ₂	.56	.34	.40	.32	.44	1.06	1.00	.90	1.06
Al ₂ O ₃	15.70	15.30	15.10	16.30	10.50	17.60	14.50	8.10	6.50
FeO	5.19	4.80	5.10	5.03	6.70		9.34 ^a	6.00	
Fe ₂ O ₃	2.74	1.27	1.34	1.02	1.06	12.20	5.09 ^a	2.14	9.50
MnO	.135	.115	.13	.115	.165	.12	.27	.165	.185
MgO	10.80	13.30	11.50	12.70	15.70	6.70	7.00	13.90	15.50
CaO	14.20	15.10	13.90	15.40	12.30	17.60	11.40	13.00	14.30
Na ₂ O	1.36	1.14	1.12	.93	.78	1.03	1.93	1.12	1.09
K ₂ O	.09	.12	.07	.08	.05	.32	.14	.87	.50
SrO	.105	.08	.09	.115	.035	.043	.03	.055	.03
BaO	.015	.005	.008	.10	.006	.005	.215	.035	.012
Cr ₂ O ₃	.06	.15	.03	.12	.14	.005	.01	.14	.065
NiO	.015	.06	.005	.022	.005	.01	.015	.005	.06
Total	98.17	99.27	95.59	99.35	98.08	98.39	100.75	94.92	98.30
Felsic Index	9.1	7.7	7.9	6.2	6.3	7.1	15.4	13.4	10.0
Mafic Index	44.0	33.2	37.8	34.2	35.1	64.5	68.0	38.4	38.0
Solid. Index	52.1	68.8	58.4	62.5	62.7	33.1	29.2	56.1	58.3
Olivine	7.4		7.5	12.2					
Plagioclase	44.0An ₈₀	An ₇₀	39.2An ₇₉	40.8An ₈₄	31.5An ₈₁		53.4An ₆₈		
Clinopyroxene	36.3		36.6	36.7	34.1		25.8		
Orthopyroxene	3.3		9.3	2.5	21.9		15.2		
Opaque Mineral	7.8		.3	trace	.8		5.6		
Brn. Amphibole	3.6		3.3	4.8	11.7				
Biotite	.3								
Grn. Amphibole	3.3		3.8	3.0					
Other									
Total	100.00		100.00	100.00	100.00		100.00		
Spec. Gravity	3.09	3.05	3.07	3.09	3.25	3.09	3.00	2.94	3.04

^a FeO and Fe₂O₃ determined by J. B. Bodkin, Mineral Constitution Laboratory, The Pennsylvania State University

cross the shear planes without offset.

Euhedral, green to black pyroxene and amphibole are conspicuous in hand specimens. These are set in a light-green to chalky-gray matrix of green amphibole, epidote and altered plagioclase. Thin-sections reveal the additional presence of chlorite and apatite. A more detailed description is found in Medlin (1968).

The relict texture and mineralogy of this rock, along with its chemistry (Table 1), suggest that it was originally similar to the gabbros that crop out in its vicinity. If this is correct, then the alteration of this rock took place subsequent to the magma consolidation.

Hornfels - Contact metamorphism of the country rocks extends for a distance of 400 to 800 feet outward from the pluton (Figure 2). Pyroxene hornfels occurs in the northwestern and southeastern portions of the complex as a compact, black, fine-grained rock. Weathered specimens and thin-sections reveal a faint foliation or layering, and a composite mineral assemblage of plagioclase (An₅₅-An₃₃ - 7 samples), orthopyroxene, clinopyroxene, with accessory magnetite, apatite, green amphibole, calcite, pyrite and sphene. The usual assemblage consists mainly of the first four listed minerals.

Except for lack of olivine, the rocks characterized as pyroxene hornfels have mineralogies identical to gabbro. Their textures, however, suggest that they are similar to hornfels discussed and pictured by Hess (1960) for the Stillwater complex in Montana, and by Matthews (1967) for the Gladesville norite in Georgia. Modal and chemical data for samples U185-29, U125-11 and U125-21 are listed in Table 1.

Biotite-Quartz Monzonite - Biotite-quartz monzonite bounds the mafic complex on the southwestern and northeastern margins; and it apparently occurs as three "island-like" masses within the northwestern portion of the complex, where the pluton "necks" down to its smallest width (Figure 2). Whether these masses are connected is not known, but they occur as mounds, as large as 200 feet in diameter and rising 10 to 15 feet above the relatively flat surrounding topography. The westernmost of these "islands" is cut by muscovite-quartz-feldspar pegmatite dikes.

No special effort was made to trace out the extent of the monzonite except where it is in contact with the gabbro. However, topographic and vegetational differences visible on aerial photographs allow some extrapolation of its extent. This rock is classified as a biotite-quartz monzonite (Table 2) despite a wide mineralogical variation between outcrops.

The rock weathers readily into a gravel-like saprolite that can be delineated very precisely, but makes sample collecting for petrographic and other studies difficult. The biotite-quartz-monzonite is gray to purplish-blue and porphyritic. It contains cryptoperthitic feldspar phenocrysts, ranging up to 1 cm by 2 cm in size, that have a rather "wormy" appearance and commonly are twinned. Thin-section studies show additionally that microcline, plagioclase, quartz and biotite, with

Table 2. Chemical and Modal Analyses of Biotite-Quartz-Monzonite.

<u>Oxides</u>	<u>U125-2</u>	<u>U208-9</u>
SiO ₂	64.00	
TiO ₂	.64	
Al ₂ O ₃	18.10	
Fe ₂ O ₃	3.50	
MnO	.055	
MgO	1.10	
CaO	1.20	
Na ₂ O	4.35	
K ₂ O	<u>6.25</u>	
Total	99.19	
 <u>Minerals</u>		
Cryptoperthite	34.6	46.0
Microcline	1.6	7.0
Plagioclase	37.4	25.0
Quartz	12.6	7.7
Biotite	8.4	3.6
Pyroxene	.5	---
Amphibole	---	1.4
Myrmekite	3.1	8.3
Opaque Minerals	<u>1.8</u>	<u>1.0</u>
Total	100.0	100.0

lesser amounts of apatite, opaque minerals, green amphibole, clinopyroxene, muscovite and sphene are present. The feldspars commonly contain patches of sericite, rare epidote and several types of perthite. Some samples contain veinlets of epidote and patches of green amphibole and chlorite. Table 2 lists modal analyses of two of the samples and chemical analysis of one. Additional data can be found in Medlin (1968).

The biotite-quartz monzonite contains dikes of diabase, muscovite-quartz-feldspar pegmatite, quartz-feldspar aplite, hornblende-plagioclase and gray to black, fine-grained, biotite-quartz-feldspar rock; xenoliths of a fine-grained biotite-quartz-feldspar gneiss similar to a gneiss in the northwest part of the area occur locally. At locality U185-12, a mafic dike, similar in composition to the nearby gabbro, cuts the biotite-quartz monzonite with little discernible alteration, suggesting that the biotite-quartz monzonite is older than the gabbro. White to pink quartz-feldspar pegmatite dikes cut through both rocks.

If alteration occurred in the vicinity of the mafic rock-quartz

monzonite contact, it apparently was not extensive. Mafic rocks exposed within 50 feet of the contact with the "island-like" masses are not noticeably changed in mineralogy. At locality U185-12, the mafic rock within 25 feet of the contact displays no noticeable change in its mineralogy or textural features other than a slight change in the amount of opaque grains exsolved from the olivine. However, the biotite-quartz monzonite here contains xenoliths of the biotite-quartz-feldspar gneiss. Individual xenoliths contain from scarce to abundant blue to gray, cryptoperthitic phenocrysts, suggesting that these xenoliths were being converted to biotite-quartz monzonite, or that a metamorphic rock was being changed to biotite-quartz monzonite.

There is no conclusive field evidence to indicate that petrogenetic or intrusive sequence between the biotite-quartz monzonite and mafic rocks; several indirect aspects are discussed by Medlin (1968), and the alternatives are summarized below:

1. The absence of intermediate rocks between the gabbro and biotite-quartz monzonite and the step-like character of their mineralogies and chemistries suggest that they are unrelated petrologically.

2. Conversely, the lack of sufficient exposure, the recognized occurrence of gabbroic and quartz monzonitic rocks together (Best, 1963; Best and Mercy, 1967), the absence of conclusive alteration features and the occurrence of biotite-quartz monzonite as island-like masses in the gabbro would support a petrologic link.

3. A third possibility is that the biotite-quartz monzonite was formed from pre-existing rock by the intrusion of the mafic mass. Supporting this are xenoliths that contain phenocrysts of cryptoperthitic K-feldspar, the existence of a possible source rock in the nearby area, the intrusion of a magma of sufficient temperature to transform such rocks, and finally the distribution of the biotite-quartz monzonite partially around, and as masses within, the mafic rocks. However, the xenoliths may be mere "stoped" inclusions of foreign rock that have not been transformed and, at present, evidence is not available to confirm the possible source rock.

The age relationship between the biotite-quartz monzonite and mafic rock is not known; however, the step-like chemistry and mineralogy, absence of numerous dike types in the mafic rocks, the local presence of cataclastic features in the biotite-quartz monzonite, the occurrence of biotite-quartz monzonite as masses in the mafic rocks, and the presence of a cross-cutting gabbro dike imply the biotite-quartz monzonite is older.

MINERALOGY

The chemical composition of coexisting minerals and their variation both within a given rock type and between rock types afford an insight into the petrologic history of the complex, especially when combined

with petrographic studies and whole-rock chemistry.

The compositions of olivine, orthopyroxene and clinopyroxene were determined using electron-microprobe and x-ray techniques. The electron-microprobe analyses are partial: only MgO and iron were determined for olivine; MgO, iron and CaO, for the pyroxenes. Two olivine compositions were determined by x-ray techniques described by Yoder and Sahama (1957).

In all, six thin-sections were used in the microprobe work (Table 3). In three, coexisting olivine, orthopyroxene and clinopyroxene compositions were determined. In the others, coexisting olivine and clinopyroxene compositions were measured. The probe samples were selected to include coexisting minerals along the major (northwest-southeast) and minor (northeast-southwest) axes of the mafic complex to see if any trends exist for a single mineral species, as well as for coexisting minerals. Each composition is an average value based on measurements of three or more grains per section. At least five measurements were obtained on edge to edge traverses across each grain. The variation between grains within a section may be as much as ten percent; within a single grain it may be up to four percent.

In order to ascertain distribution patterns of the minerals and oxides (MgO, Al₂O₃, Fe₂O₃ and CaO) within the complex, trend surface and isochemical diagrams were computed. A general discussion of the theory and practical application of trend surfaces can be found in Krumbein and Graybill (1965) and Ondrick (1968). Results of these analyses were contrasted with modal mineralogies, other oxides and solidification index (Kuno, 1959).

The microscopic character, compositional variation and abundance of the major minerals are discussed below.

Olivine

Olivine is colorless, mostly round to elongate anhedral chadacrysts, with the following minerals rarely developed along partings: opaque grains, yellow flakes (iddingsite or talc), reddish brown stripes and specks (hematite), colorless to light-green flakes (serpentine), light-brown flakes (iddingsite), and olive-green flakes (chlorite). It occurs as chadacrysts in plagioclase, rarely as oikocrysts. Brown amphibole locally is found within olivine, commonly with opaque minerals concentrated either along the contact or within the amphibole. In the finer-grained rocks, olivine shows a tendency toward flattening or elongation.

Coronas develop between olivine and plagioclase in all the olivine-bearing rocks, regardless of their position in the pluton. The following sequences of minerals are present in these coronas:

- 1) olivine — orthopyroxene — fibrous brown amphibole + green spinel — plagioclase,
- 2) olivine — orthopyroxene — plagioclase,

Table 3. Chemical Composition of Clinopyroxene, Orthopyroxene and Olivine in Buffalo Complex.

Sample No.	Clinopyroxene Atomic Comp.	Orthopyroxene Atomic Comp.	Olivine Atomic Comp.
U127-9	Wo _{41.5} En ₄₄ Fs ₁₄	En ₇₅	Fo ₇₆
U127-13	Wo ₄₂ En ₄₅ Fs ₁₃	En ₇₈	Fo ₇₄
U127-17	Wo ₄₂ En ₄₆ Fs ₁₂		Fo ₇₇
U127-30	Wo ₄₂ En ₄₄ Fs ₁₄		Fo ₇₉
U127-3	Wo ₄₄ En ₄₁ Fs ₁₅	En ₇₆	Fo ₆₈
U125-1	Wo ₄₄ En ₄₃ Fs ₁₃		Fo ₇₆
U127-4			Fo _{80.6*}
U127-6			Fo _{76.5*}

*Composition determined by x-ray diffraction using d 130 (Yoder and Sahama, 1957).

- 3) olivine — orthopyroxene — brown amphibole — plagioclase,
- 4) olivine — orthopyroxene — clinopyroxene + brown amphibole — plagioclase,
- 5) opaque grains + tan to light yellow flakes (probably once olivine) — orthopyroxene — plagioclase,
- 6) olivine — fibrous green amphibole — plagioclase,
- 7) olivine — brown amphibole — fibrous green amphibole — plagioclase,
- 8) olivine — brown amphibole — plagioclase,
- 9) olivine — opaques + brown iddingsite(?) — orthopyroxene — fibrous green amphibole — plagioclase,
- 10) lemon-colored flakes (probably once olivine) — orthopyroxene — fibrous green amphibole — plagioclase,
- 11) opaque minerals + green spinel (probably once olivine) — orthopyroxene — fibrous green amphibole — plagioclase.

The number, type, width, completeness and mineralogy of these coronas display much variation within and between thin-sections (Figure 3A, 3B), but rarely exceed seven different types and 1 mm in width within a thin-section. Thus, the coronas apparently reflect a gradation in the replacement of the olivine, from none to complete, commonly by orthopyroxene and a fibrous green amphibole-green spinel symplektite.

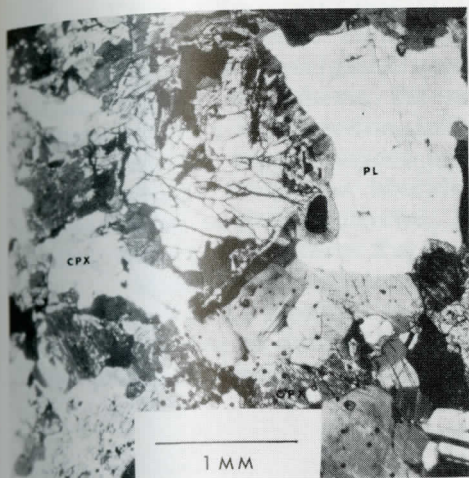


Figure 3A. Olivine corona (U127-17). Olivine grain rimmed by orthopyroxene and green amphibole-spinel symplectite. Other discernible minerals include plagioclase (PL), clinopyroxene (CPX), orthopyroxene (OPX), and opaque minerals. X-nicols.

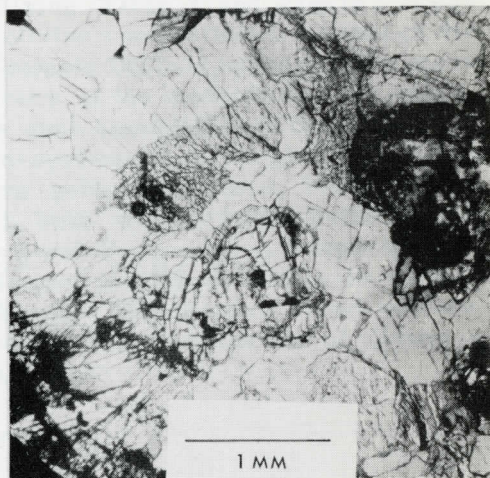


Figure 3B. Olivine corona (U185-30). Olivine grain (center) encircled by orthopyroxene. Other larger minerals include orthopyroxene, clinopyroxene and opaques. X-nicols.

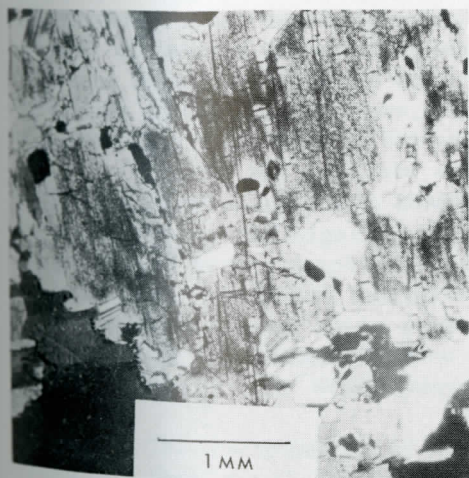


Figure 3C. Poikilitic amphibole (U185-43). Amphibole containing much opaque dust and enclosing plagioclase. Intergrowths of biotite are evident in lower left. X-nicols.

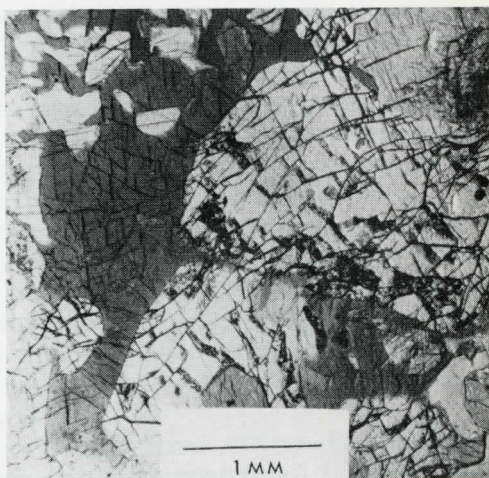


Figure 3D. Intergrowths of brown amphibole and orthopyroxene; amphibole encloses orthopyroxene in places (U127-29). Uncrossed nicols.

Regardless of corona type, almost all the olivine shows vermicular opaque masses within it as well as in the near vicinity. The coronas are thought to result from slow cooling of the pluton, and hence are sub-solidus and subsequent to the consolidation of the magma.

Irrespective of rock type, the olivine from eight samples ranges from Fo₆₈ to Fo_{80.6} (Table 3). Utilizing Table 3 and Figure 2 it can be seen that no trend exists in the olivine compositions. Although the olivine from the marginal rocks (sample U127-3) appears to be more iron-rich than the others, the accuracy of the methods and the variability within a thin-section indicate such compositional trends are not real, and that the variations are non-systematic in the pluton. Olivine compositions from the Mecklenburg gabbro (Hermes, 1970) and the Gladesville norite (Matthews, 1967), both similar to the Buffalo complex, show a range from Fo_{60.1-75.2} and Fo_{79.5-85.5}, respectively.

Modally, the olivine ranges from 0 to 20.9 percent in the norites and from 0 to 23 percent in the gabbros. Third-order trend-surface analyses using 34 data points indicate that the distribution of olivine in the pluton is random (Medlin, 1968).

Orthopyroxene

Pink to gray-green orthopyroxene is present as anhedral to euhedral, mostly fine to medium grains which occur both as partial to complete rims around olivine and as separate disseminated grains, which locally are poikilitic. The rims commonly are intergrown with either fibrous green amphibole or a green, fibrous, amphibole-spinel symplektite. Commonly, the separate orthopyroxene grains show "Bushveld type" exsolution of clinopyroxene in host orthopyroxene, as well as minute acicular opaque and red-brown inclusions in (100). These latter two features were not observed in the orthopyroxene rims around olivine; furthermore, the rims in rare examples have positive optic signs in contrast to the negative optic sign of the separate grains.

Locally the separate orthopyroxene grains have reciprocal enclosing relationships with olivine, clinopyroxene and plagioclase. In places, along the edges and cleavages of the orthopyroxene are intergrowths of brown amphibole and red-brown biotite and rarely green amphibole, which are interpreted here as a replacement texture because of the opaque oxide grains which occur as splotches and minute rods in the amphibole (Figure 3C). Sparse talc or yellow serpentine flakes, intergrown with opaque grains, occur in some orthopyroxene grains.

The composition of three orthopyroxenes (discrete crystals rather than rims) for norites and gabbros range from En₇₅ to En₇₈ (Table 3). Their magnesium and iron content are similar; calcium varies. Two of the three orthopyroxenes are slightly more magnesium-rich than their coexisting olivine. Hermes (1970) reports orthopyroxene values from the Mecklenburg gabbro of Wo_{1.4-2.6}En_{65.2-74.6}Fs_{23.8-32.6}; Matthews' (1967) values for orthopyroxene from the

Gladesville range from Eng₂ to Eng₃.

The origin of a pseudofibrous texture in orthopyroxene, thought to result from exsolution of clinopyroxene in host orthopyroxene was not resolved by the electron-microprobe. The calcium variation within each grain and between grains is so erratic that no accurate concentration values of calcium could be obtained for the mineral; this strongly favors an exsolution hypothesis for the pseudofibrous texture.

The orthopyroxene ranges from 12 to 42.2 percent in the norites and from 0.6 to 21.9 percent in the gabbros. Third-order trend-surface analysis of the modal data (34 samples) shows an elliptical "high" in orthopyroxene concentration trending northwest along the north-central margin of the pluton. The orthopyroxene content decreases toward the southwestern margin of the pluton, reaching very low concentrations near that margin. The higher orthopyroxene concentrations correspond in a general way to the coarse- to very coarse-grained rocks, where, in places, orthopyroxene is poikilitic and is probably a cumulus mineral.

Comparison of orthopyroxene trends with those of olivine, clinopyroxene, brown amphibole and plagioclase reveals several general relationships.

1) Orthopyroxene concentrations show no correlation with either olivine or with plagioclase compositions.

2) The orthopyroxene and clinopyroxene trends are similar, with a general decrease toward the pluton margin. A poor correlation exists with brown amphibole concentrations.

3) Orthopyroxene concentrations increase as plagioclase concentrations decrease.

4) Orthopyroxene and solidification index (SI) trends show a remarkable similarity: those rocks containing the largest amount of orthopyroxene have the highest solidification index. This correlation is so good that in the Buffalo complex, the SI is a measure of the orthopyroxene concentrations.

Clinopyroxene

Clinopyroxene is mostly fine- to medium-grained and anhedral to subhedral. In some grains, exsolution lamellae in (100) of the host clinopyroxene occur as thin discontinuous, rod-like, string-like and bleb-like masses, which display optical properties characteristic of orthopyroxene; however, in places amphibole intergrowths have a similar appearance. Commonly, opaque to reddish-brown acicular and rod-inclusions occur in (100) and rarely in (001). In the larger clinopyroxene grains these opaque grains in (100) become sufficiently concentrated to define a concentric zoning in the host clinopyroxene.

Commonly chadacrytic, the green to gray-green clinopyroxene is intergrown with and enclosed by brown amphibole and, rarely, by green amphibole and biotite. The intergrowths, along with minute

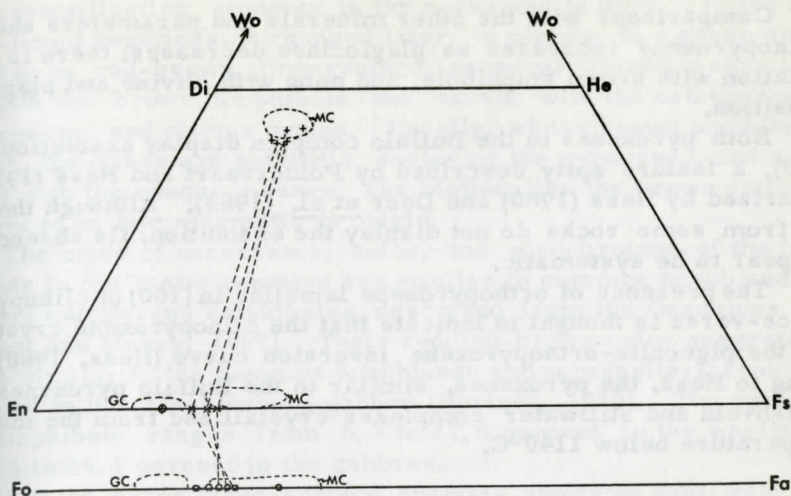


Figure 4. Plot of coexisting clinopyroxene (plus signs) and orthopyroxene (crosses) in pyroxene quadrilateral connected with coexisting olivine (filled circles) for Buffalo rocks. Enclosed dash line areas are: Mecklenburg complex (MC) clinopyroxenes, orthopyroxenes and olivines; Gladesville norite (GC) orthopyroxenes and olivines.

specks of opaque and red-brown mineral and anhedral green spinel, commonly give the clinopyroxene a cloudy appearance. Thin partial (or rarely complete) orthopyroxene rims with sharp contacts occur in places; locally brown amphibole separates the pyroxenes. Rarely, clinopyroxene rims olivine and in places show a reciprocal enclosing relationship with plagioclase.

Clinopyroxene from six localities range in composition from $Wo_{41.5}En_{44.5}$ to $Wo_{44}En_{41}$ (Table 3), and the abundance in 34 samples ranges from 7.2 to 36.3 percent in the norites and from 7.8 to 58 percent in the gabbros. The Fe/Mg relationships between coexisting clinopyroxene, orthopyroxene and olivine are shown in Figure 4, and indicate that little fractionation occurred during the crystallization of these minerals. This conclusion is consistent with that of Hermes (1970) for the Mecklenburg gabbro where the clinopyroxenes have a composition range of $Wo_{43.3-45}En_{41.6-44.8}Fs_{10.1-15.2}$.

Third-order trend surfaces of clinopyroxene show the highest concentration to be in the southeast part of the complex; this decreases to lower values to the northeast and northwest. This distribution pattern indicates that clinopyroxene is the dominant cumulus mineral only in places; a conclusion not consistent with Butler and Ragland (1969, p. 178).

Comparisons with the other minerals and parameters show that the clinopyroxene increases as plagioclase decreases; there is a poor correlation with brown amphibole, and none with olivine and plagioclase composition.

Both pyroxenes in the Buffalo complex display exsolution grains in (100), a feature aptly described by Poldervaart and Hess (1951) and summarized by Hess (1960) and Deer *et al.* (1963). Although the pyroxenes from some rocks do not display the exsolution; its absence does not appear to be systematic.

The presence of orthopyroxene lamellae in (100) of clinopyroxene and vice-versa is thought to indicate that the clinopyroxene crystallized below the pigeonite-orthopyroxene inversion curve (Hess, 1960). According to Hess, the pyroxenes, similar to the Buffalo pyroxenes, from the Bushveld and Stillwater complexes crystallized from the magma at a temperature below 1140°C.

Brown Amphibole

Tan to brown to olive-brown amphibole occurs as anhedral intergrowths, irregular rims (1.5 to 2 mm wide), and subhedral to euhedral poikilitic grains (Figure 3C). The intergrowths (Figure 3D) occur predominantly along cleavages and edges and in the interiors of the orthopyroxene, clinopyroxene and, to a lesser extent, olivine grains. Opaque mineral and green-spinel grains are associated with these intergrowths. In contrast, the rims, which are usually only partial, occur around olivine, orthopyroxene, clinopyroxene and the opaque minerals. The oikocrysts, 8 to 10 mm in diameter, enclose olivine, orthopyroxene, clinopyroxene, and plagioclase; commonly they contain intergrowths of red-brown biotite flakes.

Rarely, a light-green amphibole is intergrown with the brown amphibole. They are optically continuous but are distinguishable by color and texture. In places the light-green amphibole apparently has replaced all the ferromagnesian minerals, including the red-brown amphibole. In these places the green amphibole occurs as rosettes and swirl patches. This widespread apparent replacement is accompanied by sericitization of the plagioclase and the development of calcite and probably is related to deuteric processes. A similar amphibole has been reported by Hermes (1970) in the metagabbros of the Mecklenburg complex to be common hornblende on the basis of tetrahedral Al vs Na+K.

No attempt was made to determine the composition of the brown amphibole. Its color, pleochroism, habit (in some cases), and occurrence would seem to indicate that it is primary, especially along the northwestern part of the pluton, where the brown amphibole becomes the dominant ferromagnesian mineral and occurs as large poikilitic grains and rarely as subhedral to euhedral phenocrysts (samples U125-32p, U125-19, U125-23). It is interpreted to have formed near the end

of the crystallization sequence in the rocks and to have resulted from crystal-liquid reactions; it changed later, in places, to a green amphibole. Tan to red-brown, fine-grained, anhedral biotite occurs intergrown with the brown amphibole and rarely with the orthopyroxene, clinopyroxene, and olivine grains. Locally, where brown amphibole is abundant, the flakes are poikilitic, enclosing the pyroxenes, and are associated with the opaque masses. The biotite, like the brown amphibole, is thought to be late magmatic in origin.

The mode of occurrence, color, and pleochroism of the brown amphibole in the mafic complex are similar to common hornblendes or aluminous calciferous amphiboles described from the Guadalupe complex (Best and Mercy, 1967). Their reddish-brown color suggests that they are similar to the common hornblende and paragasite-ferrohasting-site found in the Mecklenburg gabbro (Hermes, 1970). The content of brown amphibole ranges from 6.3 to 41.5 percent in the norites and from 0.3 to 24.4 percent in the gabbros.

A third-order trend-surface analysis shows an east-west trend with a maximum concentration of amphibole in the north-central to west-central portion of the pluton. The amphibole gradually decreases in amount to the north and south of this band.

Comparing amphibole distribution with the other parameters shows the following: modal amphibole shows a poor, generally positive correlation with modal orthopyroxene and clinopyroxene, and a negative correlation with modal plagioclase and olivine and plagioclase composition.

Plagioclase

Gray, generally fresh plagioclase occurs mainly as fine- to medium-grained anhedral to subhedral laths and equant grains, or less commonly as interstitial intercumulus material; most of these grains contain minute opaque and red-brown inclusions which impart a turbid appearance to the mineral. Normal zoning occurs, especially along the outer 0.1 mm of the grains. Locally, sericite and calcite patches are abundant in the plagioclase, adjacent to veinlets, and near green amphibole resulting from the alteration of ferromagnesian minerals. Whereas the larger plagioclase grains may enclose olivine and the pyroxenes, the smaller ones occur interstitially within, and partially enclosed by, the ferromagnesian minerals. Rarely, the plagioclase grains tend to be curved, fractured, or granulated and to show undulose extinction.

The plagioclase composition is variable, ranging from An₆₇ to An₈₀ in the norites and from An₆₀ to An₈₆ in the gabbros, but the compositional variations are not systematic within the pluton. Plagioclase constitutes from 10.4 to 52.6 percent of the norites and from 9.7 to 67.2 percent of the gabbros.

The plagioclase increases in amount from the southwest to northeast, but its composition does not reflect this variation. Most of the

plagioclase is lath-like except in those rocks where it is least abundant. In these it occurs mostly as irregular platelets of probable intercumulus origin. The latter is prevalent in the coarser-grained rocks. The plagioclase in the finer-grained rocks located in the southwest portion of the complex occurs as laths, which in thin-section appear to have a preferred planar orientation, possibly due to a cumulus origin.

Trend-surface analysis of the modal plagioclase indicates lower concentrations in the north-central part of the complex and a higher amount in and along the south and southwestern margins. A comparison with other mineral concentrations and compositions yields:

- 1) low modal plagioclase corresponds to high orthopyroxene and clinopyroxene, and vice-versa;
- 2) modal plagioclase shows no correlation with modal olivine and plagioclase compositions.

Accessory Minerals

Fine-grained opaque grains are variable in shape and association. They are associated with the replacement intergrowths of amphibole and olivine, clinopyroxene and orthopyroxene; they occur as dust-size specks along olivine cleavages and as large blebs within olivine. Commonly opaque grains are intergrown with green spinel (hercynite?), but also occur as partial rims around olivine-orthopyroxene coronas, intergrown with fibrous green amphibole. Euhedral grains are interstitial to plagioclase, and in places brown amphibole, and to a lesser extent green amphibole, forms rims around the opaque grains, all suggesting late stage or subsolidus origins.

Calcite grains are rare and occur interstitially to and within plagioclase grains. Trace amounts of apatite and pyrite are present.

Mineral Paragenesis

From textural criteria, the paragenesis in the finer-grained mafic rocks in the southwest portion of the complex is olivine and plagioclase, the plagioclase was overlapped by clinopyroxene. In the medium- to coarse-grained rocks orthopyroxene overlapped olivine and clinopyroxene with local interstitial plagioclase. Brown amphibole is later than all of the above, but was followed by biotite; these were the last two minerals to crystallize in both the fine- and coarse-grained rocks. Their crystallization was followed by corona development and the synchronous or subsequent development of the transecting veinlets. There is little evidence that large amounts of opaque minerals were ever precipitated directly from the melt.

PETROCHEMISTRY

Ideally the chemistry of a rock, or a suite of rocks, should reflect the identity and concentration of those minerals composing it. In most rocks no more than six minerals (commonly four or less) control the major-oxide concentrations. In this study spectrochemical analyses of approximately 40 samples were made in an effort to:

- 1) correlate the mineralogy of the rocks with the major and minor oxides,
- 2) classify the rocks in one of three basaltic rock series (Kuno, 1959), and
- 3) obtain information as to the crystallization trend of the rocks.

The chemical analyses of the rocks and such parameters as the felsic, mafic and solidification indices and bulk density are listed in Table 1. Trend surfaces, isochemical diagrams, and AMF triangular diagrams are used to correlate the chemistry of the rocks with their modal concentrations.

In 1921 Holmes classified igneous rocks into an alkaline and a calc-alkaline series. Peacock (1931) proposed a four-fold classification on the basis of the alkali-lime index. Kuno (1959) introduced a new usage of the term alkali-lime index, proposing that it be derived from the solidification index (SI), which is the ratio: $MgO \times 100 / (MgO + FeO + Fe_2O_3 + Na_2O + K_2O)$. This ratio invariably decreases as fractionation proceeds and expresses numerically what the ternary diagram $MgO - (FeO + Fe_2O_3) - (Na_2O + K_2O)$ expresses graphically, because on such a diagram, as fractionation in a magma occurs, the composition of the liquid moves away from the MgO vertex.

The mafic index (MI) (Simpson, 1954) is an expression of relative iron enrichment in the derivative melts as a result of crystallization of the mafic minerals. The felsic index (FI) (Simpson, 1954) reflects the progressive enrichment of the feldspars in alkalies as crystallization proceeds.

Plots of the mafic versus felsic indices for the Buffalo complex are too clustered to provide a conclusive trend. The low value and small range of the mafic (35-45) and felsic (7-19) indices indicate that little fractional crystallization occurred in the Buffalo rocks and substantiates the high MgO and CaO, and low alkali, values for these rocks. The lack of marked differentiation is shown also in Figure 5, a triangular plot of the $MgO - (FeO + Fe_2O_3) - (Na_2O + K_2O)$, and for the alkali-lime index, which shows a scatter of CaO and $Na_2O + K_2O$ when plotted against SI (for values see Table 1). However, the values obtained indicate that the rocks have a tholeiitic to calc-alkaline affinity. Figure 6, a plot used by Kuno (1959), shows that the rocks have a tholeiitic affinity.

According to Kuno (1959) the tholeiitic, alkaline and calc-alkaline rock series are defined mineralogically by the mafic minerals and the reactions between them. He recognized two sequences (in order of appearance) in the tholeiitic intrusive rocks: Mg-olivine -- bronzite-

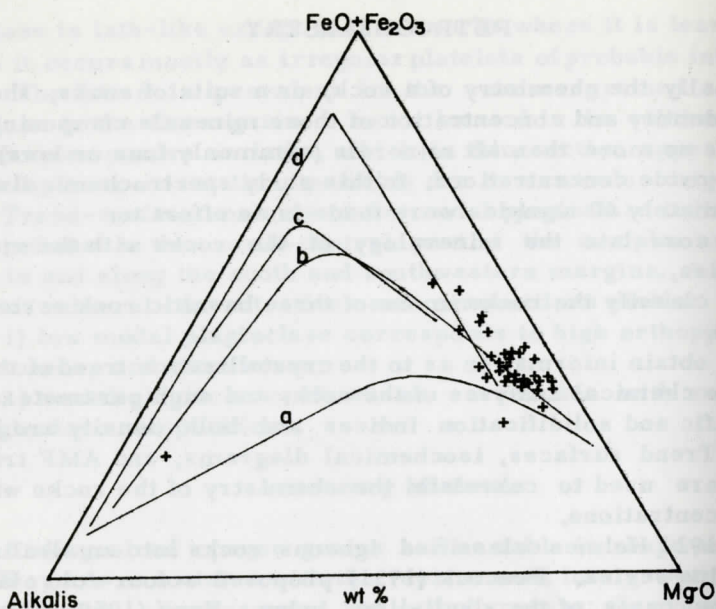


Figure 5. Plot of total alkalis ($\text{Na}_2\text{O}+\text{K}_2\text{O}$) : MgO : total iron ($\text{FeO}+\text{Fe}_2\text{O}_3$) for Buffalo (plus signs) rocks. Anomalous value near alkalis' apex is from Buffalo biotite-quartz monzonite. For comparison Daly's calc-alkaline (a), Karroo (b), Dillsburg (c) and Skaergaard (d) rock trends have been included (after Hess, 1960, plate 11).

hypersthene -- pigeonite -- Fe-olivine, and diopside -- augite -- ferro-augite. The pigeonite commonly inverts to hypersthene, the Fe-olivine may not appear, and ferro-pigeonite may be the only mineral to crystallize in the latest stages. He believes that the absence of OH-bearing minerals, except in the latest stages, is an outstanding feature of this series. According to Best and Mercy (1967) the pyroxenes from tholeiitic magma types consist of a calcium-rich clinopyroxene and a calcium-poor orthopyroxene, both exhibiting exsolution lamellae, crystallizing together in the early and middle stages of differentiation. A monoclinic pigeonite crystallizes in the middle stages but may subsequently invert to orthopyroxene.

The volcanic alkaline rock series is characterized by Mg-olivine and augite (locally titaniferous), crystallizing together (Kuno, 1959), and the complete absence of hypersthene and pigeonite. According to Best and Mercy (1967) only one pyroxene crystallizes from alkali basaltic magmas, a very calcic clinopyroxene with little or no iron

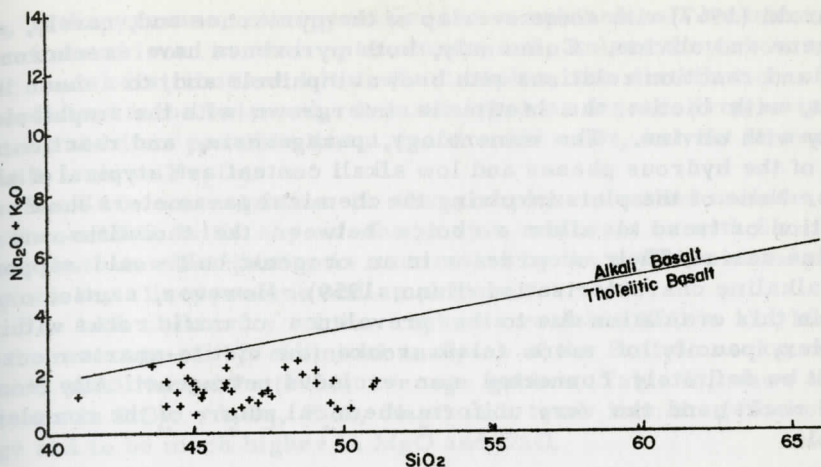


Figure 6. Plot of $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 (weight percent) for the Buffalo mafic rocks. The line separating alkaline and tholeiitic basalts is taken from Kuno (1959).

enrichment and lacking exsolution lamellae.

The volcanic equivalent of the calc-alkaline series, the hypersthene rock series (Kuno, 1959), has the following evolution of the mafic minerals:

PHENOCRYSTS: Mg-olivine -- hypersthene -- hornblende --
biotite augite

GROUNDMASS: hypersthene -- ferrohypersthene
augite --

In the hypersthene rock series the hypersthene has exsolution lamellae of augite parallel to (100). Inverted pigeonite is absent; however, interprecipitate hornblende is common. According to Yamazaki (1967), in the intrusive rocks of this series augite has no reaction with hornblende but reacts with the liquid to form hypersthene, which in turn may react with the melt to form hornblende. The latter may react with the liquid to form biotite. However, Yamazaki found the reaction relations among the mafic minerals of the Koyama calc-alkaline complex to be different from the above; he attributed these differences to different water contents in the magmas. Best and Mercy (1967) found that the rocks of the calc-alkaline Guadalupe igneous complex consist of co-existing clinopyroxene and orthopyroxene, lack essential pigeonite, and have higher values of Ca than other pyroxenes from documented igneous trends such as the tholeiites.

The Buffalo ferromagnesian minerals, Mg-olivine, calcium-rich clinopyroxene and calcium-poor(?) orthopyroxene, along with brown amphibole and biotite, follow the calc-alkaline sequence outlined by

Yamazaki (1967) with some overlap of the pyroxenes and, rarely, orthopyroxene and olivine. Commonly, both pyroxenes have exsolution textures and reaction relations with brown amphibole and, to a much lesser extent, with biotite; the biotite is intergrown with the amphibole and rarely with olivine. The mineralogy, paragenesis, and reaction relations of the hydrous phases and low alkali content are atypical of alkalic rocks. None of the plots involving the chemical parameters show enough variation or trend to allow a choice between the tholeiitic and calc-alkaline series. Their occurrence in an orogenic belt would support a calc-alkaline characterization (Kuno, 1959). However, caution must be used in this evaluation due to the prevalence of mafic rocks within the complex, paucity of more felsic rocks (the biotite-quartz monzonite cannot be definitely connected nor excluded petrogenetically from the mafic rocks), and the very uniform chemical nature of the complex as a whole.

OXIDE DISTRIBUTION IN THE COMPLEX

The mineralogy of igneous rocks is determined, aside from temperature and pressure, by the chemical composition of the magma from which they crystallize. Thus the mineralogy of a suite of rocks should be reflected in the concentration of the major and minor oxides. These latter should also provide useful information about the crystallization trend of the rocks. To ascertain if this was correct, trend-surface analyses* and isochemical diagrams were computed for MgO, Al₂O₃, Fe₂O₃, and CaO and compared with variables that exist in the Buffalo rocks.

The MgO increases from the margins to a high in the north-central part of the complex. This correlates with SI values and high concentrations of orthopyroxene and clinopyroxene, and less so with amphibole and olivine. This is to be expected because the modal amounts of orthopyroxene plus clinopyroxene compose over 50 percent of the rock in places (Table 1), and they vary sympathetically.

The Al₂O₃ content in the rocks exhibits a strong correlation with the modal plagioclase. The amount of Al₂O₃ decreases from the southwest margin of the complex to a low in the north-central portion, which corresponds to high amounts of orthopyroxene, clinopyroxene and amphibole. This trend is expected because plagioclase, and to some extent amphibole, are the only major phases in the rocks that contain Al₂O₃.

The total iron as Fe₂O₃ does not correlate well with any of the major phases present in the rocks. This results from the fact that

*Copies of the trend-surfaces for MgO, Al₂O₃ and plagioclase can be obtained from J. H. Medlin upon request.

olivine, orthopyroxene, clinopyroxene and amphibole all contain variable amounts of iron in their structures; and, unless they show sympathetic modal distributions within the complex, the distribution of the iron would tend to be balanced. Iron analyses as ferrous iron were not computed, and it is possible that a trend could exist, but not be recognized with iron as Fe_2O_3 .

The CaO concentration in the complex is rather uniform and does not correlate with any of the major minerals or with Si. This suggests that its distribution is controlled by the relative abundances of plagioclase, clinopyroxene and amphibole.

The bulk chemistry of the rocks is variable (Table 1). Moreover, the average of 32 chemical analyses of the Buffalo rocks, when compared with Nockold's (1954) average gabbro (Table 4), show them to be lower in SiO_2 , Al_2O_3 , K_2O , Na_2O and total iron than Nockold's average and to be much higher in MgO and CaO.

PETROGENESIS OF THE BUFFALO COMPLEX

The petrogenesis and subsequent geologic history of any rock must be considered in terms of its compositional data, chemistry, mineralogy and textures of coexisting mineral phases. The data described in the previous sections seem to indicate the following petrogenesis.

The Buffalo mafic rocks were derived from a silicate magma that was relatively high in MgO and CaO, and low in SiO_2 , Al_2O_3 , K_2O and Na_2O . The presence of pyroxene hornfels, cross-cutting relationships and lack of deformational features indicate that the magma was emplaced and crystallized in situ. Thin-section studies indicate that olivine, plagioclase, and clinopyroxene were the first minerals to crystallize from the magma. These minerals are found in the fine-grained rocks now exposed along the southwestern margin of the complex. The fine-grained rock unit was not encountered along the northeast margin of the complex. This implies that: 1) it is simply not exposed; 2) it never existed; or 3) it has been removed by erosion. Progressing inward from the southwestern margin, the rocks are typified by a gradual increase in grain size and by the appearance of orthopyroxene as discrete grains. There is also a systematic variation in the proportion of plagioclase, clinopyroxene and orthopyroxene toward the major axis of the pluton with an inverse relationship between plagioclase and the orthopyroxene and clinopyroxene. All of these variations are characteristic of cumulates of layered mafic and ultramafic bodies described by Wager (1968). Brown amphibole and, to a lesser extent, biotite become abundant inward in the coarser-grained rocks. The mineral distribution and petrochemistry suggest a similar, but less obvious, variation inward from the northeastern margin to the major axis, possibly indicating large-scale chemical zoning within the pluton. Most

of the rocks contain brown amphibole, as intergrowths within the ferromagnesian minerals; but amphibole maximum concentration, as poikilitic grains, is mostly in the rocks exposed near the center of the complex and in those rocks along the northwestern margin. In all rocks it is a late phase followed by biotite in the paragenesis; possibly both are intercumulus minerals.

After the final consolidation of the magma and during the cooling stages, the coronas formed around olivine. This was followed by formation of veinlets that caused the slight hydration of the ferromagnesian minerals and plagioclase to green amphibole and sericite, respectively. Probably concomitant with this hydration was the alteration of the gabbro to the altered mafic rock, the emplacement of the fine-grained mafic dikes, and then the emplacement of mica-quartz-feldspar aplites and pegmatites. The latter caused much alteration of the mafic rock.

The lack of microscopic and mesoscopic deformational features in the rocks, as well as the relative paucity of metamorphic (and retrograde metamorphic) minerals, indicates that the pluton has undergone little, if any, deformation at moderate or elevated temperature conditions since its emplacement. These features along with the presence of hornfels suggest that the pluton and the hornfels have not been subjected to any intense metamorphism subsequent to emplacement.

The opaque minerals do not occur in large concentrations in the rocks studied in thin-section, nor do they occur as distinct bands in the field. This implies that they were never precipitated in large concentrations, or that field exposures are inadequate. Thin-section studies indicate that most of the opaque minerals resulted from the breakdown of olivine, clinopyroxene, orthopyroxene and amphibole.

The near absence of layering, at least in hand specimens and outcrops, suggest that: 1) exposures are inadequate for its recognition; or 2) it does not exist. However, the distribution of the minerals and the petrochemistry indicate that zoning does exist in the complex and is expressed inward from the northwestern and possibly northeastern margins. The textural and mineralogical (modal) variation can be explained readily by mineral accumulation during the cooling of the magma. The mineral compositions indicate essentially no differences between the minerals occurring along the margins and those occurring elsewhere in the complex. The latter suggests that little if any fractional crystallization occurred in the magma. The low mafic and felsic indices and the total alkalis: MgO : total iron plot support this conclusion.

CONDITIONS OF CRYSTALLIZATION OF THE BUFFALO COMPLEX

Rocks crystallize over a range of temperatures, with the actual beginning of crystallization, the paragenesis, and the end of crystallization depending upon the magma composition, pressure (both load and water) and oxygen fugacity. In recent years, experimental studies on

both synthetic and natural compositions have been carried out in an effort to evaluate the effect of the above variables on rock crystallization.

The role of oxygen fugacity in the crystallization trend of iron-bearing silicate systems has been extensively studied at one atmosphere pressure by Muan and Osborn (1956), Muan (1958), Osborn (1959, 1962), Speidel (1964), Fudali (1965), Roeder and Osborn (1966) and Presnall (1966). These studies have attempted to quantitatively define the "Bowen" and "Fenner" trends presented by Kennedy (1948) for simplified basalt systems. The "Bowen" trend, in which the residual melt is enriched in silica and the alkalis, is thought by Osborn (1959, 1962) to result from a fairly high or increasing oxygen fugacity. Presnall (1966) has shown that the silica enrichment can be produced in the late stages of fractional crystallization of a basaltic magma if the oxygen fugacity is buffered at a sufficiently high value that spinel has a large primary field or volume. However, Presnall points out that the composition, and not oxygen fugacity, is more important in the initial stages of crystallization. In reference to buffering, Osborn (1962) emphasized that the fractional crystallization of a basalt toward a high-silica residual was possible if the magma had a significant water content, allowing a continuous replenishment of oxygen (oxygen fugacity remains constant or increases slightly).

The "Fenner" trend is characterized by a high FeO content in the residual magma. Presnall (1966) attributes this trend to conditions in which the oxygen fugacity is buffered at a sufficiently low value that the primary field or volume of spinel is either very small or absent. This would allow the ferrous iron to remain in solution, rivaling Mg, Ca, Al, and others in forming the ferromagnesian silicates. This trend is thought by Osborn to be characteristic of the layered gabbro intrusions such as the Skaergaard. Osborn defined this trend as being due to decreasing oxygen fugacity, or total composition remaining constant during crystallization.

Most gabbroic rocks are thought to crystallize under oxygen fugacity conditions that exist between the quartz-magnetite-fayalite and hematite-magnetite buffer conditions (Hamilton *et al.*, 1964). Work on Hawaiian tholeiitic basalts by Sato and Wright (1966) and Fudali (1965), using different methods, arrived at a value of $10^{-8.3}$ atm for the oxygen fugacity at which those basalts crystallized.

The role of load and water pressures on the crystallization of basaltic magmas has recently been evaluated by Yoder and Tilley (1962), Hamilton *et al.* (1964), Cohen *et al.* (1967), Green and Ringwood (1967), Burnham (1967), and others. Results of load-pressure studies indicate that increased pressure raises the temperature of the appearance of a particular mineral in dry systems. The effect of high water pressure in these studies is to lower the temperature of appearance of minerals. Eugster (1957), Eugster and Wones (1962), Wones and Eugster (1964), Hamilton *et al.* (1964) and Tuthill (1968) stress the control of oxygen fugacities with buffers at high water pressures and show that the

presence or absence of a particular mineral, the paragenesis, and the melting temperature are affected by this variable.

According to Buddington (1959), the metamorphic environment in which a granitic pluton is found allows an estimate of the depth of intrusion and hence the load (total) pressure at which it crystallized. Although the Buffalo complex is not a granite pluton, the criteria proposed by Buddington are useful for a first approximation. The occurrence of pyroxene hornfels, lack of discernible deformation, and the cross-cutting relationship indicate that the magma was emplaced into metamorphic rocks assigned to the greenschist facies by Overstreet and Bell (1965a). This would place the pluton in the mesozone as defined by Buddington (1959), corresponding to depths of 4 to 10 miles (6.4 to 16.0 km) and a temperature of 250°C to 500°C, depending on the depth. This depth would correspond to a load pressure ranging from 2 to 5 kbs. Winkler (1965) suggests that the pyroxene hornfels begins at 630±20°C at 2 kbs load pressure and that a gabbroic magma intruded at 1200°C, at 5 to 6 kms, into rocks with an initial temperature of 150°C, will produce a contact temperature of 875°C. If Winkler's data are extrapolated to 11 kms, the contact temperature would be 1000°C and the pressure 3 kbs, an intermediate value for mesozonal plutons. The contact temperature is derived by assuming an initial value for the surrounding rocks and the magma, thus this temperature would depend largely on the crustal depth, the time relationship between plutonic intrusion and regional metamorphism and water pressure and magma temperature. If the intrusion is synchronous with the regional greenschist metamorphism, the contact temperature would be significantly higher. Winkler (1965) places the temperature span of the greenschist facies from 400°C to 550°C. If the mesozonal assignment is correct, the load pressure was not much greater than 4 kbs.

Observations on basaltic lavas indicate that their liquidus temperatures are generally less than 1200°C. Experimental data suggest that the range is from 1200°C to 1000°C and that usually the first phase crystallizes at about 1100°C. However, the temperature and appearance of the first phase depend on the composition, pressure (load and water) and oxygen fugacity (Cohen, 1967, Hamilton *et al.*, 1964, Yoder and Tilley, 1962). If the exsolution lamellae in the Buffalo pyroxenes can be used as a geothermometer, then it would suggest that these minerals were precipitated from the magma at a temperature below 1140°C. This suggests that the olivine, plagioclase and clinopyroxene began crystallizing at or near 1140°C and that the orthopyroxene became stable at a lower temperature. However, this temperature is an upper limit and will be lowered when the composition, pressure and oxygen fugacity variables are considered. Yoder and Tilley (1962) found that at approximately 950°C and 5 kbs pressure ($P_{load}=P_{water}$) amphibole appears as a stable phase in tholeiites and olivine tholeiites from Hawaii. Tuthill (1968) found that amphibole was abundant in the Hawaiian basalts at 1000°C under quartz-fayalite-magnetite buffer conditions at 5 kbs

pressure ($P_{load}=P_{water}$). This pressure is within the limits proposed for mesozonal intrusions, and suggests that the amphibole in the Buffalo rocks became stable above 950°C followed by crystallization of biotite at lower temperatures. According to Yoder and Tilley (1962), the Hawaiian tholeiites, at $P_{water}=5$ kbs, crystallized over an interval of 1100°C to 775°C . Tuthill's data, using controlled oxygen fugacity conditions, indicates that the lower temperatures should be $695^{\circ}\text{C}\pm 10^{\circ}$.

If the above approximations and assumptions are correct, then the Buffalo rocks probably crystallized within the 1100°C to 695°C range. Supporting evidence of the lower temperature limit is found in O'Hara (1968, Figure 14). He depicts the assemblage forsterite, diopside, enstatite, anorthite and amphibole as being stable within the temperature and pressure range from 740°C to 890°C and 2 kb to 7 kb, respectively; a good approximation for the Buffalo assemblage.

The estimates reached above probably will be affected by the higher MgO and CaO in the Buffalo rocks; however, the oxygen fugacity and pressures might offset the compositional differences. This is likely because Yoder and Tilley (1962) neglected the control of oxygen fugacity, and Hamilton *et al.* (1964) and Tuthill (1968) found that its control lowered the temperature and changed the paragenesis.

COMPARISON WITH OTHER COMPLEXES

The Buffalo complex can be compared, to a limited extent, with the Mecklenburg complex (Hermes, 1966, 1968, 1969, 1970), Gladesville norite (Matthews, 1967, 1969), Concord complex (Legrand and Murdorff, 1954; Bell, 1960; Butler and Ragland, 1969; Cabaup, 1969) and Mt. Carmel complex (Medlin, 1968). The Buffalo, Mecklenburg gabbro and Gladesville complexes are similar in mineralogy and texture. All contain olivine, orthopyroxene, clinopyroxene, brown amphibole, biotite, and plagioclase, and exhibit coronas and textures indicating well-developed reaction relations between the mafic phases. The petrology (gabbro/gabbro-norite), mineralogy, and petrochemistry indicate little magma differentiation in these intrusions. A plot of coexisting clinopyroxene, orthopyroxene and olivine from these complexes (Figure 4) shows little compositional differences within individual complexes. As a group, the compositions of the olivine and orthopyroxene from these intrusions do show a range. However, the clinopyroxenes exhibit little range in composition.

A comparison of the Buffalo chemical analyses with those reported by Butler and Ragland (1969) for the Buffalo complex (Table 4) show significant differences in SiO_2 , TiO_2 , Al_2O_3 , CaO and K_2O . These differences probably result from the chemical and mineralogical zoning in the complex. Even larger differences in chemistry are observed between the Mecklenburg gabbro and Buffalo complexes (Table 4). Notable differences include TiO_2 , Al_2O_3 , iron, MgO, CaO, Na_2O , and

Table 4. Comparisons of Average Chemical Compositions.

No.	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O
1. Average gabbro (142 analyses), Nockold's, 1954	48.36	1.32	16.84	2.55	7.97	.18	8.06	11.07	2.26	.56
2. Buffalo complex (32 analyses)	46.30	.61	15.30	2.33	5.92	.13	12.42	13.52	1.41	.21
3. Buffalo complex (9 analyses), Butler and Ragland, 1969	47.80	1.14	16.80	9.64		.14	12.33	12.54	1.52	.16
4. Contact zone, Mecklenburg gabbro complex (4 analyses), Butler and Ragland, 1969	48.60	1.31	17.30	11.62		.16	8.71	9.95	2.04	.47
5. Mt. Carmel gabbro (13 analyses), Medlin <i>et al.</i> , in preparation	43.75	3.71	18.42	11.74		.17	5.25	11.25	2.92	.35
6. Concord gabbro pluton (5 analyses), Butler and Ragland, 1969	46.90	1.29	17.50	11.00		.13	9.15	10.55	2.50	.39
7. Border Facies, Stillwater complex, Hess, 1960, Table 47	50.68	.45	17.64	.26	9.88	.15	7.67	10.47	1.87	.24
8. Border facies, Bushveld complex, Hess, 1960, Table 47	51.45	.34	18.67	.28	9.04	.47	6.84	10.95	1.58	.14
9. Border facies, Skaergaard complex, Hess, 1960, Table 47	47.92	1.40	18.87	1.18	8.65	.11	7.82	10.46	2.44	1.35

K₂O. These can be attributed to original compositional differences as well as different geologic histories.

A comparison of the Buffalo gabbro-norite chemical analyses with the Mt. Carmel and Concord gabbros (Table 4) indicate little similarity. The reason for these differences become apparent when the petrology and mineralogy of the complexes are considered. The Mt. Carmel complex is composed of gabbro, diorite and syenite; the Concord complex of hornblende gabbro and augite syenite (Legrand and Murdorff, 1954). The Buffalo rocks contain olivine, orthopyroxene, clinopyroxene, amphibole, biotite and plagioclase. The Mt. Carmel and Concord gabbros contain olivine, titanite, brown and green amphibole and plagioclase; the Mt. Carmel gabbro contains up to 10 percent opaque minerals. The Mt. Carmel diorite, in addition to feldspar, is composed mainly of ferroaugite, ferropargasite, amphibole and biotite. Mafic phases of the Mt. Carmel syenite include ferroaugite, amphibole and biotite; the Concord syenite contains a clinopyroxene. Work by Medlin (1968) indicates that the Buffalo complex has a tholeiitic or calc-alkaline affinity; the Mt. Carmel an alkaline affinity. Probably the Concord has alkaline affinities.

Few radiometric age determinations are available on these complexes. However, Fullagar (1970) reports a Rb-Sr age of about 413 m. y. for the Concord syenite. Medlin (1968, 1969) lists three K-Ar

biotite ages from the Mt. Carmel diorite: 380, 386 and 387 m.y. If these ages accurately date the crystallization of these intrusions, then they are at least Devonian in age.

Table 4 lists chemical analyses of the Stillwater, Bushveld and Skaergaard border facies rocks, considered representative of the parent magmas of these intrusions (Hess, 1960, Table 7), along with chemical analyses of the Buffalo complex. The most obvious differences between the Buffalo rocks and the others is the lower SiO_2 , Al_2O_3 , and Na_2O and higher CaO and MgO concentrations of the Buffalo rocks.

The olivine and orthopyroxene from the ultramafic zone of the Stillwater complex (Hess, 1960) are more magnesian than those from the Buffalo rocks; however, those present in the Stillwater anorthosite zone are comparable to those in the Buffalo complex. Coexisting clinopyroxene and orthopyroxene from the Buffalo complex plot close to those initially crystallized and coexisting from the Bushveld, Skaergaard and Stillwater complexes. The amount of chemical variation of coexisting olivine-orthopyroxene-clinopyroxene is small in the Buffalo rocks and shows no discernable differentiation trend when compared to the Skaergaard trend for these minerals.

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GIBBSITE IN WESTERN NORTH CAROLINA

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ABSTRACT

Significant concentrations of gibbsite have been found in certain soils and saprolites of the Southern Appalachians and foothills in Western North and South Carolina. The gibbsite apparently has formed by direct weathering in the present environment. High rainfall and rapid removal of silica and bases from the weathering zone are suggested as key factors. Gibbsite-containing regoliths average less than eight feet in thickness.

INTRODUCTION

Gibbsite, principal mineral component of most aluminum ores or "bauxite", is a significant component of certain soils and saprolites of the Blue Ridge, intermountain valleys and foothills of Western North Carolina and South Carolina. In this report, we describe our findings of significant occurrence of gibbsite at certain locations studied in some detail, summarize gibbsite findings at other locations in this region which we and our associates have reported elsewhere as parts of other soil studies (McCracken *et al.*, 1971, Losche, *et al.*, 1970), and present hypotheses to attempt to explain these gibbsite accumulations.

Gibbsite accumulations in soils have been considered indicative of, or associated with, advanced stages of weathering (Jackson and Sherman, 1953), chiefly in tropical soils. Summaries of the geological literature on the origin and occurrences of gibbsite can be found in Harder (1949, 1952) and Gordon *et al.* (1958).

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Theoretically, a soil profile or saprolite may become enriched in gibbsite in three main ways:

(1) Primary minerals in the weathering rock may alter directly to gibbsite ("direct weathering").

(2) The minerals may first weather to silicate clays, which in turn are desilicated in additional weathering over time to produce gibbsite.

(3) The gibbsite may accumulate by translocation from above, either physically in suspension in downward percolating water or chemically in solution followed by crystallization.

The first mechanism, direct gibbsitization, has been described as a significant mechanism for origin of bauxite deposits (Gordon *et al.*, 1958) and as a source of gibbsite in saprolite and soils, followed by re-silication (Alexander *et al.*, 1941; Cady, 1951).

The second mechanism of gibbsite formation has been described from a geological point of view by Allen (1952) and as a stepwise desilication process acting on soil clays over relatively long periods of time by Jackson and Sherman (1953). There is some agreement that both processes may produce significant concentrations of gibbsite. The third mechanism has in general not been considered important, though a common secondary phenomenon. It appears that some Russian geologists may consider this translocation-precipitation mechanism as an important one in the formation of certain bauxites (Zans, 1959).

MATERIALS AND METHODS

Two sites were selected for detailed sampling of the soil profile and laboratory analyses. Both are on the eastern slope foothills of the Blue Ridge in North Carolina. One site was in Caldwell County, approximately eight miles east of Lenoir, North Carolina. The second site was located in Rutherford County, west of Rutherfordton, on Bill's Creek Road north of Highway 74.

A brief, condensed description of each soil profile follows:

Site I, Caldwell Co.

Soil horizon	Depth, In.	
A1	0-2.5	Dark grayish brown fine sandy loam
A2	2.5-7	Strong brown fine sandy loam
B2	20-48	Red clay
B3	48-62	Red clay with yellow and gray spots
C	62-90	Red, yellow, gray friable clay loam or loam, saprolite; weathered mica-hornblende gneiss.

This is about the midpoint of a long, gentle concave slope.

Site II, Rutherford Co.

Soil horizon	Depth, In.	
A	0-4	Strong brown loam
B	4-28	Red clay loam
C1	28-38	Yellowish red, red and gray loam saprolite; weathered mica gneiss.

This site is on the upper part of a convex hillslope, on the Blue Ridge scarp.

Samples were collected by subhorizons and analyzed according to standard methods of soil characterization (Black, 1965). Mineralogical methods were those described by Kunze and Rich (1959). Gibbsite percentages were determined by differential thermal analyses (DTA) for both clay ($<2\mu$ diameter) and soil <2 mm. Based on determinations of gibbsite in standard reference samples with our DTA apparatus, we estimate that we could measure percentages of this mineral in the soil and saprolite samples with an accuracy of ± 2 percent, over the range with which we were working. X-ray diffraction was used to estimate percentages of other clay minerals. Thin sections of the soils were prepared to study gibbsite occurrence in situ with a polarizing microscope.

RESULTS AND DISCUSSION

Gibbsite percentages in the clay fraction (particles $<2\mu$ dia.) and the "whole soil" (<2 mm dia.) of selected layers or horizons are presented in Table 1. Also included in this table are data on percentages of clay and "free" iron (iron not combined with silicates, occurring as hydrous oxides).

In thin sections, gibbsite pseudomorphs after feldspar and other minerals were observed in the B3 and C soil horizons. Gibbsite crystals were observed lining some voids in B3 horizon thin sections.

We conclude that the gibbsite in the soil and saprolite of these two sites formed chiefly as a primary weathering product, by direct gibbsitization, in the present weathering environment. We base this conclusion on these lines of evidence:

- (1) The thin section observation of gibbsite pseudomorphs after primary feldspar minerals;
- (2) Our X-ray and DTA patterns indicate that particle size and/or crystallinity of the gibbsite decreases toward the soil surface, and more gibbsite occurs in the nonclay fractions in the lower B and C horizons (saprolite) than in the upper fractions;
- (3) There is a decrease in total gibbsite in the surface soil

Table 1. Gibbsite, Clay and Free Iron Percentages of Selected Soil Horizons.

Soil Subhorizon	Depth Inches	Percent Gibbsite		Clay %	Free Fe % (whole soil)
		Whole Soil	Clay		
Site I, Caldwell County					
A2	2.5-7	4	35	9	0.76
B12	12-20	12	42	30	1.4
B22	26-36	32	65	62	7.4
B32	54-62	40	86	54	8.0
C1	62-72	34	70	27	7.0
C2	72-82	38	nd	35	6.4
C3	82-90	30	nd	9	4.2
Site II, Rutherford County					
A2	1-4	10	35	19	1.1
B2	9-18	16	30	50	4.4
B3	18-28	14	24	52	3.8
C	28-38	16	28	27	2.1

horizons, rather than an increase;

(4) Significant amounts of gibbsite were found in both the soil and saprolite of Site II, a site on an apparently young and retreating backslope. We do not think this gibbsite could have persisted as relicts from a previous weathering period on an older landscape (now destroyed by erosion) owing to the geomorphology at this site.

(5) Gibbsite has been found in significant amounts in similar sites in this region, but west of the Blue Ridge scarp (Losche *et al.*, 1970; McCracken *et al.*, 1971), suggesting a regional pattern of occurrence.

With respect to this latter point, we summarize as follows:

(a) Losche *et al.* (1970) found 34-54 percent gibbsite in the coarse clay of soils on south-facing slopes in a steeply sloping landscape in Transylvania County, N. C.

(b) McCracken *et al.* found maxima of 26 and 67 percent gibbsite in the clay fractions of B horizons of each of two Hayesville soils in Oconee County, S. C. and maxima of 16 and 36 percent gibbsite in each of two Hayesville soils in Clay County, N. C. (Both of these areas are in steeply sloping landscapes immediately west of the Blue Ridge scarp in the Southern Appalachians).

(6) We believe our data correlate with the measurements and observations of LeGrand (1958) that silica content of ground water (associated with rocks of similar composition) decreases with increasing

steepness and decreasing length of slope. The areas of gibbsite occurrence are mainly on relatively short and steep slopes, with shallow regolith, which indicates limited opportunity for resilication of gibbsite formed by direct weathering. The rapid rate of water percolation on these slopes, plus occurrence of a weathering front in the rocks near the surface on these landforms, means rapid rate of removal of silica and of bases thereby driving the weathering reaction directly to gibbsite formation, rather than solely to aluminosilicate clays (McCracken et al., 1971).

The soil A and B horizons plus the underlying saprolite are thin at most locations in this region where gibbsite concentrations have been found to be high. The total thickness of the weathered materials is only a few feet rather than tens of feet. The solum at site I in Caldwell County is thicker than in most soils with significant gibbsite in this area, probably because it occupies a slightly concave landform, lessening rate of erosion and causing some sideslope colluvium accumulation.) As a generalization, the gibbsite-containing regoliths are ordinarily less than eight feet thick in this area.

Most soils of this region that contain gibbsite are classed in the Hayesville soil series of the Hapludult great soil group in the new Comprehensive Soil Classification System (McCracken et al., 1971). However, the soil at site I would be classed as Cecil due to its thick solum, high clay content, and location in the lower foothills with warmer temperatures (though its gibbsite mineralogy would cause it to be designated as a "variant" or "taxadjunct" of that series).

SUMMARY

1. Certain soils and saprolites in the Blue Ridge portion of the Southern Appalachians and its eastern foothills, contain significant amounts of gibbsite - as much as 40 percent of the whole soil (particles less than 2mm diameter) and 86 percent of the clay (< 2 μ diameter).

2. This gibbsite appears to have formed by direct weathering of primary minerals, probably in the present environment.

3. The relative amounts of gibbsite in such soils appear to be determined at the time the parent rock is first weathered, possibly with some later resilication in the upper part of the soil solum.

4. Determining factors for the gibbsite formation appear to be presence of appropriate precursor primary minerals in the parent rock, steep and short slopes with rapid removal of silica and bases from the "weathering front" in the rock, presence of the weathering rock close to the surface, relatively high rainfall and general absence of resilication factors.

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A PRELIMINARY STUDY OF THE DISTRIBUTION AND MORPHOLOGY OF THE CAMINADA/MOREAU SAND RIDGES

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ABSTRACT

The Caminada/Moreau sand ridges of south Louisiana are types of accretion ridges deposited near a former outlet of the Mississippi River. The area of ridges, intervening swale lakes and marsh may be divided into four systems. The analysis of the form, orientation and sedimentary characteristics indicates a genetic difference between the oldest group of ridges and the remainder. The former are distributary mouth accretion ridges; the latter beach ridges. Although the age of the ridge systems, in relation to the chronological sequence of Mississippi lobes and sub-deltas, is open to discussion, the field evidence would suggest a relatively recent origin. The effect of the ridge topography on the drainage and lake patterns of the Caminada/Moreau deltaic plain is also discussed.

INTRODUCTION

The Caminada sand ridges occupy an area approximately nine miles long and a maximum of three miles wide between Bayou Lafourche and Caminada Pass in south Louisiana. The ridges have been specifically mapped by Fisk (1955) and described as part of a delta-front sheet sand body, the upper surface of which carries a series of spit-like sand ridges which have now sunk as a result of regional subsidence to sea level. Morgan (1970) accepts the term "delta front sheet sands" and considers these accretion ridges at Caminada to represent the reworking by wave and current action of the coarse materials being brought down by active distributaries.

Although described locally as cheniers, it is clear that by comparison with definitions and descriptions given by Price (1954), Russell and Howe (1935) and Byrne, LeRoy and Riley (1959) the term is inappropriate. The smaller size, greater ridge density, variations in altitude in both seawards and landwards directions and according to Fisk (1955, p. 389) an absence of marsh and bay silts at shallow depths beneath the ridge sequence, preclude the use of the genetic term chenier to these ridges.

Methods of Study

Using 1956 Coast and Geologic Survey aerial photographs and recent 1970 and 1971 NASA remote sensing images (black and white infra-red, color infra-red and thermal day and night infra-red) the precise distribution of the ridges east of Bayou Moreau has been mapped. The area was field checked and sampled, and the results of these investigations indicate that on the basis of distribution and morphology, certain hypothesis may be advanced for the formation of the ridges. Some comments may also be made on the role played by these ridges in shaping the adjacent marsh topography.

Acknowledgments

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Particular thanks are rendered to J. M. Coleman of Coastal Studies Institute who provided the impetus for the project and revised the original manuscript.

DISTRIBUTION OF RIDGES

The basic distribution of ridges east of Bayou Moreau are shown on Figure 1. Additional ridges are also found west of Bayou Moreau and were mapped by Fisk (1955). The latter ridges are much less conspicuous and cannot be interpreted with reliability from existing aerial photographs. Greater subsidence and greater disruption of the natural environment by pipeline canals, as well as the destruction of the natural drainage system, renders meaningful deductions impossible. The fact that some ridges do extend westwards of Bayou Moreau as far as the delta lake depression of east Timbalier Bay, is nevertheless important. An inset on Figure 1 (Figure 1a) show the distribution of sand ridges west of Bayou Moreau as deduced from Fisk (1955) and photo-interpretation. The seaward ridges of this system is, however, quite clear and shows a landwards curvature with three distinctive hooks.

Similarly, ridges of comparable size, separation and morphology are found to the east on Grand Isle. The sand size is similar and the general northeast to southwest orientation and presence of recurved

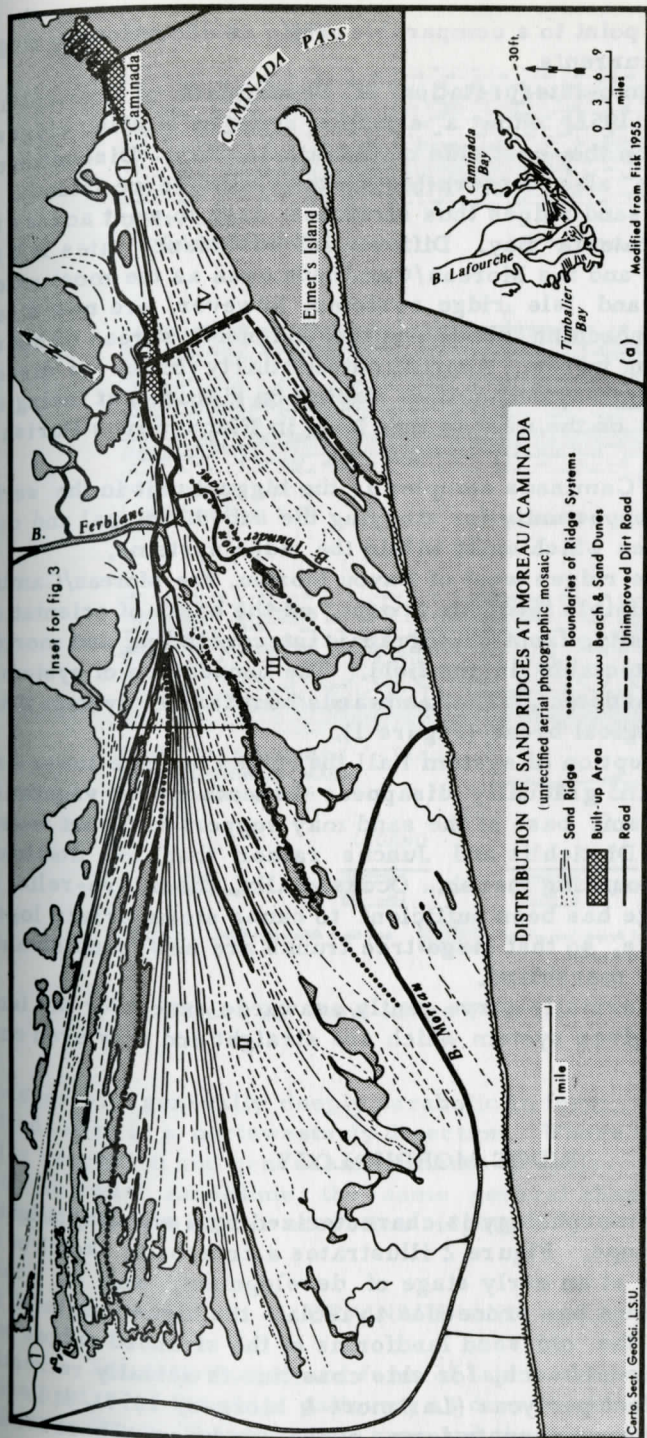


Figure 1. Distribution of sand ridges at Moreau/Caminada. (Unrectified aerial photographic mosaic.)

and spit-like ridges point to a comparable origin as accretionary ridges built-up by littoral currents.

Additional photo-interpretation of Department of Agriculture aerial photographs (1952) show a smaller area of similar ridges at Cheniere Ronquille on the east side of Barataria Bay. Discontinuous ridge fragments are also discernible on the Grand Terre Islands. A broken chain of low sand ridges thus stretches east to west across the entire width of Barataria Bay. Differential subsidence rates left the ridges of Grand Isle and the Moreau/Caminada area as the most extensive areas. The Grand Isle ridge systems, however, are now disappearing beneath encroaching dunes and the extensive spread of houses and other engineering works. Approximately thirty miles of discontinuous, broken ridge systems thus stretch on a southeast facing arc from Timbalier Bay on the west to Bay Long in Plaquemines Parish on the east.

The Moreau/Caminada complex is the biggest unit in the series and offers the best opportunity for studying the sub-divisional and morphological differences which exist within the ridge systems.

Excepting the ridges west of Bayou Moreau, the Moreau/Caminada ridge system is divisible into four groups, on the basis of orientation, distribution and density, from photographic interpretation, and morphological changes detectable in the field. The boundaries of system III are most difficult to detect. The landwards boundary of system II is a prominent morphological break (Figure 1).

With the exception of system I all the ridges become lower westwards, flatten-out and gradually disappear beneath marsh vegetation. The firmer, drier soil base of the sand may support different vegetation with species of Distichlis and Juncus rather than the dominant Spartina of the surrounding marsh. Occasionally, the micro-relief and firmness of the ridge has been sufficient to cause storm driven logs to ground on its surface, so that large tree trunks are now found stranded in the midst of level marshland.

The ridges normally curve gently seawards except for the innermost ridges of any given system which are straight and may even curve slightly landwards.

RIDGE MORPHOLOGY

The surface morphology is characterized by a smooth, slightly steeper seawards slope. Figure 2 illustrates a composite sketch of the average morphology at an early stage of development, and at a later stage when subsidence has proceeded to isolate the higher parts of the ridge. Comparing the old sand landforms of the stranded ridges with the encroaching present beach, for this coastline is actually retreating at an estimated 62 feet per year (Larimore & Morgan, 1957), it is possible to pick out the low, cusped forms of the low dunes. These aeolian

SAND RIDGE FORMS

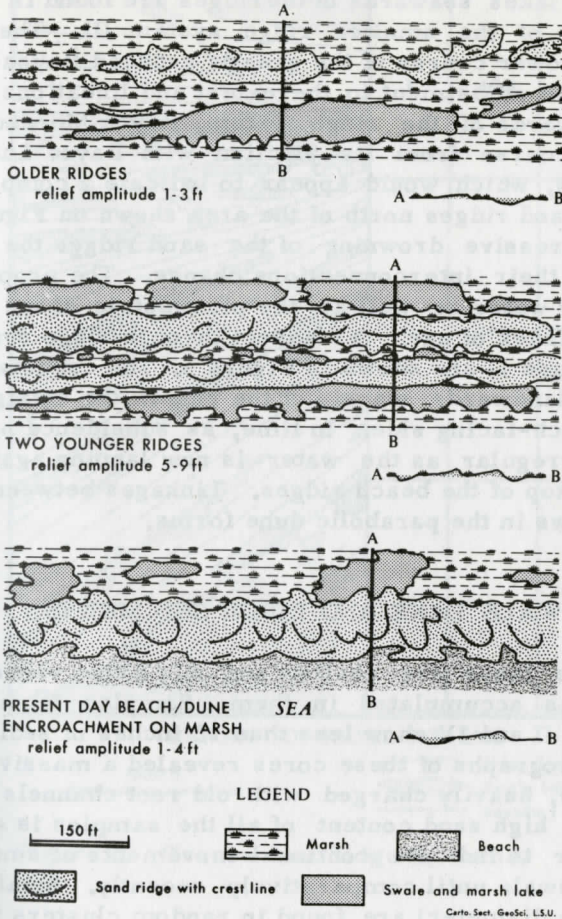


Figure 2. Sand Ridge Forms.

ridges are generally deeply parabolic in form, with the horns pointing landwards in a northwesterly direction. Where it is possible to do so, for vegetation and general movements of surface materials have reduced the relief amplitude, the same general shapes and trends are discernible on the higher accretion ridges.

DRAINAGE AND LAKES

The distribution and form of lakes and drainage channels reflect the control of the ridge patterns, as does the inner coastline at Elmer's Island. Bayou Thunder von Tranc and the lower part of Bayou Ferblanc

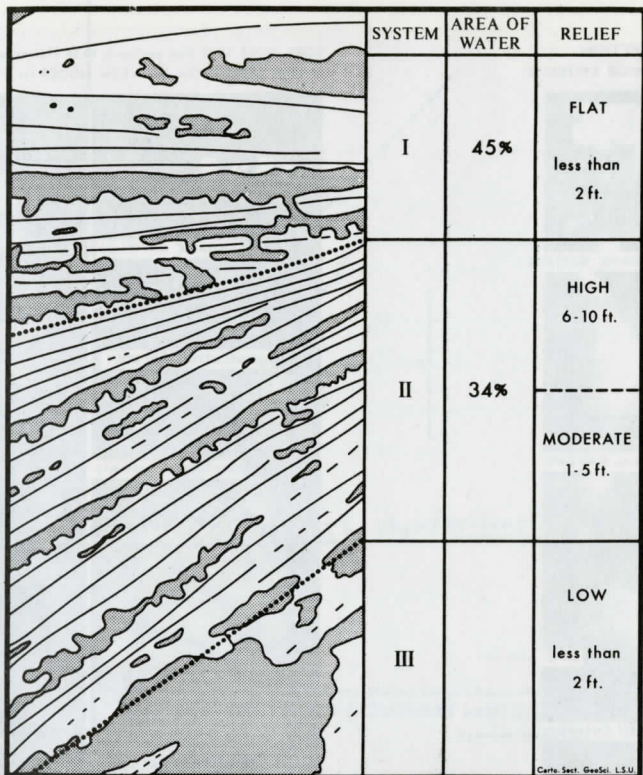
follow the low axis between ridge systems II, III and IV (Figure 1). The larger number of lakes seawards of the ridges are found in association with the margins of the smaller ridge system III. The pronounced meander of the lower course of the greatly reduced Bayou Moreau demonstrates a close adjustment to the extent and lineations of the ridge systems. Northwards of the ridge system, Bayou Moreau flows in an almost straight course from its junction with Bayou Lafourche, six miles to the north, which would appear to indicate a complete absence of diversionary sand ridges north of the area shown on Figure 1.

With progressive drowning of the sand ridges the shape of the swale-lakes and their interconnections change. The proportion of the area occupied by water as well as the planimetric shape of the banks provide a useful index of age. A transect through ridge systems I and II (located in Figure 1) are shown in Figure 3. The crenelate edge corresponds to the landwards margin of the ridge: the straighter edge to the seawards, beach-facing side. In time, as subsidence proceeds, both edges become irregular as the water is now lapping against the sub-aerial ridges on top of the beach ridges. Linkages between lakes are made between lows in the parabolic dune forms.

SWALE LAKE AND RIDGE SEDIMENTS

The swale lakes are shallow and only a thin veneer of organic and silty clay has accumulated in them. Samples from short cores taken in systems II and IV show less than 12 inches of sedimentation on top of sand. Radiographs of these cores revealed a massive, structureless stratigraphy, heavily charged with old root channels and burrows (Figure 4). The high sand content of all the samples is apparent and this would appear to indicate continual movements of small quantities of sand into the swale until comparatively recently. Small iron oxide nodules (1/8 inch diameter) are found in random clusters in the cores. The relative absence of silts and clays must represent the enclosed nature of many of these swale lakes. The lakes are not connected to the marsh distributary system and have only rarely been covered by flood waters of any kind. A similar observation is made by Rosen (1971) who notes the increase in fine-grained sediment in the inter-ridge sediments as compared with the ridge sediments in Grand Isle. Rosen (1971) attributes the presence of silt and clay to lagoonal waters introduced during storms and abnormally high tides. The shallow depth of supra-sand sedimentation in the swales need not therefore be taken as evidence of the recent origin of these features. The lack of a thick organic layer in this admittedly vigorous zone of vegetational growth, however, is more conclusive in suggesting that these are comparatively recent landforms.

The sand of all the ridges is fine textured and well sorted with a median size range from 0.12 to 0.15 mm. Surface samples show



TRANSECT OF RIDGE SYSTEMS

1000 ft.


— Ridge Crests
 Swale Lake
 Boundary of System

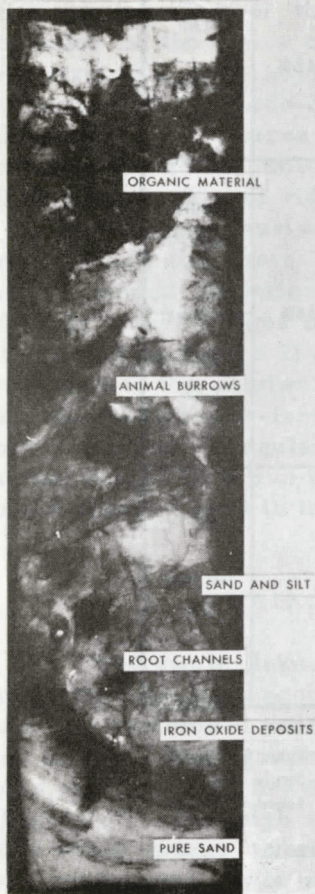
Figure 3. Transect of Ridge Systems.

considerable variation in color due to differences in wetness and organic staining but are generally of a light brown or grey color. Shells are rare. A few intact shells (Modiolus Demissus and Noetia Ponderosa) were recovered from system III. In this respect the ridges are similar to the present beach which is also relatively deficient in shells.

THE PHYSIOGRAPHY OF THE RIDGE SYSTEMS

Figure 1 indicates the separation of the ridge systems into four generations. It can be demonstrated that system I is different from systems II, III and IV by comparing the changing number of individual ridge crests as one proceeds westwards along each system. This is represented by the regression lines of Figure 5 that are constructed by counting the number of ridges at regular intervals along each system.

BORE HOLE TOP SECTION
SWALE LAKE IN RIDGE SYSTEM II



BORE HOLE 1538 (top section), 1539 (lower section)
MARSH DEPRESSION BETWEEN RIDGES IN SYSTEM IV

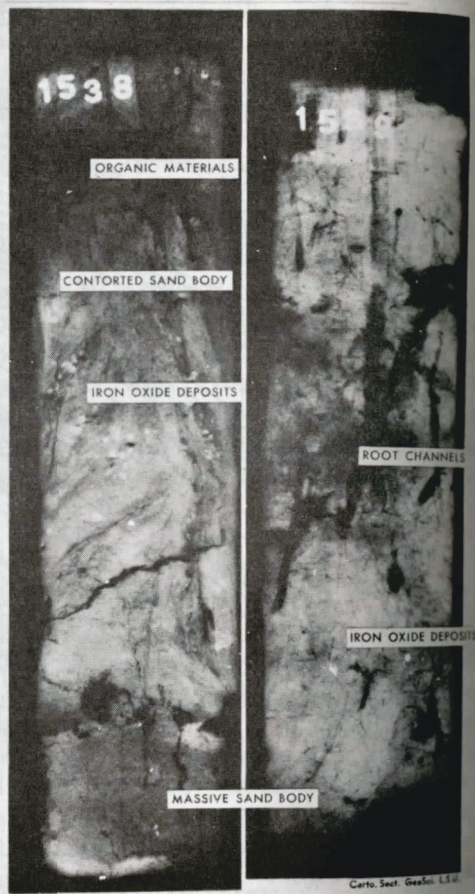


Figure 4. Radiographs of sediment cores from a swale depression lake and marsh area.

Considering also morphology and comparing the distribution pattern with other sand ridges and cheniers as described by Morgan (1970), Hoyt (1969), Fisk (1955), Russell and Howe (1935) and Kolb and Van Lopik (1966) it would appear that system I consists of distributary mouth bar ridges. These ridges would be created when Bayou Lafourche was active. The ridges of system I represent local reworking of the heavier sediment to either side of the prograding distributary lobe. Some of the ridges west of Bayou Moreau are also related to this early phase of development. Systems II, III and IV represent reworking by littoral drift processes. Longshore drift appears to have proceeded from a northeast to southwest direction. The spit-like extensions of the beach

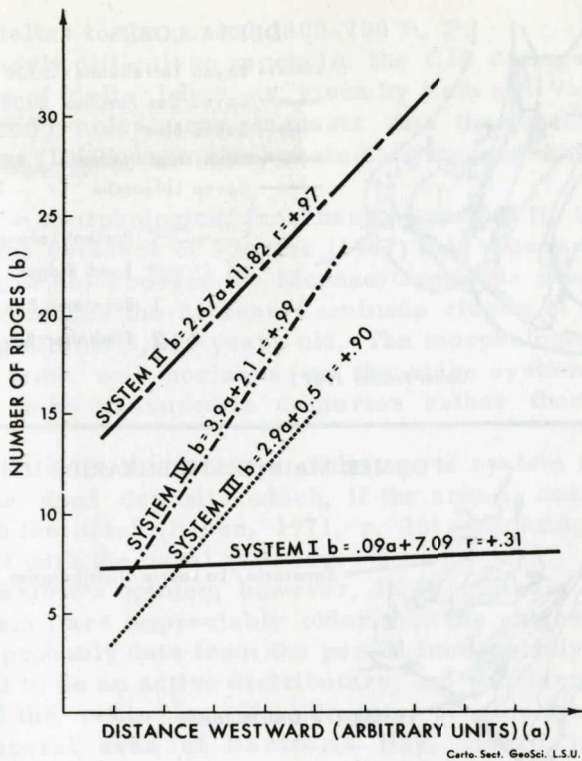
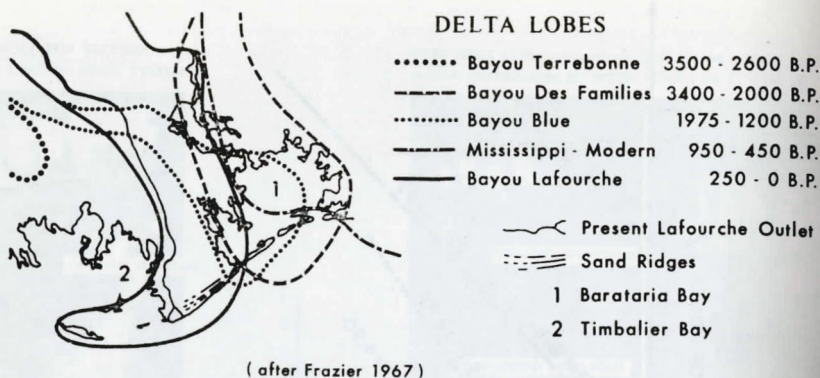


Figure 5. Linear regression lines. Number of Ridge Systems against distance along system.

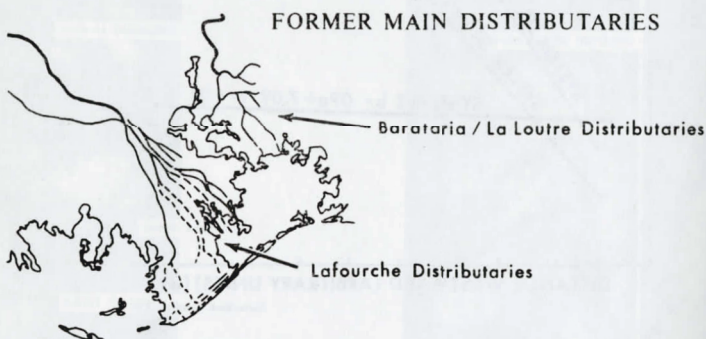
ridges of system II can be seen to have truncated the seawards extensions of the distributary mouth beach ridges (system I). Alternatively, it is possible that system I passes under system II. There is, nevertheless, a basic formative and physiographic difference between system I and systems II, III and IV. The sand of systems II, III and IV could be either local reworking of earlier sand ridges, or a longshore translation of sand as a result of the coastal erosion of similar distributary front sand deposits to the northeast, in the general area of what is now west Caminada Bay.

THE ORIGIN AND DEVELOPMENT OF THE RIDGE SYSTEMS

Figures 6a and 6b which are reproduced from Kolb and Van Lopik (1958 and 1966) and Frazier (1967) indicate the complexity of the chronological and spatial order of the distributary outlets and delta lobes in this area. Using the nomenclature of Frazier (1967), it can be postulated that lobes 7, 10 and 13 as well as the recent Lafourche, lobe 15,



A



Carto. Sect.
GeoSci. I. S. U.

(after Kolb and Van Lopik 1966)

B

Figure 6. Delta lobes (a). Former main distributaries (b).

could be regarded as sources of sand for littoral reworking. Many distributary systems waxed and waned during the last 2,000-3,000 years when the Lafourche was the active delta of the Mississippi. It is probably inadvisable to specify a particular area or distributary as being the definitive source of sediment for the Moreau/Caminada area.

The only direct evidence of age comes from C14 dates of Otvos (1969) which are taken from "autochthonous peat layers interbedded with beach sand" and were found in the sand quarries near the main highway La. 1 and are thus from the oldest part of system I. The samples are dated at 2940 ± 120 and 2340 ± 130 years B. P. Otvos (1969) infers that "the cheniers were contemporaneous with the Bayou Terrebonne Lobe which is known to be active farther north between 3500 and 2000 B. P." (p. 2354). Comparisons with the chronology given by Frazier (1967, p. 308) also shows these dates to be contemporary with Lobe 7 which was active between 3400 and 2000 B. P. and much closer to the Caminada/Moreau area.

Kolb and Van Lopik (1966, p. 23) date the active phase of the

Lafourche subdeltas to the period 1800-700 B. P.

It is clearly difficult to reconcile the C14 datings with the ages and distribution of delta lobes as given by Kolb and Van Lopik (1966) and Frazier (1967) unless one suggests that the "autochthonous peat layers" of Otvos (1969) were older materials incorporated into younger stratigraphy.

Given the morphological freshness of system II, in particular, and accepting the evidence of Frazier (1967) that recent delta lobes i. e. 15 and 10 (Figure 6b) crossed the Moreau/Caminada area, it is almost impossible to consider the Moreau/Caminada ridges of systems II, III and IV as being almost 3,000 years old. The morphological identity and the absence of thick soil horizons on the ridge systems II, III and IV suggest an age to be measured in centuries rather than thousands of years.

The solution to the problem of the age of system I requires coring through the sand deposits which, if the area is analogous to Grand Isle, are 20-30 feet thick (Rosen, 1971, p. 26) and dating organic materials in contact with the basal sand layers.

In the writer's opinion, however, it is unlikely that the sand ridges of system I are appreciably older than the ridges of systems II, III and IV and probably date from the period immediately after the Lafourche ceased to be an active distributary, and subsidence, with coastal retreat and the reworking of sediments, became the dominant process in the general area of Barataria Bay. Hence, to be consistent with the ages given by Kolb and Van Lopik (1966), Frazier (1967) and Rosen (1971) for Grand Isle these Caminada/Moreau sand ridges of systems II, III and IV cannot be older than 700 B. P. at a maximum.

GENERAL CONCLUSION

In general, however, this study demonstrates local changes in the relative dominance of distributary outflow and coastal currents. Changes in the amount of sediment input and alterations in the direction of coastal processes also affect the size, spacing and distribution of the ridge systems. The relative positions of systems II, III and IV (Figure 1) show a progressive shift eastwards and suggest a littoral current coming from the northeast with a sediment source being located in the general vicinity of west Barataria Bay. Alternatively, the sediment source could have been highly localized near the present outlet of Bayou Lafourche and Bayou Moreau and brought to the coastline from the east as a result of a local cell-circulation such as exists at present near the Mermentau River (Coleman, personal communication, 1971) in southwest Louisiana, and demonstrated diagrammatically in Figure 7. The present beach cuts the older ridge systems at an angle of approximately 30° and represents current processes and a recent history of subsidence and large scale coastline retreat in this area of south Louisiana.

MODEL OF TWO POSSIBLE SOURCES OF CAMINADA/MOREAU RIDGES

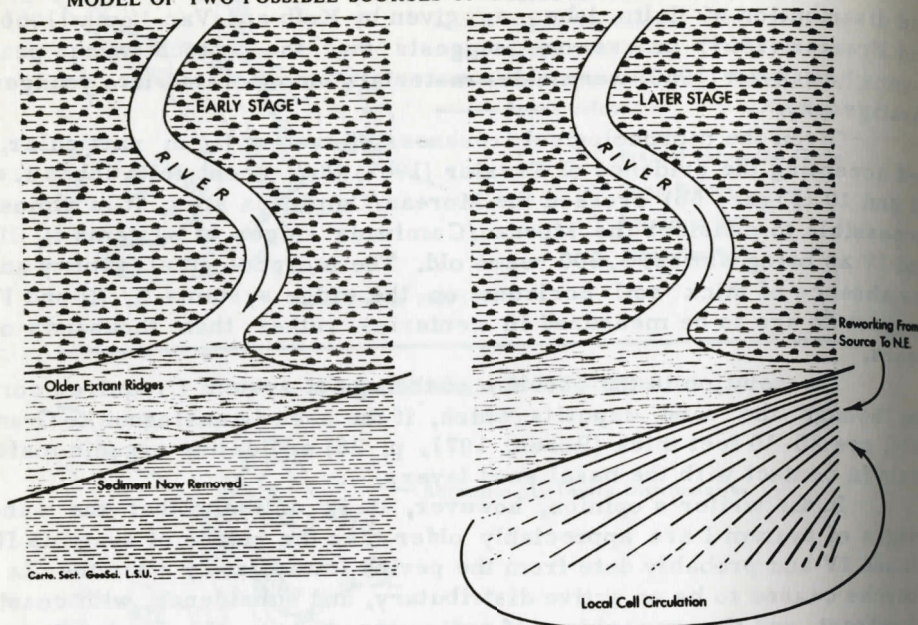


Figure 7. Model of two possible sources of Caminada/Moreau ridges,

This coastal retreat is symptomatic of an insufficient input of beach materials to keep pace with deltaic plain subsidence. In this respect the modern Caminada/Moreau beach ridge contrasts with the older abandoned "cheniers", which represent a period of excess sedimentary influx when the main Mississippi delta terminated naturally, and nearby, in relatively shallow water.

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INTRODUCTION

General description probably crystals are present in the host... contact with the... of the... of Kings Mountain, Cleveland County, North Carolina. The host rocks rarely contain... minerals; however, quartz is present.

The purpose of this note is to describe the problems and to appraise and review data on the crystals found in the... The... warrants description, but are probably simply... crystals... aggregates,... which... and multiple overgrowths are typical... (Bassett, 1961; Aggregates of subparallel plates... (Richardson, Jr., 1937).

GENERAL GEOLOGY

The geology of the... and the surrounding... (Kolb, 1958). At the... this... quartzite... This... of quartzite... The... are very irregular.

PREHNITE CRYSTALS FROM KINGS MOUNTAIN,

NORTH CAROLINA

By

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ABSTRACT

Clear transparent prehnite crystals up to 2 cm long are present in the amphibolites near their contact with spodumene pegmatites at the Foote mine, Kings Mountain, North Carolina. Prehnite rarely develops euhedral crystals. Optical and x-ray data are presented for this unusual occurrence. The prehnite probably originated when hydrothermal solutions from the pegmatites permeated fractures in the amphibolites, forming massive replacement veinlets and prehnite crystals in druses of the incompletely filled fractures.

INTRODUCTION

Unusual transparent prehnite crystals are present in the host amphibolites near their contact with the spodumene pegmatites at the Foote Mineral Company mine at Kings Mountain, Cleveland County, North Carolina. The host rocks rarely contain reaction minerals; however, prehnite is present.

The purpose of this note is to describe the prehnite and to present optical and x-ray data on the crystals found in the vugs. The unusual occurrence warrants description, because prehnite rarely occurs as single crystals; barrel-shaped aggregates, rosettes, tabular groups, reniform globular masses and multiple over-growths are typical (Deer, Howie, and Zussman, 1962). Aggregates of subparallel plates have been reported (Richmond, Jr., 1937).

GENERAL GEOLOGY

The geology at the Foote mine and the surrounding area has been described by Kesler, 1961. At the Foote mine, thin layered amphibolites and muscovite schist containing thin layers of quartzite are cut by a swarm of spodumene pegmatites. The pegmatites are very irregular

and vary in thickness and width. The largest is over 300 meters long and 75 meters thick. Seven smaller dikes 6-33 meters thick are also exposed in the mine. The dikes strike NNE parallel to the schistosity of the enclosing rock, and dip steeply east and west. The pegmatites are uniform in composition, containing 32 percent quartz, 41 percent feldspar (microcline and plagioclase), 20 percent spodumene, 6 percent muscovite and a wide variety of late stage hydrothermal and secondary minerals deposited in druses and fractures.

Over eighty minerals have been identified (White, 1969); these include pyrite, albite, quartz, apatite, vivianite, tourmaline, fairfieldite, rhodochrosite, siderite, laumontite, roscherite, eosphorite, milarite and lithophilite. Several rare minerals such as eucryptite and bikitaite (Leavens, Hurlbut, and Nelen, 1968), and lithophosphate (White, 1969) are present. Two new minerals, switzerite (Leavens and White, 1967) and eakerite (Leavens and White, 1970) have been described.

OCCURRENCE

Veinlets (5-10 cm) of massive white translucent prehnite cut the amphibolites near the contact with a spodumene pegmatite. The veinlets contain vugs in which individual terminated transparent crystals project from a more dense interlocking crystal mosaic deeper in the druse. White translucent calcite crystallized on the prehnite and partially fills the vug.

PHYSICAL DESCRIPTION

Individual prehnite crystals vary from 2-4 mm long by 0.5-1.0 mm wide and 0.25-0.50 mm thick; however, a few crystals are 2 cm long. The crystals are transparent, crystal-clear, adamantine, and consist of flattened blade-like prisms. The prisms are terminated by a dome whose apex in some cases is cut by a small pedion. Terminal faces are smooth; however, the prism sides always show abundant growth striations.

Optical properties, determined from crushed crystals, are: biaxial positive, $2V=50-60$, negative elongation, parallel extinction, and indices $\alpha = 1.616(5)$, $\beta = 1.628$, $\gamma = 1.646$.

X-RAY DATA

The optical identification was confirmed by x-ray powder diffraction. Three samples prepared from separate picked crystals were x-rayed using Ni filtered Cu radiation. The Kings Mountain material gives a pattern almost identical to the calculated pattern data (Borg and

Table 1. X-ray Diffraction Data for Prehnite from Kings Mountain, North Carolina.

I		II		I		II	
$d(\text{obs})\text{\AA}$	I/I ₀	$d(\text{obs})\text{\AA}$	I/I ₀	$d(\text{obs})\text{\AA}$	I/I ₀	$d(\text{obs})\text{\AA}$	I/I ₀
9.26	W	9.24	W	2.32	12	2.32	23
5.27	14	5.26	17	2.31	15	2.31	12
4.63	20	4.63	30	2.20	10	2.20	13
4.14	W	4.14	10	2.12	W	2.12	W
3.54	40	3.54	42	2.07	20	2.07	24
3.47	90	3.47	100	2.05	W	2.05	15
3.30	40	3.30	63	2.015	W	2.015	W
3.27	40	3.27	39	1.990	W	1.990	W
3.07	90	3.07	98	1.940	10	1.936	13
2.81	40	2.81	42			1.923	W
2.74	W	2.74	W	1.850	10	1.848	W
2.63	10	2.63	20	1.777	15	1.776	24
2.56	90	2.56	89	1.769	20	1.769	75
2.38	W	2.38	W			1.767	W
2.36	40	2.36	34	1.755	W	1.754	W
2.34	15	2.34	23				

a Diffractometer: nickel filtered CuK α 1 radiation, tube at 30 Kv, 15 ma, 1 $^{\circ}$ divergence slit, 0.01 inch receiving slit, scanning speed 1 $^{\circ}$ per 20 minute, W = less than 10.

I Average of three different analyses, Kings Mountain, North Carolina

II Calculated x-ray powder pattern for prehnite, Borg, I. Y., and Smith, D. K., 1969, p. 600.

Smith, 1969). Two weak reflections at 1.923 Å and 1.767 Å present in the calculated pattern were not recorded in the Kings Mountain material (Table 1).

ORIGIN

Prehnite is found most commonly as hydrothermal veinlets and groups of tiny crystals forming globular masses in amygdules of basic igneous rocks. It also occurs as hydrothermal veins in diorites and granites. Prehnite is also a widespread product of low grade regional metamorphism and is sometimes produced in contact zones (Deer, Howie and Zussman, 1962).

At Kings Mountain, North Carolina, the prehnite is a product of hydrothermal solutions. The solutions moved out from the pegmatites along joints into the host amphibolites and apparently filled the fissures

and replaced the amphibolites. Most fractures are small and contain only massive prehnite; however, in the larger openings associated with brecciation, filling was incomplete. Individual terminated crystals, smaller interpenetrating clusters, and parallel crystal groups were thus able to form on the massive prehnite which is present at the contact with the amphibolite. Calcite partially covers the prehnite.

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