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Abstract

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SOUTHEASTERN GEOLOGY

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Vol. 11, No. 4

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DAMMING ACTION OF XENOLITHS AND SEGREGATES

WITHIN THE MOUNT AIRY "GRANITE"

(North Carolina, U. S. A.)

By

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ABSTRACT

Mafic-poor microphaneritic zones up to a few centimeters thick occur directly beneath several of the xenoliths and some of the biotite-rich segregates within the Mount Airy granodiorite mass. Petrographic and bulk chemical analyses show that compared to the typical granodiorite most of the mafic-poor zones have finer grain size; lower percentages of feldspars as well as of mafic minerals; higher percentages of quartz; lower Al, Ti, Fe, Mg, Ca, Sr, and Mn contents; and higher Si contents. In addition, both the mafic-poor zones and all other analyzed rocks of the mass have Rb/Sr and Ca/Sr ratios that are notably less than those recorded for most granitic rocks.

Electron microprobe analyses for 20 elements of several grains of coexisting quartz, plagioclase, microcline, biotite, and muscovite from one mafic-poor zone and from the typical granodiorite show that, with some notable exceptions, most minerals do not range significantly in composition. The exceptions are Ba in biotite and microcline, Na in muscovite, Sr in oligoclase, and Ti in biotite and muscovite all of which exhibit systematic trends across the mafic-poor zone. In addition, the partition coefficients $Ba_{K-feld.}/Ba_{muscov.}, Ba_{biot.}/Ba_{muscov.},$

Na_K-feld. $/Na_{muscov.}$, and $Ti_{biot.}/Ti_{muscov.}$ also range systematically across the zone. Furthermore, several of the concentration and partition coefficient values for the samples of the mafic-poor zone differ rather markedly from those of the analyzed specimen of the typical granodiorite.

The mafic-poor zones are interpreted to represent differentiated material the upward movement of which was stopped by damming action of the xenoliths and segregates. The systematic trends in mineral compositions and partition coefficients across the mafic-poor zone are attributed to changes in physiochemical conditions that took place during the differentiation process.

INTRODUCTION

The Mount Airy "granite," actually a leucogranodiorite, exhibits an interesting phenomenon that was tentatively interpreted to reflect "the damming action of inclusions" (Dietrich, 1961). Briefly, mafic-poor zones, most of which are much finer grained than the typical granodiorite at the present level of exposure, occur directly beneath several of the relatively large xenoliths (Figure 1) and some of the biotite-rich segregates within the mass. The relationship was so interpreted because these mafic-poor zones have been found to occur only beneath xenoliths and segregates within the mass, i.e., such mafic-poor rocks have not been found to surround inclusions like the rather widely reported "reaction haloes" do. This note summarizes the results of an intensive study of the petrography and chemical composition of one of the mafic-poor zones.

PETROGRAPHY

The petrography of the Mount Airy granodiorite has been dealt with at length by Dietrich (1961, p. 8-27). The main rock, hereinafter called simply granodiorite, is a relatively homogeneous, medium- to coarse-grained leucogranodiorite that consists of quartz (~21 per cent), oligoclase (~55 per cent), microcline (~20 per cent), biotite (~2.5 per cent), and several accessory minerals. The investigated fine-grained mafic-poor zone consists of the same major constituents but with different mineral percentages — the quartz percentage is higher, the feldspar percentages are slightly lower, and the rest of the minerals including biotite constitute less than one per cent. Also, as the above designations indicate, there are marked differences in grain size and in fabric: whereas the typical granodiorite has a hypidiomorphic granular texture, the mafic-poor zone has a very fine-grained panallotriomorphic texture. The former may be characterized as a medium- to coarse-grained phanerite, the latter as a saccharoidal microphanerite.

UP
↑

Figure 1. Mafic-poor microphan-
eritic zone beneath a xenolith,
which is constituted by biotite
gneiss plus a recrystallized
epidote- and sphene-bearing
quartz vein, from the Mount
Airy mass. (Another picture
of this phenomenon is shown
as Fig. 18-10 in Leet and Jud-
son, 1965. The orientation of
that picture, which is of a hori-
zontal surface, is bottom of
xenolith toward upper left of
photograph.)

CHEMICAL COMPOSITION

Bulk chemical analyses were made of 18 specimens representa-
tive of the granodiorite at the current level of exposure and of one
sample of the mafic-poor zone. The analyses were made by utilizing
atomic absorption and x-ray spectrometry. Analytical procedures,

Table 1. Chemical Analyses — Analyses by P. C. Ragland

	leucocratic zone (sample 15)	"typical" granodiorite (sample 17)	average of 18 granodiorites	extreme values of 18 rocks in "ave."	
SiO_2	76.4	73.2	72.99	74.9	71.4
TiO_2	0.03	0.20	.19	0.23	0.16
Al_2O_3	14.1	15.3	15.3	16.2	14.1
Fe as FeO	0.44	1.22	1.31	1.60	1.09
MgO	tr.	0.37	0.37	0.45	0.32
CaO	1.37	1.62	1.55	1.80	1.41
Na_2O	4.70	5.00	4.90	5.37	4.37
K_2O	2.86	3.04	2.92	3.33	2.57
Total	99.90	99.95	99.53		
ppm Rb	63	66	62.4	70	52
ppm Sr	520	660	682	750	600
ppm Zr	100	200	204	240	190
ppm Mn	120	190	173	190	160
K/Rb	380	380	387	450	340
Rb/Sr	.121	.100	.092	.110	.069
Ca/Sr	19	18	16.4	18	15

precision, and accuracy have been discussed by Ragland et al. (1968, p. 78).

Two of the individual analyses and an average of the 18 analyses of the typical granodiorite are given in Table 1. The leucocratic zone has a higher SiO_2 content and lower TiO_2, Al_2O_3, Σ Fe(FeO), MgO, CaO, Sr, and Mn contents than the typical granodiorite, whereas the Na_2O, K_2O, and Rb contents fall within the range of those for the typical granodiorite.

In addition, the Rb and Sr contents for all analyzed rocks of the mass are notably lower and higher, respectively, than those thus far reported for most "granitic" rocks. Consequently, the Rb/Sr ratio for the Mount Airy mass is extremely low as compared to the same ratio reported for most granitic rocks. Furthermore, the Ca/Sr ratios for the Mount Airy rocks analyzed in this investigation are all less than 20, which is less than half of the ~50 typical of most previously reported granitic rocks.

Calculations of normative albite and anorthite for both the rock of the mafic-poor zone and the typical granodiorite yield an An-content
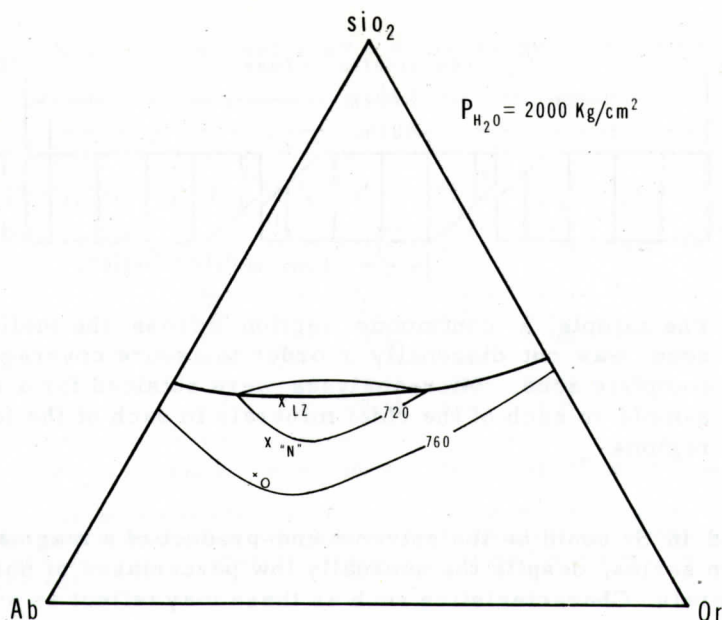
Figure 2. Plot of Q-Ab-Or ratios for typical grano-
diorite ("N"), a mafic-poor zone (LZ), and an
other differentiate (0) on the Tuttle and Bowen
(1958) $P_{H_2O} = 2000$ kg/cm^2 diagram. The "other
differentitate (0)" is from a dikelet within the
mass. The mineralogical and chemical com-
positions of the dikelet and of the main mass,
including their similarly atypical Rb/Sr and Ca/Sr
ratios, attest to their consanguinity. It appears
that the magma that consolidated to form the
dikelet probably represents differentiated mate-
rial that remained mobile after and was injected
upwards into the rock material at the present
level of exposure subsequent to the latter's be-
coming essentially solidified.

of 14.5 $\pm$ 0.5, which is very close to the average content of the zoned
plagioclases within the rock. Calculations of the normative albite,
orthoclase, and quartz give the two points, LZ (mafic-poor) and "N"
(normal), plotted on the Q-Ab-Or diagram (Figure 2).

It is noteworthy that, as previously mentioned (Dietrich, 1961,
p. 30-32), the Mount Airy rock appears to be alien to any "normal" in-
trusive series. Along with the already mentioned extraordinary Rb/Sr
and Ca/Sr ratios, the rocks are oversaturated with silica but also have
normative corundum. Moreover, the normative quartz and feldspars
constitute approximately 98 per cent of the total analyses (after recal-
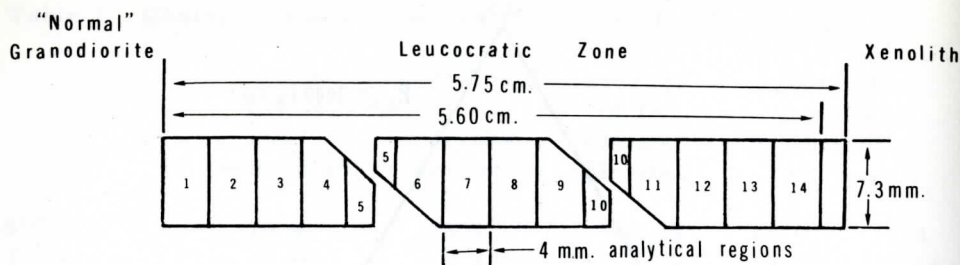culation to 100 per cent). It seems highly unlikely, e. g. , that a magma

Figure 3. The sample, a continuous section across the mafic-poor zone, was cut diagonally in order to assure coverage of the complete zone. Microanalyses were obtained for a random sample of each of the chief minerals in each of the fourteen regions.

so enriched in Sr could be the extreme end-product of a magmatic differentiation series, despite the unusually low percentages of normative femic minerals. Characteristics such as these may reflect an anatectic origin for the magma that consolidated to form the mass.

Microprobe analyses were made at the Virginia Polytechnic Institute for twenty elements (Si, Al, Mg, Na, K, Ca, Ti, Mn, Fe, Sr, Ba, Pb, Ge, F, Rb, P, B, Zr, Cl, and V) of the five main mineral constituents (quartz, plagioclase, microcline, biotite, and muscovite) within diverse parts of the pictured mafic-poor zone (Figure 1) and of a specimen of the typical granodiorite. The sample of the mafic-poor zone was prepared as indicated in Figure 3. The $\underline{x}$-ray data were corrected for deadtime, background, and drift and were averaged for each sample. Corrections were made for absorbtion, according to Philibert's (1965) formula; for fluorescence on the basis of a modified combination of Castaing's (1951) and Wittry's (1962) suggestions; and for the atomic number affect, utilizing Thomas' (1963) equation. The stopping power of the i'th element which was used is from Nelm's (1956, 1958) data. Calculations were carried out on the Virginia Polytechnic Institute's IBM 360 computer system.

A summary of the analyses is given as Table 2. Many of the trace elements were below detection for most mineral grains. Complete data for all minerals is available from CCM Information Sciences, Inc. (22 W. 34th St., New York 10001)

Except for Ba, Na, Sr, and Ti, each of the elements is present in essentially the same concentrations in all analyzed grains of the individual minerals from both the typical granodiorite and diverse parts of the mafic-poor zone. The exceptions are: from typical granodiorite toward xenolith, Ba in biotite and microcline, Na in muscovite, and Sr in oligoclase decrease, whereas Ti in both muscovite and biotite increases (Table 3). In addition, the two partition coefficients $Ba_{K-feld.}/Ba_{muscov.}$ and $Ba_{biot.}/Ba_{muscov.}$ decrease, whereas the coefficients

Table 2. Microprobe Analyses* - analyses by M. R. Hall

	Quartz		Oligoclase		Microcline		Biotite		Muscovite	
	D-ave[1]	Typ gd[2]	D-ave	Typ gd	D-ave	Typ gd	D-ave	Typ gd	D-ave	Typ gd
SiO_2	99.79	98.21	65.51	64.19	64.39	63.18	36.34	35.39	46.65	45.88
TiO_2	-	-	-	-	-	-	3.48	2.84	1.74	1.73
Al_2O_3	0.036	0.004	22.76	22.66	18.84	18.66	15.37	15.27	28.01	28.30
Fe as Fe_2O_3	0.016	0.008	0.040	0.012	0.019	0.018	25.64	24.52	6.76	5.89
MnO	-	-	-	-	-	-	0.25	0.25	-	-
MgO	-	-	0.008	<0.005	0.007	0.009	8.57	8.26	1.87	2.20
CaO	-	-	3.04	3.29	<0.002	0.005	0.032	0.025	-	-
BaO	-	-	0.019	0.036	0.56	0.71	0.166	0.205	0.104	0.106
SrO	-	-	0.074	0.047	0.017	0.022	-	-	-	-
Na_2O	-	-	9.88	9.58	0.75	0.88	0.050	0.054	0.227	0.116
K_2O	-	-	0.16	0.15	15.62	15.23	9.26	9.25	10.65	10.53
Total	99.84	98.22	101.49	99.96	100.21	98.93	99.16[4]	96.06[4]	96.01[4]	94.75[4]

* Additional analyses were made for Pb, Ge, F, Rb, P, B, Zr, Cl, and V. F at a level of "< 500 ppm" is reported for biotite and muscovite; Rb at a level of "< 500 ppm" is reported for oligoclase and microcline; P at a level of "< 100 ppm" is present in each of the minerals on the table.

1. Average within mafic-poor zone (D).
2. In typical granodiorite.
3. - means none detected.
4. Plus unanalyzed H_2O.

Table 3. Concentration Values and Partition Coefficients. ($\overline{X}$ - arithmetic means; S - best linear estimate of σ, standard deviation. For Samples 1-4: $S = .4539 (X_4-X_1) + .1102 (X_3-X_2)$, where $X_1 < X_2 < X_3 < X_4$; for samples 5-7: $S = .5908 (X_3-X_1)$. See Dixon & Massey, 1957, Table A8b(6), p. 407).

| Variate	Samples 1-4 $\overline{X} \pm S$	Samples 5-7 $\overline{X} \pm S$	Granodiorite
BaO - biotite	.260 + .092	.116 + 0.59	.205
BaO - microcline	.718 + .337	.409 + .063	.926
TiO_2 - muscovite	1.67 + 0.95	1.85 + 0.09	1.73
TiO_2 - biotite	2.80 + 0.53	3.90 + 0.14	2.84
Na_2O - muscovite	2.75 + 0.23	.164 + .063	.116
SrO - oligoclase	.105 + .027	.070 + .021	.047
$\dfrac{\text{Ba - microcline}}{\text{Ba - muscovite}}$	7.00 + 2.68	3.78 + 0.89	8.74
$\dfrac{\text{Na - microcline}}{\text{Na - muscovite}}$	2.89 + 0.35	4.87 + 1.99	7.57
$\dfrac{\text{Ti - biotite}}{\text{Ti - muscovite}}$	1.68 + 0.30	2.11 + 0.10	1.64
$\dfrac{\text{Ba - biotite}}{\text{Ba - muscovite}}$	2.58 + 0.90	1.09 + 0.62	1.93
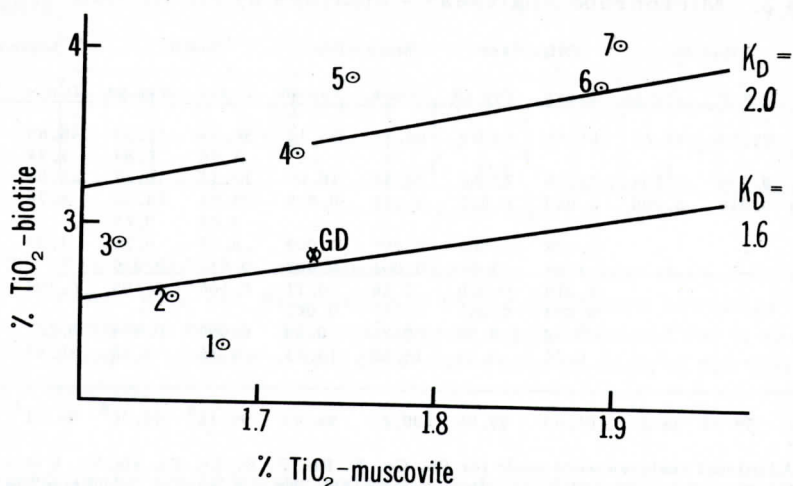
Figure 4. Plot of percentage TiO_2 in muscovite: TiO_2 in biotite. Note that there is a systematic change in both the absolute concentration and the partition coefficient across the mafic-poor zone. (Explanation: GD is the typical granodiorite; the numbers correspond to sample numbers given in Table 2; the solid, sloping lines represent constant partition coefficient (K_D) values.)

$Na_{K-feld.}/Na_{muscov.}$ and $Ti_{biot.}/Ti_{muscov.}$ increase toward the xenolith (Table 3). Figure 4 shows that there is a systematic change in the absolute concentration of TiO_2 in biotite and muscovite, as well as in the partition coefficient.

Some of the concentration values and partition coefficients for the "typical" granodiorite differ markedly from the correlative values for the mafic-poor zone. Note in particular 1) Ba in microcline, 2) Na in muscovite, 3) Sr in oligoclase, 4) $Ba_{K-feld.}/Ba_{muscov.}$, and 5) $Na_{K-feld.}/Na_{muscov.}$ (Table 3).

The reason that data for only samples 1-7, rather than for all samples, were used in these calculations is that the other samples (8-14) lack muscovite and, thus, have different coexisting phases.

DISCUSSION

The absence of muscovite in that part of the mafic-poor zone closest to the xenolith (samples 8-14) may have important implications so far as changes that might have taken place in physiochemical conditions during the differentiation process. The mafic-poor zones appear most likely to have formed as a result of a damming action of xenoliths

(and segregates) on the upward movement of a mobile, probably volatile-rich differentiated phase. Assuming $P_{vapor} < P_{total}$ during most of the crystallization of the granodiorite, the formation of this mobile phase very likely would have occurred at the stage when $P_{vapor} = P_{total}$ and some vapor phase(s) separated. Thus the muscovite-free zone closest to the xenolith may have formed earlier than the zone farther away, at a time of lower vapor pressure and possibly at a slightly higher temperature.

This interpretation could be extended to account for even those samples in which muscovite does coexist as a stable phase (samples 1-7), in that those samples may reflect progressively changing conditions of temperature and vapor pressure with higher vapor pressures and slightly lower temperatures of crystallization farther from the xenolith. Such variations in physicochemical conditions of crystallization with space and presumably also with time could then explain all of the systematic changes in partition coefficients noted in Table 3 and Figure 4. Since the composition of such an ascending mobile phase would also probably undergo changes, the progressive compositional variations in individual minerals across the mafic-poor zone might be interpreted to be indicative of those changes.

It can be predicted qualitatively that a given partition coefficient between coexisting phases will approach 1.0 as temperature increases. This, then, could explain the Ba partitioning, but not necessarily that of Na or Ti, shown on Table 3. As no experimental data are available, it is not now possible to evaluate the relative effects of temperature, vapor pressure, and/or some other variable(s), such as composition, on these latter partition coefficients.

CONCLUSIONS

The data presented in this paper and all other pertinent data are considered to support the following conclusions: 1) the Mount Airy mass was consolidated from a magma that was probably formed by the melting of some preexisting rock; 2) the magma underwent differentiation during consolidation; 3) the fine-grained mafic-poor zones beneath several of the relatively large xenoliths and some of the biotite segregates represent differentiated material the upward movement of which was stopped by the damming action of those xenoliths and segregates, i.e., the xenoliths and segregates acted as local ceilings beneath which there is a telescoping of the results of differentiation; and 4) systematic trends in mineral compositions and partition coefficients across the mafic-poor zones may reflect changes in physiochemical conditions during differentiation.

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PHYSICAL STRATIGRAPHY OF THE ORDOVICIAN MARTINSBURG

FORMATION AND ITS LATERAL EQUIVALENTS IN

SOUTHWESTERN VIRGINIA

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ABSTRACT

The Ordovician Martinsburg Formation in southwestern Virginia is a shallow water geosynclinal unit consisting of three lithofacies. A lower limy facies of limestone and calcareous siltstone occupies the western and central outcrop belts of the Valley and Ridge province. It is replaced to the east and vertically by a silty facies of siltstone and shale. The upper sandy facies consists mostly of sandstone and siltstone and underlies the littoral and nonmarine redbeds of the Juniata Formation. The Martinsburg sediments were deposited in a shallow marine environment with a diverse benthonic fauna. The change from a carbonate depositional environment to nonmarine conditions occurred during Middle (Trentonian) to Late (Maysvillian) Ordovician time as terrigenous material spread southwestward across the southern Appalachian geosyncline. The source area was produced by Taconic orogenic activity that culminated in nonmarine conditions in the geosyncline.

INTRODUCTION

The Martinsburg Formation in southwestern Virginia consists of three facies that were deposited in environments which progressed in a southwestward direction during middle and late Ordovician time (Figure 1). Land associated with Taconic orogenic activity was the source for terrigenous material that spread across the Appalachian geosyncline and the Martinsburg Formation records the change from a marine carbonate depositional environment to one of littoral or nonmarine sand and shale.

The Martinsburg crops out within the Valley and Ridge province from eastern Tennessee to New York. The name is actually applied to a stratigraphic interval including several facies rather than to a body
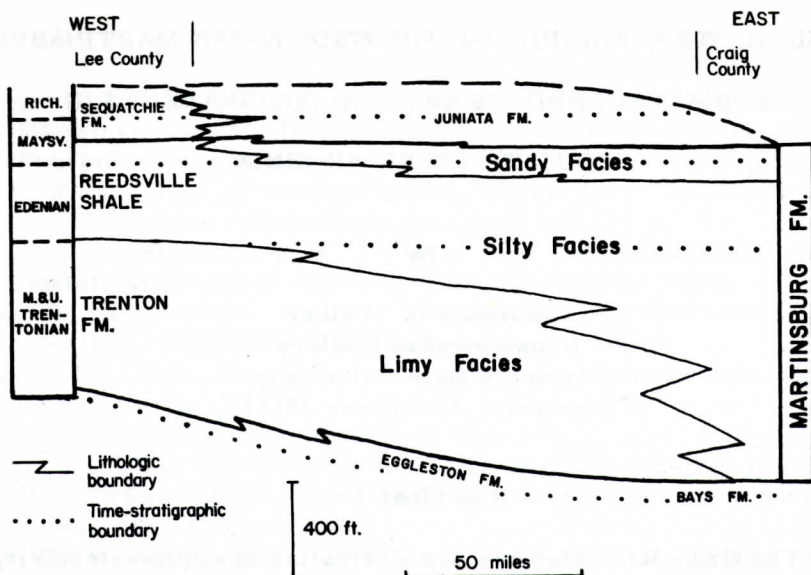
Figure 1. Schematic cross-section across southwestern Virginia showing facies within the Martinsburg Formation and the equivalent formations, the Trenton Formation and the Reedsville Shale, in western outcrop belts. Time-stratigraphic boundaries are approximate.

of rock characterized by lithologic homogeneity. From west-central Virginia northeastward to New York, the Martinsburg is primarily shale and graywacke and is interpreted as a thick sequence of pelagic shale and turbidite deposits for which the source area lay to the southeast (McBride, 1962). Estimates of thickness range from 2500 to 9000 feet. The type section near Martinsburg, West Virginia, contains the shale and graywacke lithologies (Geiger and Keith, 1891).

The Martinsburg in southwestern Virginia differs markedly from strata at the type section. Three lithofacies can be recognized (Figure 1): (1) a limy facies, which is predominantly argillaceous limestone and calcareous siltstone; (2) a silty facies, mostly of siltstone and shale; and (3) a sandy facies consisting of sandstone, siltstone, and shale. The formation iu this area has a much greater amount of carbonate than to the northeast, it is less than 2000 feet thick, fossil benthos are abundant, and clastic beds show little or no evidence of turbidite origin. It was deposited in a shallow water environment that was favorable to a diverse fauna and under conditions that permitted accumulation of large amounts of shell debris by winnowing of muds and transportation of the residual shells.

In western outcrop belts in the southern Appalachians, the stratigraphic interval approximately equivalent to the Martinsburg is divided

into a lower limestone unit, the Trenton Formation, and the Reedsville Shale. These formations correspond to the limy and silty facies, respectively (Figure 1).

Five well-exposed and reasonably complete sections were selected for detailed study (Figure 2). They are geographically situated to illustrate the east to west facies changes within the Martinsburg. Several other partial exposures of the formation were examined. All the principal sections are within different thrust blocks and their present positions with respect to one another are not the same as at the time of deposition. Locations of these sections are given in the appendix.

Acknowledgements

PHYSICAL STRATIGRAPHY

Adjacent Formations

In the southern Appalachians the Martinsburg Formation overlies either sandstone of the Bays Formation or argillaceous limestone and shale of the Moccasin or Eggleston Formations, which are at least partial equivalents to the Bays. The bulk of terrigenous material in these units was derived from a southeasternly source associated with the Blountian phase of the Taconic orogeny during the Middle Ordovician (Rodgers, 1953). Probably part of the clastics within the Martinsburg came from this source. The area of maximum subsidence and accumulation as well as the primary source area apparently shifted northward after deposition of the Bays and allowed the thicker accumulation of Martinsburg sediments in the central Appalachians.

The Martinsburg is overlain in most sections by the Juniata Formation, a sequence of red sandstone, shale and argillaceous limestone interpreted as littoral and nonmarine deposits (Rodgers, 1953; Thompson, 1969). These redbeds and the overlying nonmarine Tuscarora sandstone of Silurian age were formed during westward encroachment of land associated with the Taconic Orogeny. The Reedsville shale is overlain by the marine Sequatchie Formation, a red and gray argillaceous limestone considered to be an offshore equivalent of the Juniata.
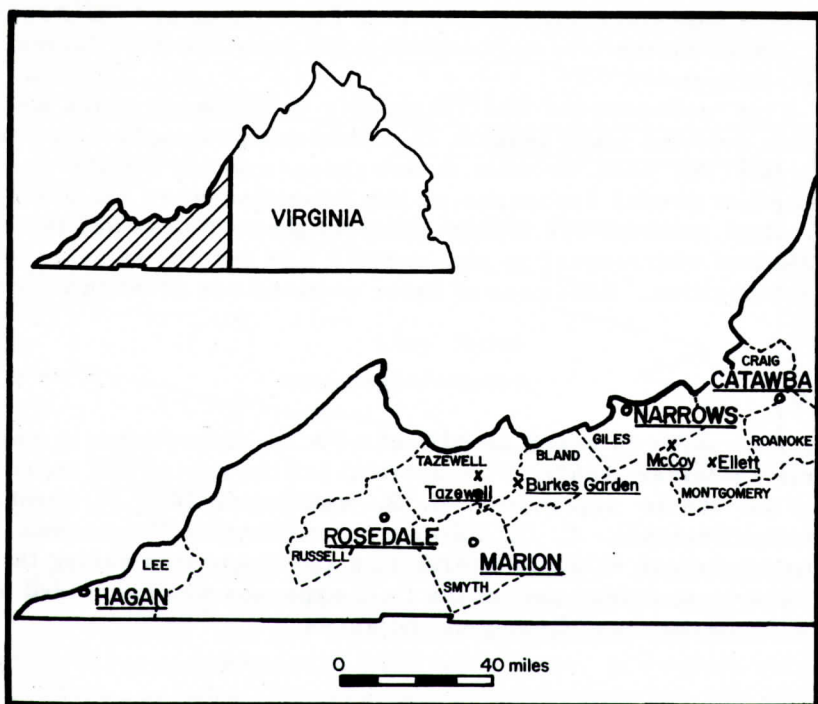
VIRGINIA

CRAIG
CATAWBA
NARROWS
GILES   ROANOKE
TAZEWELL   BLAND   McCoy xEllett
    X   X Burkes Garden
   Tazewell   MONTGOMERY
        ROSEDALE   o
    RUSSELL   MARION
LEE   SMYTH
  HAGAN

0   40 miles

Figure 2. Index map of southwestern Virginia showing lo-
cations of principal sections (upper case letters,
underlined) and of important supplementary expo-
sures (lower case letters, underlined).

The Juniata is thin or missing in some southeastern sections
and the Tuscarora rests on the Martinsburg. B. N. Cooper (1961)
pointed out that the Martinsburg, Juniata, and Tuscarora Formations
all thin from the central into the southern Appalachians, implying a
northerly or northeasterly source for the bulk of the clastics, or pos-
sibly less subsidence in the southern geosyncline.

Martinsburg Facies and Equivalent Formations

Common practice has been to subdivide the Martinsburg into
Trenton, Eden, and Maysville "divisions" or "members" on the basis
of fossil taxa (Butts, 1940; B. N. Cooper, 1944; Woodward, 1951; Rader
and Ryan, 1965). However, in the writer's opinion members should be
defined on the basis of lithology and this work was pointed toward a
lithologic subdivision of the Martinsburg. Perhaps expanded work in
other areas will show the practicality of redefining as members the
three facies described herein.

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Lithologic descriptions are based on field observations and examination of hand specimens in the laboratory. The samples were examined with a hand lens and binocular microscope; no thin sections were made. Limestone and calcareous siltstone and sandstone specimens were etched with dilute hydrochloric acid.

Base of the Martinsburg. Coquinoid limestone beds containing more than 70 per cent megascopic fossil fragments are interbedded with argillaceous limestone, calcareous siltstone, and shale strata in the lower Martinsburg. In central and western outcrop belts the presence of numerous coquinoid beds is an obvious lithologic difference between the basal Martinsburg and the underlying Eggleston Formation. The Eggleston, named by Mathews (1934), is a sequence of sparingly fossiliferous, gray argillaceous limestone and shale with interbedded bentonites. Some silicified limestone beds commonly display "cuneiform jointing" (described by Rosenkrans, 1936; B. N. Cooper, 1944). The Eggleston ranges from about 25 to 125 feet thick. It overlies the Moccasin Formation, which consists mainly of red calcareous mudstone. The bentonites in the Eggleston are probably the same as those in the upper Bays Formation, a sandstone unit that underlies the Martinsburg in southeastern exposures (Rosenkrans, 1936; B. N. Cooper, 1961). A thin transitional sequence occurs at the top of the Bays. Sandy coquinoid limestones or shelly sandstones are found within this interval and the lowest is chosen as the basal bed of the Martinsburg.

The coquinoid beds weather to a pink crumbly clay that is distinctive in outcrops. The beds are obviously lenticular and the lowest bed in one section is probably not the same as the lowest in any other section, and they almost certainly are not contemporaneous. However, they do represent an environment different from that which prevailed during the deposition of the underlying units.

In extreme southwestern Virginia, the base of the Trenton Formation is likewise placed at the lowest coquinoid bed in a sequence that contains numerous such beds as well as medium and coarsely crystalline limestone. It overlies beds of argillaceous limestone and calcareous mudstone that have been called the Eggleston Formation (Butts, 1940; Miller and Fuller, 1954). B. N. Cooper (in Twenhofel and others, 1954) pointed out that the term "Eggleston" is possibly not appropriate in the area, as the unit below the Trenton Formation contains fossils considered definitely to be older than that in the type Eggleston.

Limy Facies. The limy facies of the Martinsburg Formation consists of argillaceous limestone, calcareous siltstone, coquinoid limestone, and calcareous shale interbeds. The proportions of each rock type vary between any given stratigraphic intervals. For the facies as a whole, approximate percentages would be: argillaceous limestone and calcareous siltstone 60, coquinoid limestone 20, and shale 20 per cent.

The argillaceous limestone and calcareous siltstone are similar in appearance and differ only in the proportions of constituents, the

limestone having more than 50 per cent calcium carbonate and the silt-stone more than 50 per cent detritus exclusive of calcium carbonate. The fresh rock is gray or dark gray and weathers to a light to moderate brown color. The calcium carbonate matrix is mostly very finely crys-talline. Coarser sparry calcite is common within articulated brachio-pods, in beds with abundant fossils, and in the numerous fracture fill-ings and vugs. The estimated percentage of carbonate in a single bed was seldom greater than 85 per cent and was as low as 10 to 15 per cent.

Clay and silt are the predominant terrigenous constituents. They tend to be concentrated in parallel and cross laminae and in cross beds up to 3 inches thick. Very fine to fine-grained, mostly subangular sand is common within siltstone beds, but sandstone beds are rare. Small amounts of mica can be found in siltstone beds. Limonite is pre-sent in the matrix of some beds, particularly those with higher amounts of clay.

Beds range in thickness from 1 to 14 inches and commonly thin by as much as one-half within a distance of several feet. Bedding sur-faces are irregular, with some having ripple marks and casts of de-pressions in underlying beds. Many of the depressions appear to be due to scour and fill--laminae are truncated and the fill includes graded fossil fragments.

The fossil content of the limestone and siltstone is quite vari-able. The fossils are commonly concentrated in thin layers at the top or bottom of beds. Finely-ground fossil debris probably comprises much of the matrix material.

In the coquinoid limestones, the fossils commonly have a ran-dom orientation and brachiopods are rarely articulated. These beds contain fragments of shale up to 5 inches long. The beds are 1 to 16 inches thick and most pinch and swell, some as much as 90 per cent within a distance of several feet. Bedding surfaces are irregular and show evidence of scour and fill. Terrigenous material is present in most of the coquinoid beds and under weathering they become a crumbly, silty, pink clay. In a given interval of 20-30 feet, these coquinoid beds may make up as much as 50 per cent of the thickness.

Gray, dark gray and green calcareous shale occurs as shaly partings and interbeds. Shale intervals are up to 18 inches thick, but most are less than 4 inches thick. The boundaries between shale inter-vals and adjacent argillaceous limestone or calcareous siltstone beds are commonly gradational and many of these intervals include beds of calcareous siltstone less than 1 inch thick.

The upper boundary of the facies is gradational and must be ar-bitrarily placed somewhere within the interval between predominantly limestone and calcareous siltstone and the overlying predominantly shale and siltstone sequence. Figure 3 illustrates the thickness of all the facies in southwestern Virginia. At Narrows the lower facies is about 670 feet thick. It is no more than 500 feet thick at Marion; how-
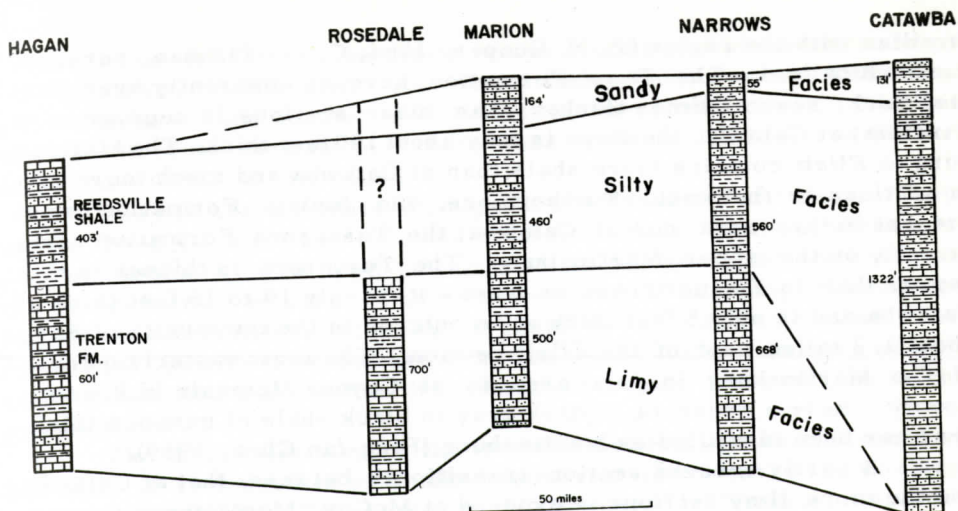
Figure 3. Columnar sections showing the thicknesses of facies within the Martinsburg Formation and thicknesses of the Trenton and Reedsville Formations in the Hagan section.

ever, elements of the brachiopod and trilobite fauna are found unusually low in the section and perhaps indicate the presence of a fault in covered intervals in the lower 500 feet. The complete facies might be as much as 700 feet thick here. At Rosedale only the lower 700 feet of the Martinsburg was measured because of the deformation and poor exposure of the rest of the formation. The entire measured interval belongs to the limy facies.

The facies is not distinguishable in the eastern outcrop belts of the Martinsburg, in which the Catawba section is located. Here the same rock types are present but in different proportions. Shale and siltstone are the prevalent lithologies, making up about 80 per cent of the lower half of the formation. Thin bedded, very argillaceous limestone and sandy coquinoid beds are present and abundant within limited intervals. Both calcareous and non-calcareous shale and siltstone are present here.

Twenty-five miles southwest of Catawba, near the town of Ellett, Montgomery County, the lower Martinsburg is also shaly. The outcrop is located on strike with the Catawba section. On County Road 641 about 0.8 mile east of Ellett, approximately the lower 500 feet of the formation is well-exposed and measureable. Virtually the entire section is silty, calcareous shale. The fresh color is gray to dark gray and weathers olive-gray. Siltstone and very fine to fine-grained sandstone make up about 10 per cent of the sequence, and coquinoid beds less than 1 per cent.

This area is considered to be somewhat anamolous by geologists

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familiar with the region (B. N. Cooper, 1964; C. G. Tillman, personal communication). The Bays Formation here is apparently over 1000 feet thick, several times thicker than most sections in southwestern Virginia; at Catawba the Bays is only about 100 feet thick. The Martinsburg at Ellett contains more shale than at Catawba and much more than in sections to the west. Furthermore, the Juniata Formation is not present in this area nor at Catawba; the Tuscarora Formation rests directly on the upper Martinsburg. The Tuscarora is thinner in this region than to the northeast or west - it is only 10 to 15 feet thick at Catawba and is only 5 feet thick at an outcrop in the community of Fagg, about 1.2 miles east of the Ellett section. The most easterly outcrop of the Martinsburg in this area is at Coyner Mountain in Roanoke County. Here a sequence of dark gray to black shale of unknown thickness has been identified as Martinsburg (Ping-fan Chen, 1959).

A partly covered section transitional between that at Catawba and the more limy sections is exposed at McCoy, Montgomery County, along the old Virginian Railway tracks. The lower 650 feet to 700 feet of the Martinsburg consists primarily of shale and siltstone. However, calcareous siltstone, argillaceous limestone, and coquinoid limestone are more abundant than to the east. Shale occurs mostly as thin interbeds in the lower 150 feet, and coquinoid limestone beds up to 8 inches thick are numerous.

Trenton Formation. The approximate stratigraphic equivalent of the limy facies of the Martinsburg in Lee and Wise Counties is called the Trenton Formation (Butts, 1940). This unit is present in the Hagan section, which is located in the most northwestern outcrop belt of the Ordovician in this region. The name Trenton Formation will be used whenever the rock unit in southwest Virginia is discussed in order to avoid confusion with the time implications of the term Trentonian.

The Trenton Formation is composed of two main types of limestone: a coarsely-crystalline, light gray coquinoid limestone like that found in the Martinsburg and a fine to medium-crystalline grey limestone. These beds all contain some clay and silt, but on the whole they have much less terrigenous material than do the beds in the Martinsburg, the amount being generally less than 30 per cent and perhaps as low as a few per cent. Some beds contain parallel or cross laminations of silt or very fine sand. Silicification is common in these limestones and silica occurs in the matrix, in chert lenses or thin layers and in partially silicified fossils.

Beds range in thickness from 1 inch to generally less than 10 inches, and rarely up to 2 feet. Argillaceous limestones tend to be evenly bedded; coquinoid beds show both irregular and regular surfaces.

Gray calcareous shale interbeds are generally less than 3 inches thick; some are up to 12 inches thick, particularly within the upper 50 feet of the formation. In this upper interval, some shale is brown and olive gray and only slightly calcareous, being very similar to that in the overlying Reedsville Shale.

A one-foot thick bentonite occurs at 72 feet above the base of the formation. A probable bentonite, 7 inches thick, occurs at 31 feet above the base, and another of 2 inches thickness was found 12 feet below the top of the formation. These were discussed in detail by Miller and Fuller (1954, p. 127-128).

Fossils are abundant throughout the Trenton Formation. They can be found in most argillaceous limestone beds, some of which are almost coquinoid. The high concentrations of fossils tend to occur at the top or bottom of beds. The great majority of brachiopods, excluding small round forms such as Zygospira, are disarticulated. The formation contains taxa that are common within the Martinsburg as well as species that occur within the Cincinnati region and are rare in the geosyncline.

The total thickness of the formation at Hagan was measured at 601 feet. It is underlain by a unit considered equivalent to the Carters Limestone of central Tennessee (B. N. Cooper, in Twenhofel and others, 1954). The upper part of this unit consists of finely crystalline shaly limestone and calcareous mudstone with fewer fossils. The base of the Trenton Formation is placed at the lowest coquinoid limestone (photograph of contact in Miller and Fuller, 1954, pl. 22C). The base is probably slightly older than the base of the Martinsburg and is equivalent in age to strata within the Eggleston Formation (see Page 219). The contact of the Trenton Formation with the overlying Reedsville Shale is placed at the top of a 12-foot interval of limestone and at the base of beds composed predominantly of shale (photograph in Miller and Fuller, 1954, pl. 25B). The age of the upper Trenton beds could be late Trentonian, or early Edenian.

Silty Facies. This facies is composed mainly of siltstone and shale (70 to 85 per cent) with thin argillaceous limestone beds, sandy coquinoid beds, and thin sandstone beds. It is well-exposed at Marion where it was measured to be 460 feet thick; the facies is possibly as much as 60 feet thicker here because it is immediately underlain by a covered interval. At Narrows, the facies is 560 feet thick and there the lower half is mostly covered. The silty facies is above the measured section at Rosedale. At Catawba, where the limy facies is absent, the silty facies is 1322 feet thick.

The siltstone and shale are medium gray to dark gray and olive gray in color. They are mostly calcareous, with the olive gray beds being slightly calcareous or noncalcareous. They weather as porous mudstone fragments or silty shale chips of gray or light brown color. Small scale cross bedding or cross laminations are present in many beds. The shale and siltstone are generally micaceous (mostly muscovite) and limonite is common in the siltstone and sandstone, so that some leached beds are "rusty." Beds are usually less than 8 inches thick, rarely as much as several feet thick. Pinching and swelling are common but are in part due to structural deformation of these relatively incompetent beds.

Very argillaceous limestones are thin-bedded and make up only a small proportion of the member. They are more abundant in the western outcrop belts, comprising as much as 20 per cent of the unit in places (Cooper, 1944). Coquinoid limestone beds are common in some intervals but do not constitute over 10 to 15 per cent of the member; they are scarce in the upper 50 to 100 feet. They contain varying amounts of sand and silt and some would be properly termed coquinoid sandstones or siltstones.

Thin sandstone beds, with fine to medium subangular grains are medium gray to olive gray in color, calcareous or noncalcareous, and weather to "rusty" brown porous beds. Sandstone content increases in the upper one-third of the unit.

Most of the Martinsburg at Catawba is similiar to the silty facies in other sections. The lower half of this facies at Catawba contains more argillaceous limestone than the upper half. Sand is more abundant in the 400 feet below the upper member in this section than elsewhere.

Sandy Facies. Everywhere at the top of the Martinsburg is a sequence of gray, greenish-gray and reddish-gray siltstone and sandstone beds with shale interbeds. This interval is consistently differentiable and is referred to as the sandy facies. The thickness in southwestern Virginia varies between about 50 and 165 feet. Siltstone is the most abundant lithology, comprising about 50 per cent of the member. Proportions of sandstone and shale are variable and either may be dominant lithology in intervals up to 30 feet thick.

The siltstone and sandstone weather to an olive-gray or brown color and are commonly "rusty." Beds near the top are red in color, which may be in part due to staining by waters percolating downward from the overlying Juniata redbeds. Most of the sandstone is very fine to medium-grained; coarser grained particles as large as 5 mm are rarely found. Grains are generally subangular; the minor coarse sand and granules are subround to round. Sorting is only moderate in most beds, especially in beds containing medium or coarser sized grains. Cross laminations and cross bedding up to 4 inches thick are common. Mica (mostly muscovite) is generally present. Both slightly calcareous and noncalcareous beds are found, with the latter prevalent in the upper half of the facies. In the slightly calcareous beds, the carbonate may be from dissolution of shells contained therein.

Shale partings and interbeds are gray, greenish-gray, and reddish gray in color, slightly calcareous or noncalcareous, and generally silty or sandy.

A transitional interval exists between the silty facies and the sandy facies, and the contact is arbitrarily placed at a horizon above which there are considerably more sandy beds than below.

In most places, red clastics of the Juniata Formation overlie the Martinsburg. A gradational sequence between these two units includes gray and greenish-gray beds like those in the Martinsburg and

red siltstone and sandstones like those in the Juniata.

Fossils in the sandy facies include some of the species common to the middle member, but the most abundant fossils are the brachiopods Orthorhynchula linneyi and Lingula and small gastropods (Liospira?). Beds containing Lingula and Liospira? alternate with those containing Orthorhynchula and beds apparently barren of fossils and similar to the overlying Juniata redbeds. The top of the Martinsburg is placed at the top of the highest fossiliferous and/or calcareous beds.

The Martinsburg is directly overlain by white sandstones of the Tuscarora Formation in the Catawba section. Mostly reddish and greenish-gray beds containing Lingula and Liospira? are found below the white orthoquartzite. The actual contact is obscured by float, but there appears to be no angular discordance.

Reedsville Shale. In Lee and Wise Counties, the shaly sequence overlying the Trenton Formation and underlying the red and gray mudstones of the Sequatchie Formation is called the Reedsville Shale, named by Ulrich (1911) from exposures near Reedsville, Pennsylvania, and recognized in Virginia by Butts (1940). This unit is considered to be the lithologic equivalent of the silty facies. It is well exposed in the Hagan section, where it was measured as being 403 feet thick. Approximately the lower 50 feet is mainly shale, above which shale and limestone occur in about equal proportions. Gray calcareous siltstone and sandstone beds are common but make up a small percentage of the formation.

The shale is gray to greenish-gray and olive gray in color and weathers light brown and gray. Most of the shale is calcareous; the olive gray shale intervals are only slightly calcareous or noncalcareous. Except for the lower part of the formation, most shale intervals are less than 8 inches thick. These commonly contain thin beds of calcareous siltstone or argillaceous limestone.

Limestone in the Reedsville Shale is like that found in the Martinsburg; gray to dark gray, finely crystalline argillaceous or sandy limestone and light gray to gray, coarsely crystalline coquinoid limestone.

Fossils are abundant throughout the Reedsville at Hagan; these include species common in the Martinsburg as well as forms found only to the north and west. Orthorhynchula linneyi is abundant in the uppermost beds, but no specimens of Lingula were found in this section. The age of the Reedsville is Edenian to early Maysvillian.

The lower contact with the Trenton Formation is at the sharp change from a limestone sequence to shale beds. The contact with the overlying Sequatchie Formation is likewise sharp and is placed at the change from Orthorhynchula-bearing shaly limestone to red and gray calcareous mudstone (photograph in Miller and Fuller, 1954, pl. 27A).

Interpretation of Physical Stratigraphy

The limy facies of the Martinsburg and the Trenton Formations comprise a calcareous lithosome that is present in the central and western outcrop belts. This is a shallow geosynclinal facies transitional between cratonic carbonate strata of the midcontinent and predominantly terrigenous deposits of eastern areas. Currents were sufficient to distribute fine grained material throughout the basin and occasionally strong enough to cause accumulation of large amounts of shell debris and to erode the bottom sediments. Increase of terrigenous material toward the northeast indicates that the primary source lay in that general direction; this interpretation is supported also by the nature of the overlying facies and the succeeding formations.

The silty and sandy facies represent an increased supply of detritus, possibly associated with migration of the source area in a westerly or southwesterly direction. The coarser-grained sandy facies suggests further shallowing and westward migration of the shoreline. The Reedsville Shale was formed in an area farther removed from the source, judging from the greater amount of carbonate contained therein as well as the absence of the sandy facies. The red and gray calcareous mudstone of the Sequatchie Formation may be considered as the offshore equivalent of the littoral and/or nonmarine Juniata redbeds.

AGE OF THE MARTINSBURG

The Martinsburg ranges from Middle Trentonian to Maysvillian, or from late Wilderness to Barneveld, in age. A detailed study of brachiopods and trilobites (Walker, 1967) corroborates the age range given by Twenhofel and others (1954). Interpretations of the biostratigraphy of the Martinsburg will be published in the future.

Both the base and top of the Martinsburg appear to be time-transgressive. In a study of bentonites in the basal Martinsburg and underlying units, Rosenkrans (1936) considered that the contact (as defined herein) varies in age, becoming younger in the northerly sections. This age difference is not large, for the contact is always found either a little above or below bentonite V-13 according to Rosenkrans. Brachiopod and trilobite taxa are not suitable to check such detailed correlation. The age of the base probably does vary, as the lowest coquinoid bed in any one section is almost certainly not the same as in other sections.

The gradational nature of the upper contact, with interbedding of fossiliferous beds with nonfossiliferous beds of typical Juniata lithology, suggests fluctuating conditions at the margins of two lithotopes. Considering distribution of lithofacies, the top of the Martinsburg probably becomes younger in a general westerly direction.

Detailed correlation of the Martinsburg with type sections of the Middle and Upper Ordovician in North America is at present impossible. The possible equivalence of the type Eden of the Cincinnati area with part of the type Trenton of New York has been discussed by Schopf (1966). Weiss and others (1965) stated that the standard fossil zones of the Upper Ordovician were founded essentially on data from the Cincinnati region, and that the relationship between faunal zones and rock facies is not fully known; the possible result is that the "index" species used for correlation may have little time value. In Figure 1, the approximate time-stratigraphic boundaries are given with respect to the old Trentonian-Edenian-Maysvillian concept.

APPENDIX

Locations of Stratigraphic Sections

In all sections, outcrops were measured with a Jacob staff and covered intervals by the tape and compass method.

1. Catawba Section: located along US Highway 311 where it crosses the western flank of Catawba Mountain, in Roanoke County. It is approximately 9 miles north of Salem and one half mile east of the town of Catawba. The formation crops out from just above the base of the mountain up to the crest of the ridge. Exposures are present along the new U. S. Highway 311 and remnants of the former highway. About 20 per cent of the Martinsburg is covered.
   1. Tuscarora Formation
   2. Martinsburg Formation ------ 1453 feet thick
      Sandy facies ------------- 131 feet
      Silty facies ------------- 1322 feet
   3. Bays Formation
      Discrepancies between my measurements and earlier ones by Woodward (1932: 1028 feet) and Ryan (1963: 1361 feet) are probably due to the fact that much of the lower 950 feet of the formation is covered.
2. Narrows Section: located on the northeast side of U. S. Highway 219 about 1.5 miles north of the town of Narrows, Giles County. It is in the water gap cut by the New River through Peters Mountain. About 20 per cent of the Martinsburg is covered.
   1. Juniata Formation
   2. Martinsburg Formation ------ 1283 feet thick
      Sandy facies ------------- 55 feet
      Silty facies ------------- 560 feet
      Limy facies ------------ 668 feet
   3. Eggleston Formation

3. Marion Section: exposed along Virginia Highway 16 on the northwest slope of Walker Mountain, 7 to 8 miles north of the intersection of Highway 16 and U. S. Highway 11 at the town of Marion, Smyth County. About 8 per cent of the Martinsburg is covered.

   1. Juniata Formation
   2. Martinsburg Formation ------ 1124 feet thick
       Sandy Facies ------------- 164 feet
       Silty Facies -------------- 460 feet
       Limy Facies ------------ 500 feet
   3. Bays Formation

4. Rosedale Section: located on Virginia Highway 80 approximately 0.8 mile northwest of the crossroads at Rosedale, Russell County. The lower 700 feet of the Martinsburg is well-exposed on the southeastern flank of a syncline. Folding and poor exposures above this level necessitated limiting the interval measured. All of the lower 700 feet belong to the limy facies. The Martinsburg is underlain by the Eggleston Formation.

5. Hagan Section: the Trenton and Reedsville Formations are well-exposed along the Louisville and Nashville Railroad switchback at Hagan, Lee County. The switchback is parallel to County Road 621, and the outcrop is about 0.5 mile north of U. S. Highway 58. The entire Reedsville Shale is well-exposed. About 170 feet of the upper Trenton Formation is mostly covered.

   1. Sequatchie Formation
   2. Reedsville Shale ------------ 403 feet thick
   3. Trenton Formation ---------- 601 feet thick
   4. "Carters Limestone equivalent"

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CONFIGURATION OF THE HOLOCENE PRIMARY BARRIER

CHAIN, OUTER BANKS, NORTH CAROLINA

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ABSTRACT

The presence of a well-developed soil zone under parts of the Outer Banks and its absence under other parts permit a reconstruction of the initial shape of the coastline during the latter part of the Holocene rise in sea level to near its present level.

The primary barrier lay to the east of the present coast in places and to the west of the present shore in others. The present coastline has developed from this original shape.

INTRODUCTION

Barrier islands, or their counterpart mainland beaches, are present on many of the stable coasts of the world. Because of their location at the juncture of the sea and land, both scientific and general public interest has been high. This has resulted in much work on evaluating the changes, on reversing or delaying the normal evolution, on the formation of barrier islands, and the evolution of such features.

A preliminary study of the samples from 33 bore holes on the Outer Banks of North Carolina, both north and south of Cape Hatteras, has recently been completed (Figure 1). The interpretation of the environment of deposition of the samples permits a delineation of the initial shape of the primary barrier formed during the latter stages of the Holocene rise of sea level, subsequent changes in this barrier, and development of other barriers.
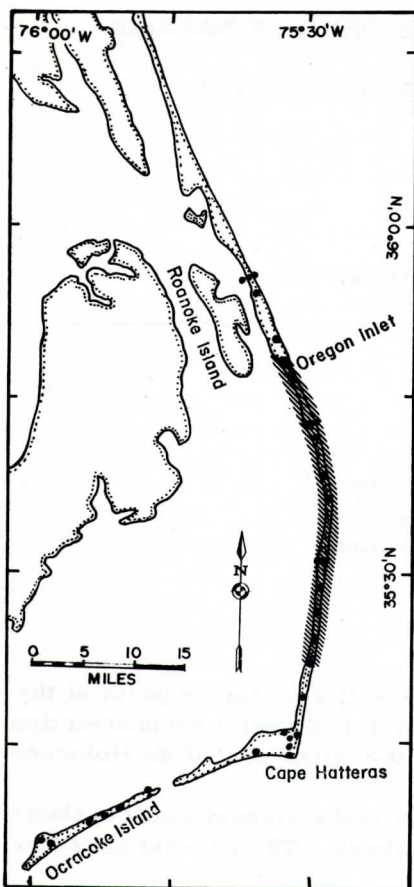
Figure 1. Map of part of the Outer Banks, North Carolina. Location of the borings is shown by black dots. Area where the soil zone is present is cross-hatched.
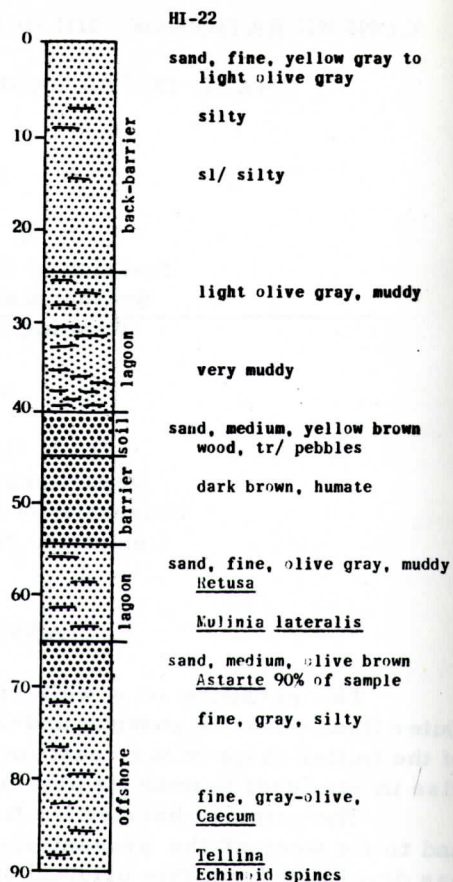

Figure 2. Log of boring HI-22, located on the back-barrier on the south shore of Oregon Inlet. Elevation of the surface is 5 feet.

Acknowledgments

STRATIGRAPHY

The sequence of sediments expected in a bore hole drilled through a barrier undergoing retrogression is for barrier sediments to overlie sediments deposited in a lagoon. Of the thirty-three holes drilled on the Outer Banks, twenty-seven encountered this expected sequence. Two of the remaining holes were located in areas where inlets have been known to occur within historical times. The last four, which did not encounter the normal stratigraphic sequence, were drilled in the area of Cape Hatteras.

Several types of material were encountered below this barrier-lagoonal sequence. The borings around Cape Hatteras went either from barrier sediments or lagoonal sediments into what is interpreted as marine sediments. The interpretation is based on contained fauna, texture of the sediment, and regional distribution of the sediment types. Where shell material was present, the interpretation relied heavily upon the fauna.

The borings on Ocracoke Island and north of the north shore of Oregon Inlet penetrated, below barrier sediments, material that is representative of low-energy depositional conditions. Part of this section undoubtedly represents a lagoonal environment while the lower part may represent a partially-barred embayment.

Eleven of the fourteen holes drilled along the stretch of coast from 10 1/2 miles north of Cape Hatteras to Oregon Inlet encountered a reddish brown, yellow, or buff colored sediment below the gray lagoonal sediments (Figure 1). The change in color is striking.

Near the southern end of its area of occurrence and again near Oregon Inlet, the bright colored sediments consist of medium to coarse sand with pebbles (Figure 2). The coloration is due to iron stain. Below the iron stained sand at Oregon Inlet is a humate coated sand. Both sands appear to have the same textural properties. Iron staining and humate impregnation are nearly mutually exclusive in occurrence in the samples. The humate may be a lower (B) soil horizon although soil development in the immediate area of occurrence is not necessary for the formation of humate, because of its mobility (Swanson and Palacas, 1965).

Between the two extreme points of occurrence, silty clays and sandy silts make up the material that is iron stained. In none of these samples were there any indications of calcareous shell material, either in the medium sands, the silty clays, or the sandy silts.

Where the bright colored sediment consists of sand, depth is 22 feet (-16 feet) at the south end and 23 feet (-17 feet) near Oregon Inlet.

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The depth increases from both north and south, reaching a maximum of about 50 feet (-41 feet) near the center of its area of occurrence.

The three holes in which the iron stained sediments were not present are believed to have been located in an area where inlets have removed the normal section. One of the three was in an area where an inlet is shown on the 1890 edition of the Coast and Geodetic Survey chart. The other two had sediments quite similar to the material found in known inlet-fill deposits.

INTERPRETATION

This area, as well as the adjoining continental shelf, was exposed to subaerial weathering during the time span between about 40,000 years BP and 6,000 years BP, depending on which sea level curve is used. This span is the time of the maximum Wisconsin glaciation and prior to the return of the sea to near its present level in the Holocene.

Iron stained sediments are prevalent on the middle and outer continental shelves and are considered to have been deposited and/or weathered during this lower stand of sea level. Some evidence of this period of weathering should exist under the Outer Banks unless it has been removed by later erosion. The material upon which the weathering effects are imposed would obviously predate the Wisconsin sea level regression.

The iron stained sediments underlying the barrier islands between Oregon Inlet and about 10 1/2 miles north of Cape Hatteras are interpreted as part of a soil zone formed at this time of lower sea level. Consideration was given to the possibility that this iron stained zone could be a coloration caused by ground water action. This was rejected because it is believed that ground water effects would occur at nearly a constant elevation, if caused by water table effects, or follow similar porous material if caused by water in an aquifer. The iron stained zone under the barrier islands does neither, ranging in elevation from -16 feet to -41 feet and involving pebbly sands to silty clays.

Emergence of the land exposes the sediments to subaerial weathering and erosion while submergence subjects the deposits to reworking (Colquhoun, 1965, p. 28-31; Hoyt and Henry, 1967, p. 84). Thus, any extended area where the soil is absent suggests removal rather than non-formation inasmuch as the entire coast was equally exposed to weathering. The areas around Cape Hatteras, Ocracoke Island, and north of Oregon Inlet do not have the soil zone developed. Absence from extended areas is most easily attributable to marine processes.

The sea, then, must be excluded, during the later part of the Holocene rise, from an area where the soil zone is present and permitted to cover the areas where the soil zone is absent. This permits the reconstruction of the location of the primary barrier that formed
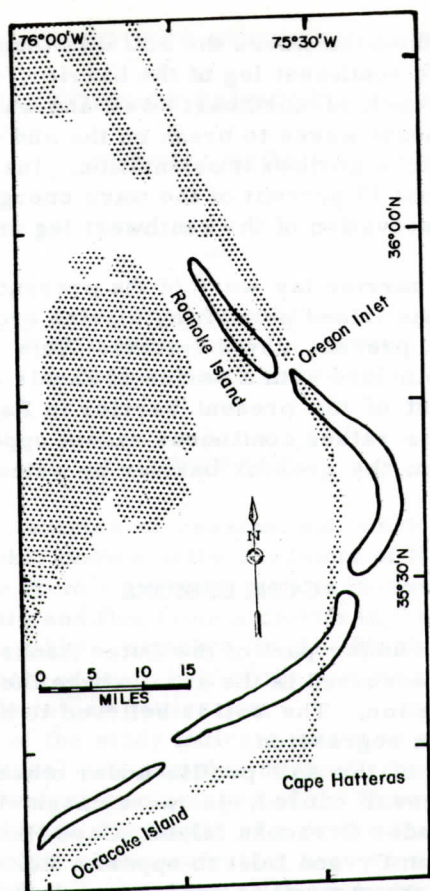
Figure 3. Postulated configu-
ration of the primary Holo-
cene barrier along part of
the Outer Banks.

when sea level approached the present level, near the end of the Holo-
cene sea level rise (Figure 3).

The primary barrier (and shoreline) lay west of the present
coast north of Oregon Inlet and is believed to have coincided with Roa-
noke Island, the sediments of which display characteristics of material
deposited in various barrier environments. The primary barrier would
cross the present coast line near Oregon Inlet and about 10 1/2 miles
north of Cape Hatteras or at the two ends of the area of occurrence of
the soil zone.

This barrier extended an unknown distance seaward. The exact
distance will undoubtedly never be known because of removal of the
necessary evidence by marine processes. It is believed that the south-
west leg may coincide with a series of shoals that join the coast near

the southern end of the area where the soil was found. The orientation of the shoals, and the southwest leg of the barrier, is nearly normal to the direction of approach of northeast swell and waves. This orientation causes the northeast waves to break on the end of the shoals rather than over a shoal with a northwest orientation. Inasmuch as northeast waves contribute about 37 percent of the wave energy to the coast, this accounts for the preservation of the southwest leg and destruction of the northwest leg.

The primary barrier lay west of the present coast around Cape Hatteras and Ocracoke Island permitting marine erosion to remove the soil zone under the present day barriers. This primary barrier is postulated to have coincided with a series of shoals and islands lying about 3 1/2 miles west of the present barrier in Pamilico Sound. The shoal island trend is rather continuous except opposite present inlets and is separated from the present barrier by water depths of up to 13 feet.

CONCLUSIONS

A soil profile under part of the Outer Banks indicates that marine erosion has not occurred in the area, where the soil profile is present, since its formation. The soil is believed to have developed during the last Wisconsin regression.

The absence of the soil profile under other parts of the Outer Banks indicates removal caused, in most cases, by marine erosion. The soil is absent under Ocracoke Island, Cape Hatteras, and most of the barrier chain from Oregon Inlet to opposite Roanoke Island.

The areas where the soil is present or absent permit a delineation of the shape of the primary barrier formed during the latter stages of the Holocene sea level rise. This primary barrier did not coincide with the present coastline. In places the barrier lay east of the present coast while in others it lay to the west.

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DETERMINATION OF SOURCE ROCKS FROM HEAVY MINERALS

IN RESIDUAL SOILS OF SAPROLITE PIEDMONT OF

NORTH CAROLINA

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ABSTRACT

Twenty-five samples of residual soil were taken from five plutonic and metamorphic bodies in the Piedmont of North Carolina. Five of the samples were from a gneissic granite in Vance County, five from an albite granodiorite and five from a diorite in Orange County, and ten from a gabbro-metagabbro complex in Mecklenburg County. Heavy minerals were recovered from half of the samples by panning and from the other half by laboratory separation.

The results of the study indicate that it is possible to differentiate between the residual soil samples derived from different rock types on the basis of the weight percentages of the heavy minerals and/or by certain minerals that are diagnostic of a particular rock type. Panning is as effective a method of concentrating heavy minerals as the laboratory technique, and for zircon appears to be more effective.

INTRODUCTION

Purpose

Crystalline rocks in the Piedmont of the southeastern United States are generally obscured by a layer of residually weathered material known as saprolite. This residuum commonly extends from the surface to depths greater than 50 feet and may reach 200 feet. The present investigation was undertaken to determine if it is possible, through the use of heavy accessory minerals in residual soil overlying saprolite, to differentiate between several plutonic and metamorphic bodies from widely separated places in the Piedmont of North Carolina. Also, this study allowed a comparison of the effectiveness of panning and of the laboratory separation in the recovery of heavy minerals from residual soil.

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Acknowledgements

The author would like to thank personnel of the Department of Geology, University of North Carolina, for their help in this investigation, particularly V. I. Mann for his suggestions concerning the problem. Thanks are also due to William C. Overstreet for his extensive technical review of the manuscript. Appreciation is extended to Charles Cazeau for his advice on the analytical methods.

Previous Work

Saprolite differentiation. Saprolite was defined by George F. Becker in his study of Appalachian goldfields (1895, p. 289) as a general name for thoroughly decomposed, earthy, but untransported rock. Derby (1891, p. 200) was one of the first American geologists to recognize that the saprolite of plutonic and metamorphic rocks may be correlated with the source rock by the presence of certain characteristic minerals or by peculiarities such as the form or color of a certain mineral species. Pardee and Park (1948, p. 25) noted work done on saprolite and residual soil in Georgia by L. M. Prindle and W. A. White in 1935, whose study indicated that the examination of residual minerals may often help to identify the underlying rocks.

More recently Hurst (1951, p. 244) used heavy minerals from saprolite to differentiate among some crystalline rocks in Georgia. He drew two conclusions (Hurst, 1951, p. 261): (1) each rock type in the Crystalline Belt of Georgia contains a distinctive assemblage of heavy minerals, and (2) saprolite derived from different rocks may be distinguished by the differences in the assemblages. Overstreet, Yates and Griffiths (1963, p. 2) made a distinction between saprolite and residual soil in their study of heavy minerals in the Shelby Quadrangle, North Carolina. Their conclusions (1963, p. 27) indicated that the identification of the source rock from the heavy minerals in the saprolite was uncertain, and in only two cases were the mineralogical assemblages characteristic of the source rock. Reports of similar research may be found in Yoho (1952), Mertie (1953; 1954), and Overstreet (1962).

Panning vs. laboratory technique. Only two papers comparing the effectiveness of the panning and heavy liquid separation techniques were found in the author's search of the literature. These articles were by Smithson (1930) and Ewing (1931). However, several articles (Mertie, 1954; Theobald, 1957; and Overstreet, 1962) have been published on the effective use of the gold pan.

Smithson (1930, p. 135) published the results of his comparative study on panning vs. heavy liquid (bromoform) separation as a series of ratios. He found that for tourmaline, staurolite and garnet the ratio of the heavies panned and separated by bromoform to the heavies elutriated and separated by bromoform was less than one. This indicated that the laboratory technique preserved a higher concentration of these

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minerals than did the panning technique. With rutile the ratio was one, indicating that panning and elutriating were equally effective. With zircon the ratio was 1.68, indicating that panning preserved a higher concentration of zircon than did the laboratory separation. The high zircon ratio in this particular study may be an indication of the loss of some of the lighter heavy minerals, or of the ineffectiveness of the bromoform separation.

Ewing's (1931, p. 137) study on the comparison of panning and the laboratory technique resulted in a ratio closer to one for zircon, rutile, staurolite, kyanite and tourmaline, than did Smithson's (1930) study. The more effective panning results in Ewing's study may be the result of a different panning technique, since Ewing used a nine inch evaporating dish and Smithson used a grooved pan.

Results of both these studies indicate that as the specific gravity of the minerals increases, the greater the recovery in the pan. Also, to obtain the greatest concentration of light heavy minerals, heavy liquids should be used.

GEOLOGIC SETTING

Locations

Samples of residual soil over a gneissic granite, albite granodiorite, diorite, gabbro, and metagabbro were collected randomly from three counties in the Piedmont physiographic province of North Carolina (Figure 1). Residual soil over a gneissic granite was sampled in the south-central to north-east corner of Vance County. Soil samples over an albite granodiorite and a diorite were taken from the southeastern corner of Orange County. Residual soils over the gabbro and metagabbro were sampled in the southern part of Mecklenburg County.

Gneissic granite. The gneissic granite extends north-north-westward as a mass about 4 miles wide in the central part of Vance County. Monazite bearing granite crops out east of the gneissic granite, and to the west a narrow strip of phyllite separates the gneissic granite from albite granodiorite. The gneissic granite consists mainly of quartz, microcline, orthoclase, plagioclase, biotite and muscovite with accessory apatite, zircon, titanite, magnitite or ilmenite, epidote and allanite (Parker, 1963, p. 8).

Albite granodiorite. The albite granodiorite is located in the southern part of the Chapel Hill Quadrangle (Mann and others, 1965), Orange County. The albite granodiorite is composed of plagioclase, microcline and orthoclase, and hornblende. The accessory minerals include magnetite and ilmenite, epidote, chlorite, biotite, sphene and zircon (Hayes, 1962, p. 37). The albite granodiorite was probably intruded into all the other rock sequences in the area, as it contains xenoliths of all the other rock types (Mann, and others, 1965, p. 18).

NORTH CAROLINA
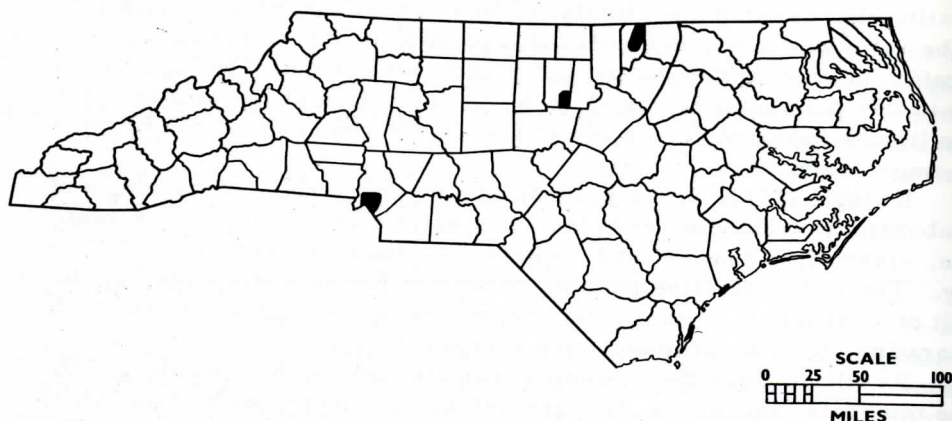
Figure 1. Index map of sample locations in Vance, Orange and Mecklenberg Counties.

Diorite. The unaltered diorites in Orange County were found by Hayes (1962, p. 20) to consist mainly of plagioclase and hornblende, with minor quartz, microcline and microperthite. The common accessory minerals are epidote, magnetite, and ilmenite, with some pyrite, sphene, apatite, and biotite.

Gabbro and metagabbro. A gabbro-metagabbro complex crops out within the Charlotte Belt in southeastern Mecklenberg County, North Carolina, and northern York County, South Carolina. The meta-gabbro is older than the gabbro as evidenced by numerous xenoliths of meta-gabbro within the gabbro (Hermes, 1966, p. 7). Outcrops of the complex are scarce, but the soil cover over part of the gabbro is only a few feet thick and residual boulders are fairly abundant (Hermes, 1966, p. 6). The metagabbro has been more extensively weathered than the gabbro, and the soil over the metagabbro generally contains greater amounts of weathered biotite.

The gabbro is mainly composed of plagioclase, hornblende, augite, hypersthene, olivine and biotite. The common accessory minerals include magnetite, ilmenite, apatite, sulfides, and locally calcite and a green spinel. The metagabbro consists of plagioclase, hornblende, clinopyroxene and biotite; accessories are magnetite, ilmenite, orthopyroxene, epidote, chlorite, zircon, sphene, calcite, and quartz. The metagabbro can be distinguished chemically from the gabbro by a higher content of SiO_2, TiO_2, Na_2O, K_2O and possibly total Fe, by lower MgO, and by possibly lower CaO content (Hermes, 1966).

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ANALYTICAL METHODS AND PROCEDURES

Field Methods

One sample of 17. 6 kilograms and four of 4. 4 kilograms were taken at random from the residual soil above each of the rock types, gneissic granite, albite granodiorite, diorite, gabbro, and metagabbro, described in the preceding section. Sample selection was made on the basis of the following geologic maps: Hermes (1966), Mann and others (1965), and Stuckey (1958). None of the 25 samples collected was obtained from localized depressions and most were located on topographic highs.

The samples were taken from a depth of 2 to 3 feet through use of an auger-type post hole digger. All samples were from below the A soil horizon. The soil sample was passed through a Tyler No. 5 mesh screen (3. 96mm) to remove coarse material; that finer than 3. 96mm was retained for further processing, and the coarse fraction was discarded.

Laboratory Procedure

The 17. 6 kilogram sample was split into eight fractions and the 4. 4 kilogram samples were split into two fractions on a Jones 1/4 inch sample splitter. Half of the samples from each locality were elutriated in a pail, and the clay and silt sized particles were decanted. The sand sized portion of the sample was dried and saved for later processing. The other half of the samples from each locality was panned in a large metal wash tub using a standard 14-inch gold pan, following the procedure recommended by Mertie (1954, p. 648) and demonstrated to the author by Henry Bell III. Panning of each sample was terminated when approximately 50 percent of the sand size fraction was composed of dark minerals. These panned samples were then dried and put aside for further processing. A flow sheet outlining the laboratory procedure can be observed in Figure 2.

RESULTS

The results of this study are discussed in two sections: (1) the determination of the source rocks from heavy minerals contained in the residual soil and (2) the comparison between panning and the laboratory technique. The section on source rock differentiation has been subdivided into two parts. The first part compares rock types by the weight percentages of particular minerals in the concentrate, and by the weight percentages of the total heavy minerals in the sample. The second part compares the rock types on the basis of the combined effect of four minerals common to all samples. To compare the relative

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Field sample (4.4-17.6 kilograms)
|
Sieved (3.96mm)
```
         ┌─────────────────┴─────────────────┐
Coarse fraction                        Fine fraction
(>3.96 mm discarded)                    (<3.96 mm)
                                            |
                                          Dried
                                            |
                                          Split
                    ┌───────────────────────┴───────────────────┐
            Weighed (2200g.)                            Weighed (2200g.)
                    |                                           |
              Elutriated                                      Panned
```
Both the elutriated and panned portions of the samples
were treated identically as follows
```
                              Dried
                                |
                             Sieved
         ┌──────────────────────┴──────────────────────┐
  Coarse fraction                               Fine fraction
  (>.149mm. stored)                             (.149-.061mm.)
                                                     |
                                          Bromoform separation
                                              (s.g. 2.89)
         ┌────────────────────────────────────────────┐
   Heavy fraction                               Light fraction
         |                                             |
  Washed (acetone)                                  Washed
         |                                             |
       Dried                                         Dried
         |                                             |
  Iron oxide removed                               Weighed
     (oxalic acid)                                    |
         |                                          Stored
      Weighed
         |
  Magnetite removed
   ┌─────────┴───────────────────────────┐
Magnetic fraction              Non-magnetic fraction
   |                           Microsplit (≈1000gr.)
Weighed                                |
   |                        Grains mounted in Canada balsam
Stored                                (n=1.537)
                                       |
                          300 grains counted per slide
```

Figure 2.   Flow sheet for the preparation
of samples for optical identification.

effectiveness of panning and laboratory separation techniques, the re-
sults obtained for each mineral by panning have been compared with the
results for that same mineral obtained by the laboratory technique.

Source Rock Differentiation

The weight percentages of the heavy minerals vary widely among
samples of residual soil over the same rock type (Figures 3 and 4).
These large variations among sample locations may be the result of
mineralogical variations within the source rock, differences in the
depth of weathering over the bodies, differential weathering of the min-
erals present, and the degree of transportation that the weathered

Weight Percent of Total Heavy Minerals
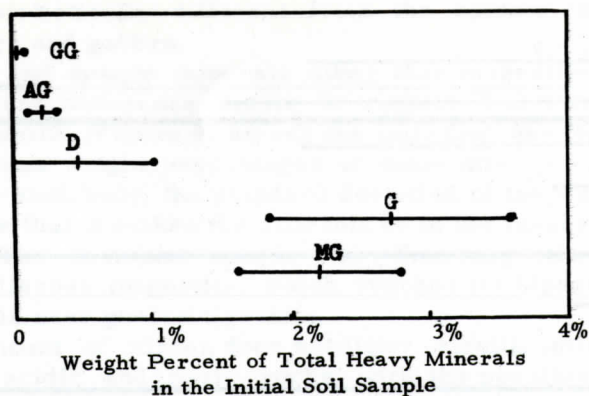in the Initial Soil Sample

Figure 3. Mean and one standard deviation for
the weight percentages of the total heavy
minerals in the initial soil sample. Sym-
bols are as follows: GG-gneissic granite,
AG-albite granodiorite, D-diorite, G-gab-
bro, MG-metagabbro, l-mean,   -marks
the boundary of one standard deviation.

material may have undergone. Nevertheless, sufficient differences
exist in the weight percentages of the heavy mineral suites among rock
types to permit differentiation among these rock types through the use
of heavy minerals in residual soil. For some rock types a few minerals
are specifically characteristic.

Total heavy minerals. The weight percentages of the total heavy
minerals in the 2200 gram soil samples were found to be characteristic
of the acidic and basic rocks. Residual soils over saprolite of gabbro
and metagabbro have a higher percentage of total heavy minerals than
the soils over the diorite, albite granodiorite, and the gneissic granite
(Figure 3). In nine out of ten samples from the gabbro-metagabbro
complex the total concentrate exceeded 1 percent of the weight of the
total soil sample, but in 14 out of 15 samples taken over the acidic
rocks the total weight of the heavy minerals did not exceed 0.6 percent
of the original sample. In addition, the five concentrates from the
gneissic granite in Vance County did not exceed 0.1 percent of the
weight of the original samples; therefore, gneissic granite could be dif-
ferentiated from albite granodiorite on this basis alone.

Amphibole. The concentration of amphibole in the residual soils
cannot be used to differentiate rock types except gabbro (Figure 4).
Soils over gabbro have the highest concentration of amphibole. Although
amphibole has a rather large standard deviation in the soils from gabbro
it does not overlap any other soils; therefore, the percentage of amphi-
bole can be used to differentiate soils over gabbro from soils over the
other rock types. The residual soils of diorite and metagabbro also
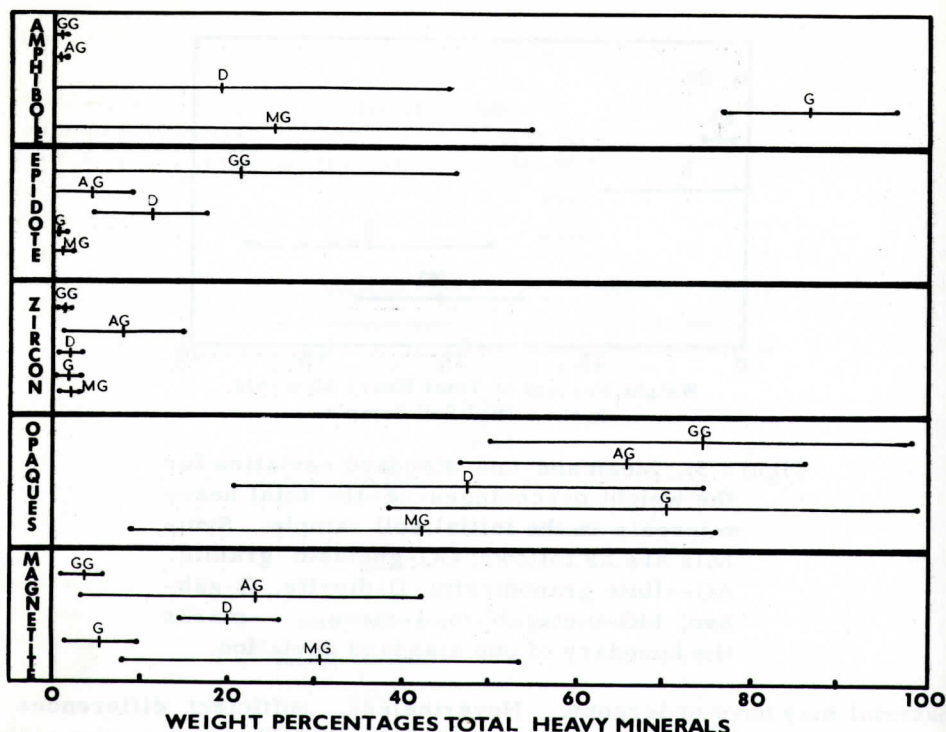
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Figure 4. Mean and one standard deviation for the weight percentages of five of the most abundant heavy minerals involved in this study. Symbols are as follows: GG-gneissic granite, AG-albite granodiorite, D-diorite, G-gabbro, MG-metagabbro, 1-mean,   marks the boundary of one standard deviation.

have high percentages of amphibole, but rather large standard deviations in percentages of amphibole make it difficult to differentiate the soils over diorite from those over metagabbro, and also to differentiate those soils from the ones over gneissic granite and albite granodiorite.

Epidote. Residual soils from the acidic and basic groups of rocks could possibly be differentiated on the basis of their content of epidote (Figure 4). Very low mean percentages of epidote characterize soils taken over the metagabbro and gabbro, but residual soil over the more acidic rocks tends to have high mean percentages of epidote. However, the standard deviation for the samples over each rock type is so large that the difference in epidote content between the metagabbro, gabbro, gneissic granite and albite granodiorite is statistically insignificant. The standard deviation for the epidote content of the diorite, although it too is quite large, does not overlap the standard deviation for epidote in residual soil over metagabbro and gabbro. Therefore, one can say that the epidote content in the residual soil over the diorite

is higher and statistically different from the epidote content of soils over metagabbro and gabbro.

Zircon and opaque minerals other than magnetite. No statistically significant difference exists in content of zircon and opaques other than magnetite (Figure 4) among the soils from the five rock types. Although the mean weight percentages of these minerals are different in the soil over each body, the standard deviation of the weight percentages is so large that it makes the differences in the means statistically insignificant. The dominant opaque other than magnetite is either ilmenite or titaniferous magnetite, which reaches its highest mean percentages in soils over gneissic granite.

The amount of zircon does not differ significantly in residual soils over the acidic and basic rocks, with the possible exception of soils from albite granodiorite, which have a higher mean percentage of zircon than other soils. However, the high mean percentage of zircon in soils from the albite granodiorite (Figure 4) probably reflects the extremely high concentration of zircon in only two of the five samples; therefore, it should not be regarded as being significant.

Magnetite. The concentration of magnetite in the soil samples could be used to differentiate the metagabbro from the gneissic granite, and to differentiate the diorite from the gabbro and gneissic granite (Figure 4). Soil from the metagabbro has the highest mean weight percentage of magnetite; despite the large standard deviation, it is significantly different from the percentage of magnetite in soils from gneissic granite. The percentage of magnetite in soils from diorite occupies a position between those of soils over gneissic granite, gabbro and metagabbro. Although the percentage of magnetite in soils over diorite is not statistically different from that of soils over metagabbro, it is significantly greater than that of soils over gabbro and gneissic granite. Therefore, the residual soils on diorite could be differentiated from those over gabbro and gneissic granite on this basis. However, in the other soils, magnetite displays such a large standard deviation that identification of the soils by magnetite alone is most uncertain.

Other minerals. Some heavy minerals not shown in Figure 4, but observed as trace amounts, were found to be quite diagnostic of the rock types from which the host soils were derived. Actinolite was found to be quite characteristic of many of the samples of residual soil over gabbro and metagabbro. Muscovite and transparent rutile were observed in most of the samples of residual soil from the gneissic granite. Large concentrations of coarse sand size muscovite were also observed in all of the gneissic granite samples when they were elutriated and panned.

Statistical interpretation. A statistical interpretation of the distributions of those minerals that appear in all the residual soil samples (magnetite, opaques other than magnetite [mainly ilmenite], amphiboles and epidote) was made by recalculating the weight percentages to 100 percent. Figures 5 and 6 are diagrams showing the composition formed
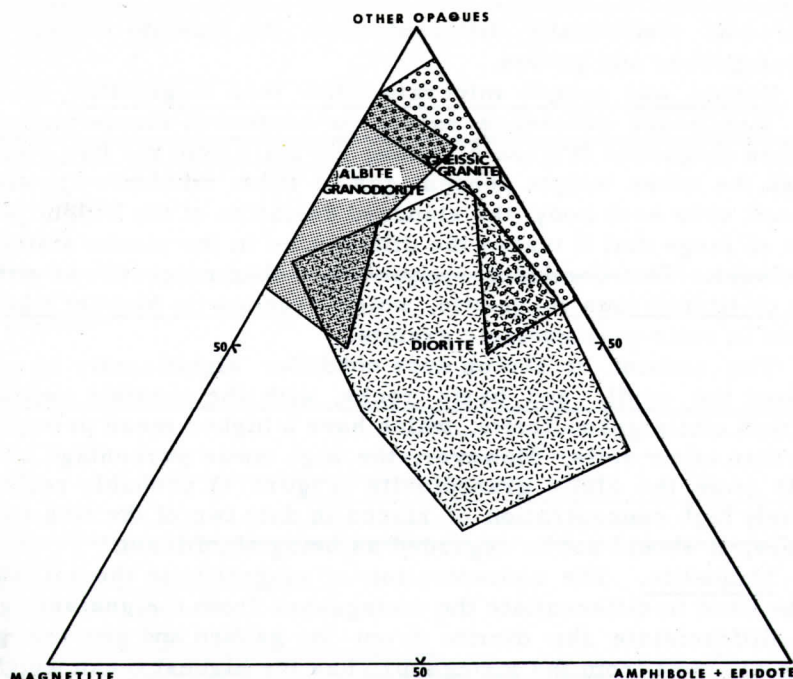
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Figure 5. Diagram of minerals in residual soils over albite granodiorite, gneissic granite and diorite based on the mean and one standard deviation (Table 1) for the concentrations of magnetite, opaques other than magnetite and amphibole + epidote.

by the means and standard deviations (Table 1) for these minerals.

Table 1. The Means and Standard Deviations of the Magnetite, Opaques Other than Magnetite, and Amphiboles+Epidote, Recalculated to Total 100 Percent.

Rock Type	Magnetite		Opaques other than magnetite		Amphiboles and Epidote	
	Mean	S. D.	Mean	S. D.	Mean	S. D.
Gneissic Granite	3. 4	+2. 4	74. 9	+26. 2	21. 7	+25. 0
Albite Granodiorite	24. 0	+19. 4	71. 3	+21. 8	4. 7	+4. 9
Diorite	20. 3	+17. 7	48. 5	+27. 6	31. 3	+30. 2
Gabbro	5. 3	+4. 1	68. 5	+33. 0	26. 2	+31. 0
Metagabbro	31. 1	+25. 6	42. 5	+34. 5	26. 4	+28. 7

OTHER OPAQUES

GABBRO

50

50

METAGABBRO

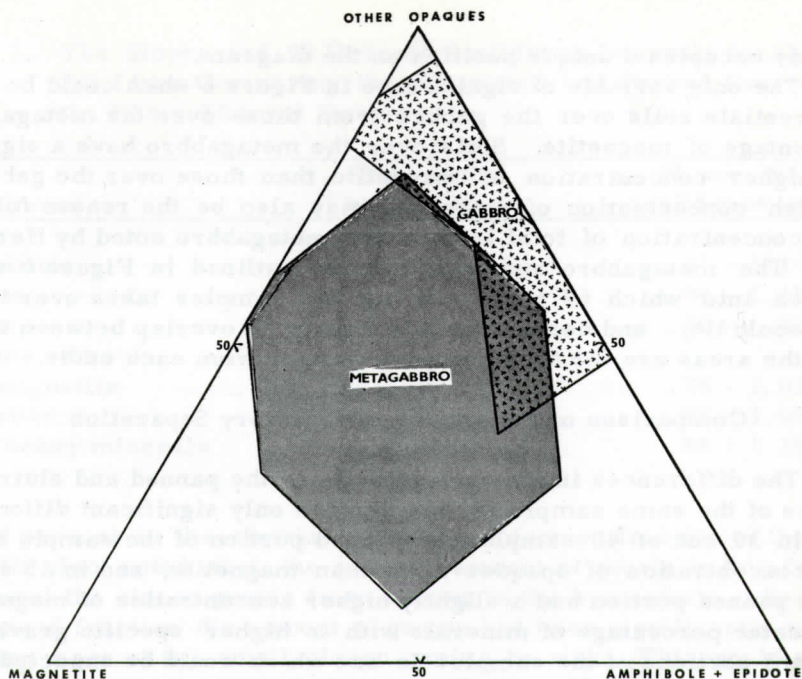MAGNETITE   50   AMPHIBOLE + EPIDOTE

Figure 6. Diagram of minerals in residual soils over gab-
bro and metagabbro based on the mean and one
standard deviation (Table 1) for the concentrations
of magnetite, opaques other than magnetite, and
amphibole + epidote.

It is difficult to differentiate between the acidic and basic rocks
on the basis of the heavy minerals in Figures 5 and 6. The large over-
lap in the standard deviations shown in Figures 5 and 6 makes any state-
ment on the differences in mineralogy of soils between acidic and basic
rocks statistically insignificant. However, Figures 5 and 6 show that
the soils over the various acidic rocks can be differentiated among
themselves, and the soils over the two basic rocks can also be dif-
ferentiated between themselves on the basis of their concentrations of
magnetite, opaques other than magnetite, and amphiboles + epidote.

The only variable of significance in Figure 5 is the magnetite
concentration in soils over the albite granodiorite, which is significant-
ly higher from the magnetite concentration in soils over the gneissic
granite. These two rocks may be differentiated on this basis. The
magnetite concentration in soils over the albite granodiorite is very
similar to that of soils over diorite, and these two rocks can not be dif-
ferentiated on this basis alone. However, 67 percent of the samples
taken over these bodies would lie within their respective areas on the
triangular diagrams, and even though there is a fair amount of overlap,

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each body occupies a unique position on the diagram.

The only variable of significance in Figure 6 which could be used to differentiate soils over the gabbro from those over the metagabbro is percentage of magnetite. Soils over the metagabbro have a significantly higher concentration of magnetite than those over the gabbro. This high concentration of magnetite may also be the reason for the higher concentration of total iron in the metagabbro noted by Hermes (1966). The metagabbro and gabbro fields outlined in Figure 6 show that area into which 67 percent of all the samples taken over these bodies would lie, and despite the fair amount of overlap between these fields, the areas are for the most part distinct from each other.

Comparison of Panning vs. Laboratory Separation

The differences in mineralogy between the panned and elutriated fractions of the same sample are slight. The only significant difference is that in 30 out of 40 samples the panned portion of the sample has a higher concentration of opaques other than magnetite, and in 25 samples the panned portion had a slightly higher concentration of magnetite. This greater percentage of minerals with a higher specific gravity in the panned portions of the samples is one which could be expected, inasmuch as a higher percentage of the lighter heavy minerals are lost during panning. Therefore, those minerals with a high specific gravity which are left comprise a larger portion of the sample.

To make a statistical comparison of the relative effectiveness of panning and the laboratory separation technique, the results obtained for each mineral by the laboratory technique have been compared with the results obtained by panning. The results from the elutriated fraction of each sample were plotted against the results from the panned fraction of the same sample, and the best fit line was calculated according to methods described by Simpson and others (1960, p. 235, 236). The slope of the best fit line is used here as a function of the reliability of the two methods. If the elutriating and panning techniques for processing the samples are equivalent, then the slope of the best fit line should be 1. In most instances the slope of the best fit line for these minerals is close to 1 (Table 2).

The most significant variation between concentrations of the elutriated and panned portions of the same sample is expressed by the slope of the best fit line for zircon. This line has a slope of 0.65 and a 95 percent confidence interval of 0.40-0.82 (Table 2), indicating that a statistically greater concentration of zircon is preserved in the panned portion of the samples. This was also found in Smithson's (1930, p. 135) and Ewing's (1931, p. 138-139) experiments. This great deviation from unity can probably be explained by the small size of the zircon grains in comparison to the larger size of the rest of the other heavy minerals. Ewing (1931, p. 138-139) noted that the comparatively small size of the zircon grains caused them to float with the large amount of

Table 2. The Slopes and 95 Percent Confidence Intervals for the Best
Fit Lines Comparing the Elutriated and Panned Fractions of
the Heavy Mineral Samples.

Mineral	Slope of the best fit line	95 percent confidence interval
Hornblende	1.03	.91 - 1.14
Epidote	1.15	.94 - 1.48
Zircon	.63	.40 - .82
Opaques other than magnetite	.99	.95 - 1.05
Magnetite	.93	.82 - 1.05
Total heavy minerals	1.12	.88 - 1.39

quartz present in the elutriated and bromoformed fraction of the sam-
ple. He also noted that the greater amount of quartz in this fraction
prevented the zircon grains from sinking in the bromoform, whereas in
the panned fraction the amount of quartz had been significantly reduced
enabling more of the small zircon grains to sink. This would explain
the greater concentration of zircon in the panned fraction. In order to
remove completely the zircon from the light fraction the layer of min-
erals on the bromoform surface should not be more than one grain thick.
Such a condition would enable all those minerals with a specific gravity
of greater than bromoform (2.89) to sink, provided the surface tension
of the shape of the mineral did not impede it from sinking.

CONCLUSIONS

(1) Major minerals in most of the residual soils are magnetite,
ilmenite, titaniferous magnetite, hornblende, actinolite-tremolite, epi-
dote and zircon. Trace amounts of muscovite, orthopyroxene, rutile,
sphene, spinel, monazite, apatite and tourmaline were found in some of
the samples.

(2) The weight percentages of the total heavy-mineral concen-
trates from the residual soils were found to be characteristic of the
acidic and basic rock types over which the soils formed: soils from the
gabbro and metagabbro have the greatest percentage of heavy minerals,
and soils from the gneissic granite the lowest.

(3) No single major mineral was found to be diagnostic for the
soil over every rock type, but some minerals are characteristic of cer-
tain rocks. Epidote is more abundant in the residual soil from the dio-
rite than it is in soils over gabbro and metagabbro. Magnetite is also
more abundant in the residual soil from the diorite than it is in soils
over gabbro or gneissic granite. Therefore, diorite can be distinguished

from these rock types by these associations of minerals in the residual soils. Gabbro can also be differentiated from the other rocks through the large amount of amphibole in its residual soil.

(4) Transparent rutile and muscovite were found to be characteristic of the concentrates from residual soil derived from gneissic granite in Vance County. A large amount of visible muscovite was also noted during the panning and elutriating processes in all of the samples taken over this body.

(5) It is difficult to differentiate between residual soils over acidic and basic rocks on the basis of their contained magnetite, opaques other than magnetite, and amphibole+epidote. However, in some instances it was found possible to differentiate between the various types of acidic and basic rocks on the basis of these heavy minerals. The concentration of magnetite in residual soil from metagabbro was found to be higher than and statistically different from the concentration of magnetite in the residual soil of gabbro. The concentration of magnetite in the residual soil from albite granodiorite was found to be higher and statistically different from that for the gneissic granite. Therefore, these rock types can be differentiated on the basis of the concentrations of magnetite in their residual soils.

(6) For every mineral except zircon the recovery rates using the panning and the laboratory techniques were not satistically different. The probable reason for the significantly greater concentration of zircon in the panned fraction of the samples is that the zircon grains tend to be smaller than the other heavy minerals, and this causes them to float with the large amount of quartz in the elutriated and bromoformed fraction of the samples. However, in the panned fraction the amount of quartz has been significantly reduced, thus enabling more of the small zircon grains to sink in the bromoform. In the interest of speed and economy the use of a panning technique to concentrate the heavy minerals before bromoforming is the best method for carrying out a heavy mineral survey.

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AGE AND STRATIGRAPHIC RELATIONS OF THE TALLADEGA

SLATE: EVIDENCE OF PRE-MIDDLE ORDOVICIAN TECTONISM

IN CENTRAL ALABAMA

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ABSTRACT

Folded and thrust faulted Lower Paleozoic rocks exposed near Sylacauga, Alabama include the Weisner, Shady, and Rome Formations of Early Cambrian age, the Conasauga Formation of Middle Cambrian age, and the Knox Group of Late Cambrian and Early Ordovician age. These strata were strongly folded and then covered by Talladega sediments in post-Early Ordovician to Devonian time. This interpretation is supported by exposures of the basal contact of the Talladega Slate at four places near Sylacauga and by coarse conglomerate with some pebbles of gneiss and granite above the unconformity in Chilton County, where the Sylacauga carbonate belt wedges out beneath the Talladega. Upper Knox strata of Early Ordovician age are the youngest rocks known to underlie the Talladega. Basal Talladega rocks, therefore, are considered post-Early Ordovician in age. Devonian fossils in the upper Talladega at Jemison in Chilton County provide an upper age limit.

Butts (1926, 1940) mapped strata in Shelby County as Talladega Slate which are in part equivalent to rocks older than Talladega at Sylacauga, Alabama. Tracing of the Sylacauga section westward indicates that the name "Talladega Slate" should be confined to rocks southwest of the village of Jumbo in Chilton County. Strata northwest of Jumbo lie below the Talladega and belong chiefly to the Weisner Formation and possibly to equivalents of the Rome Formation in part.

It is concluded that the lower part of the Talladega Slate represents a Middle Ordovician clastic wedge in Alabama and that its deposition was preceded by a period of relatively strong tectonism.

INTRODUCTION

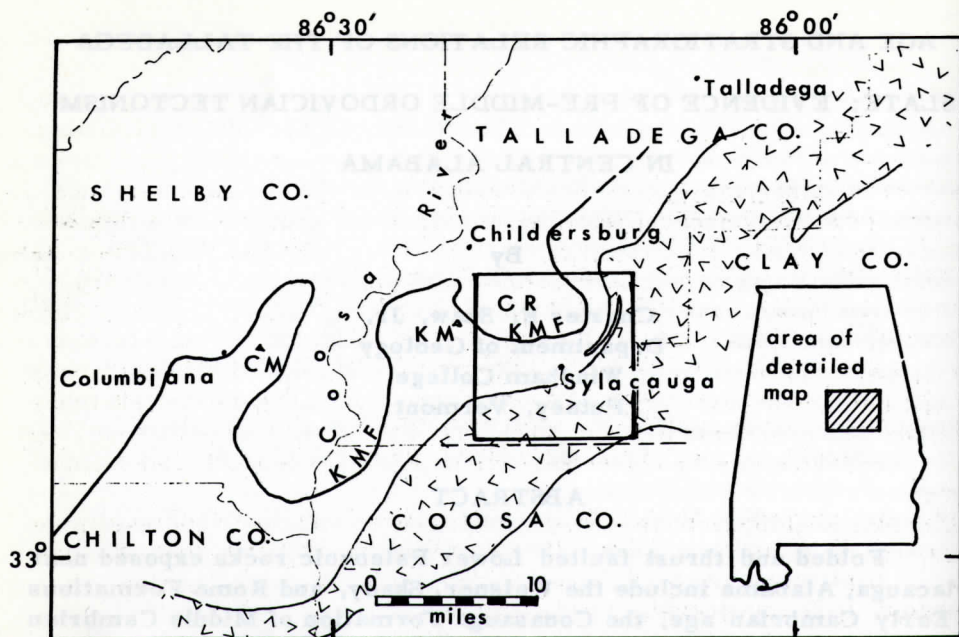This paper presents evidence that the Talladega Slate overlies

Figure 1. Index map of Central Alabama showing location of study area
and some localities referred to in the text. Area covered
by Figure 2 is indicated by heavy lines. CR, Childersberg
Reentrant; KMF, Kahatchee Mountain Fault; KM, Kahatchee
Mountain; CM, Columbiana Mountain. Pattern shows out-
crop belt of the Talladega Slate as defined in this report.

with angular unconformity a sequence of Cambrian and Ordovician
strata near Sylacauga, Alabama (Figure 1). The Talladega is thought
to represent a Middle Ordovician clastic wedge in this region. Thus,
it becomes a stratigraphic link between largely unmetamorphosed mio-
geosynclinal strata of the Valley and Ridge Province and their meta-
morphosed clastic equivalents in the Piedmont Province.

Previous Work

The distribution of clastic and carbonate rocks near Sylacauga
has been mapped generally by early workers (McCalley, 1897; Prouty,
1916; Butts, 1926). All show quartzite and phyllite underlying Kahatchee
Ridge, north of Sylacauga, as thrust northeastward onto carbonate
rocks of the Knox Group in the Coosa Valley (Figure 2). Marble Valley
to the south of Kahatchee Ridge also is underlain by carbonates, the
southeasternmost of which have been metamorphosed to form the Syla-
cauga marble of Prouty (1916) and Butts (1926). Phyllite and sandstone
of the Talladega Slate lie southeast of the marble belt. The stratigraphic

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and structural relations among these rocks have been subject to various interpretations.

The earliest discussion of the geology of the Sylacauga area known to the author is that of McCalley (1897, p. 588-593). McCalley recognized a lower sequence of clastic "Cambrian and Silurian" rocks overlain by carbonate rocks of the "Knox dolomite" and "Pelham limestone." The Sylacauga marble was described as the metamorphosed equivalent of these carbonate units. McCalley believed the Talladega Slate, which borders these rocks on the southeast, had been thrust-faulted into its present position.

Prouty (1916), who was concerned chiefly with the marble-bearing rocks, generally concurred with the structural and stratigraphic ideas of McCalley and mapped the clastic rocks that underlie Kahatchee Ridge and nearby foothills as "Cambrian sandstones and shales." He regarded the marble as consisting of rocks of several ages, but mapped the marble belt as a single unit.

A quite different interpretation of the stratigraphy and structure near Sylacauga was advanced by Butts (1926, p. 51-53). According to Butts, the clastic rocks that underlie Kahatchee Ridge belong to the Talladega Slate thrust over carbonate rocks of the Coosa Valley. The carbonate belt to the southeast of Kahatchee Ridge, with the exception of the marble belt, was interpreted as the autochthonous Coosa Valley rocks exposed in a window through overthrust Talladega Slate. Butts considered the Sylacauga marble a single stratigraphic unit all of one age and a member within the Talladega Slate.

Along strike to the southwest in Shelby County, Butts (1940, 1926) mapped all the rocks above the Kahatchee Mountain fault, except those at Columbiana Mountain, as Talladega, which he subdivided into several members. At Columbiana Mountain the Weisner, Shady, Rome, and Conasauga Formations are folded into a northeast plunging syncline.

Acknowledgements

STRATIGRAPHIC AND STRUCTURAL RELATIONS NORTHWEST OF

THE TALLADEGA SLATE BELT AT SYLACAUGA

Structural Setting

Strata exposed at Sylacauga are part of the Kahatchee Mountain thrust sheet, displaced to the northwest along the Kahatchee Mountain Fault. Both the allocthonous and autochthonous rocks have been arched by a major east-southeast plunging cross-anticline, expressed in the topography of the region by the Childersberg reentrant of the Coosa Valley (Figure 1).

The Kahatchee Mountain Fault thrusts the Lower Cambrian Weisner Formation over Cambrian and Ordovician carbonate rocks belonging to the Knox Group. Where exposed, the fault is approximately parallel to bedding in the lower slaty part of the Weisner Formation of the overriding block. The strike of bedding in the overridden rocks also is generally parallel to that of the fault, but dips are variable.

Horizontal displacement on the Kahatchee Mountain Fault is at least five miles, which is the southeastward extent of the Childersberg reentrant. Displacement indicated by a similar reentrant eight miles to the west is on the order of 15 miles. Stratigraphic displacement on the fault near Sylacauga is at least equivalent to the thickness of the entire Cambrian and most of the Lower Ordovician section, about 9700 feet.

Evidence at Sylacauga indicates the Kahatchee Mountain Fault is post-Early Ordovician in age. However, Mississippian rocks have been overridden by the fault in Shelby County to the west, indicating the thrusting is Mississippian or younger.

Cambrian and Ordovician strata in the Kahatchee Mountain thrust sheet dip gently to the southeast and succeed one another without break or important repetition to within half a mile of the Talladega Slate. Near Sylacauga a line of cherty hills and ridges formed on the lower part of the Knox Group traces out the hinge of a major fold, here named the Sylacauga syncline, which appears to plunge south or southwesterly. The southeast limb of the syncline is truncated sharply by the Talladega Slate (Figure 2).

Most rocks in the area southeast of the Kahatchee Mountain Fault have undergone low grade regional metamorphism and, in general, the intensity of metamorphism increases southeastward. Northwest of the Kahatchee Mountain Fault the few available outcrops appear unmetamorphosed.

Southeast of the fault metamorphic intensity is variable from place to place, largely due to the influence of rock type. All pelitic rocks are more or less recrystallized with white mica-quartz the dominant mineral assemblage. Grain size increases to the southeast. Carbonate rocks range from finely recrystallized beds to coarse white
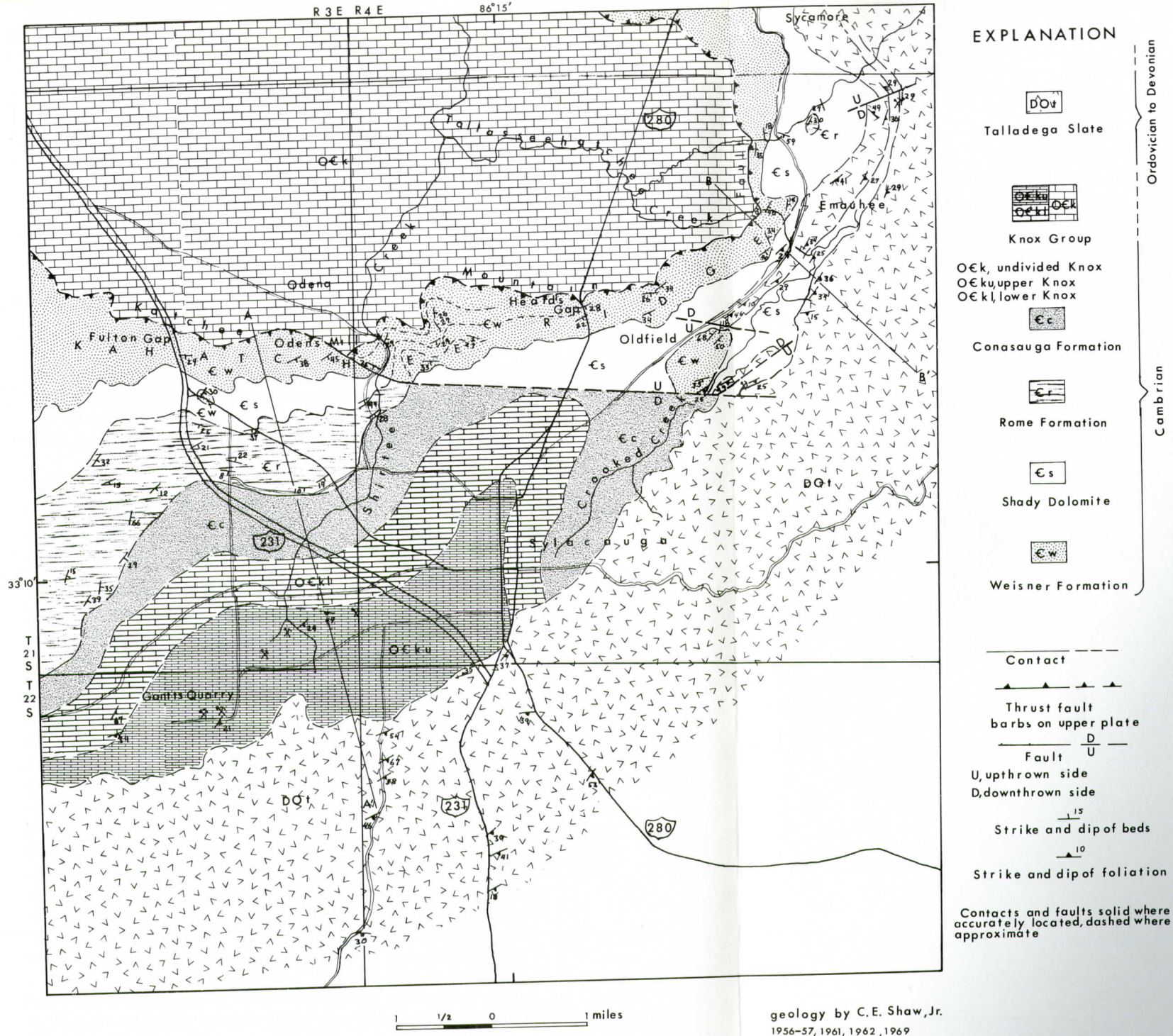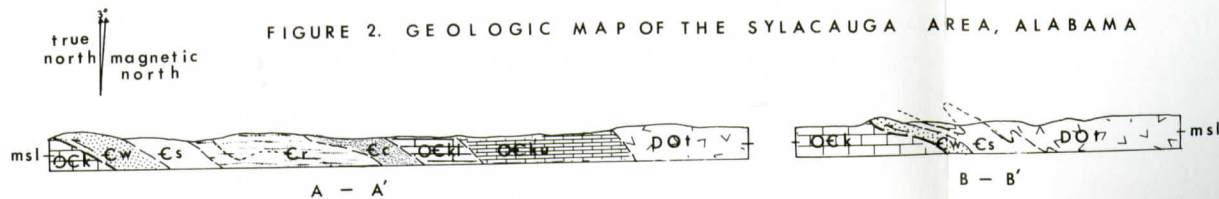
EXPLANATION

Ordovician to Devonian

D€Ov
Talladega Slate

Cambrian

O€ku | O€k
O€kl
Knox Group

O€k, undivided Knox
O€ku, upper Knox
O€kl, lower Knox

€c
Conasauga Formation

€r
Rome Formation

€s
Shady Dolomite

€w
Weisner Formation

Contact

Thrust fault
barbs on upper plate

D
U
Fault
U, upthrown side
D, downthrown side

15
Strike and dip of beds

10
Strike and dip of foliation

Contacts and faults solid where
accurately located, dashed where
approximate

geology by C.E. Shaw, Jr.
1956-57, 1961, 1962, 1969

FIGURE 2. GEOLOGIC MAP OF THE SYLACAUGA AREA, ALABAMA

true
north | magnetic
north

1   1/2   0   1 miles

msl   O€k €w €s   €r   €c   O€kl   O€ku   D€Ot   msl
A — A′

msl   O€k   €w €s   D€Ot   msl
B — B′

marble, the latter especially in the purer limestone units close to the Talladega Slate border. In general, the limestones are more recrystallized than interlayered dolomites. The apparent abrupt jump in metamorphic grade along the northwest edge of the Sylacauga marble belt is largely due to the appearance of limestone in the section at that horizon, not to increased metamorphism. Close to the Talladega Slate border, all of the carbonate rocks have been recrystallized, including the dolomites.

Mineral assemblages in the Talladega Slate immediately southeast of Sylacauga are similar to those in phyllites in the Weisner Formation to the north, essentially white mica-quartz with some feldspar and epidote. Somewhat higher in the Talladega, phyllite with the assemblage quartz-albite-white mica-chlorite-epidote is interlayered with greenish arkosic sandstone. The chlorite is pleochroic, from colorless to pale green and has an anomalous brown interference color, indicating a relative enrichment in magnesium (Albee, 1962). No thin sections were examined from the upper Talladega rocks.
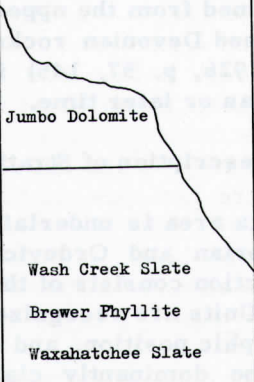
Metamorphosed Devonian rocks in the upper part of the Talladega Slate (Butts, 1926, p. 57, 145) indicate that the metamorphism took place in Devonian or later time.

Description of Stratigraphic Units

The Sylacauga area is underlain by lower Paleozoic strata belonging to the Cambrian and Ordovician systems. The lower part of the stratigraphic section consists of the Cambrian Weisner, Shady and Rome Formations. Units are recognized on the basis of their lithologic character, stratigraphic position, and continuity with these formations to the northeast. The dominantly clastic Lower Cambrian rocks are overlain by Middle and Upper Cambrian and Lower Ordovician carbonate rocks that formerly were mapped as appearing in windows through overthrust Talladega Slate (Butts, 1926). Based on ages assigned elsewhere in Alabama (Butts, 1926) and in Tennessee (Rodgers, 1953), the Conasauga Formation in the Sylacauga area is mostly Middle Cambrian. The Knox Group is Late Cambrian and Early Ordovician and at Sylacauga is separated into an upper and lower part. Formations recognized in the Knox Group elsewhere in Alabama were recognized at some localities in the Sylacauga area but are not everywhere separable.

The Sylacauga Marble Member of the Talladega Slate of earlier reports (Butts, 1926) is abandoned as a stratigraphic unit in this report. On the basis of its stratigraphic position and the sequence of mappable units within it, the marble is regarded as metamorphosed portions of the Shady and Conasauga Formations and the Knox Group, as was originally proposed by McCalley (1897, p. 588-593) and Prouty (1916). The very pure calcium carbonate marble reported by Prouty (1916, p. 24), which constitutes the bulk of the commercial Sylacauga Marble, is probably metamorphosed Newala Limestone in the upper Knox.

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Table 1. Correlation of Formations Exposed at Sylacauga, Alabama with Strata Exposed in Chilton and Shelby Counties, Alabama.

Sylacauga Area, Alabama	Chilton and Shelby Counties, Alabama (Butts, 1926, 1940)	Series	System
Lower part of the Talladega Slate	Lower part of the Talladega Slate	Middle	ORDOVICIAN
Upper, non-cherty part of the Knox (Knox Group)		Lower	ORDOVICIAN
Lower, cherty part of the Knox (Knox Group)	Strata missing due to pre-Talladega erosion	Lower	ORDOVICIAN
Conasauga Formation		Upper	CAMBRIAN
	Jumbo Dolomite	Middle	CAMBRIAN
Rome Formation		Middle	CAMBRIAN
Shade Dolomite			CAMBRIAN
Weisner Formation	Wash Creek Slate	Lower	CAMBRIAN
	Brewer Phyllite		CAMBRIAN
	Waxahatchee Slate		CAMBRIAN

The ages and correlations of Lower Paleozoic strata exposed at Sylacauga are summarized in Table 1.

Weisner Formation - The Weisner Formation near Sylacauga consists of interbedded slate, phyllite, quartzite, and feldspathic quartz-ite and reaches a thickness in excess of 1, 300 feet. A complete section is not exposed at Sylacauga because the base of the formation has every-where been cut out by the Kahatchee Mountain thrust fault.

No fossils have been found in the Weisner Formation in the Syla-cauga area or elsewhere in Alabama. However, its stratigraphic posi-tion below the Shady Dolomite and its lithologic character leave little doubt that the Weisner Formation is the lower clastic formation of the Lower Cambrian Series.

Shady Dolomite - The Shady Dolomite is exposed in the bed of Tallaseehatchee Creek near Emauhee (Figure 2) and consists of light to dark gray, fine to coarse grained banded or "ribboned" dolomite. Far-ther to the northeast, presumably higher in the section, light gray

dolomite is exposed. The implied sequence of a banded or ribboned portion of the Shady overlain by more massive light colored dolomite is similar to the sequence locally recognized in Tennessee (Rodgers, 1953, p. 42).

Weathering of limestone layers in the Shady produces seams of yellow, non-silty, "buckfat" clay. Near the base of the formation layers of the clay are interbedded with decomposed shaly material that has resulted from weathering of silty and shaly layers in the dolomite. A road cut near Emauhee exposes yellow to brown limonite and manganese-stained clay residuum of the Shady that contains large blocks of soft white chert. Where not covered by wash from the adjacent Weisner ridge, a red soil several feet deep has formed on the residuum.

The thickness of the Shady in the Sylacauga area is unknown but on the basis of width of outcrop south of Fulton Gap it probably is 500 to 1,000 feet.

Rome Formation - The Rome Formation in the Sylacauga area consists of interlayered pelites, sandstones, and carbonate rocks. The pelitic rocks have been metamorphosed to low grade slates and phyllites in most places, although apparently unaltered red and green shale has been seen in a few places in the northernmost part of the outcrop belt. The carbonate rocks, chiefly dolomite, have been recrystallized to fine-grained white to gray or pinkish dolomite marble toward the south and southwest, but are not recrystallized in the northern parts of the outcrop belt.

The thickness of the Rome in the Sylacauga area is not precisely known, but is estimated to be 1000 to 2000 feet.

Conasauga Formation - At Sylacauga the Conasauga Formation, largely covered with alluvium, underlies a low-lying belt a half mile wide northeast of the city and is inferred to underlie a part of the marble belt east of the city. It is uncertain whether the northwestern and the eastern belts connect owing to the presence of the Oldfield cross-fault in the area where they would join. The Conasauga is estimated to be 800 to 1,000 feet thick.

Knox Group - In the Sylacauga area the Knox Group is divisible into a lower, cherty part and an upper, generally non-cherty part. The lower, cherty part of the Knox crops out in a curved belt above the Conasauga Formation where it forms a discontinuous series of northeast-trending cherty ridges from the western border of the area near Gantts Quarry to the northeastern outskirts of Sylacauga. Northeast of the city the ridge line bends sharply to the south and assumes a south-southwest strike before terminating abruptly against the northeast striking outcrop of the Talladega Slate. Massive and jagged Knox chert also has been observed to underlie most of the hilly terrain north of Kahatchee Ridge.

The upper part of the Knox underlies a lowland inside the curved belt of cherty ridges formed on the outcrop of the lower Knox. The belt of upper Knox is covered by wash from the Talladega Slate near its

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southeast margin and by a deep red soil elsewhere.

Near Sylacauga the upper part of the Knox has been metamorphosed to a white marble in most places, but contain dark marble layers and laminae of green phyllite. The marble is generally medium to coarse grained and is composed of rather pure calcium carbonate (Prouty, 1916, p. 24). Tectonic thickening and thinning make estimates of stratigraphic thickness unreliable.

STRATIGRAPHIC AND STRUCTURAL RELATIONS OF THE

TALLADEGA SLATE NEAR SYLACAUGA

General Statement

The Cambrian and Lower Ordovician rocks at Sylacauga are bordered on the southeast by the Talladega Slate. The Talladega contact bevels down across the carbonate units and in different places the Talladega is in contact with the Shady Dolomite, the Conasauga Formation, and the Knox Group. These relations can be explained either by a thrust fault, as shown on the Geologic Map of Alabama (1926), or by an angular unconformity (Shaw and Rodgers, 1962).

Description of the Talladega Slate

Smith (1888) applied the name "Talladega group" to a thick section of phyllite and arkosic sandstone exposed along Talladega Creek in Talladega County, Alabama. Prouty (1923, p. 33) referred to these same rocks as the "Talladega phyllite or slate". Butts (1926, p. 49) used the term Talladega Slate for these rocks and extended the term to include strata below the type Talladega in Chilton and Shelby Counties, which he subdivided into several formations. The term Talladega Slate is used informally in this report, pending detailed stratigraphic study of the type Talladega and its subdivision into formations.

The Talladega Slate is exposed in a broad belt southeast of Sylacauga. Weathered Talladega is fairly well exposed in numerous road cuts along U. S. Route 231, and especially along U. S. Route 280, (route 241 on the Goodwater 1/62500 quadrangle), both of which cut obliquely across the entire belt. Fairly fresh exposures may be found in the beds of some streams. These rocks are some twenty miles to the southwest of the type area on Talladega Creek and presumably represent essentially the same stratigraphic interval.

Traveling southeast from Sylacauga on U. S. Route 280, the lowest exposed Talladega consists chiefly of crinkled gray, brownish- to red-weathering phyllite. This is succeeded approximately one mile southeast of the contact by phyllite interbedded with appreciable amounts of impure greenish, arkosic sandstone. This sandy or "grit"

facies of the Talladega is characteristic of much of the Talladega in Talladega, Coosa, and Chilton Counties (Shaw and Rodgers, 1962).

The Talladega also crops out in a narrow northeast-trending belt near Oldfield. The belt is separated from the main mass of the Talladega to the southeast by a narrow strike valley underlain by marble, much of it apparently calcitic. In this belt the Talladega consists dominantly of red-weathering phyllite with numerous pods of vein quartz and little or none of the sandy or "grit" facies found higher. Southeast of Sycamore the belt merges with the main Talladega outcrop belt, the lower part of which consists of the same red-weathering phyllite exposed in the westerly belt.

Contact Relations at the Base of the Talladega Slate

The Talladega Slate may be seen in contact with underlying carbonate rocks or their weathering products at four places in the vicinity of Sylacauga (see Appendix for description). In each case the Talladega overlies dolomite or calcite marble. At the Crooked Creek locality a relief of ten feet or more can be demonstrated on the surface of the marble (Figures 3 and 4). This is interpreted as an erosion surface beneath the Talladega. At two other localities the Talladega overlies the weathering products of the Shady Dolomite and the upper Knox, respectively. In no case does the contact show any evidence that can be interpreted as indicating a fault. Rocks adjacent to the contact are no more deformed than rocks elsewhere. There is no fault gouge or other evidence of movement in the rocks. Tight chevron folds found near the contact appear to be entirely the result of collapse of the phyllite into recent solution channels in the underlying carbonate rocks. On the basis of demonstrable erosional relief at the Crooked Creek locality and the lack of evidence for faulting, it is concluded that the Talladega Slate was deposited on an erosion surface cut into folded Cambrian and Lower Ordovician rocks and that the contact between the Talladega Slate and rocks to the northwest is an angular unconformity at Sylacauga.

Age of the Talladega Slate

Devonian fossils have been collected from the Jemison Chert in the upper part of the Talladega Slate in Chilton County (Smith, 1903, p. 244-246; Butts, 1926, p. 145, 219). Assuming that the lower part of the Talladega is in its proper stratigraphic position below the Jemison, the lower Talladega must be Devonian or older in age. At Sylacauga the Talladega unconformably overlies folded rocks as young as Early Ordovician. Thus, the lower Talladega must be post-Early Ordovician to Devonian in age (Rodgers and Shaw, 1962).

The pre-Talladega erosion surface at Sylacauga truncates strata as high as Knox, suggesting correlation with the unconformity above the Knox and Chickamauga in east Tennessee. The lower Talladega, there-
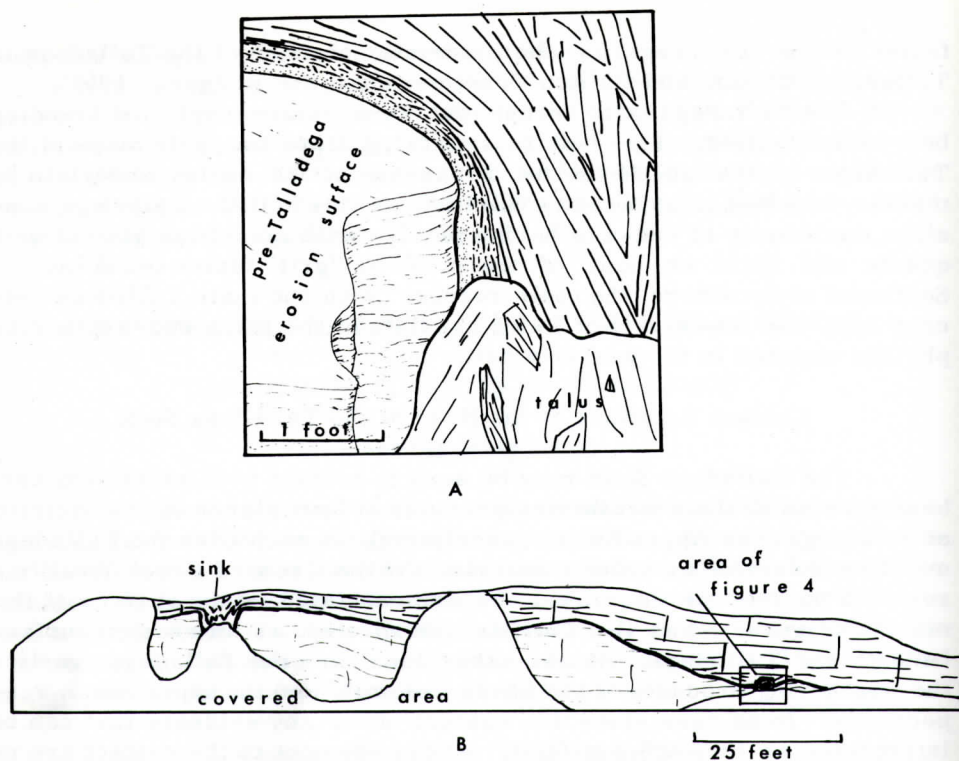
A

B

Figure 3. Sketches showing contact relations between the Talladega Slate and underlying marble. No vertical exaggeration.

    A. Detail of contact showing phyllite crumpled by collapse into a recent solution channel in the underlying marble. The eroded channel wall, typical of solution channels throughout the area, is seen left of center. Light stipple above contact indicates layer of rusty, fine, quartz sand which is probably an insoluable residue left from recent solution of marble along the contact. Sand is followed by a red-weathered zone containing fragments of phyllite. Both sand and weathered phyllite are bent downward with the phyllite into the solution channel.

    B. Generalized sketch of vertical exposure of contact at Crooked Creek locality. Phyllite, indicated by medium-weight lines representing trace of foliation, overlies dolomitic marble. Trace of bedding in the marble is shown by fine lines. Heavy line at contact indicates weathered zone. Phyllite lies on some ten feet of relief and cuts across bedding in the marble. Crumpling of foliation is confined to area of sink hole. View is to the east.
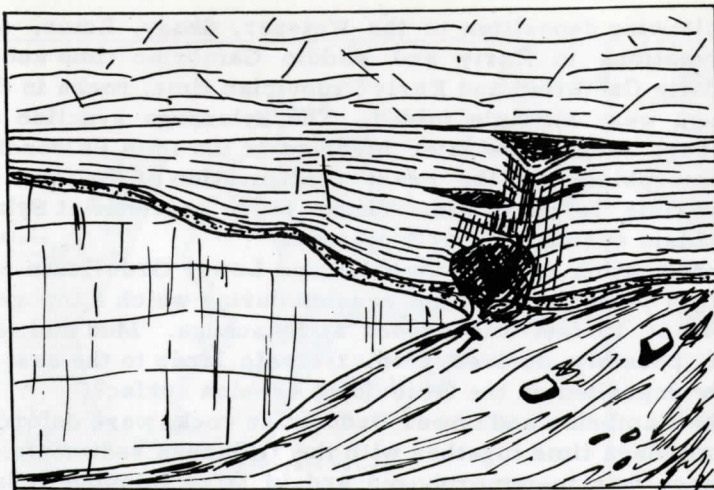
Figure 4. Vertical exposure showing unconformity beneath the Talladega Slate near Crooked Creek. A recent weathered zone is developed along the contact. Trace of bedding in the marble is approximately horizontal. Dip is 25 degrees to the east, away from the viewer. Foliation in the phyllite is approximately parallel to bedding in most places. Phyllite appears to thicken in low places. Trace of foliation above is subhorizontal and converges toward marble to the left (north). A dark tunnel entrance is seen near the contact at right of center. Tunnel is three feet in diameter.

fore, may be in the part equivalent to the Sevier clastic wedge of east Tennessee.

Conclusions

Stratigraphic and structural relationships near Sylacauga, Alabama indicate that the tectonic history of the area, and by implication that of the surrounding region, is more complex than had been thought. Earlier workers supposed, with much justification, that the folds and faults of the Valley and Ridge Province in Alabama were entirely Late Paleozoic in age (Butts, 1926, p. 229). However, the angular unconformity beneath the Talladega Slate exposed at Sylacauga indicates at least one earlier period of deformation, presumably in Middle Ordovician time. Cloud (1967, p. 16) has presented evidence pointing toward an Ordovician orogenic episode in northeast Alabama. An interpretation of the chronology of tectonic events at Sylacauga follows.

Following deposition of the Weisner, Shady, Rome, and Conasauga Formations in Early and Middle Cambrian time and the Knox Group in Late Cambrian and Early Ordovician time, rocks in the vicinity of Sylacauga were strongly folded. The Sylacauga syncline and other pre-Talladega folds were later involved in the Late Paleozoic folding and metamorphism, so the exact configuration of the folds in Middle Ordovician time is uncertain. There is no evidence at Sylacauga of metamorphism during this early episode.

Deformation of the Cambrian and Lower Ordovician rocks was followed by a period of subareal erosion during which a topographic relief of at least 10 feet was eroded at Sylacauga. Mud and sand of the Talladega, possibly derived from tectonic lands to the east of southeast, were deposited on the Ordovician erosion surface.

The Cambrian and Lower Ordovician rocks were deformed again in post-Talladega time together with the Talladega sediments. The entire sequence was metamorphosed and in Mississippian or later time was thrust relatively to the northwest along the Kahatchee Mountain Fault, bringing the Lower Cambrian Weisner Formation onto carbonates of the Knox Group.

REINTERPRETATION OF THE TALLADEGA SLATE IN SHELBY, CHILTON, TALLADEGA AND NORTHWESTERN COOSA COUNTIES

The Cambrian and Lower Ordovician section exposed at Sylacauga has been traced into southwestern Talladega County and adjacent Coosa County to the Coosa River (Figure 5) where the carbonate units are progressively truncated against the base of the Talladega Slate. Near Marble Valley the section includes, from west to east, the Weisner, Shady (a covered interval), and Rome Formations overlain by a thin belt of cherty dolomite in the stratigraphic position of the Conasauga Formation. This dolomite was mapped by Butts (1926) as the Jumbo Dolomite Member of the Talladega Slate. Although Butts shows phyllite intervening between the southwest end of the Sylacauga marble belt and the Jumbo Dolomite, a carbonate valley is continuous between the two. The Jumbo Dolomite, therefore, is the feather edge of the Sylacauga carbonate belt beveled beneath the Talladega unconformity. Rocks below (northwest) the Jumbo, in Coosa and Shelby Counties, therefore, must correspond to the Cambrian clastic sequence in the lower part of the section at Sylacauga (Table 1).

The Shady and Rome Formations have not been recognized southwest of Marble Valley or west of the Coosa River, except at Columbiana Mountain. Typical Weisner rocks are widespread in both areas, however. The ferrugenous sandstone and Brewer Phyllite of Butts (1926, 1940) probably correspond to ferrugenous layers in the Weisner at Sylacauga while the sandstone layers resemble any clean Weisner
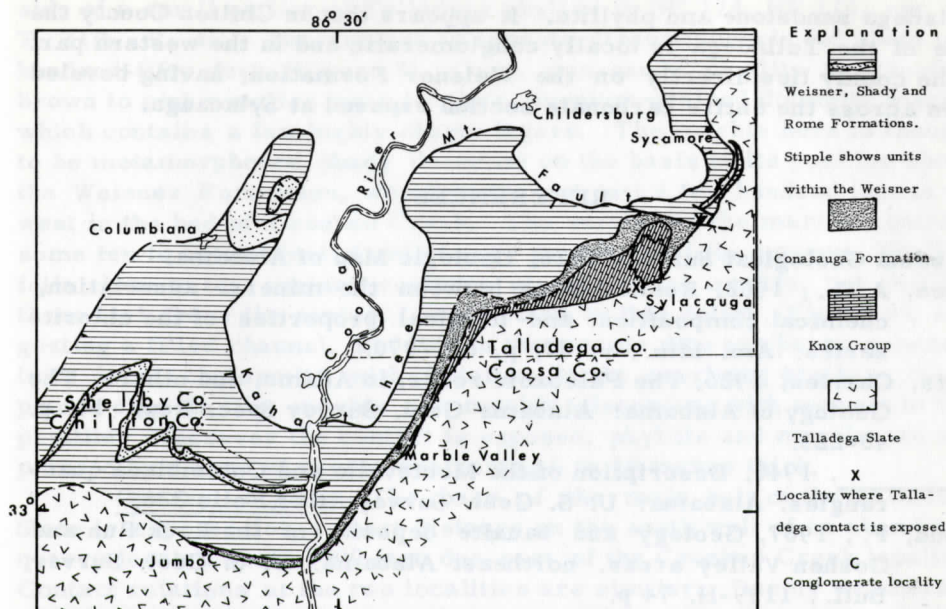
Figure 5. Reinterpretation of portions of Talladega, Coosa, Chilton, and Shelby Counties, Alabama. Southwest end of the Syla- cauga carbonate belt corresponds to the Jumbo Dolomite of Butts (1926). The Talladega Slate bevels down across the carbonate belt regionally to the southwest and in western Chilton County rests on Cambrian clastic rocks. Structures in pre-Talladega rocks in Shelby and Chilton Counties are disharmonic with respect to the overlying Talladega, as is the case at Sylacauga. Conglomerates appear at localities in Chilton and western Coosa Counties. Chilton, Shelby, and Coosa Counties modified from Butts (1926).

quartzite. These rocks are not at all like the greenish, "gritty" sand- stone found in the typical Talladega near Sylacauga. Thus, it appears that all of Butts' "Talladega" north of and stratigraphically below the Jumbo Dolomite lies below the Cambrian and Ordovician carbonate se- quence and represents a greatly thickened Weisner Formation, possibly with Rome equivalents in the upper part.

Phyllite and greenish, sandy phyllite typical of the Talladega Slate at Sylacauga overlie the Jumbo Dolomite. Butts (1926, p. 53) re- ported that the dolomite is overlain by 20 feet of coarse conglomerate with pebbles reaching six inches in diameter in a quarry near the vill- age of Jumbo in Chilton County. A similar display of pebbles, many of gneiss and granite, may be seen in the road gutter just south of the vill- age. The matrix between the pebbles consists of typical greenish

Talladega sandstone and phyllite. It appears that in Chilton County the base of the Talladega is locally conglomeratic and in the western part of the county lies directly on the Weisner Formation, having beveled down across the entire carbonate section exposed at Sylacauga.

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APPENDIX

Description of Localities Where the Basal

Talladega Contact is Exposed

Crooked Creek Locality - The basal contact of the Talladega Slate is best exposed in the valley of Crooked Creek on the northwest

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side of a small outlier of Talladega phyllite in NE 1/4, NE 1/4, sec 22, T. 21S., R. 4E. The contact is exposed discontinuously for about one hundred-fifty feet (Figure 3). Gray, non-sandy phyllite weathering brown to pink overlies gray to white, medium-grained dolomite marble which contains a few highly cherty layers. The marble here is thought to be metamorphosed Shady Dolomite on the basis of its position above the Weisner Formation, which crops out just a few hundred feet to the west in the bed of Crooked Creek. The surface of the marble contains some ten to fifteen feet of relief over the length of the outcrop. Phyllite folia close to the contact conform closely to the marble, but phyllite three feet above the contact is unaffected by this relief (Figure 4), suggesting a filled channel. Bedding in the marble dips southeast, generally in local conformity with foliation in the overlying phyllite, but in places dips in the marble are notably discordant with respect to the phyllite. Wherever the contact is exposed, phyllite and marble are separated by a zone of recent weathering 6 to 18 inches thick.

Quarry Locality - The base of the main belt of the Talladega Slate is exposed for a short distance on the south wall of an abandoned quarry located about 1500 feet due east of the Crooked Creek locality. Contact relations at the two localities are similar. Deeply weathered, red-weathering, non-sandy Talladega phyllite with even, southeast dipping foliation overlies jointed, southeast dipping, white and blue layered calcite marble for an exposed distance of about 150 feet. The marble is assumed to represent a metamorphosed limestone horizon in the Shady Dolomite.

Emauhee Locality - Red-weathering, non-sandy Talladega phyllite overlies manganese-rich residuum of apparently unmetamorphosed Shady Dolomite in a roadcut about 600 feet west-northwest of Emauhee. The contact with the red-weathering basal phyllite of the Talladega is parallel to bedding in the Shady at this locality. Both units dip eastward at about 45 degrees. There is no evidence of erosional relief over the length of the exposure, nor is there evidence of faulting or movement along the contact.

Expressway Locality - Talladega phyllite is exposed in roadcuts and in the drainage ditch along the U. S. Route 231 by-pass around Sylacauga. Brown to red-weathering phyllite with several layers of greenish sandstone is in contact with deep red, non-cherty upper Knox soil in the drainage ditch at the base of an escarpment that marks the northwest border of the Talladega Slate. The contact between soil and phyllite is sharp and the phyllite is undisturbed.

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FASSAITES FROM MADAGASCAR

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ABSTRACT

Eighteen samples of the aluminous pyroxene, fassaite, were separated from the pyroxenites of southeast Madagascar and studied optically and chemically. The optical and physical properties were determined by analysis of the indices of refraction, birefringence, extinction angles, optic axial angles, pleochroism and specific gravity. It was found that the indices of refraction vary from 1.672 to 1.724 for α, and from 1.690 to 1.743 for γ. Birefringence varies from 0.019 to 0.029. Extinction angles vary from 38° to 45°. Optic axial angles vary from 57.5° to 61° (with a positive optic sign). Specific gravity varies from 3.285 to 3.600 with the samples having the darker color and higher Fe content being the more dense. The most remarkable characteristic of these pyroxenes is their notable high content of Al_2O_3, varying from 7.00-11.35 percent. The number of Ca atoms is equal to the sum of Fe and Mg atoms. Also, the weight percentages of CaO, 23.40-25.70 percent, correspond to those which exist in true diopside having the formula $Ca(Mg, Fe)Si_2O_6$. On the contrary, however, the SiO_2 content in these pyroxenes is low, between 41.10 and 44.50 percent by weight, when it normally would be expected to fall between 49 and 55 percent according to the formula. In other words, these pyroxenes seem to have aluminum substituted for silicon.

INTRODUCTION

Fassaite is the name applied to diopsidic pyroxene relatively high in alumina and ferric oxide, practically devoid of alkalis, and having a content of CaO essentially that of the ideal diopside (Tilley, 1938).

As this pyroxene is noted in very few localities, I wish to report it briefly. This is only a report of its optical, chemical and x-ray studies.

Eighteen samples of fassaite were separated from the pyroxenites of southeast Madagascar (Pre-Cambrian). All the samples were given to me for investigation by Dr. H. de La Roche, Assistant Director of the Center of Researches in Petrography and Geochemistry,

C. N. R. S. , Nancy, France.

These pyroxenites are famous for the occurrence of phlogopite and uranothorianite deposits. Lacroix (1922) was the first to describe these rocks systematically. Besairie (1930) later studied the phlogopite occurrences. In 1941, Lacroix made a detailed survey of these phlogopitic pyroxenites and published the first chemical analysis. De La Roche (1956) was the first worker to give a detailed description and interpretation of these pyroxenites. He discussed the essential regional characteristics and the principal aspects of metamorphism, granitization and their possible mode of origin.

Lacroix (1941) divided pyroxenites chemically into two general groups. One consists essentially of calcomagnesian pyroxenes and phlogopite. Silico-aluminous minerals (feldspars or scapolites) are present in a far smaller amount, or are completely absent. Pyroxenites of the second group contain light colored minerals, i. e. silico-aluminous feldspar and scapolite, which are present in quite a considerable proportion.

For convenience, I have divided these pyroxenites mineralogically into the following three groups:
1. Pyroxenites with phlogopite
2. Feldspathic pyroxenites
3. Scapolitic pyroxenites

Pyroxenes of groups 1 and 3 are nonpleochroic, while those of group 2 are the pleochroic varieties. Phlogopites are pleochroic from yellowish to greenish or reddish brown in color, with shining splendent luster. Plagioclase feldspars (An$_{52-62}$) are greyish white in color, turbid and have a vitreous luster. Scapolites are whitish grey, granular and appear to show a more or less greasy luster. They exhibit a transition from plagioclase feldspars, evidenced by the presence of pseudomorphous twin-lamellae.

Diverse ideas concerning the origin of these rocks have been discussed in previous regional studies (Lacroix, 1941; Besairie, 1948, 1956; De La Roche, 1956, etc.). In general, most workers have agreed that these rocks are of metamorphic origin.

Acknowledgments

EXPERIMENTAL RESULTS

Mineral Separations

Pyroxene fractions of 99.5 percent purity were separated by a number of methods. The major part of pyroxene separation was performed with Forrer's Paramagnetic Separator called 'Electrona'. In cases where a completely pure separate was not obtained, 'L' électro-à-pointes' (a point magnetic source) was used for final purification. In cases where even this method did not yield 99.5 percent pure material, heavy liquids were used for the final separations.

Optical Properties

Optical properties of the fassaites were systematically determined. Results are tabulated in Table 1.

The ranges of the indices of refraction were determined by the 'Immersion Media' method. Ordinary immersion methods were used with a set of organic liquids. After a liquid or a mixture of liquids matched the index of refraction of the fassaite, its index of refraction was determined with a Jolley's micro-refractometer. The value, α, varies from 1.672 to 1.724 and γ, varies from 1.690 to 1.743. The normal accuracy for index determinations is estimated to be $\pm$ 0.0005 except for material of variable composition. It is $\pm$ 0.001 for such material.

An estimate of the birefringence was made using Michel Levy's principle and it was found to vary from 0.019 to 0.029.

In determining the extinction angles, methods described by Turner (1942) and Kleeman (1952) were used. They vary between 38° and 45°.

All measurements of optic axial angles were made on the 'Universal Stage' by direct rotation from one optic axis to the other. This method, according to Hess (1949), is more accurate than measurements of optic axial angles by rotation from one optic axis to a bisectrix. Optic axial angles vary from 57.5° to 61° with a positive optic sign. The normal accuracy, for both extinction angles and optic axial angles, is estimated to be $\pm$ 0.5°.

The specific gravity measurements were made by means of a pycnometer. Nearly 5 gms of the sample grains, 25 ml pycnometer bottle and distilled water were used. Results indicate that samples having a darker color and higher Fe content have higher densities. The densities vary from 3.285 to 3.600. The normal accuracy is estimated to be $\pm$ 0.01. All the observations were made at room temperature and are uncorrected for atmospheric pressure.

Table 1. Optical Properties of Fassaites from Southeast Madagascar.

Sample No.	Sp. Gravity	α	Y	Biref.	Extinction Angles	2V	Pleochroism
1	3.312	1.682	1.709	0.027	39°	58.5°(+)	Nonpleochroic
2	3.311	1.678	1.706	0.028	40°	58° (+)	Nonpleochroic
3	3.301	1.682	1.709	0.027	40°	59°(+)	Nonpleochroic
4	3.300	1.682	1.709	0.027	40°	59°(+)	Nonpleochroic
5	3.300	1.682	1.710	0.028	39°	58.5°(+)	Nonpleochroic
6	3.390	1.693	1.718	0.025	42°	60°(+)	Pleochroic (weakly)
7	3.300	1.681	1.709	0.028	40°	59°(+)	Nonpleochroic
8	3.285	1.672	1.701	0.029	38°	57.5°(+)	Nonpleochroic
9	3.600	1.700	1.729	0.029	43°	61°(+)	Pleochroic
10	3.489	1.724	1.743	0.019	45°	59°(+)	Pleochroic
11	3.391	1.699	1.722	0.023	41°	60°(+)	Pleochroic
12	3.382	1.700	1.724	0.024	42°	60°(+)	Pleochroic
13	3.295	1.680	1.707	0.027	40°	58.5°(+)	Nonpleochroic
14	3.300	1.680	1.708	0.028	40°	58.5°(+)	Nonpleochroic
15	3.300	1.679	1.707	0.028	39°	58.5°(+)	Nonpleochroic
16	3.310	1.684	1.711	0.027	40°	59.5°(+)	Nonpleochroic
17	3.312	1.680	1.707	0.027	41°	58.5°(+)	Nonpleochroic
18	3.345	1.695	1.720	0.025	41°	59.5°(+)	Nonpleochroic

Samples Nos. 1- 8 - Pyroxenites with phlogopite
9-12 - Feldspathic pyroxenites
13-18 - Scapolitic pyroxenites

Spectrochemical Analysis

The partial chemical analysis of the fassaites were carried out with an Applied Research Laboratories 'Optical Emission Quantometer'. The samples were crushed to 200 mesh size. Then 300 mg of these powdered samples were mixed with 600 mg of lithium carbonate and 3000 mg of boric acid. The material was mixed thoroughly and fused in a platinum crucible until it became glass. This glass was ground to 200 mesh size in a mechanical ball crusher. Then 600 mg of this powdered glass was mixed with 200 mg of nickel oxide (internal standard) and 1200 mg of graphite. After thorough mixing, small tablets were prepared.

Before analysing these samples in the quantometer, synthetic standards were prepared by mixing pure chemicals in approximately the same amounts expected in the natural samples of fassaite. Tablets of these standards were prepared and analysed in the quantometer.

Table 2. Partial Chemical Analysis of Fassaites from Southeast Madagascar.

	1 Ranopiso	2 Mafilefy	3 Beampin-garatra	4 Beampin-garatra	5 Beampin-garatra	6 Ambotaka	7 Unknown	8 Ambia	9 Beampin-garatra
SiO_2	44.30%	44.00%	42.30%	42.60%	41.10%	42.30%	41.60%	44.20%	41.50%
Al_2O_3	10.10	7.40	8.85	8.20	11.10	7.70	11.35	7.00	8.85
Fe_2O_3*	5.15	5.00	7.50	7.10	5.00	11.80	6.10	3.30	15.60
CaO	25.20	25.10	23.90	24.60	25.00	24.41	24.73	24.70	23.50
MgO	14.80	15.40	15.50	14.40	14.35	12.00	14.00	17.20	9.00
MnO	0.04	0.33	0.10	0.09	0.10	0.40	0.05	0.14	0.26
TiO_2	0.42	0.43	0.38	0.36	0.67	0.24	0.58	0.30	0.18
Na_2O	0.61	0.61	0.35	0.27	0.56	0.60	0.66	0.45	0.64

	10 Amboalengo	11 Androy Man-drareen	12 Ambo-taka	13 Anara	14 Anara	15 Ambinden-drakemba	16 Ranopiso	17 Mafilefy	18 Mafilefy
SiO_2	41.20	43.30	43.52	43.10	41.28	41.25	44.00	44.50	42.30
Al_2O_3	7.45	7.10	7.44	9.30	10.15	10.10	6.75	9.20	9.88
Fe_2O_3*	23.00	12.80	12.97	5.10	5.60	5.60	7.15	5.00	9.00
CaO	23.60	24.30	24.51	24.73	24.40	24.40	25.70	23.40	24.20
MgO	3.60	9.50	9.51	13.10	16.85	17.00	13.10	13.20	10.53
MnO	0.40	0.12	0.31	0.08	0.06	0.10	0.09	0.32	0.43
TiO_2	0.58	0.76	0.33	0.65	0.74	0.73	0.59	0.33	1.62
Na_2O	0.68	0.52	0.44	0.77	0.86	0.85	0.57	0.62	0.53

*Total Fe as Fe_2O_3

All of the analyses were done in quadruplicate and the average is given (except for samples 12 and 18, which were analysed 16 times).

Working curves were drawn using these readings. Tablets of the fassaites were then analyzed in the quantometer and the percentages of various oxides were determined from the working curves.

The conditions of the quantometer used for the analysis were as follows:

Source	A.R.L. High Precision Source (4700)
Capacity	30 F
Inductance	360 H
Resistance	45 ohms
Voltage	950 v
Frequency	50 cps

Time of Excitation 21-22 seconds.

The results are tabulated in Table 2. Table 3 contains the average values and analytical dispersion for sample 18, which has been analyzed 16 times. From this the analytical dispersion can be judged for the other analysis.

Figure 1 gives the variations of CaO, MgO and Fe_2O_3 in fassaites. Figure 2 gives the relation of physical and optical properties and chemical composition in fassaites. The specific gravity, the refractive indices, the birefringence, the extinction angles and the optic axial angles are plotted against the percent Fe atoms in the total Ca, Mg and Fe. Figure 3 gives the three-point diagram Al - Fe - Mg. The diagram illustrates that, in fassaites from the feldspathic pyroxenites, the Fe/Al ratio is greater than 1, and in the other fassaites, this ratio

Table 3. Precision Data for Sample 18.

Elements	n	Mean	Relative Deviation (per cent)
SiO_2	16	42.30%	1.06
Al_2O_3	16	9.88	8.13
Fe_2O_3	16	9.00	11.06
CaO	16	24.20	6.14
MgO	16	10.53	2.66
MnO	16	0.43	2.22
TiO_2	16	1.62	7.30
Na_2O	16	0.53	5.88
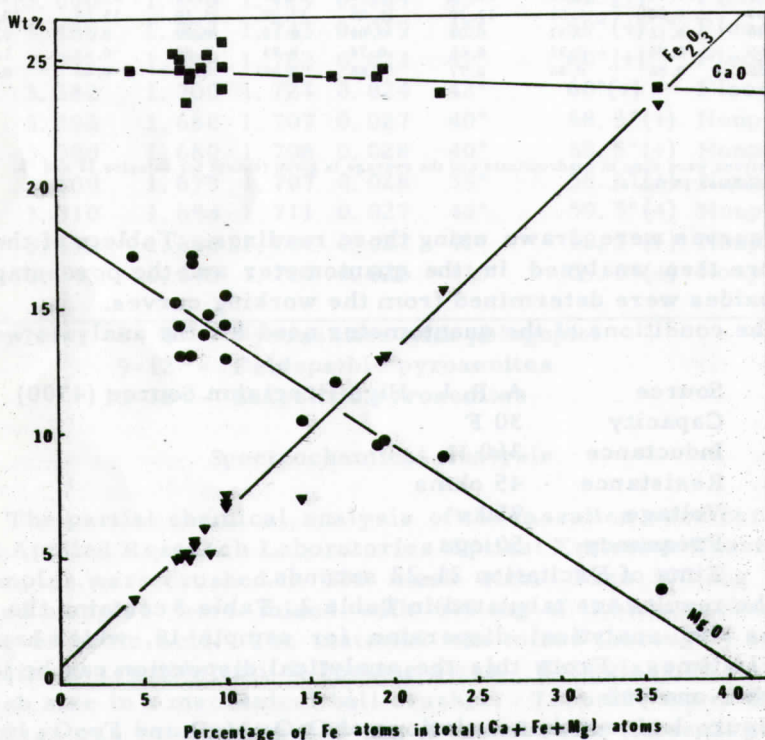
Figure 1. Variations of CaO, MgO and Fe_2O_3 in fassaites.

is less than 1. A distinct straight line can be drawn joining the vertex of Mg and Al_{50} in order to distinguish the fassaites of the feldspathic pyroxenites from the other types.
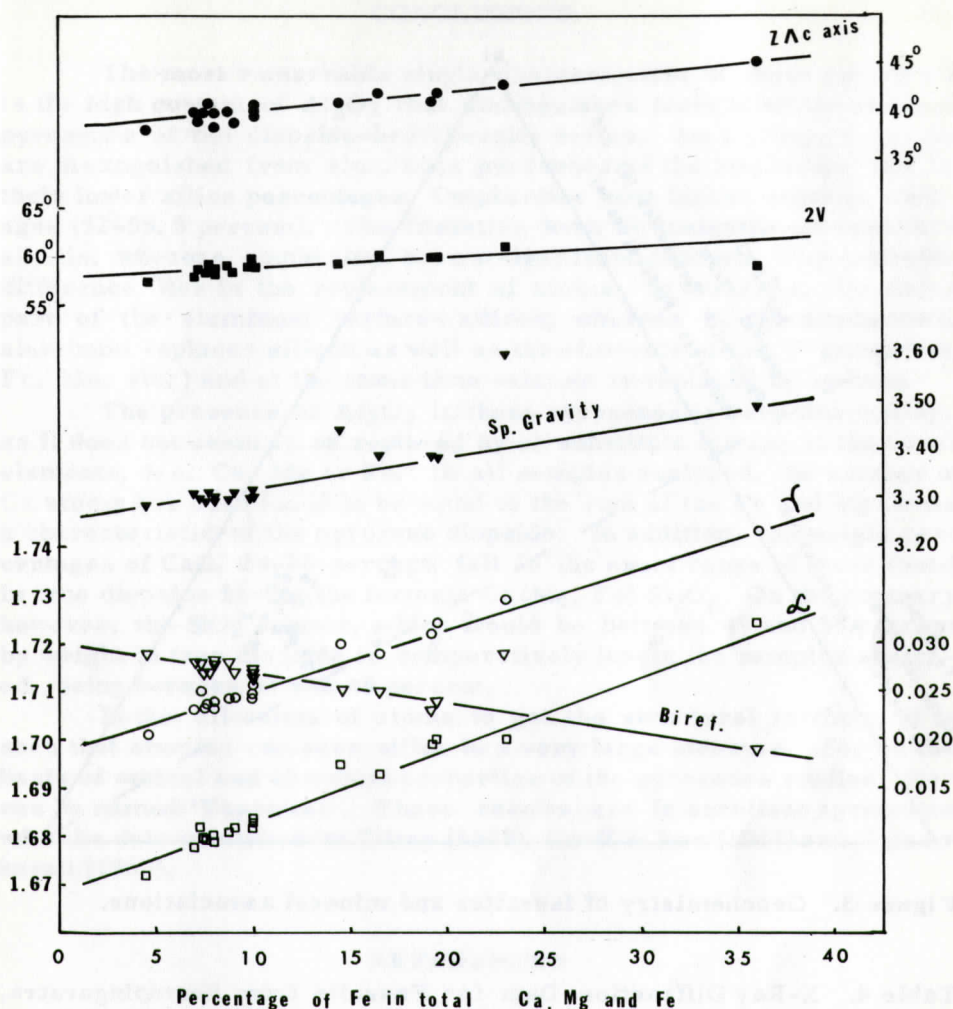
Figure 2. Relation of physical and optical properties and chemical composition in fassaites.

X-Ray Diffraction Pattern

The x-ray powder pattern for the representative sample was obtained with both CuKα and FeKα radiations using 114.6 mm diameter camera on a Philips generator. The results are shown in Table 4. These results are found to be very satisfactory when compared with the data published by Clark (1946) and Sakata (1957).
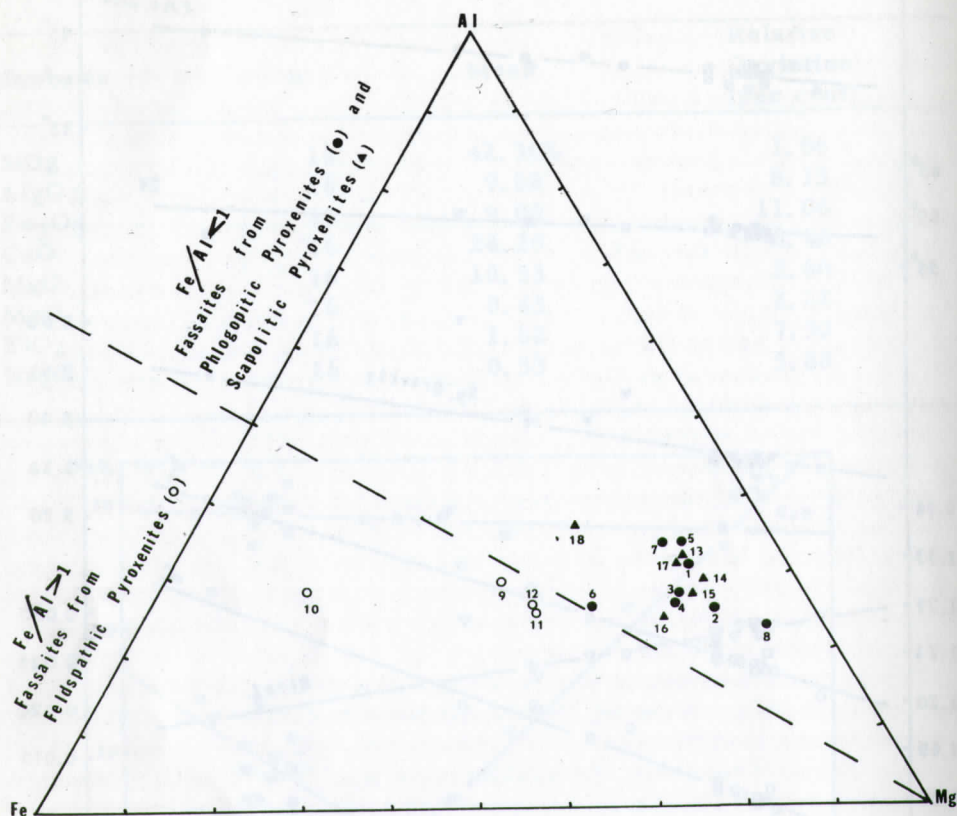
Figure 3. Geochemistry of fassaites and mineral associations.

Table 4. X-Ray Diffraction Data for Fassaite from Beampingaratra, Southeast Madagascar.

d(Å)	I	d(Å)	I	d(Å)	I
3. 330	10	1. 973	10	1. 486	10
3. 224	20	1. 831	10	1. 421	40
3. 000	80	1. 778	20	1. 408	20
2. 890	40	1. 669	10	1. 391	10
2. 720	10	1. 626	60	1. 375	10
2. 530	100	1. 583	10	1. 326	10
2. 299	10	1. 563	10	1. 292	10
2. 207	10	1. 546	10	1. 280	10
2. 128	10	1. 528	10	1. 244	10
2. 023	10	1. 504	10	1. 071	20

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CONCLUSIONS

The most remarkable single characteristic of these pyroxenes is the high content of Al_2O_3 that distinguishes them from the common pyroxenes of the diopside-hedenbergite series. As a group, fassaites are distinguished from aluminous pyroxenes of the omphacite type by their lower silica percentages. Omphacites have higher silica percentages (52-55.5 percent). The fassaites have an insignificant content of alkalis, whereas omphacites have a significant content. The noticable difference lies in the replacement of atoms. In fassaites, the major part of the aluminum replaces silicon, whereas in the omphacites, aluminum replaces silicon as well as the elements of the 'Y' group (Mg, Fe, Mn, etc.) and at the same time calcium is replaced by sodium.

The presence of Al_2O_3 in these pyroxenes is very interesting, as it does not seem to be replaced by or substitute for any of the usual elements, i.e. Ca, Mg or Fe. In all samples analyzed, the number of Ca atoms has been found to be equal to the sum of the Fe and Mg atoms, a characteristic of the pyroxene diopside. In addition, the weight percentages of CaO, 24-26 percent, fall in the exact range of those found in true diopside having the formula $Ca(Mg, Fe) Si_2O_6$. On the contrary, however, the SiO_2 content, which would be between 49 and 55 percent by weight in true diopside is comparatively low in the samples analyzed, being between 41 and 45 percent.

In the allotment of atoms to get the structural formula, it is seen that alumina replaces silica in a very large measure. So, on the basis of optical and chemical properties of the pyroxenes studied, they can be named 'Fassaites'. These results are in excellent agreement with the data published by Tilley (1938), Knoff & Lee (1957) and Rajasekaran (1960).

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