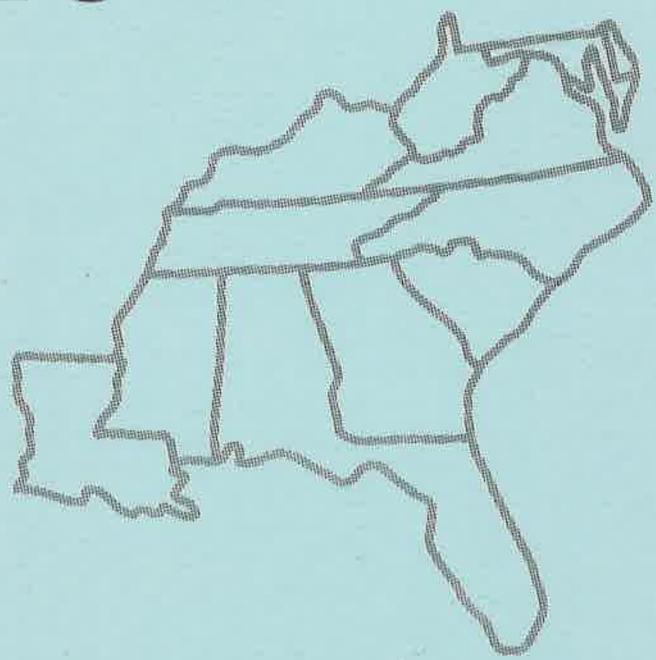
SOUTHEASTERN GEOLOGY



PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 19 NO. 1 DECEMBER, 1977

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

This journal welcomes original papers on all phases of geology, geophysics, and geochemistry as related to the Southeast. Transmit manuscripts to S. DUNCAN HERON, JR., BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA. Please observe the following:

(1) Type the manuscript with double space lines and submit in duplicate.
(2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
(3) Submit line drawings and complex tables as finished copy.
(4) Make certain that all photographs are sharp, clear, and of good contrast.
(5) Stratigraphic terminology should abide by the Code of Stratigraphic Nomenclature (AAPG, v. 45, 1961).

Proofs will be sent authors.

Reprints must be ordered prior to publication. Prices are available upon request.

Subscriptions to Southeastern Geology are $6.00 per volume Inquiries should be addressed to: SOUTHEASTERN GEOLOGY, BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA. Make checks payable to Southeastern Geology.

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 19, No. 1

1977

IMPLICATIONS OF K/Ar AGE DETERMINATIONS TO THE CHRONOLOGY OF MOUNTAIN BUILDING IN THE CENTRAL APPALACHIANS

By

Douglas G. Mose
Department of Chemistry
George Mason University
Fairfax, Virginia 22030

ABSTRACT

Numerous radiometric age determinations have been obtained by the K/Ar technique (a) in the Blue Ridge of North Carolina and Tennessee, and (b) in the Reading and Manhattan Prongs in southeastern New York and New Jersey. The data indicate that there were major metamorphic events at about 1100, 450 and 350 million years ago. There is also increasing evidence of another event at about 900 million years ago.

INTRODUCTION

The chronology of Precambrian (older than 570 million years) and Paleozoic (570 to 225 million years) metamorphic events that involved the crystalline rocks of the Appalachians has been one of the most actively studied geologic problems in eastern North America. The purpose of this paper is to examine the metamorphic histories of two extensively studied areas in the Appalachians: the Blue Ridge in North Carolina and Tennessee (Figure 1) and the Reading and Manhattan Prongs in southeastern New York, New Jersey and Pennsylvania (Figure 2).

Many of our ideas about the geologic history of these areas have come from field studies and from Rb/Sr and U/Pb age determinations. In an ideal case, a Rb/Sr whole-rock age determination derived from a collection of cogenetic igneous or meta-igneous rocks is usually considered to be the time of crystallization of the igneous rock. With sedimentary rocks, the Rb/Sr whole-rock age is usually considered to be the time of sediment deposition or lithification if the rock is a shale; coarser clastic rocks such as sandstones appear to yield Rb/Sr whole-rock ages closer to the provinance age (Faure and Powell, 1972). For metamorphic rocks of sedimentary parentage, the Rb/Sr whole-rock age is usually considered to be the time of metamorphism. A U/Pb

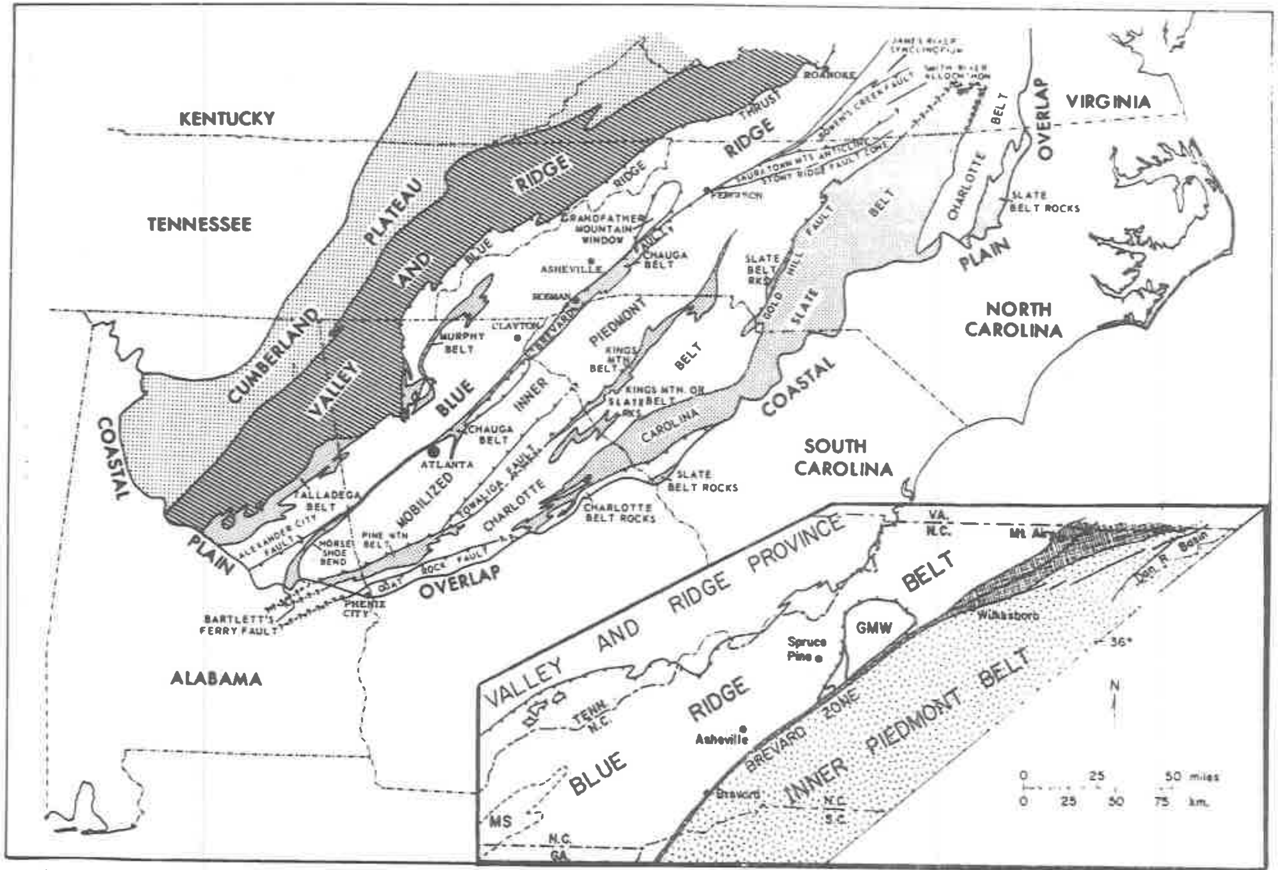


Figure 1. Generalized geologic map of the Blue Ridge and adjacent provinces in the southern Appalachians. Large map from Hatcher (1975). Inset is of the area from which the K/Ar ages discussed in this paper were obtained. Inset from Butler (1973).

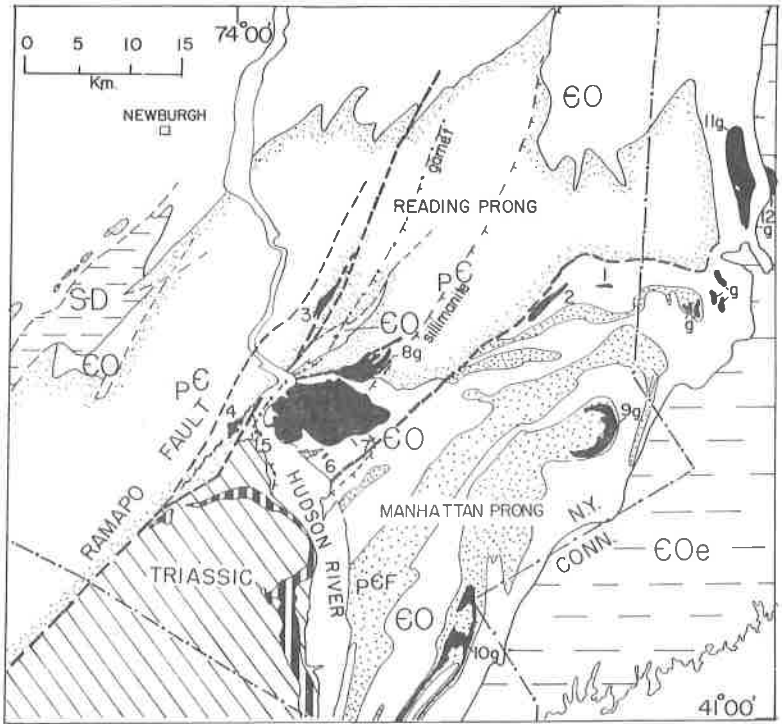


Figure 2. Generalized geologic map of the northern Manhattan Prong and the eastern Reading Prong in New York and New Jersey. Also shown are the basic plutons (1. Peach Lake, 2. Croton Falls, 3. Canopus, 4. Rosetown, 5. Stony Point, 6. Torment Hill, 7. Cortlandt) and the granitic plutons (8g. Peekskill, 9g. Pound Ridge, 10g. Yonkers, 11g. Candlewood Lake, 12g. Brookfield). Key to symbols: p€. Grenville basement gneiss of the Reading Prong, p€f. Grenville basement gneiss (Fordham Gneiss) of the Manhattan Prong, €O. Cambrian and Ordovician miogeosynclinal rocks, €Oe. Cambrian and Ordovician eugeosynclinal rocks, SD. Silurian and Devonian sedimentary rocks. Map from Mose and others (1976).

age determination obtained from a collection of cogenetic zircons is considered to be the time of zircon crystallization. In U/Pb work, the zircon morphology is crucial in evaluating an age determination in terms of host rock chronology (Doe, 1970).

Many Rb/Sr whole-rock and U/Pb zircon age determinations have been used to determine the chronology of Precambrian rock units in the Blue Ridge of North Carolina and Tennessee (Fullagar and Odom, 1973; Davis and others, 1962; Rankin and others, 1969; Tilton and others, 1960). The Blue Ridge rocks have been grouped into older Precambrian basement rocks (Grenville gneisses) which were metamorphosed during the Grenville event about 1100 m. y. ago, and younger Precambrian (700 to 800 m. y. old) rocks which were not (Rankin, 1970). Both the older and younger Precambrian rocks in the Blue Ridge were affected by Paleozoic metamorphism.

Similar radiometric work has been done in the Reading and Manhattan Prongs to determine the chronology of rocks which existed before the Paleozoic metamorphism. The Reading Prong, except for a few inliers of Paleozoic rocks, is made of Precambrian rocks metamorphosed during the Grenville event (Tilton and others, 1960; Ratcliffe and others, 1972; Hall and others, 1975). Only east of the Ramapo fault zone does the Reading Prong appear to have been affected by Paleozoic metamorphism (Long and Kulp, 1962; Dallmeyer and others, 1975).

The Manhattan Prong, which lies entirely east of the Ramapo fault zone, is composed partly (about 50 percent) of a Precambrian rock known as the Fordham Gneiss. Based on U/Pb studies of detrital and syn-metamorphic zircons, Grauert and Hall (1973) have shown that the Fordham Gneiss is between 1300 and 980 m. y. old. The Fordham Gneiss is overlain by a Cambro-Ordovician sequence including the Lowerre Quartzite, Inwood Marble and Manhattan Schist. All these rock units including the Fordham Gneiss were metamorphosed in Paleozoic time (Hall, 1968).

Late Precambrian rocks have been found in this part of the northern Appalachians. The Canada Hill Granite in the north-central Reading Prong formed by partial melting of the Grenville metasediments about 915 m. y. ago (Hall and others, 1975; Mose and Helenek, 1976). The Fordham Gneiss in the Manhattan Prong contains intrusive granitic gneisses of Avalonian age (600 to 550 m. y. old) which were also probably formed by partial melting (Long, 1969; Mose and Hayes, 1975). The significance of these late Precambrian events and their areal extent are poorly understood.

Acknowledgments

I am pleased to acknowledge the work of many geologists whose efforts over the past 15 years have made a summary paper such as this possible. I am also indebted to A. L. Odom (Florida State University), P. D. Fullager (University of North Carolina), R. D. Dallmeyer (University of Georgia), H. L. Helenek (Bradley University), L. M. Hall (University of Massachusetts) and N. M. Ratcliffe (City College, CUNY) whose comments and enthusiasm initiated this effort.

CHRONOLOGY OF PRECAMBRIAN METAMORPHISM

The timing of the Precambrian metamorphic event which occurred in the Appalachians has been of special interest to geochronologists.

Although much of our understanding has come from Rb/Sr and U/Pb studies, additional but conflicting interpretations have come from K/Ar age determinations. Much of the problem is due to the Paleozoic metamorphism of Precambrian rocks. The Paleozoic metamorphic zones in the Reading and Manhattan Prongs are shown in Dallmeyer and others (1975) and in Mose and others (1976). The Paleozoic metamorphic zones in the Blue Ridge have been outlined by Carpenter (1970), Hadley and Nelson (1971), Rankin (1971) and Rankin and others (1972). Table 1 is a compilation of all published K/Ar mica age determinations from Grenville rocks in the Blue Ridge of North Carolina and Tennessee and in the Reading and Manhattan Prongs in New York and New Jersey.

It is apparent from Table 1 that micas from Grenville gneisses from both the Blue Ridge and the Reading and Manhattan Prongs yield ages over a range of about 500 m. y. Long and Kulp (1962) and Dallmeyer and others (1975) observed that if only K/Ar ages from samples in the western Reading Prong are considered, (this is an area of only minor and sporadic Paleozoic metamorphic effects), K/Ar and $^{40}Ar/^{39}Ar$ ages are more closely grouped, and average about 200 to 300 m. y. less than the 1100 m. y. Grenville event when the micas are thought to have formed. Offield (1967) and Baker and Buddington (1967) suggested that these K/Ar ages are the result of a post-1100 m.y. event at about 900 m. y. ago. Recently, 800 to 900 m. y. old igneous and volcanic rocks have been found in the central Appalachians (Rankin and others, 1969; Hall and others, 1975; Mose and Helenek, 1976; Sinha, 1976). The possible significance of these rocks is that they were produced by a distinct thermal event, the areal extent of which is now unknown. If so, the K/Ar ages from the western Reading Prong may record this event.

Long and Kulp (1962), Dallmeyer (1974) and Dallmeyer and others (1975) suggested that these closely grouped K/Ar ages from the western Reading Prong are post-Grenville uplift ages produced when biotite cools to a retention temperature estimated by Dallmeyer to be about 300 to 350° C. Dallmeyer suggested that biotite in these Grenville rocks remained at a temperature in excess of about 300° C for 200 m. y. as they were slowly uplifted in post-Grenville time. Long and others (1959), Tilton and others (1959) and Dallmeyer (1975b, c) observed that micas from retrograded gneisses of Grenville age in the southern Appalachians also yield K/Ar ages of about 200 to 300 m. y. younger than the Grenville event. Dallmeyer (1975a) compared this pattern to the western Reading Prong and has suggested that these are retention ages produced as the rocks were uplifted and cooled.

Although the uplift or retention age model is an attractive interpretation, it is not yet clear that an estimate of 300 to 350° C for the biotite retention temperature is correct. This estimate is based on a study of contact metamorphism by Hanson and Gast (1967). They found that the 2600 m. y. old Snowbank Stock biotite lost all its radiogenic argon where the stock was heated for more than 100,000 years to at

Table 1. A compilation of published K/Ar age determinations on micas from Grenville gneisses in the Reading and Manhattan Prongs in New York and New Jersey and K/Ar age determinations on micas from Grenville gneisses in the Blue Ridge of North Carolina and Tennessee. Symbols: R-Reading Prong gneiss; M-Manhattan Prong gneiss; B-Blue Ridge gneiss. (*) indicates that the age determination is an $^{40}Ar/^{39}Ar$ total-gas age which is analogous to a conventional K/Ar age. The age determinations come from these sources:

1. Clark and Kulp, 1968.
2. Davis and others, 1968.
3. Dallmeyer, 1975a.
4. Dallmeyer, 1975c.
5. Dallmeyer and Sutter, 1976.
6. Dallmeyer and others, 1975.
7. Kinkel and others, 1965.
8. Kulp and Eckelmann, 1961.
9. Long and others, 1959.
10. Long and Kulp, 1962.
11. Thomas, 1963.
12. Tilton and others, 1959.
13. Tilton and others, 1960.

INTENSITY OF METAMORPHISM DURING THE PALEOZOIC	GRENVILLE GNEISSES READING - MANHATTAN PRONG	GRENVILLE GNEISSES BLUE RIDGE
Less Than Biotite Grade	810 - Biot, R, 5, *	790 - Biot, B, 4
	795 - Biot, R, 6	775 - Biot, B, 4
	785 - Biot, R, 6	755 - Biot, B, 4
	770 - Biot, R, 6	680 - Biot, B, 4
	770 - Biot, R, 6	795 - Biot, B, 4, *
	800 - Biot, R, 6	775 - Biot, B, 4, *
	815 - Biot, R, 6, *	705 - Biot, B, 4, *
	810 - Biot, R, 6, *	635 - Biot, B, 4, *
	805 - Biot, R, 6, *	520 - Biot, B, 4, *
	775 - Biot, R, 6	500 - Biot, B, 4, *
	780 - Biot, R, 6	760 - Biot, B, 4, *
	820 - Biot, R, 6	810 - Biot, B, 4, *
	795 - Biot, R, 6	695 - Biot, B, 9
	840 - Biot, R, 10	800 - Biot, B, 9
	810 - Biot, R, 10	675 - Biot, B, 9
	820 - Biot, R, 10	650 - Biot, B, 9
	905 - Phlog, R, 10	640 - Biot, B, 9
	795 - Biot, R, 10	525 - Biot, B, 9
	830 - Biot, R, 10	780 - Biot, B, 12
	840 - Biot, R, 10	660 - Biot, B, 12
	850 - Biot, R, 10	770 - Biot, B, 13
	870 - Biot, R, 10	
	765 - Biot, R, 10	
	690 - Biot, R, 10	
	740 - Biot, R, 10	
	735 - Biot, R, 10	
Biotite Grade	710 - Biot, R, 5, *	295 - Biot, B, 4, *
	800 - Biot, R, 10	285 - Biot, B, 4
	810 - Biot, R, 10	325 - Biot, B, 4, *
Garnet Grade	455 - Biot, M, 1	345 - Biot, B, 4
	390 - Biot, R, 10	340 - Biot, B, 4, *
	380 - Musc, M, 10	355 - Biot, B, 9
More Than Garnet Grade	370 - Biot, M, 1	320 - Biot, B, 2
	310 - Biot, M, 1	360 - Biot, B, 8
	315 - Biot, M, 1	345 - Biot, B, 8
	360 - Biot, M, 1	305 - Biot, B, 8
	335 - Biot, M, 1	355 - Biot, B, 9
	400 - Biot, M, 1	440 - Biot, B, 9
	395 - Biot, R, 5, *	320 - Biot, B, 12
	380 - Biot, R, 5, *	
	450 - Biot, R, 10	
	360 - Biot, R, 10	
	380 - Biot, M, 10	
	405 - Biot, M, 10	
	445 - Biot, M, 10	
	360 - Biot, M, 10	
	480 - Biot, M, 10	

least 350° C by the intrusion of the Duluth Gabbro at 1050 m. y. ago. However, Hart and others (1968) found that total loss of argon from biotite occurred in the country rocks around the Eldora Stock in Colorado only where temperatures reached at least 500° C. Secondly, Hanson and Gast (1967) found that the country rock around a basic dike in the Beartooth Mountains lost only part of the biotite's argon immediately adjacent to the 15 m wide dike, even though they were heated to temperatures in excess of 700° C.

A reasonable conclusion from these studies is that the duration of elevated temperatures is as important as the temperature reached during the contact metamorphic event. However, it seems reasonable that the termal effect of a large intrusion such as the Duluth Gabbro best simulates the thermal effect of regional metamorphism. This would lead to the conclusion that the outgassing temperature of Grenville micas heated by Paleozoic metamorphism ought to be about 350° C.

An examination of the K/Ar age determinations obtained from Grenville rocks (Table 1) shows that, in fact, the outgassing temperature is not well defined. In the less-than-biotite grade of Paleozoic metamorphism, the K/Ar ages reveal that the Grenville (about 1100 m.y. old) micas lost varying amounts of their radiogenic argon. In the garnet and higher grades, the ages indicate that the Paleozoic metamorphism outgassed the Grenville micas or produced a new generation of mica. However, of the six K/Ar age determinations from biotite grade rocks, three age determinations suggest that biotite grade metamorphism is high enough to reset the K/Ar age while three age determinations suggest that it is not high enough. Phase equilibria studies (Fyfe and others, 1958; Winkler, 1967) and $^{18}O/^{16}O$ studies (Garlick and Epstein, 1967; Epstein and Taylor, 1967 O'Neill and Ghent, 1975) indicate that temperatures in excess of about 550° C are reached in the staurolite and higher grades of metamorphism. Temperatures in exeess of about 450° C are reached in the garnet zone. Temperatures in excess of about 300° C are reached in the biotite zone.

In summary, it appears that during Paleozoic regional metamorphism of these Grenville rocks, garnet grade metamorphism (more than 450° C) was enough to outgas the micas or to form new micas. Less than biotite grade metamorphism (less than 300° C was not enough to outgas the micas. Biotite grade metamorphism (300 to 450° C) sometimes was and sometimes was not high enough.

Even if it were possible to estimate the outgassing temperature, it is not clear that an outgassing temperature obtained in contact or regional metamorphism studies should be the same as the retention temperature below which a mineral retains radiogenic argon, such as when minerals are uplifted following regional metamorphism. The uplift chronologies based on K/Ar ages Dallmeyer, 1975a; Dallmeyer and others, 1975) contain the implied assumption that the outgassing temperature estimate obtained by Hanson and Gast (1967) is the same as the retention temperature.

There are no studies that define the post-metamorphic uplift or retention temperature for biotite or any other mineral. A consideration of reaction rates suggests that the outgassing temperature should be higher than the retention temperature. Alternatively, there are reasons to suspect that the outgassing temperature (during contact metamorphism) should be lower than the retention temperature (during uplift after regional metamorphism). Hydrothermal fluids associated with contact metamorphism might facilitate argon loss at temperatures lower than the temperature at which argon could be retained during regional uplift. Also, the magma chamber could act as an argon sink in areas of contact metamorphism while the partial pressure of argon might be quite high during regional metamorphism.

In any case, prior to this year, it appeared that the 800 to 900 m. y. K/Ar ages obtained from the Appalachian Grenville rocks were probably uplift (also called gas retention) ages (Dallmeyer and others, 1975). Now we have some information about igneous activity in the north central Appalachians at about 900 m. y. ago (Hall and others, 1975; Sinha, 1976; Mose and Helenek, 1976). It is now not clear what the significance is of these post-Grenville ages. *

CHRONOLOGY OF PALEOZOIC METAMORPHISM

It has long been known that the development of westward directed Paleozoic clastic wedges are related to uplifts within the Appalachian orogenic belt. The southern Appalachians (south of Virginia) produced a major clastic wedge in the Middle Ordovician and Pennsylvanian while the northern Appalachians (north of Virginia) produced clastic wedges in the Late Devonian and Late Ordovician. The central Appalachians produced part of the Middle Ordovician clastic wedge and part of the Late Devonian clastic wedge. In the northern Appalachians, Pavlides and others (1968) discussed the stratigraphic evidence for the Taconic orogeny and Boucot (1968) discussed the stratigraphic evidence for the Acadian orogeny. In the central and southern Appalachians, Colton (1970) and Meckel (1970) have discussed many important stratigraphic sequences.

However, it has been difficult clearly to correlate surficial folding, erosion and deposition within the upper few km of a geosyncline

*Recently, very significant papers concerning the interpretation and reliability of K/Ar ages determined by the $^{40}Ar/^{39}Ar$ technique have been published. A review of these papers is beyond the scope of this paper. The discussions are contained in Simmons and Hanson (1974), Hanson and others (1975), Dallmeyer (1975c), McLelland and Bence (1976) and Lanphere and Albee (1974).

Table 2. A compilation of published K/Ar age determinations on (a) micas from rocks of early Paleozoic age in the Reading and Manhattan Prongs in New York and New Jersey and (b) micas from rocks of younger Precambrian age in the Blue Ridge in North Carolina and Tennessee. All of these rocks were formed after the Grenville event and metamorphosed during the Paleozoic. Symbols: S-Manhattan schist; I-Inwood marble; OS-Ocoee series; BG-Beech granite; AF-Ashe formation. See the caption under Table 1 for the sources of the K/Ar data. (*) indicates $^{40}Ar/^{39}Ar$ total-gas ages.

INTENSITY OF METAMORPHISM DURING THE PALEOZOIC	ROCKS OF EARLY PALEOZOIC AGE READING - MANHATTAN PRONG	ROCKS OF YOUNGER PRECAMBRIAN AGE BLUE RIDGE
Less Than Biotite Grade	(No Data)	(No Data)
Biotite Grade	(No Data)	435 - Biot, OS, 9
Garnet Grade	405 - Biot, S, 10 410 - Biot, S, 10	380 - Biot, BG, 12
More Than Garnet Grade	330 - Biot, S, 1 440 - Biot, S, 1 345 - Biot, S, 1 360 - Biot, S, 10 380 - Biot, S, 10 355 - Biot, S, 10 350 - Phlog, I, 10 355 - Phlog, I, 10	345 - Biot, OS, 3, * 350 - Biot, OS, 3, * 350 - Biot, OS, 3 345 - Mica, AF, 7 335 - Mica, AF, 7 380 - Biot, OS, 8 375 - Biot, OS, 8 365 - Biot, OS, 8 325 - Musc, OS, 8 345 - Biot, OS, 9 325 - Musc, OS, 9 340 - Mica, OS, 9 465 - Mica, AF, 11

with metamorphic activity occurring at depths of more than 10 km. Rodgers (1971) has pointed out that the thick sequence of Middle Ordovician sedimentary rocks west of the Blue Ridge contains fragments of Cambrian to early Middle Ordovician limestones and sandstones that were deformed and metamorphosed prior to their deposition in the clastic wedge. This clearly indicates that in this area, some metamorphic activity occurred in early Middle Ordovician time. In general, the chronology of surficial events has not been closely correlated with the chronology of metamorphism. Instead, surficial events which include folding, uplift and erosion are regarded as possibly occurring after a metamorphic event.

It is apparent from Table 2 that micas from rocks which were first metamorphosed during the Paleozoic yield K/Ar ages over a range of about 100 m.y. The central portion of the Blue Ridge is the most intensely metamorphosed and has yielded many samples which have K/Ar mica ages of about 350 m.y. Many workers have concluded that these micas lost all their radiogenic argon during a metamorphic event and that this age most closely records the time of metamorphism (Long and others, 1959; Kulp and Eckelmann, 1961). A similar pattern in the northeastern Reading Prong and in the Manhattan Prong has been interpreted by others to be the result of a Late Ordovician event (Taconic orogeny, about 470 m.y. ago) and a Late Devonian event (Acadian, about 360 m. y. ago) (Long and Kulp, 1962; Clark and Kulp, 1968). The older Taconic ages were found in lower grade metamorphic rocks in the Manhattan Prong which were considered to have been outgassed during the Taconic event but not during the Acadian event; the younger Acadian ages are found in higher grade rocks and were considered to be from rocks outgassed by both the Taconic and Acadian events (Long and Kulp, 1962; Clark and Kulp, 1968).

Recently, an alternative interpretation has come into favor. Butler (1972) pointed out that K/Ar age determinations may yield the time of Paleozoic metamorphism only in a rather restricted case. He suggested that since the younger Precambrian sediments in the Blue Ridge were first progressively metamorphosed in Paleozoic time, the biotite from these progressively metamorphosed rocks from within the biotite zone should yield an age closest to the time of Paleozoic metamorphism. Biotite in this zone formed as a result of biotite grade metamorphism and presumably cooled down below the argon retention temperature relatively quickly after the event compared to mica from higher metamorphic zones. Following this interpretation, the Paleozoic metamorphism which formed the metamorphic rocks in the Blue Ridge occurred prior to the Middle Ordovician, at least 430 m.y. ago (Butler, 1972). K/Ar age determinations from micas which grew during prograde metamorphism in the garnet and higher zones are interpreted as minimum ages for the Paleozoic event because micas in these zones continued to lose radiogenic argon until the area was uplifted and cooled. Dallmeyer (1975c) interpreted $^{40}Ar/^{39}Ar$ ages to indicate that the rocks were metamorphosed by a single event about 480 m. y. ago.

Dietrich and others (1969) and Odom and Fullagar (1973) found ages of about 330-360 m. y. for sheared rocks from fault zones in the Blue Ridge. If these ages are the times of faulting, and if the faulting produced uplift which cooled the Blue Ridge, the Paleozoic progressively metamorphosed rocks in the garnet and higher zones must have remained in excess of the argon retention temperature for about 100 m.y.

A similar interpretation can be made about the Reading and Manhattan Prongs. In this area, K/Ar ages from progressively metamorphosed rocks in the garnet zone of Paleozoic metamorphism (no biotite zone data) indicate that the metamorphism occurred at least 410 m. y. ago. McLelland and Bence (1976) made the suggestion, based on $^{40}Ar/^{39}Ar$ ages on biotite from the Dutchess County area just north of the eastern Reading Prong, that the metamorphism was a single event which occurred about 435 m. y. ago. Dallmeyer and Sutter (1976) also used $^{40}Ar/^{39}Ar$ data from the eastern Reading Prong to suggest that the metamorphism occurred as a single event at about 470 m. y. ago.

Mose and others (1976) studied the 371 ± 14 m. y. old Peekskill pluton (Figure 2) which was emplaced along the western end of a major fault zone between the Reading and Manhattan Prongs. The pluton was emplaced in the fault zone but was not recrystallized by the Paleozoic metamorphism. Their study showed that the metamorphic rocks along the western edge of the Manhattan Prong cooled earlier than the rocks along the center of the area. K/Ar ages along the center of the Manhattan Prong are about 350 m. y. These ages, and the 410-480 m. y. estimate for the time of metamorphism, indicate that the high-grade rocks in the Manhattan Prong, like those in the Blue Ridge, remained at a temperature in excess of the argon retention temperature for about 100 m. y.

Many geologists now regard the metamorphic activity which affected the Manhattan Prong, eastern Reading Prong and Blue Ridge to be a single Middle Ordovician event, commonly called the Taconic orogeny. It is, however, not clear to what extent the Devonian age Acadian orogeny (which is recorded in the rocks of New England and in the Piedmont of the central and southern Appalachians) was felt in these areas.

A major debate has developed concerning the evidence of an Acadian event in the Manhattan Prong, and because of its proximity, in the eastern Reading Prong. Based on $^{40}Ar/^{39}Ar$ ages, Dallmeyer and Sutter (1976) and McLelland and Bence (1976) support the concept of a single metamorphic event which occurred in Ordovician time (Taconic orogeny). However, others such as Fisher and Warthin (1976) support the concept of a second event which occurred in Devonian time (Acadian orogeny).

There are several factors which, when taken together, suggest that the Acadian orogeny that was strongly felt in New England was accompanied by significant activity in the Manhattan Prong. The Peekskill pluton (Mose and others, 1976), the Bedford, New York pegmatite and the Branchville, Connecticut pegmatite (Clark and Kulp, 1968) all formed in Devonian time, indicating that the Manhattan Prong and probably the eastern Reading Prong were at high enough temperatures to generate igneous intrusions. The Cortlandt complex (Figure 2) intruded Cambrian to Middle Ordovician metasedimentary rocks in the Manhattan Prong about 440 m. y. ago (Long and Kulp, 1962; Dallmeyer, 1975d). Since it truncates refolded isoclinal folds in the metamorphosed country rocks, it was clearly emplaced after the Taconic metamorphic event (Ratcliffe, 1968). However, the eastern end of the complex has subsequently been folded (Tracy, 1967). Chase and Brock (1976) have shown that the Paleozoic sillimanite isograd in the Croton Falls area (north central Manhattan Prong) appears folded, indicating that structural deformation occurred after the Taconic orogeny. The Croton Falls complex (Figure 2) also intruded the early Paleozoic metasedimentary rocks in the Manhattan Prong and was subsequently intruded by granite veins and plutons at 407 ± 25 m. y. (95% confidence level; Brock and Mose in prep.). Field relations reveal that the complex and the granites have been metamorphosed during a clearly post-Taconic event.

Post-Taconic deformation and igneous activity have clearly occurred in the Manhattan Prong and probably also in the eastern Reading Prong. However, it is not yet possible to decide if this activity represents a distinct Acadian event, preceded by a post-Taconic cooling, or if the activity merely represents adjustments which occurred during a long mid-Paleozoic interval of slow post-Taconic uplift and cooling. Most workers today favor the second interpretation, relating the uplift to the folding of Silurian and Devonian rocks west of the Reading Prong. However, it appears more likely that the metamorphism of Silurian and Devonian sedimentary rocks east of Camerons Line in New England was accompanied by recrystallization of metamorphic rocks in the Manhattan Prong and eastern Reading Prong.

CONCLUSIONS

Three generalizations may be made about argon outgassing of micas in Precambrian rocks due to Paleozoic metamorphism. Temperatures in excess of about 450° C are clearly high enough for outgassing biotite around large intrusive bodies and in areas of regional metamorphism. Second, temperatures of less than 300° C are clearly not enough to cause complete outgassing. Third, temperatures of about 300 to 350° C may or may not be high enough to outgas biotite. In short, garnet grade metamorphism is high enough to effect complete outgassing, less than biotite grade is not high enough, and biotite grade may be high enough.

In general, the argon outgassing temperature during contact and regional metamorphism is similar, though probably not identical, to the argon retention temperature during uplift after regional metamorphism. Micas in rocks probably retain radiogenic argon after they cool to the temperatures found in the biotite grade of regional metamorphism. An examination of the Appalachian K/Ar data suggests that the retention temperature during uplift varies between 300 to 450° C.

Dallmeyer (1975a) and Dallmeyer and others (1975) have used K/Ar age determinations to produce what is called an uplift chronology (the time at which biotite and hornblende in metamorphic rocks cool enough to retain radiogenic argon). However, it is important to note that the uplift chronology is based on a knowledge of the retention temperatures of biotite and hornblende. Estimates of uplift chronology based on K/Ar age determinations assume that the retention temperatures following regional metamorphism are the same as the outgassing temperature of biotite (300° C) and hornblende (550° C) obtained from a study of contact metamorphism (Hanson and Gast, 1967). This assumption may be incorrect.

Several observations result from an examination of the available radiometric data concerning the Precambrian metamorphic history of the Blue Ridge in North Carolina and Tennessee and the metamorphic history of the Reading and Manhattan Prongs in New York and New Jersey. U/Pb and Rb/Sr data have clearly shown that there was a period of widespread igneous and metamorphic activity between about 1050 and 1150 m.y. ago (the Grenville orogeny). Recent U/Pb and Rb/Sr data suggest that plutonic and volcanic activity also occurred in the central Appalachians between about 800 and 900 m.y. ago, but the extent and intensity of metamorphic activity at the time is not known. The 800 to 900 m.y. K/Ar age determinations from New York, New Jersey, North Carolina and Tennessee have been interpreted as evidence for a thermal event at this time. Other workers have suggested that these ages record the cooling of rocks as they were uplifted after the Grenville orogeny. At the present time, the uplift-and-cooling model is favored, pending a clearer understanding of the 800 to 900 m.y. apparent ages.

K/Ar age determinations on micas from post-Grenville rocks in the Blue Ridge, eastern Reading Prong and Manhattan Prong reach minimums of about 350 m.y. in the high grade zones of Paleozoic metamorphism. These were at one time interpreted to mean that a metamorphic event occurred at this time. A more recent interpretation using only data from lower grade zones of metamorphism (biotite or garnet) is that the progressive metamorphism occurred at least 435 m.y. ago in the Blue Ridge and at least 410 m.y. ago in the Manhattan Prong (Taconic orogeny). Following this interpretation, K/Ar ages from high grade rocks and studies of faulting chronology suggest that rocks in the higher zones remained in excess of their argon retention temperature for micas until about 350 m.y. ago. However, it is suggested in this paper that, at least in the Manhattan and eastern Reading Prongs, there is clear evidence of a distinct post-Taconic and probably Aradian orogeny. This evidence includes the presence of Devonian intrusive activity, refolding of Taconic folds, and recrystallization of post-Taconic igneous rocks.

REFERENCES CITED

Baker, D. R., and A. F. Buddington, 1970, Geology and magnetite deposts of the Franklin quadrangle and part of the Hamburg quadrangle, New Jersey, U. S. Geol. Survey Prof. Paper 638, 73 p.

Boucot, A. J., 1968, Silurian and Devonian of the northern Appalachians, p. 83-94 in E. Zen, W. S. White, J. B. Hadley, and J. B. Thompson, Eds. Studies of Appalachian Geology: Northern and Maritime. John Wiley-Interscience, New York.

Butler, J. R., 1972, Age of Paleozoic regional metamorphism in the Carolinas, Georgia and Tennessee, southern Appalachians. Am. Jour. Sci., v. 272, p. 319-333.

Butler, J. R., 1973, Paleozoic deformation and metamorphism in part of the Blue Ridge thrust sheet, North Carolina. Am. Jour. Sci., v. 273A, p. 72-88.

Carpenter, R. H., 1970, Metamorphic history of the Blue Ridge province of Tennessee and North Carolina. Geol. Soc. Am. Bull., v. 81, p. 749-762.

Chase, P. J., and P. W. G. Brock, 1970, Sillimanite and sillimanite-orthoclase isograds in the Croton Falls area, southeast New York (abs.). Geol. Soc. America Ann. Mtg., Northeastern-Southeastern Section, v. 8, no. 2, p. 149-150.

Clark, G. S., and J. L. Kulp, 1968, Isotopic age study of metamorphism and intrusion in western Connecticut and southeastern New York, Am. Jour. Sci., v. 266, p. 865-894.

Colton, G. W., 1970, The Appalachian Basin - its depositional sequences and their geologic relationships. p. 5-48 in G. W. Fisher, F. J. Pettijohn, J. C. Reed, and K. N. Weaver, eds. Studies in Appalachian Geology: Central and Southern. John Wiley-Interscience, New York.

Dallmeyer, R. D., 1974, Metamorphic history of the northern Reading Prong, New York and New Jersey. Jour. Petrology, v. 15, p. 325-359.

__________, 1975a, Incremental $^{40}Ar/^{39}Ar$ ages of biotite and hornblende from retrograded basement gneisses of the southern Blue Ridge: their bearing on the age of Paleozoic metamorphism. Am. Jour. Sci., v. 275, p. 444-460.

__________, 1975b, $^{40}Ar/^{39}Ar$ age spectra of biotite from Grenville basement gneisses in northwest Georgia. Geol. Soc. Bull., v. 86, p. 1740-1744.

__________, 1975c, $^{40}Ar/^{39}Ar$ ages of biotite and hornblende from a progressively remetamorphosed terrane: their bearing on interpretation of release spectra. Geochim. et Cosmochim. Acta, v. 39, p. 1655-1669.

__________, 1975d, $^{40}Ar/^{39}Ar$ release spectra of biotite and hornblende from the Cortlandt and Rosetown plutons, New York and their regional implications. Jour. Geology, v. 83, p. 629-643.

Dallmeyer, R. D., J. F. Sutter, and D. J. Baker, 1975, Incremental $^{40}Ar/^{39}Ar$ ages of biotite and hornblende from the northeastern Reading Prong: their bearing on late Proterozoic thermal and tectonic history. Geol. Soc. Am. Bull., v. 86, p. 1435-1443.

Dallmeyer, R. E. and J. F. Sutter, 1976, $^{40}Ar/^{39}Ar$ incremental-release ages of biotite and hornblende from variably retrograded basement gneisses of the northeasternmost Reading Prong, New York: Their bearing on early Paleozoic metamorphic history. Am. Jour. Sci., v. 276, p. 731-747.

Davis, G. L., G. R. Tilton, L. T. Aldrich, and G. W. Wetherill, 1958, The age of rocks and minerals. Carnegie Institution of Washington Yearbook, v. 57, p. 176-181.

Davis, G. L., G. R. Tilton, and G. W. Wetherill, 1962, Mineral ages from the Appalachian province in North Carolina and Tennessee: Jour. Geophys. Res., v. 67, p. 1987-1996.

Dietrich, R. V., P. D. Fullagar and M. L. Bottino, 1969, K/Ar and Rb/Sr dating of tectonic events in the Appalachian of southwestern Virginia. Geol. Soc. Am. Bull. v 80, p. 307-314.

Doe, B. R., 1970, Lead isotopes. Springer-Verlag, New York, 137 p.

Epstein, S., and H. P. Taylor, 1967, Variation of $^{18}O/^{16}O$ in minerals and rocks. p. 29-62 in Abelson, P. H., editor. Researches in Geochemistry, John Wiley and Sons, Inc., New York.

Faure, G., and J. L. Powell, 1972, Strontium isotope geology. Springer-Verlag, New York, 188 p.

Fisher, D. W., and A. S. Warthin, 1976, Stratigraphic and structural geology in western Dutchess County, New York. New York State Geological Association Field Guide Book, B-6-1 to B-6-36.

Fullagar, P. D., and A. L. Odom, 1973, Geochronology of Precambrian gneisses in the Blue Ridge province of northwestern North Carolina and adjacent parts of Virginia and Tennessee. Geol. Soc. Am. Bull., v. 84, p. 3065-3080.

Fyfe, W. S., F. J. Turner, and J. Verhoogen, 1958, Metamorphic reactions and metamorphic facies. Geol. Soc. of America Memoir 73.

Garlick, G. D., and S. Epstein, 1967, Oxygen isotope ratios in coexisting minerals of regionally metamorphosed rocks. Geochim. et Cosmochim. Acta, v. 31, p. 181-214.

Grauert, B., and L. M. Hall, 1973, Age and origin of zircons from metamorphic rocks in the Manhattan Prong, White Plains area, southeastern New York. Carnegie Institution of Washington Yearbook, v. 72, p. 293-297.

Hadley, J. B., and A. E. Nelson, 1971, Geologic map of the Knoxville quadrangle, North Carolina, Tennessee and South Carolina. U. S. Geol. Surv. Misc. Geol. Inv. Map I-654, scale 1/250,000.

Hall, L. M., 1968, Times of origin and deformation of bedrock in the Manhattan Prong, p. 117-127 in E. Zen, W. S. White, J. B. Hadley, and J. B. Thompson, eds. Studies of Appalachian Geology: Northern and Maritime, John Wiley-Interscience, New York.

Hall, L. M., H. L. Helenek, R. A. Jackson, K. G. Caldwell, D. G. Mose and D. P. Murray, 1975, Some basement rocks from Bear Mountain to the Housatonic Highlands. New York Intercollegiate Geological Conference Guidebook, City College, CUNY, p. 1-29.

Hanson, G. N., and P. W. Gast, 1967, Kinetic studies in contact metamorphic zones, Geochim. et Cosmochim. Acta, v. 31, p. 1119-1153.

Hanson, G. N., K. R. Simmons and A. E. Bence, 1975, $^{40}Ar/^{39}Ar$ spectrum ages for biotite, hornblende and muscovite in a contact metamorphic zone. Geochim. Cosmochim. Acta, v. 39, p. 1269-1277.

Hart, S. R., G. L. Davis, R. H. Steiger and G. R. Tilton, 1968, A comparison of the isotopic mineral age variations and petrologic changes induced by contact metamorphism. p. 73-110 _in_ E. I. Hamilton and R. M. Farquhar, eds., Radiometric Dating for Geologists. John Wiley-Interscience, New York.

Hatcher, R. D., 1975, Second Penrose Conference: The Brevard Zone. Geology, v. 3, p. 149-152.

Kinkel, A. R., H. H. Thomas, R. F. Marvin and F. G. Walthall, 1965, Age and metamorphism of some massive sulfide deposits in Virginia, North Carolina and Tennessee. Geochim. et Cosmochim. Acta, v. 29, p. 717-724.

Kulp, J. L., and D. L. Eckelmann, 1961, Potassium-argon isotopic ages on micas from the southern Appalachians. Ann. N. Y. Acad. Sci., v. 91, p. 408-419.

Lanphere, M. A., and A. L. Albee, 1974, $^{40}Ar/^{39}Ar$ age measurements in the Worcester mountains: Evidence of Ordovician and Devonian metamorphic events in northern Vermont. Am. Jour. Sci., v. 274, p. 545-555.

Long, L. E., J. L. Kulp and F. D. Eckelmann, 1959, Chronology of major metamorphic events in the southeastern United States. Am. Jour. Sci., v. 257, p. 585-603.

Long, L. E., and J. L. Kulp, 1962, Isotopic age study of the metamorphic history of the Manhattan and Reading Prongs. Geol. Soc. Am. Bull., v. 73, p. 969-996.

Long, L. E., 1969, Whole-rock Rb-Sr age of the Yonkers Gneiss, Manhattan Prong. Geol. Soc. Am. Bull., v. 8, p. 2087-2090.

McLelland, J. M. and A. E. Bence, 1976, Progressive metamorphism in Western County, New York. New York State Geological Association Field Guide Book, B-7-1 to B-7-27.

Meckel, L. D., 1970, Paleozoic alluvial deposition in the central Appalachians: A Summary, p. 49-68 _in_ G. W. Fisher, F. J. Pettijohn, J. C. Reed and K. N. Weaver, eds. Studies of Appalachian Geology: Central and Southern. John Wiley-Interscience, New York.

Meyerhoff, H. A., 1972, Postorogenic development of the Appalachians. Geol. Soc. Am. Bull., v. 83, p. 1709-1728.

Mose, D. G., and J. Hayes, 1975, Avalonian igneous activity in the Manhattan Prong, southeastern New York. Geol. Soc. Am. Bull. v. 86, p. 929-932.

Mose, D. G., N. M. Ratcliffe, A. L. Odom and J. Hayes, 1976, Rb-Sr geochronology and tectonic setting of the Peekskill pluton, southeastern New York. Geol. Soc. Am. Bull., v. 87, p. 361-365.

Mose, D. G. and H. L. Helenek, 1976, Origin, age and mode of emplacement of the Canada Hill granite, Hudson Highlands, New York (Abs.): Geol. Soc. Am. Abstracts with Programs, v. 8, p. 233-234.

Odom, A. L. and P. D. Fullagar, 1973, Geochronologic and tectonic relationships between the inner Piedmont, Brevard Zone and Blue Ridge Belts, North Carolina. Am. Jour. Sci., v. 273A, p. 133-149.

Offield, T. W., 1967, Bedrock geology of the Goshen-Greenwood Lake area, New York, N. Y. State Mus. Map and Folio Series, No. 9, 77 p.

O'Neill, J. R. and E. D. Ghent, 1975, Stable isotope study of co-existing metamorphic minerals from Espanade Range, British Columbia. Geol. Soc. Am. Bull., v. 86, p. 1708-1712.

Pavlides, L., A. J. Boucot and W. B. Skidmore, 1968, Stratigraphic evidence for the Taconic orogeny in the northern Appalachians. p. 61-82 in E. Zen, W. S. White, J. B. Hadley and J. G. Thompson, eds. Studies of Appalachian Geology: Northern and Maritime. John Wiley-Interscience, New York.

Rankin, D. W., T. W. Stern, J. C. Reed and M. F. Newell, 1969, Zircon ages of felsic volcanic rocks in the upper Precambrian of the Blue Ridge, Appalachian Mountains. Science, v. 166, p. 741-744.

Rankin, D. W., 1970, Stratigraphy and structure of Precambrian rocks in northwestern North Carolina. p. 227-245 in G. W. Fisher, F. J. Pettijohn, J. S. Reed and K. N. Weaver, eds. Studies of Appalachian Geology: Central and Southern. John Wiley-Interscience, New York.

________, 1971, Guide to the geology of the Blue Ridge in southwestern Virginia and adjacent North Carolina. Guidebook to Appalachian tectonics and sulfide mineralization in southwestern Virginia: Va. Polytech. Inst., Dept. Geol. Sci., v. 5, p. 39-86.

Rankin, D. W., G. H. Espenshade and R. B. Neuman, 1972, Geologic Map of the west half of the Winston-Salem Quadrangle, North Carolina, Virginia and Tennessee. U. S. Geol. Surv. Misc. Geol. Inv. Map, I-709-A, scale 1/250,000.

Ratcliffe, N. M., 1968, Contact relations of the Cortlandt complex at Stony Point, New York, and their regional implications. Geol. Soc. Am. Bull, v. 79, p. 777-786.

Ratcliffe, N. M., R. L. Armstrong, H-t Chai and R. G. Seneshal, 1972, K-Ar and Rb-Sr geochronology of the Canopus pluton, Hudson Highlands, New York, Geol. Soc. Am. Bull. 83:532-530.

Rogers, J., 1971, The Taconic orogeny. Geol. Soc. Am. Bull., v. 82, p. 1141-1178.

Simmons, K. and G. N. Hanson, 1974, $^{40}Ar/^{39}Ar$ spectrum ages for biotite, muscovite and hornblende in a contact metamorphic zone. Trans. Am. Geophys. Union, v. 55, p. 466.

Sinha, A. K., 1976, Timing of metamorphic and igneous events in the central Piedmont and Blue Ridge (Abs.): Geol. Soc. America Abstracts with Programs, v. 8, p. 267.

Thomas, H. H., 1963, Isotopic ages on coexisting hornblende, mica and feldspar. Trans. Am. Geophys. Union, v. 44, p. 110-111.

Tilton, G. R., G. L. Davis, G. W. Wetherill, L. T. Aldrich and E. Jager, 1959, The ages of rocks and minerals. Carnegie Institution of Washington Yearbook. v. 58, p. 170-178.

Tilton, G. R., G. W. Wetherill, G. L. Davis and M. N. Bass, 1960, 1000 million year old minerals from the eastern United States and Canada. Jour. Geophys. Res., v. 65, p. 4173-4179.

Tracy, R. J., 1970, The petrology of the eastern end of the Cortlandt complex, Westchester County, New York, M. S. thesis (unpub.), Brown University, 45 p.

Winkler, H. G. F., 1967, Petrogenesis of metamorphic rocks. Springer-Verlag, New York, 237 p.

FLINT HILL GNEISS - A DEFINITION

By

Michael T. Lukert
Edinboro State College
Edinboro, Pennsylvania 16412

E. B. Nuckols
Consolidated Gas Supply Corporation
Clarksburg, West Virginia 26301

and

James W. Clarke
U. S. Geological Survey
Reston, Virginia 22092

ABSTRACT

The Flint Hill Gneiss is a strongly foliated biotite meta-igneous rock, which has now been mapped in the Flint Hill, Linden, and Orlean quadrangles in the core of the Blue Ridge anticlinorium in northern Virginia. A type locality is designated and described. Veins of blue quartz are common within this unit but do not occur in other units of the areas mapped.

Zircon from the Flint Hill Gneiss has yielded a concordant age of 1081 m. y. This date may represent the time at which the original igneous rock was emplaced and metamorphosed, or it may represent a time when the uranium and thorium clocks for the zircon were reset completely. Igneous events after the imposition of the strong foliation on the Flint Hill Gneiss were: (1) intrusion of Robertson River Formation, which has a zircon date of 732 m. y.; and (2) intrusion of basaltic dikes (now greenstone), which may be Catoctin.

INTRODUCTION

A distinctive biotite gneiss is abundant in the older Precambrian core of the Blue Ridge anticlinorium in Rappahannock and Fauquier Counties, Virginia. The unit has been mapped in the Flint Hill and Linden quadrangles by Michael T. Lukert and E. B. Nuckols for the Virginia Division of Mineral Resources, and in the Orlean quadrangle by James W. Clarke for the U. S. Geological Survey. This biotite gneiss, described in this paper, is hereby designated Flint Hill Gneiss.

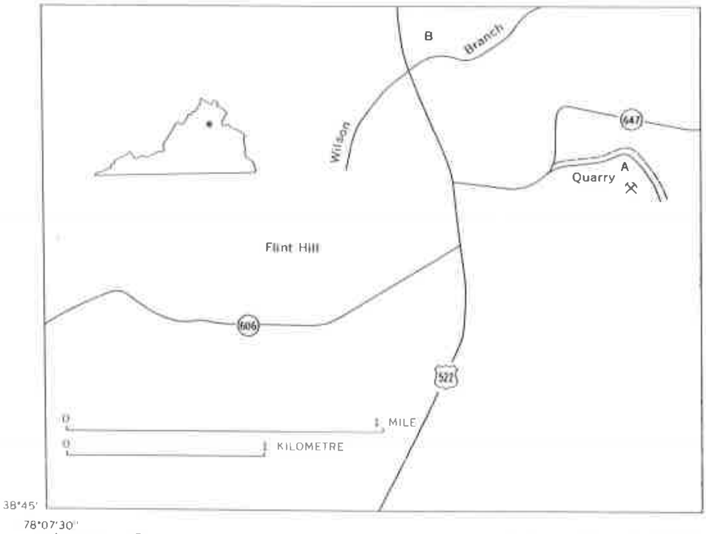


Figure 1. Location of exposures of the Flint Hill Gneiss. A - type locality; b - excellent natural exposures.

TYPE LOCALITY

The type locality for the Flint Hill Gneiss is a large quarry half a mile east of the community of Flint Hill in the southwest quadrant of the Flint Hill quadrangle in Rappahannock County, Virginia. The U.T.M. coordinates of this quarry are 752500E, 4294500N (Figure 1, area A). Excellent natural exposures are just east of U.S. Route 522 north of Wilson Branch half a mile north of Flint Hill (Figure 1, area B).

AREAL DISTRIBUTION

The distribution of the Flint Hill Gneiss within the Flint Hill, Linden, and Orlean quadrangles is shown on Figure 2.

Within the Linden and Flint Hill quadrangles, the Flint Hill Gneiss underlies an area of about 15 square miles (38 sq km). It was designated in this area as "layered gneiss" by Lukert and Nuckols (1976). In these two quadrangles the outcrop area is a northeast-trending strip having a maximum width of approximately 3.4 miles (5.4 km). The Flint Hill Gneiss of the Linden and Flint Hill quadrangles is separated from that of the Orlean quadrangle by a belt of gneissic and granitic rocks. In the Orlean quadrangle, the Flint Hill Gneiss underlies five separate areas, which together have a total area of about 4 square miles (10 sq km). Where outcrops are lacking, the presence of Flint Hill Gneiss beneath the soil cover can be inferred in many places by chips of blue quartz on the surface (veins of blue quartz are distinctive in the

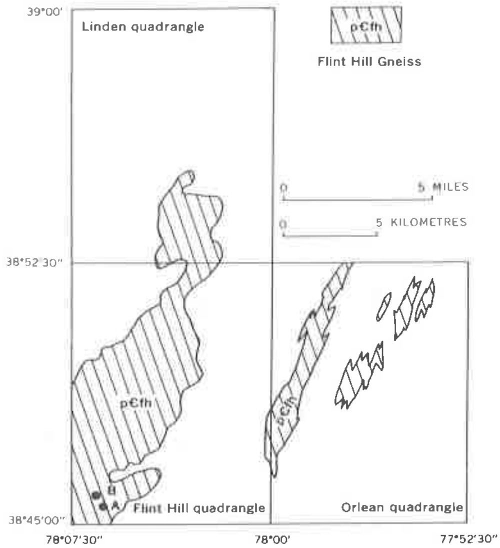


Figure 2. Distribution of Flint Hill Gneiss in Linden, Flint Hill, and Orlean quadrangles, Virginia. A-type locality; B - excellent natural exposures.

Flint Hill). Also, the Flint Hill Gneiss commonly yields redder soil than that derived from the surrounding rocks.

APPEARANCE IN OUTCROP

The Flint Hill Gneiss crops out as low exposures on the tops and sides of hills; it also is found commonly as pavement outcrops. The natural exposure in Figure 3 is such a pavement outcrop.

The fresh rock is irregularly layered at the type locality and in most outcrops. Dark layers rich in biotite are generally 1-2 mm thick and generally continuous, although these dark layers compose less than half the rock. Light layers are discontinuous aggregates of quartz and feldspar 2-10 mm thick and 5-30 mm long. In some places this layering is distorted. Figure 4 shows typical layering on a quarried block of fresh rock from the type locality. Other examples of layering can be seen in Figures 3 and 5. In many outcrops, the layering is more persistent, giving the rock a banded appearance. In other places, the light discontinuous aggregates grades into augen.

Most weathered surfaces are differentially etched, sialic aggregates 2-5 mm in diameter protruding above the mafic layers.

Figure 3. Natural exposure of Flint Hill Gneiss. Dark (D) and light (L) thick layers of discontinuously banded material alternate. Crosscutting the foliation of the gneiss are bodies of blue quartz-bearing leucocratic granite (G). Exposure is just east of U. S. Route 522 north of Wilson Branch half a mile north of Flint Hill (Figure 1, area B).

Veins of blue quartz are common; these are generally 10-25 cm wide, 25-100 cm long, and irregular along their edges. Generally one such blue-quartz vein is found for each 20-30 m^2 of outcrop.

PETROGRAPHY

In thin section, the Flint Hill Gneiss shows strong foliation (Figure 6).

Quartz is irregular in outline where it is in juxtaposition with other quartz grains. Grain size is 0.1 - 1.0 mm. Individual grains are highly strained.

Microcline occurs as equant to elongate irregular grains 1-3 mm in diameter. Most grains show grid twinning and are microperthitic. Microcline is commonly rimmed by albite.

Plagioclase is oligoclase, An-25 to An-28, on the basis of

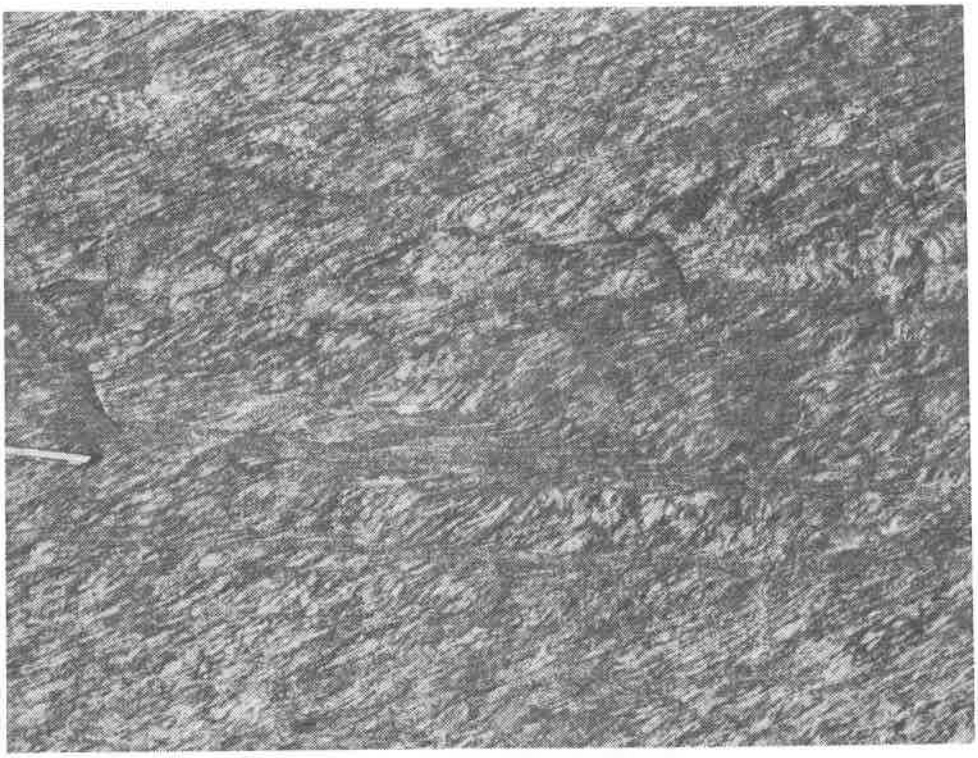


Figure 4. Typical fresh Flint Hill Gneiss at the type locality at the Flint Hill quarry half a mile east of the community of Flint Hill in the Flint Hill quadrangle.

extinction angle. Zoning is absent. Twinning is predominantly of the albite and Carlsbad variety, although pericline twins also occur. Most plagioclase grains are strongly sericitized.

Biotite is pleochroic from pale brownish green to colorless; flakes are rarely longer than 0.3 mm.

Sericite is abundant both as a replacement of plagioclase grains and of plagioclase exsolution lamellae in microcline.

Ilmenite grains 0.2 mm in diameter surrounded by coronas of sphene 0.2 mm wide are common (Figure 7).

Garnet is common along the western border of the Flint Hill Gneiss in the Linden and Flint Hill quadrangles.

Chlorite is present in some samples, probably formed by retrograde metamorphism of garnet in many instances.

Accessory minerals include zircon, apatite, and epidote.

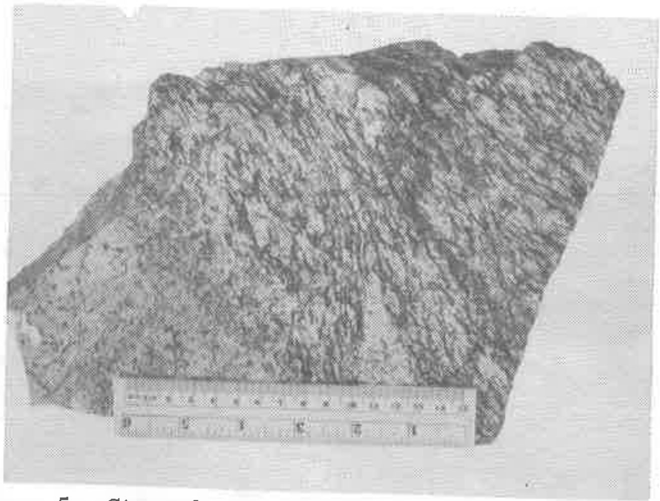


Figure 5. Strongly foliated Flint Hill Gneiss is invaded by granite, the Robertson River Formation of Allen (1963). The sample is from an area about a mile northwest of Jerrys Shop in the Orlean quadrangle.

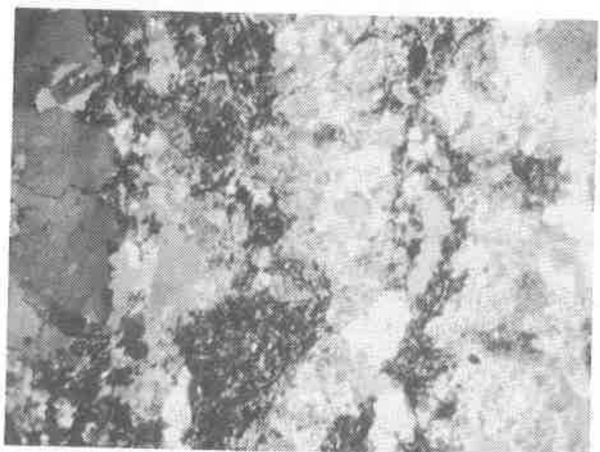


Figure 6. Photomicrograph of Flint Hill Gneiss from type locality (20x). The dark layers are composed largely of biotite; abundant quartz and minor microcline, sphene, ilmenite, apatite, epidote, and zircon are also present. The light layers consist of quartz, microcline, and plagioclase.

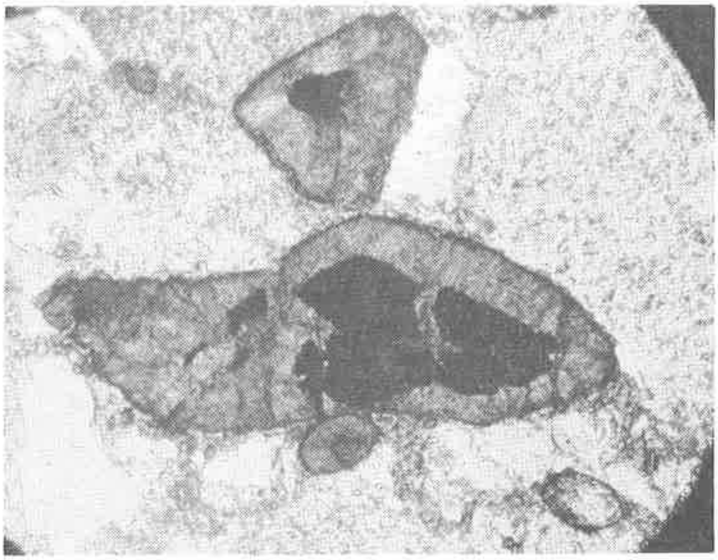


Figure 7. Photomicrograph of Flint Hill Gneiss showing coronas of sphene around ilmenite, (66 x).

Within the western outcrop belt, cataclasis has had considerable effect on the Flint Hill Gneiss. Mortar quartz, bent and deformed plagioclase twin lamellae, and fractured garnet are common. Cataclastic material is a conspicuous component of samples from this area. The abundance and mineralogy of the cataclastic material is quite variable. Granulated quartz and fine green biotite flakes are consistently present, and varying amounts of sericite and epidote are commonly present.

Modes of several samples of Flint Hill Gneiss are listed in Table 1.

AGE RELATIONSHIPS

The relative homogeneity of the Flint Hill Gneiss and its granitic composition suggest that it was originally an igneous rock. If so, an orogenic event after its emplacement resulted in the fine layering seen in Figure 4.

The thick layers shown in Figure 3 appear to have been caused by migmatization that postdated formation of the fine layering; dark septa of the more mafic thick layers fray out into material of the more silicic thick layers. Formation of the fine and of the thick layering, however, may have been part of the same event.

The next event was deformation of the layering (Figure 4). In the southwest part of the Orlean quadrangle, strong deformation and

Table 1. Modal Analyses (in volume percent) of the Flint Hill Gneiss (X, trace amounts; --, not present).

Sample Repository Number ( ) (Virginia Division of Mineral Resources)	R-5034	R-5037	R-5061	R-5066	R-5092	R-5093
Quartz	29.7	33.0	30.9	29.0	26.0	22.4
Potassium feldspar	42.7	36.3	35.8	37.2	24.2	26.7
Plagioclase	9.6	2.9	18.1	17.0	27.2	15.8
Cataclastic material	1.1	21.4	8.7	8.6	5.0	26.0
Biotite	3.5	X	5.1	7.3	14.5	7.4
Muscovite and sericite	11.3	--	X	X	1.6	X
Epidote	X	X	X	X	X	X
Apatite	X	X	X	--	X	X
Opaques Minerals	1.3	X	X	X	X	X
Sphene	--	X	X	X	X	X
Zircon	X	X	X	X	X	X
Chlorite	X	4.1	--	X	--	X
An content of plagioclase	25	28	28	28	25	28

Sample localities

R-5034	1/10 mile (0.16 km.) due south of Harrell's Corner (Linden quad.).
R-5037	1 mile (1.61 km.) northwest of Belle Meade and 3/10 mile (0.48 km.) north of State Rte. 55 (Linden quad.).
R-5061	3/5 mile (0.97 km.) north of Flint Hill on U.S. Rte. 522 (Flint Hill quad.).
R-5066	1/10 mile (0.16 km.) west of the intersection of State Roads 630 & 628 (Flint Hill quad.).
R-5092	Flint Hill quarry (Flint Hill quad.).
R-5093	3/5 mile (0.97 km.) west of Hitch on State Road 635 (Flint Hill quad.).

brecciation of the layered gneiss is associated with a change in regional strike of the Flint Hill Gneiss; consequently, this event may represent a separate orogeny.

Blue-quartz-bearing granite and pegmatite and blue-quartz veins were emplaced after the layering had been deformed. These blue-quartz-bearing bodies may be associated with the Old Rag Granite of Furcron (1939) from which zircon ages of 1140 m. y. (Grenville orogeny) have been obtained by the senior author (M. T. L.). Consequently, the original crystallization of a granitic magma, the formation of its strong foliation, its migmatization, and, finally, the deformation of the layering could have taken place prior to 1140 m. y. A date of 1081 m. y. has now been obtained on the Flint Hill Gneiss, however. This suggests that the entire process may have taken place during the Grenville, or that the uranium-thorium clocks of the rocks were completely reset during the Grenville and that the Flint Hill Gneiss may actually be older.

Subsequent events in the geologic history of the Flint Hill Gneiss are: (1) Intrusion of Robertson River Formation of Allen (1963), (Figure

5). Zircons from the Robertson River Formation have been dated by the senior author at 732 m.y. (2) Intrusion of basaltic dikes that are now greenstone. These dikes are probably part of the Catoctin igneous event.

UNIFYING CHARACTERISTICS

The unifying characteristics of the rock designated here as Flint Hill Gneiss are its finely layered structure and granitic composition. This gneiss thus contrasts with the contiguous Marshall Granite of Jonas (1928) and Robertson River Formation of Allen (1963), which are weakly foliated to isotropic. The Flint Hill Gneiss is also in justaposition to an augen gneiss and to the Pedlar Formation of Bloomer and Werner (1955). The augen gneiss, however, is coarsely layered, in contrast to the fine layering of the Flint Hill; the Pedlar, although finely layered near its contact with the Flint Hill, is mineralogically different.

Veins of blue quartz are distinctive in the Flint Hill Gneiss. No other unit in the region has such veins.

ECONOMIC GEOLOGY

The type locality of the Flint Hill Gneiss is the only active quarry in the basement rocks of the Flint Hill, Linden, and Orlean quadrangles. Operations began here in 1965 on the site of a small quarry; since then, approximately 700,000 tons has been crushed and shipped (George A. Romer, 1974, personal communication).

Pavementlike outcrops are extensive in the area of the quarry, and overburden between outcrops is only 10-15 feet (3-5 m) thick. The rock beneath this overburden is generally fresh. Weathered rock is scarce.

The hardness of the Flint Hill Gneiss at the quarry is satisfactory for use in highway construction. The rock is classified as Grade A; it has a Los Angeles Rattler Test of 26, and the minimum value for this classification is 43. The rock is very abrasive, however, because of the high quartz content; consequently, drilling is expensive.

Quarrying of dimension stone from the Flint Hill Gneiss is not possible because of pervasive cleavage in the rock. Other quarries for crushed stone are not likely to be opened within this unit, because the 20-25 mile (32-40 km) economic shipping distance of crushed stone brings the entire area of the Flint Hill Gneiss within the region supplied by the quarry at Flint Hill.

REFERENCES CITED

Allen, R. M., Jr., 1963, Geology and mineral resources of Greene and Madison Counties: Virginia Div. Min. Resources Bull. 78, 102 p.

Bloomer, R. O., and Werner, H. J., 1955, Geology of the Blue Ridge region in central Virginia: Geol. Soc. America Bull., v. 66, no. 5, p. 579-606.

Furcron, A. S., 1939, Geology and mineral resources of the Warrenton quadrangle, Virginia: Virginia Geol. Survey Bull. 54, 94 p.

Jonas, A. I., 1928, Preliminary edition of geologic map of Virginia: Virginia Geol. Survey.

Lukert, Michael T., and Nuckols, E. B., 1976, Geology of the Linden and Flint Hill quadrangles, Virginia: Virginia Div. Min. Resources Rept. Inv. 44, 83 p.

THE GENUS *ANARTIOCRINUS* (CRINOIDEA: INADUNATA)

By

H. L. Strimple
T. J. Frest
The University of Iowa
Iowa City, Iowa

and

J. F. Miller
Southwest Missouri State University
Springfield, Missouri

ABSTRACT

The genus *Anartiocrinus* is poorly represented among prolific crinoid faunas of the Chesterian (Upper Mississippian) but is of more than passing interest because of the hypertrophied arms in *B* and *E* rays. Our purpose here is to discuss inequalities among arms of the five rays, as well as to report new occurrences of the genus *Anartiocrinus* in Kentucky and Alabama.

Both previously know species (*A. lyoni* Kirk and *A. maxvillensis* (Whitfield)) are redescribed on the basis of new material. Hypertrophy is believed to compensate for structural weakness in the *B* & *E* rays, which apparently are ontogenetically the last rays to develop arms. The rugged nearly inflexible *B* & *E* ray arms of *Anartiocrinus* may also have protected the delicate arms of the remaining rays and the anal tube. *Anartiocrinus* is probably ancestral to *Cromyocrinus* of Pennsylvanian age which has five "hypertrophied" arms.

The rare Chesterian inadunate crinoid genus *Anartiocrinus* was originally erected by Kirk (1940) on the basis of half a dozen specimens from Kentucky and Ohio. According to Kirk *Anartiocrinus* is represented by two species, the type species *A. lyoni* Kirk (1940), and *A. maxvillensis* (Whitfield, 1882); both are superficially similar in appearance. The holotype and one paratype of *A. lyoni* are from the Glen Dean Formation of Sloans Valley, Pulaski County, Kentucky. One paratype is from "the Chester" near Stephensport, Breckinridge County, Kentucky, and another from "the Chester" of Grayson Springs, Grayson County, Kentucky. Two specimens (cotypes) of *A. maxvillensis* are from the Maxville Limestone (Chester, Upper Mississippian) of White Cottage, Muskingum County, Ohio. These are the only specimens of the genus recorded prior to the present study. Four additional specimens with more precise locality and stratigraphic data form the basis of this study. These specimens, which are apparently divided equally between the two species, have led us to question some features of previous descriptions.

Kirk (1940, p. 50) stated for A. lyoni "... and the RR proportionately lower than in A. maxvillensis." This apparently was a mistake because the RR are proportionately higher in A. lyoni than in A. maxvillensis; this is readily discernable from Kirk's illustrations and is confirmed by our specimens. A quick computation based on measurements provided by Kirk verifies that A. maxvillensis has proportionately lower radials. The point is belabored because of the loose similarity of the two species; there is evidently a need for corroborating morphologic features to determine whether they are two discrete species and, if so, how they differ.

Another statement made by Kirk (1940, p. 49) needs some clarification; he states that in A. lyoni, "The IAx (primaxils) are wider than high and closely appressed laterally, so the sides are straight instead of being somewhat hourglass-shaped as in Anartiocrinus maxvillensis." Based on Kirk's illustrations not all primaxils in A. maxvillensis are hourglass-shaped; this is verified by material at hand. Some primaxils of A. lyoni show enough median constriction to void any meaningful use of this character as far as separation of these two species are concerned.

A third statement made by Kirk (1940, p. 50) that we question is "... it is noteworthy that the right tube plate in A. lyoni consistently lies at a consideraly lower level in the cup than in A. maxvillensis." This condition is probably true in the material available to Kirk, but is not the condition found in the material available to us. The right tube plate is well down in the cup of all four specimens, whatever the species. Evidently the position of this plate is somewhat variable, and hence is not a good taxobasis at the specific level.

With clarification of the above points, which could cloud the issue if left unattended, a precise definition of both species can still be attempted. A. lyoni has a high, essentially cone-shaped cap (Kirk says turbinate). A. maxvillensis has a medium globe-shaped cup (depressed turbinate of Kirk). All cup plates are proportionately somewhat longer in A. lyoni and lower cup plates (infrabasals and basals) appear to be more so than is actually the case because of their more erect positions as compared to those of A. maxvillensis. We do not believe A. maxvillensis to be the progenitor of A. lyoni despite the latter's occurrence in younger rocks because the tall more nearly cone-shaped cup shape of the latter is a more primitive trait.

Two specimens that are reported on here are from the Bangor Limestone, Chesterian, near Moulton, Alabama. One is conspecific with A. lyoni and the other with A. maxvillensis. Also reported are two specimens from Walche's Railroad Cut south of Princeton, Kentucky One, which is from the Menard Formation, Chesterian, is identified as

A. maxvillensis; the other, from the Clore Formation, Chesterian, is A. lyoni.

There appears to be a close relationship between Anartiocrinus and Agassizocrinus with the former retaining a column and developing two hypertrophied arms and the latter separting from the column and developing a fused infrabasal cone for a bottom-sitting life style (Strimple, 1962; Ettensohn, 1975; Strimple, 1976). Anartiocrinus is sparsely represented in the Chester but Agassizocrinus proliferated during the same period. Both genera have comparable but not closely similar cup shapes and arms. Older species of both have in common upflared infrabasals and a conical cup shape, but Anartiocrinus retains small infrabasals whereas those of Agassizocrinus become prominent and greatly enlarged, eventually overrunning the stem in some species (Strimple, 1962), and fusing to form a solid cone that commonly makes up more than half the cup height. Both genera, as will be detailed below, share a disposition toward unequal arm development and reduction in width of some radials, but each has exploited this tendency in a different way. Neither is likely to have been ancestral to the other, but we suspect that both diverged from a common ancestor, perhaps near the beginning of the Chester.

SYSTEMATIC DESCRIPTIONS

Subclass INADUNATA Wachsmuth & Springer, 1885

Order CLADIDA Moore & Laudon, 1943

Suborder POTERIOCRININA Jaekel, 1918

Superfamily AGASSIZOCRINACEA S. A. Miller, 1890

Family AGASSIZOCRINIDAE S. A. Miller, 1890

Genus ANARTIOCRINUS Kirk, 1940

Type species. - Anartiocrinus lyoni Kirk, 1940, p. 49.

Diagnosis. - Crown elongate, moderate height. Cup cone-shaped to medium globe-shaped, five infrabasals plainly visible in side view of cup. Articular facets of radials plenary (fill full summit width of plates). Three anal plates in CD interray usually in normal (primitive) arrangement. Ten uniserial arms, small in all rays except B and E, in which one-half ray (that facing posterior interray) is exceptionally large. Anal sac small cylindrical, recurved, extending about two-thirds the length of arms. Column transversely cylindrical, relatively large, heteromorphic.

Range. - Chesterian, U. S. A. (Ohio, Kentucky, Alabama).

ANARTIOCRINUS LYONI Kirk, 1940

Plate 1, Figures e-g

Diagnosis. Cup cone-shaped, infrabasals erect.

Discussion. - Anartiocrinus lyoni is readily distinguishable because of the more nearly regular cone-shape of the cup with very slight constriction near the cup summit and the proportionately tall cup plates. As previously discussed, A. maxvillensis has a medium globe-shaped cup with pronounced constriction toward the cup summit and cup plates are proportionately shorter.

Material studied. - Two specimens of the species are available, one from Moulton Quarry in Alabama and one from Walche's Railroad cut in Kentucky.

Description of Moulton Quarry specimen. - Indiana University, Geology Department catalogue number IU 8961-17 (Plate 1, figures e-f).

Dorsal cup small, cup height 9. 0 mm, diameter 10. 0 mm. The three anal plates are in normal contact with each other, but the radianal is not in contact with the BC basal plate. This condition has been discussed in the introduction. Anal X and right tube plate are missing. Width of stem facet 2. 1 mm.

Although somewhat smaller and presumably younger than the specimen of A. maxvillensis from the same locality the difference in size is not judged to be sufficient to allow for modifications in cup plates that would be needful to change the cup shape enough to attain the semi-globose contour of A. maxvillensis.

The specimen was collected by the Spencer Waters family of Moulton, Alabama from the now inactive state quarry about 1. 5 miles northwest of downtown Moulton, Lawrence County, Alabama (N 1/2 SE 1/4 NE 1/4 sec. 30, T6S, R7W). It is probably from the black arenaceous shale and limestone of the Bangor Formation formerly exposed in the lowest portions of the quarry. A middle Chesterian age is suggested.

Plate 1. a-d, Anartiocrinus maxvillensis Whitfield. a, lateral view of partial crown (SUI 43303), centered on CD interray. b, side view of SUI 43303 showing hypertrophied arm (B ray). Note smooth articulating surface of uppermost brachial. C, IU 896-16, lateral view centered on B ray. d, same specimen as preceeding, basal view. e-g, Anartiocrinus lyoni Kirk. e-f, Moulton Quarry specimen (IU 8961-17), lateral view centered on posterior interray and basal view respectively. g, SUI 43302, crown showing hypertrophied and normal arms, anal sac, and part of column. Lateral view centered on D ray. All figures whitened with ammonium chloride sublimate. Magnifications: a, x1. 4; b, x3. 5; c-g, x2. 8.

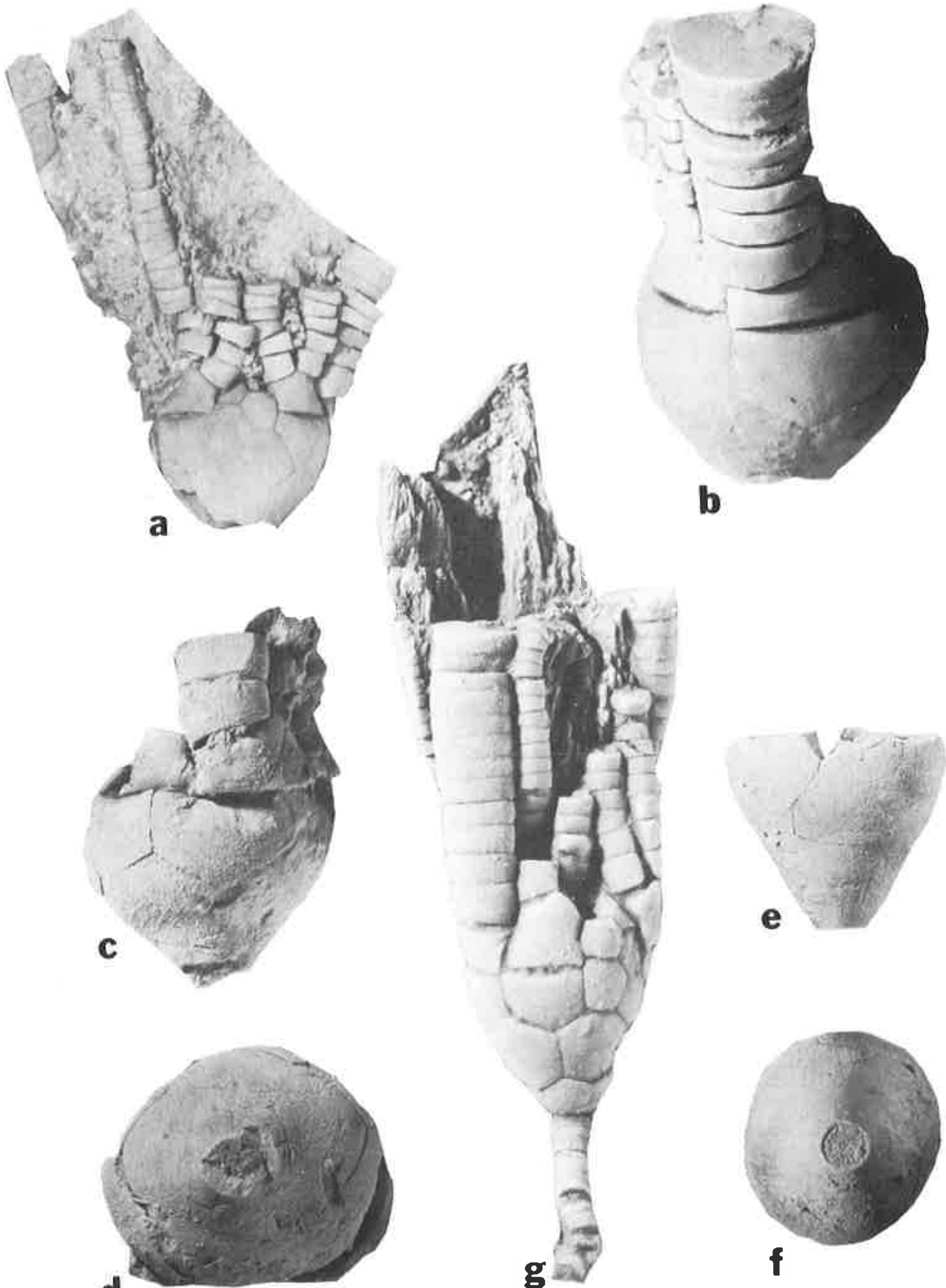


Plate 1.

Springer (1926, p. 60) had to do with inequality in the summit widths of radial plates. Slight differences do exist in Agassizocrinus and are reflected in the arms, as well as in all cup plates except anal plates, but the differences do not even approach the gross differential afforded by the two hypertrophied half-arms in Anartiocrinus. In addition the anterior radial is by Springer's calculations one of the widest in Agassizocrinus. In Anartiocrinus the E, A, and B radials are nearly equal in width, with the anterior ray radial (A) being slightly narrower. These minor differences aside the characteristic prominent modified infrabasal circlet and stemless condition of Agassizocrinus are without parallel in Anartiocrinus.

Material studied. - Two specimens of the species are available to us, one from Moulton Quarry in Alabama and one from Walche's Rail- road cut in Kentucky.

Description of Moulton Quarry specimen. - Indiana University, Geology Department catalogue number IU 8961-16 (Plate 1, figures c- d).

Small incomplete crown free of matrix with most of the arms and the proximal tips of infrabasals missing. The infrabasals are more erect than found in the holotype of the species, but a pronounced outward flexing of the cup just above the infrabasals and subsequent constriction of the distal portion of the cup is typical. Three anal plates are in normal (primitive) position. Only three primaxils are preserved. Those of A and C rays are smaller than primibrach 1 of B ray. Posi- tion of the hypertrophied arm B primaxil is marked by an enlarged posterior facet. Two secundibrachs are preserved in each half ray of the B ramus. Cup slightly crushed: maximum width 12.9 mm, mini- mum width 10.2 mm; height 10.3 mm.

The specimen was collected by the Spencer Waters family of Moulton, Alabama. Locality and stratigraphic information is identical with that previously given for the Moulton Quarry Anartiocrinus lyoni.

Description of Walche's Railroad cut specimen. - The University of Iowa, Geology Department, Geology Department Repository catalogue number SUI 43301 (Plate 1, figures a-b).

Incomplete crown relatively free of matrix with infrabasals and most of DE basal missing, but many of the arms are partially preserv- ed. The specimen is the largest representative of the genus found to date. Cup configuration is compatible with holotype of the species. Radial plates are proportionately longer and there is less constriction at the cup summit than is typical of the species. Three anal plates are in normal (primitive) arrangement. Primaxils of C and D rays and ap- preciable portions of the arms, which are only small when compared to the hypertrophied half rays, are preserved. Height and width of primaxils are equidimensional. Proximal secundibrachs are almost as tall as wide, but subsequently the brachials become low and have oblique facets with the alternating pinnular bearing sides higher. In the hypertrophied arms primaxils are low and wide with the posterior facets considerably wider than the anterior facets. The lower two secundibrachs in the enlarged arms are relatively tall but thereafter the brachials are very short. The arms expand until they attain a width equal to that of the radial plate of the ray. Greatest cup width is about 16.9 mm: minimum cup height 11.2 mm. Width of large arm brachial (B ray, seventh SBr) 7.6 mm; small arm brachial width (measured on sixth SBr of proximal half-ray, C ray) 3.8 mm.

The specimen was collected by Miller from Walche's Railroad cut west southwest of Princeton, Kentucky in the Menard Formation, Elvirian Stage, Chesterian (37° 6' N latitude, 87° 55' E longitude, Princeton West 7 1/2' quadrangle).

HYPERTROPHIED ARMS

It was noted by Strimple (1938, 1959) that B (right anterior) and E (left anterior) primibrachs in Graphiocrinus carbonarius, now Apographiocrinus typicalis (Moore & Plummer), and in juvenile Erisocrinus typus Meek & Worthen were shorter than primibrachs in the other three rays. The same pattern usually prevails in order inadunate crinoids in which there is appreciable inequality in length of primibrachs between the five rays. As noted by Burdick & Strimple (1969) arm characters are normally paired with features of B ray, a mirror image of E ray and those of C ray matched by D ray. Contrary to those author's statement (1969) that "abnormalities may occur in the A ray," the A ray frequently exhibits characters different from other rays; these are not, however, necessarily abnormal.

In many forms that entirely suppress two arms, e.g., the cladids Indocrinus and Tribrachyocrinus and the hybocrinids Baerocrinus, Cornucrinus, and Tripatocrinus, the B and E arms are lost. In other cases not only the corresponding arms but the B and E radials themselves are absent (e.g., the disparid Holynocrinus). This condition together with the occasional smaller or shorter stature of primibrachs in B and E rays has led to the conclusion that the two rays are "weaker" (i.e., more subject to morphological variation) than A, C, and D rays. Alternatively it is possible, particularly for those forms with less than the normal complement of radials, that the observed "modification" represents a reversion to an ancestral triradiate condition. The problem is far too complex and controversial for an attempt to be made at its resolution in the present study. For our purposes the formula C+D-A-B+E will serve. We are inclined to believe the most primitive condition consists of two arms analogous to C and D rays. This condition may reflect the fundamental bilateral symmetry of the presumed echinoderm ancestor, or it may be a secondary feature. A arm appears to be independent and may appear before or, less likely, after B and E arms are developed. B arm is a branch of C and E arm is a branch of C, in our opinion.

A most unusual expression of the pattern C+D-A-B+E is found in the genus Anartiocrinus and in Pentaramicrinus bimagnaramus Burdick & Strimple, both of Chesterian age, wherein hypertrophied arms are found in B and E rays. Production of robust arms might have been a response to frequent damage to the arms because of turbulent conditions, with the "weaker" rays,which are a mirrored-pair (enantiomorphs), becoming enlarged because they were the most susceptible to damage. In both P. bimagnaramus and in Anartiocrinus the posterior half-rays of B and E rays are twice as large as other arms. Variation in arm size has been reported in other Agassizocrinidae. Ettensohn (1975) has noted that some arms in Agassizocrinus are somewhat shorter than others and has suggested the shorter ones might have acted as props against the substrate with the longer arms forming a vertical filtration fan held upright into the current or surf. We have been unable to verify a difference in arm lengths in specimens of Agassizocrinus available to us but we do not reject the hypothesis.

The two large (hypertrophied) arms of Anartiocrinus have the same general structure and are comparable in size to those found in Cromyocrinus. Species of this genus have reduced the total number of arms from ten to five but all five are robust. The ambulacral grooves and the pinnules are larger in the robust arms of Anartiocrinus than are those of the smaller arms, which may indicate that the larger arms have a capability of capturing and retaining larger food particles. With this arrangement eight arms could act as a filtration fan for smaller food particles and the two robust arms could act independently as erect elements capturing either smaller or larger prey. The usual thickness of individual ossicles in the hypertrophied arms and lack of articular fossae (Plate 1, figure b) indicates an almost total lack of flexibility of these elements, other than in the lowermost segments. It thus appears they served yet another function which was probably protective. With the arms closed the two rugged arms would serve as protection for the more delicate ones as well as shielding the delicate anal tube.

It is suggested here that the modification of the two half-rays in Anartiocrinus established a disposition to produce very large arms which became a feature of the lineage which led to Cromyocrinus. In this latter genus the trend is carried to the ultimate extreme; Cromyocrinus has five instead of ten arms, all "hypertrophied."

A similar evolutionary trend can be documented in the genus Pentaramicrinus. One older ten-armed species, P. bimagnaramus, has two hypertrophied half-rays, predictably in the B and E rays.

Pentaramicrinus bimagnaramus Burdick and Strimple (1971) is from the Beech Creek Formation, lower Hombergian, middle Chesterian of Illinois. P. gracilis Wetherby, 1880, type species of the genus, has reduced the number of arms to five and is from the Glen Dean, or Tar Springs Formation, upper Hombergian of Kentucky. The five arms in P. gracilis are moderately robust and attain a considerable length but the species is relatively small. It probably represents the end member of the divergent arm modification first appearing in P. bimagnaramus.

REFERENCES CITED

Burdick, D. B., and H. L. Strimple, 1969, Revision of some Chesterian inadunate crinoids: Univ. of Kansas, Pal. Contr., Paper 40, 14 p.

__________, 1971, Faunal studies of the type Chesterian, upper Mississippian of southwestern Illinois, part 2, Crinoids from the Beech Creek Limestone, lower Golconda group, St. Clair County, Illinois: Univ. of Kansas, Pal. Contr., Paper 51, p. 15-47.

Ettensohn, R., 1975, The autecology of Agassizocrinus lobatus: Jour. Paleontology, v. 49, p. 1044-1061.

Kirk, E., 1940, Anartiocrinus, a new crinoid genus from the Mississippian: American Jour. Sci., v. 238, p. 47-55.

Springer, F., 1926, Unuaual forms of fossil crinoids: U. S. National Mus., Proc., v. 67, 137 p.

Strimple, H. L., 1938, A group of crinoids from the Pennsylvanian of northeastern Okalahoma: 14 p. Bartlesville, Oklahoma (privately printed).

__________, 1959, Crinoids from the Missourian near Bartlesville, Oklahoma:: Oklahoma Geology Notes, v. 19, p. 115-127.

__________, 1962, Associated Cryphiocrinus and Agassizocrinus: Oklahoma Geology Notes, v. 22, p. 183-188.

__________, 1976, Two new species of Agassizocrinus from northern Alabama: Southeastern Geol., v. 18, p. 29-35.

Whitfield, R. P., 1882, Descriptions of new species of fossils from Ohio, with remarks on some of the geological formations in which they occur: Ann. New York Acad. Sci., v. 2, p. 193-244.

PETROLOGY OF THE DUNITE EXPOSED AT THE MINCEY MINE, MACON COUNTY, NORTH CAROLINA[1]

By

K. R. Hahn*
and
R. A. Heimlich
Department of Geology
Kent State University
Kent, Ohio 44242

ABSTRACT

The dunite body is elliptical in plan view, roughly 200 by 375
feet in maximum dimensions, and discordant relative to the attitude of
foliation in the country rock. It was emplaced in a terrain dominated
by quartzofeldspathic gneiss characterized by foliation that strikes
northeast-southwest and dips steeply to the northwest. The gneiss was
formed under conditions of middle-grade regional metamorphism.
Evidence of contact metamorphism is lacking.

The major primary minerals of the dunite are olivine (Fo89),
enstatite (En 90), and chromite. Texturally, much of the dunite con-
sists of unstrained polygonal grains of olivine possessing triple point
junctions which meet at angles of 120°. Locally, porphyroclasts of
larger strained olivine grains are surrounded by mosaics of smaller
polygonal grains or by granulation zones. Undulatory extinction and
strain bands characterize the porphyroclasts. The body is progres-
sively more altered toward its margins. A largely peripheral altera-
tion zone consists of an intergrowth of vermiculite, talc, chlorite, and
serpentine minerals. In addition, small amounts of tremolite and
anthophyllite occur throughout the body. Chemically, the dunite is
characterized by high concentrations of chromium and nickel uniformly
distributed throughout. Silica and alumina increase slightly while
magnesia decreases slightly toward the margins.

Of three samples sectioned for petrofabric analysis, two exhibit
no obvious preferred orientation. However, the third, characterized
by polygonal texture and located centrally within the pluton, has a

---

[1] Contribution No. 142, Department of Geology, Kent State University

*Current address: Department of Geology, Rice University, Houston, Texas 77001

well-defined fabric dominated by a strong Z maximum (exceeding 9 percent) normal to an incomplete X girdle. Weaker point maxima exist for Y in this sample.

Collectively, the field and laboratory data suggest vertical emplacement of solid mantle material that underwent syntectonic recrystallization in the source region prior to emplacement. During emplacement, the dunite was altered to hydrous minerals resulting from chemical interchange with the country rock. The alteration process involved addition of water, silica, and alumina with loss of magnesia and some iron oxide. A second episode of alteration is represented by the development of small amounts of secondary amphiboles distributed throughout the mass.

INTRODUCTION

The study of alpine ultramafic plutons has grown in importance during the past few years because of their relevance to plate tectonics theory and the information they provide regarding the nature of the mantle and mantle processes. Many alpine ultramafic bodies are thought to represent upper mantle material diapirically emplaced into the continental crust above areas of active plate subduction (Chidester and Cady, 1972; Stevens et al , 1974). Others may be portions of ophiolite sequences thrust laterally onto continental margins during episodes of plate obduction (Moores, 1970; Coleman, 1971; Moores and MacGregor, 1972). Debate exists, as well, regarding the common serpentinization of these rocks. Some workers believe that the plutons were emplaced as fresh ultramafic material which was serpentinized either during or after emplacement (Ragan, 1963; Condie and Madison, 1969), while others believe that they were intruded as serpentinite which was later dehydrated (Astwood, Carpenter, and Sharp, 1972). An additional question concerns whether or not serpentinization was a volume-for-volume metasomatic process (Thayer, 1966, 1967; Page, 1967; Condie and Madison, 1969).

The object of this investigation was to determine the petrology, petrography, petrofabrics, and geochemistry of a single alpine ultramafic body representative of those in the southern Appalachian Mountains. For this purpose, we chose the dunite exposed at the Mincey Mine, some nine miles east of Franklin, North Carolina (Figure 1). Brief reference to this body is made by Hunter (1941) and by Conrad et al (1963).

Acknowledgments

We are extremely grateful to the Office of Earth Resources, North Carolina Department of Natural and Economic Resources for a grant which helped to defray costs relative to thin-section preparation

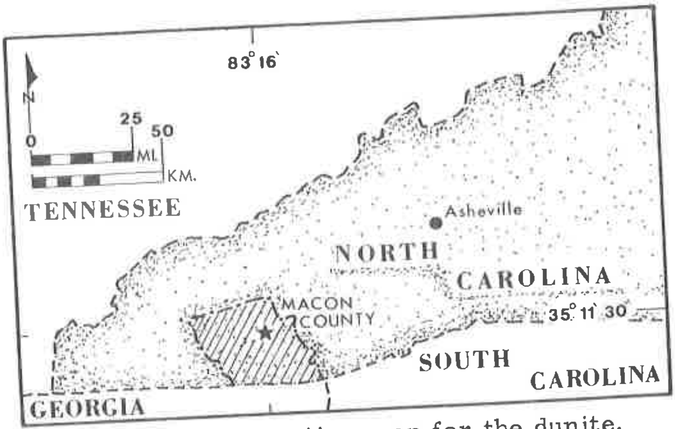


Figure 1. Location map for the dunite.

and field work. The study would not have been possible without the help of Mr. W. Reynolds, owner of the Mincey Mine, who gave permission to work on his property and who assisted freely during the field work.

We thank E. H. Carlson and D. F. Palmer for their advice during the laboratory studies and for critical reading of an earlier draft of the paper.

METHODS OF STUDY

Hand specimens of both the dunite and the country rock were collected for laboratory analysis. Although the central portion of the body is well exposed due to past mining operations, spoil piles prevent sampling in several areas. Dunite sample localities were located by use of the pace-compass method and steel tape where possible, so that a sample was collected every 15 feet across the body. Spot sampling filled in the network (Figure 2). The country rock was sampled wherever outcrops permitted. Sample localities and structural information were plotted on a topographic base map (scale 1:3000) with the aid of aerial photographs (scale 1:3960).

Selected samples of both the dunite and the country rock were prepared for thin-section study. Based on 1200 points, modes were obtained for all thin-sections. Chemical analyses, for major and selected trace elements, were determined by atomic absorption spectrophotometry. Precision, based upon duplicate analyses of all samples including whole rock standards DTS-1, PCC-1, and GSP-1, is better than 5 percent. Accuracy was determined by comparing analyses obtained for the standards with their accepted values. Olivine compositions were determined by the X-ray diffraction method of Yoder and Sahama (1957). The accuracy of this method is ± 3 percent forsterite.

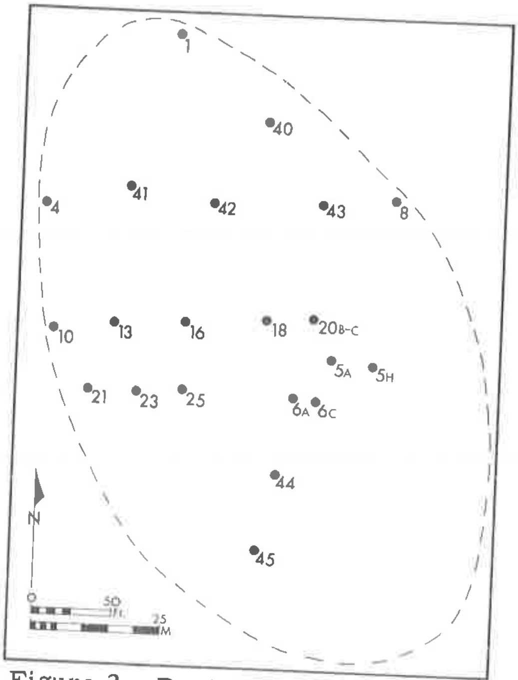


Figure 2. Dunite sample localities.

Three oriented dunite samples were analyzed using a 5-axis universal stage, following the procedure outlined by Emmons (1943), to determine the presence or absence of preferred orientation of the olivine grains. For each thin section, the orientation of 150 grains was noted and pole positions were plotted on an equal area stereographic net.

Identification of alteration products and determination of pyroxene and plagioclase compositions were done by means of critical optical properties and use of appropriate determinative charts. Some alteration products were also identified by use of X-ray diffraction.

COUNTRY ROCK

The dunite occurs within a terrain of well-layered gneiss. The attitude of foliation in the gneiss is relatively uniform throughout the study area (Figure 3), with an average strike of N60°E and a dip of 82°NW. The body is situated on the eastern limb of a northeasterly plunging synform, whose axis lies west of the map area. The long axis of the mass is oriented at right angles to the strike of foliation in the gneiss. Based on observed mineralogy, the country rock lies in the amphibolite facies of regional metamorphism (Turner, 1968).

Petrographically, the country rock can be divided into two major lithologies in this area: 1) quartzofeldspathic gneiss, the dominant

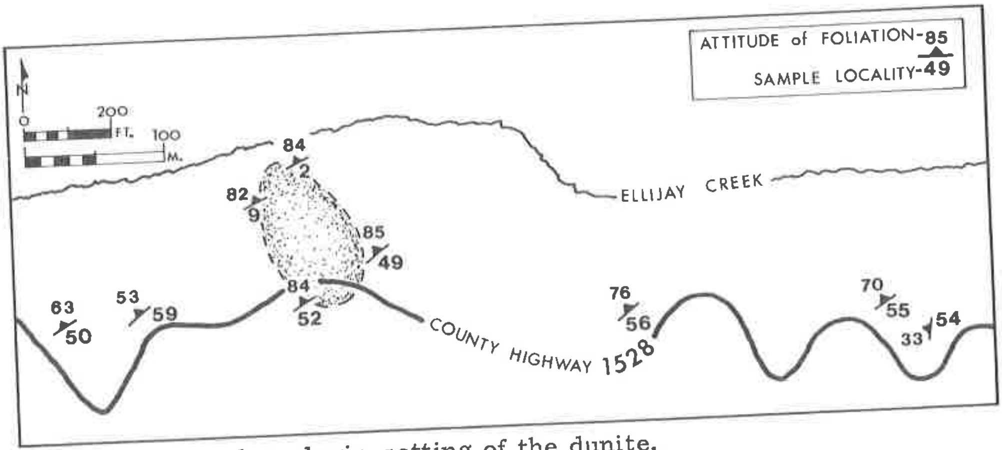


Figure 3. Areal geologic setting of the dunite.

lithology and the one adjacent to the pluton, and 2) an amphibolitic gneiss, found further from it. Dominant minerals of the quartzofeldspathic gneiss are plagioclase, microcline, quartz, biotite, and muscovite. The amphibolitic gneiss consists primarily of hornblende, plagioclase, microcline and quartz (Table 1). For both types of gneiss, plagioclase is commonly the most abundant feldspar, occurring as subidioblastic grains with polysynthetic twinning. Plagioclase compositions, as determined by the Michel-Levy method, range from An22 to An32, with an average composition of An28 in the quartzofeldspathic gneiss. The amphibolitic gneiss has slightly more calcic plagioclase (An36). Although most of the grains are relatively fresh, some are either completely or partially altered to sericite. Typically fresh and containing epidote and quartz inclusions, microcline grains are commonly subidioblastic and smaller than the surrounding grains. In both lithologies quartz is xenoblastic and shows some degree of undulose extinction and sutured boundaries.

In the quartzofeldspathic gneiss the micas are biotite and muscovite. Both occur as subidioblastic flakes showing some variation in grain size. Slightly more abundant than muscovite, the biotite is pleochroic (light green to dark brown) and contains abundant inclusions of magnetite and zircon. Although typically fresh, specks of secondary magnetite have developed along cleavage traces in some of the biotite. The muscovite occurs as small laths commonly bent around adjacent grains.

Hornblende occurs as pleochroic grains (X=yellow green, Y=olive green, Z=dark green) with well-developed cleavage and rare twinning. Most grains are fresh, although there is some minor alteration to secondary magnetite. Inclusions of opaque grains and zircon are common.

The more common accessory minerals include epidote and garnet, in the quartzofeldspathic gneiss, and zircon and magnetite in both lithologies.

Table 1. Modal and Mineralogical Data for the Country Rock.

Sample Number	9	2	49	52	50	59	56	55	54
Plagioclase	1.2	11.9	35.6	19.0	35.2	46.9	16.8	23.8	17.8
Microcline	1.2	25.0	22.9	24.4	29.3	25.5	16.9	30.7	23.8
Quartz	4.2	15.8	17.9	17.7	23.3	19.6	16.7	20.3	15.7
Muscovite	14.1	13.2	12.1	6.7	1.9	3.5		17.3	14.8
Biotite	26.5	14.7	10.9	31.3	8.8	1.0		0.3	20.1
Hornblende									
Chlorite	13.3	Tr	Tr	Tr			45.6***		
Talc	15.8		Tr						
Epidote	15.3	8.8			1.5	3.2		7.5	2.2
Opaques	5.2	1.8	0.6	0.4	Tr	Tr	3.5	0.1	2.1
Nonopaques*	3.2	8.8	Tr	0.5	Tr	0.3	0.5	Tr	3.5
Plagioclase An content**		30	26	32	28	28	36	29	22

*Includes garnet & zircon
**Minimum of six extinction angle measurements per thin section.
***2V = 60°; Z $\wedge$ c = 15°

Table 2 gives chemical analyses of the quartzofeldspathic gneiss. Samples nearest the dunite (numbers 2, 9, 52) are generally lower in silica, alumina, potash, and soda content but higher in magnesia and lime as compared with those farther away. These oxide variations constitute some evidence for chemical reaction between the siliceous country rock and the magnesian dunite. However, mineralogic evidence for thermal metamorphism is lacking.

DUNITE

In map view the dunite body is elliptical in shape, measuring roughly 200 by 375 feet in maximum dimensions (Figure 4). No data are available regarding the dip of its contact with the country rock or its subsurface configuration.

The body consists of a relatively unaltered dunite core which grades progressively outward to a moderately altered rim roughly 20 feet thick. In the field, the fresh dunite appears homogeneous; layering, preferred orientation of minerals, and shear zones are absent. However, the body is cut by a set of well-defined fractures which have roughly the same strike as that of foliation in the country rock. Many of these fractures have been filled by secondary alteration products such as talc, vermiculite, tremolite, and anthophyllite. At least one is occupied by a pegmatite dike.

Petrography

The main lithology of the body is an enstatite-bearing dunite composed of olivine, enstatite (typically less than 10 percent), and chromite as the primary minerals (Table 3). The olivine occurs either as anhedral grains, varying greatly in size (0. 1 to 2. 5 mm) and shape,

Table 2. Chemical Analyses of the Quartzofeldspathic Gneiss.

Sample Number	9	2	49	52	59	50	55	54
SiO_2	51.86	53.33	56.26	54.30	63.01	59.93	59.08	53.58
Al_2O_3	11.75	12.40	17.51	13.95	19.18	14.50	15.76	13.41
Fe_2O_3*	15.31	12.78	11.25	16.98	3.12	10.84	10.03	14.05
MgO	12.43	10.88	3.86	4.74	2.30	3.79	4.61	8.02
CaO	5.48	5.44	1.03	2.67	1.46	1.09	0.77	2.44
K_2O	0.50	4.61	4.31	4.14	4.43	5.41	6.05	4.58
Na_2O	0.44	1.32	3.94	2.10	5.19	3.89	2.63	1.97
	97.77	100.76	98.16	98.88	98.69	99.45	98.93	98.05

*Total Fe as Fe_2O_3.

Table 3. Modal and Mineralogic Data for the Dunite.

Sample Number	1	4	6A	6C	8	10	13	16	18	20B
Olivine	57.5	59.8	82.3	77.3	77.9	26.1	69.8	62.5	84.8	78.8
Enstatite	9.2	4.2	13.2	8.3	7.3	Tr.	4.8	2.6	9.5	4.6
Chromite	2.2	0.3	0.8	0.6	0.4	2.0	0.9	1.1	0.8	0.6
Talc	3.6	10.7	0.2	0.7	4.2	15.5	4.0	10.3	2.4	3.2
Vermiculite	10.1	15.1	1.5	5.3	6.4	13.4	11.7	17.8	Tr.	6.1
Chlorite	14.1	9.7	2.0	4.2	2.3	9.3	6.7	2.5	1.7	3.0
Secondary Amphiboles	1.0	0.2		1.2	0.7	7.0	1.8	0.6	0.8	2.2
Augite						26.7				
Serpentine*	2.3			2.4	0.8		0.2	2.6	0.8	1.2
Olivine %Fo	88	92	88	88	90	89	89	87	89	91
Enstatite 2V	86°	75°	84°	83°	85°		87°	85°	83°	86°
Enstatite %MgSiO3	89	94	90	91	90		88	89	91	89

Sample Number	20C	21	23	25	40	41	42	43	44	45
Olivine	82.5	79.0	76.5	88.6	80.7	76.4	79.3	80.9	94.7	87.1
Enstatite	10.5	14.6	2.9	2.4	5.0	2.3	5.8	9.6	2.0	8.7
Chromite	0.9	0.5	1.2	0.6	0.9	1.6	1.0	1.0	0.4	1.5
Talc	1.9	1.6	3.0	0.2	2.1	3.3	1.4	0.9	0.2	0.2
Vermiculite	1.2	1.8	10.3	1.8	4.6	3.7	4.3	3.9	Tr	
Chlorite	3.0	1.7	2.3	1.7	4.1	3.2	1.4	1.8	1.8	1.8
Secondary Amphiboles			2.0	0.7	0.8	0.8	0.3	0.5	Tr	0.3
Serpentine*		0.7	1.6	4.0	1.8	8.7	6.5	1.4	0.8	0.4
Olivine %Fo	91	88	87	89	88	87	92	89	88	88
Enstatite 2V	85°	82°	77°	84°	85°	83°	88°	85°	80°	82°
Enstatite %MgSiO3	89	91	93	90	89	91	87	90	92	91

*includes secondary magnetite.

or as polygonal equant grains possessing triple point junctions which meet at 120°, typical of the textures of recrystallized metamorphic rocks (Spry, 1969). Many of the anhedral grains exhibit undulose extinction, fairly well-developed parting, and some degree of alteration. Nearest the margins of the body, the dunite shows extensive deformation expressed by strain banding of the olivine and by granulation. In these areas the olivine is also commonly elongated. Olivine porphyroclasts, surrounded by mosaics of smaller recrystallized grains, also occur near the margins. The polygonal grains, typical of the recrystallized dunite, are confined largely to the central portion of the body. These

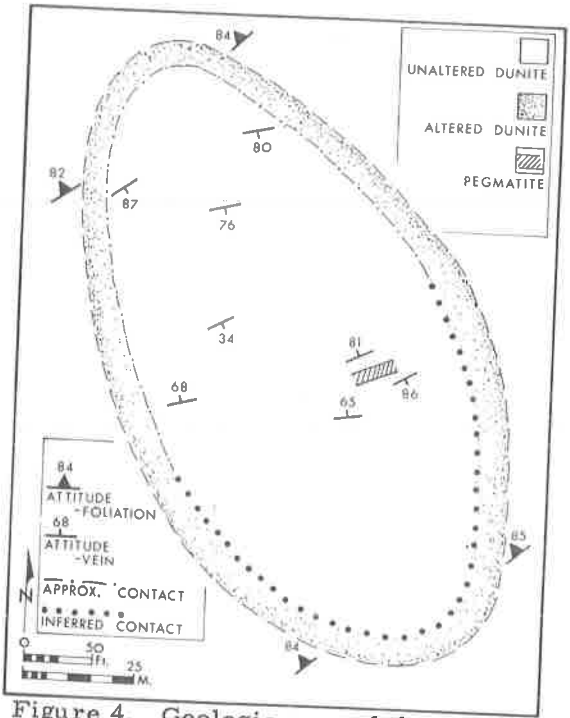


Figure 4. Geologic map of the dunite.

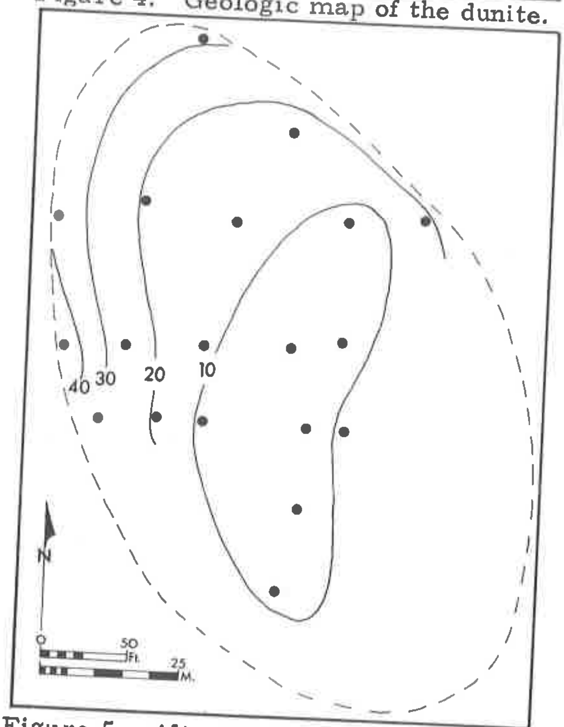


Figure 5. Alteration map of the dunite.

grains exhibit none of the deformational textures seen in the anhedral grains, and they lack parting as well. X-ray analysis indicates that the olivine is uniformly magnesium-rich, with an average forsterite content of 89 percent (Table 3).

Enstatite is more common within the central portion of the body. It occurs as bronze-colored, nonpleochroic, subhedral grains typically smaller than the surrounding olivine grains. It possesses well-developed cleavage and a blocky habit. Near the margins, the enstatite tends to occur as elongated poikilitic laths which enclose smaller anhedral olivine grains. Enstatite composition, expressed in weight percent $MgSiO_3$, was determined by 2V measurements on the universal stage and checked with indicies of refraction using the determinative charts of Poldervaart (1950). The enstatite is uniformly magnesium-rich, with an average value of 90 percent $MgSiO_3$ (Table 3). Chromite occurs as subhedral, equant grains which are disseminated throughout the pluton.

Augite, present in only one thin-section, occurs as anhedral columnar grains. Cleavage and fracturing are prominent. Most grains are fresh, but some alteration to chlorite can be seen locally.

Modal analysis indicates that both olivine and enstatite reach their highest percentages near the center of the dunite. Both decrease in abundance gradually toward the margins as a function of increasing alteration in that direction (Figure 5).

The most common alteration products are talc, vermiculite, and chlorite. Both talc and vermiculite occur as irregular masses between grains of the primary minerals or as patches replacing both olivine and enstatite. They are rare as cross-cutting fracture fillings. Chlorite occurs as subhedral laths, typically replacing chromite along parting planes.

Small amounts of two secondary amphiboles are also present. The most common of these is tremolite occurring as acicular grains which cross-cut and replace olivine. Anthophyllite, replacing both olivine and enstatite, occurs as shorter grains showing typical parallel extinction.

Other alteration products include the serpentine minerals chrysotile and antigorite (identified petrographically by texture), along with secondary magnetite. Chrysotile is confined to fractures which cut olivine and enstatite, while antigorite occurs as highly irregular patches along the boundaries of adjacent olivine grains. Secondary magnetite occurs as tiny grains disseminated throughout the serpentine minerals.

Chemical Data

Although the dunite is relatively uniform chemically, there is a slight decrease in magnesia and corresponding slight increases in silica and alumina as the margins are approached (Figure 6).

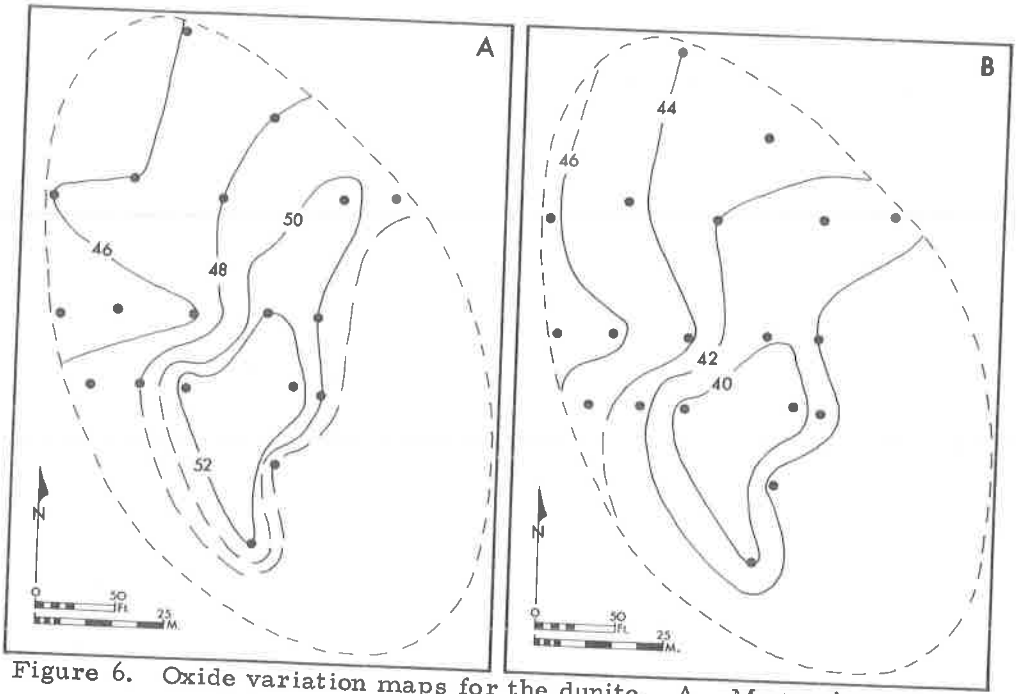


Figure 6. Oxide variation maps for the dunite. A. Magnesia content. B. Silica content.

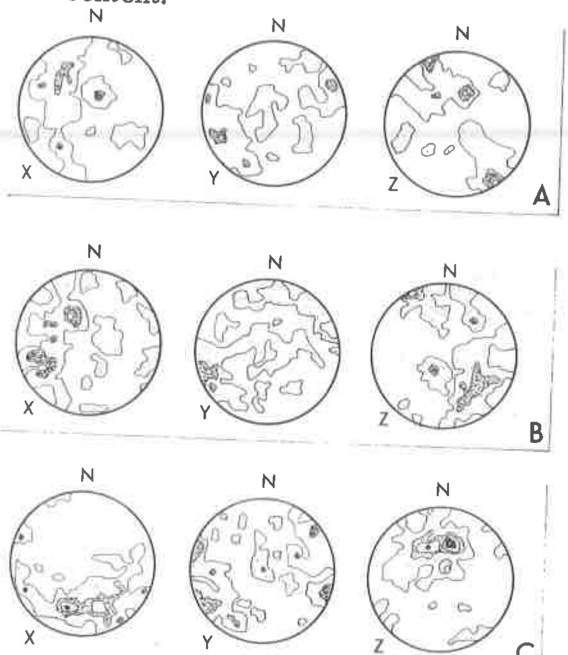


Figure 7. Petrofabric diagrams for samples 4 (A), 40 (B), and 41 (C). Contours at 1, 3, 5, 7, and 9 percent.

Table 4. Chemical Data for the Dunite.

Sample Number	1	4	6A	6C	8	10	13	16	18	20B
SiO_2	44.29	46.35	38.23	40.46	41.08	45.60	46.18	44.84	40.26	43.73
Al_2O_3	1.47	1.39	0.92	0.65	1.63	2.94	2.40	0.91	0.93	1.76
Fe_2O_3*	6.41	5.89	6.51	8.89	7.62	6.51	7.28	8.29	6.52	6.87
MgO	47.29	45.97	54.22	49.79	49.17	44.65	44.07	45.91	51.99	47.52
Ni (ppm)	2238	2197	2309	2456	2217	2226	2409	2409	2399	2432
Cr (ppm)	3009	3173	3098	3011	3102	3119	3206	3206	3197	3119

Sample Number	20C	21	23	25	40	41	42	43	44	45
SiO_2	39.96	44.89	42.79	39.40	42.75	44.97	42.02	41.78	42.87	40.22
Al_2O_3	1.21	0.61	1.18	0.77	1.48	0.69	1.46	0.20	1.32	0.60
Fe_2O_3*	6.31	7.61	7.82	7.16	7.67	7.49	7.84	6.65	6.79	7.62
MgO	52.29	46.76	47.46	52.65	48.00	46.28	48.23	51.35	48.38	52.03
Ni (ppm)	2209	2298	2345	2334	2365	2405	2388	2239	2398	2407
Cr (ppm)	3208	3349	3220	3316	3301	3285	3476	3199	3244	3356

*Total Fe as Fe_2O_3

The trace elements, nickel and chromium, exhibit no systematic variation throughout the body (Table 4). Average values for these elements are 2334 ppm and 3210 ppm, respectively. Both values are higher than those given for typical ultramafic rocks which analyze 1500 ppm nickel and 2400 ppm chromium (Goles, 1967).

Petrofabric Data

Petrofabric analysis was made of three oriented specimens of the dunite, one from the margin and two from the interior. In two of the samples (Figures 7A and B) there is no obvious preferred orientation. However, for the sample located furthest in from the margins (Figure 7C), there is a well-defined fabric dominated by a strong Z maximum (exceeding 9 percent) normal to an incomplete X girdle. Weaker point maxima exist for Y.

Textures of the oriented samples range from the polygonal type, characterized by fairly uniform grain size, to an irregular type characterized by relatively large elongate olivine grains surrounded by smaller equant olivines. The sample with polygonal texture (Figure 7C) shows the strong Z maximum, whereas those with irregular textures are characterized by the more diffuse fabrics. No aspects of the preferred orientation appear to be systematically related to the shape (plan view) of the pluton or to the attitude of foliation in the country rock.

DISCUSSION AND CONCLUSIONS

Field mapping indicates that the dunite was emplaced in a gneissic terrain along the eastern limb of a northeasterly plunging synform. The body is clearly discordant; its long axis cuts squarely across the strike of foliation in the gneiss, suggesting that the dunite and country rock are not related petrogenically. Unlike some of the ultramafic bodies in the northern Appalachian Mountains in Newfound- land (Church and Stevens, 1971; Dewey and Bird, 1971; Stevens, Strong, and Kean, 1974), this dunite is not part of an ophiolite sequence, a con- clusion supported by the absence of other units typical of ophiolites (Church, 1971; Hutchison, 1975). The lack of layering and megascopic preferred orientation of minerals also argue against an ophiolite inter- pretation.

Textures of the dunite are clearly deformational in nature. That with polygonal texture has undergone extensive recrystallization where- as portions with prophyroclastic and strain-banded olivine are incom- pletely recrystallized (Ragan, 1969; Nicolas, et al., 1971). The granu- lation textures and elongation of olivine and enstatite grains indicate solid state movement (Ragan, 1963). Based upon these features and their interpretation, we conclude that the dunite has undergone recrystal- lization and was emplaced as a solid.

The recrystallization could have occurred before (Ave'Lallemant and Carter, 1970), during (Ragan, 1969), or after (Spry, 1969) emplace- ment. Petrofabric studies show that dunite with polygonal texture has a distinct preferred orientation which is unrelated to the shape of the pluton and to the attitude of foliation in the country rock. On the other hand, its fabric is similar to that reported for some peridotite xeno- liths in basalt (Brothers and Rodgers, 1969; Mercier and Nicolas, 1975). Fabrics dominated by a Z maximum occur in some of these xenoliths characterized by both porphyroclastic and "mosaic equigranular" (poly- gonal) textures (Mercier and Nicolas, 1975). The fabric of these mantle- derived xenoliths clearly pre-dates their upward transport in the host basaltic liquid. By analogy it would appear that the fabric of the pre- sent dunite is also inherited from its mantle source region, where the material underwent syntectonic recrystallization, as suggested by the experimental results of Ave'Lallemant and Carter (1970). Supporting this conclusion is a lack of evidence (in the form of a contact meta- morphic aureole) for the high temperatures required for recrystalliza- tion (Himmelberg and Loney, 1973; Ave'Lallemant and Carter, 1970) during intrusion and lack of a geometric relationship between preferred orientation in the dunite and country rock foliation.

A mantle source if well-established for this dunite. In most respects the body is similar to other alpine ultramafic masses (Jack- son and Thayer, 1972) in the southern Appalachian Mountains (Carpenter and Phyfer, 1975; Yurkovich, 1976; Swanson and Raymond, 1976; Dribus et al., 1976) and elsewhere (Loney et al., 1971; Ehrenberg, 1975). In turn these alpine peridotites are extremely uniform in terms of their occurrence in the cores of folded mountain belts, recrystallization textures and tectonite fabrics, bulk chemical composition, and highly magnesian mineral compositions. Moreover, they are similar, in de- tail, to ultramafic xenoliths associated with basalts (Mercier and Nicolas, 1975) generated by partial melting within the upper mantle.

The progressive increase in alteration from the center of the pluton to its margins, reflects either inward hydrous alteration of dunite or outward dehydration of serpentinite. In this case, the former process is indicated because of the preferred orientation of the olivine crystallographic axes, a feature incompatible with the concept of serpentinite intrusion and subsequent dehydration (Astwood, Carpenter, and Sharp, 1972).

Serpentine mineral constitute a small portion of the alteration mineral assemblage (Table 3). The dominant minerals are vermiculite and chlorite with small amounts of talc and serpentine. Because of its largely peripheral location and the intimate intergrowth of the four minerals, as seen in thin-section, we believe that this mineral assemblage constitutes a single episode of hydrous alteration of the dunite body. Because the extent of the alteration decreases gradually inward (Figure 4), it would appear that this episode of alteration occurred during emplacement and that the source of the water was the adjacent country rock, as proposed for other alpine ultramafic bodies (Burch, 1968; Coleman, 1966). Chemical analyses (Tables 2 and 4) suggest that during this process silica and alumina were introduced, as well as water, while magnesia and some iron was removed from the dunite (Condie and Madison, 1969; Thayer, 1966, 1967; Page, 1967).

A second episode of alteration is represented by the development of amphiboles that constitute typically less than 2 percent of almost all the dunite samples examined in thin-section (Table 3). Prisms of tremolite and anthophyllite transect patches of two or more of the minerals representing the major alteration assemblage. Except for addition of water, this alteration probably operated in a closed system. Calcium required for development of the very small amount of tremolite was undoubtedly provided by breakdown of enstatite and possibly diopside (now lacking in the dunite).

REFERENCES

Astwood, P. M., Carpenter, J. R., and Sharp, W. E., 1972, A petrofabric study of the Dark Ridge and Balsam Gap dunites, Jackson County, North Carolina: Southeastern Geol., v. 14, p. 183-194.

Ave'Lallemant, H. G., and Carter, N. L., 1970, Syntectonic recrystallization of olivine and modes of flow in the upper mantle: Geol. Soc. America Bull., v. 81, p. 2203-2220.

Brothers, R. N., and Rodgers, K. A., 1969, Petrofabric studies of ultramafic nodules from Auckland, New Zealand: Jour. Geol., v. 77, p. 452-465.

Burch, Stephen H., 1968, Tectonic emplacement of the Burro Mountain ultramafic body, Santa Lucia range, California: Geol. Soc. America Bull., v. 79, p. 527-544.

Carpenter, J. R., and Phyfer, D. W., 1975, Olivine compositions from southern Appalachian ultramafics: Southeastern Geol., v. 16, p. 169-172.

Chidester, A. H., and Cady, W. M., 1972, Origin and emplacement of alpine-type ultramafic rocks: Nature Phys. Science, v. 240, p. 27-31.

Church, W. R., 1971, Ophiolite: Its definition, origin as oceanic crust, and mode of emplacement in orogenic belts, with special reference to the Appalachians: Canada Dept. Energy, Mines and Resources, Earth Physics Br. Pub. 42, p. 71-86.

Church, W. R., and Stevens, R. K., 1971, Early Paleozoic ophiolite complexes of Newfoundland Appalachians as mantle-ocean crust sequences: Jour. Geophys. Res., v. 76, p. 1460-1466.

Coleman, R. G., 1966, New Zealand serpentinites and associated metasomatic rocks: New Zealand Geol. Survey Bull., n. s., 76, 100 p.

__________, 1971, Plate tectonic emplacement of upper mantle perdotites along continental margins: Jour. Geophys. Res., v. 76, p. 1212-1222.

Condie, K. C., and Madison, J. A., 1969, Compositional and volume changes accompanying progressive serpentinization of dunites from the Webster-Addie ultramafic body, North Carolina: Am. Mineralogist, v. 54, p. 1173-1177.

Conrad, S. G., Wilson, W. F., Allen, E. P., and Wright, T. J., 1963, Anthophyllite asbestos in North Carolina: North Carolina Dept. Cons. Devel. Bull. 77, 61 p.

Dewey, J. F., and Bird, J. M., 1971, Origin and emplacement of the ophiolite suite: Appalachian ophiolites in Newfoundland: Jour. Geophys. Res., v. 76, p. 3179-3207.

Dribus, J. R., Godson, W. L., Hahn, K. R., Kammer, T. W., Lawton, D. L., Pesch, H. L., Schick, J. T., and Tata, S. S., 1976, Comparative petrography of some alpine ultramafic plutons in western North Carolina: Compass, v. 53, p. 33-45.

Ehrenberg, S. N., 1975, Feather River ultramafic body, northern Sierra Nevada, California: Geol. Soc. America Bull., v. 86, p. 1235-1243.

Emmons, R. C. 1943, The universal stage: Geol. Soc. Amer. Memoir 8, 205 p.

Goles, G. G., 1967, Trace elements in ultramafic rocks in Wyllie, P. J., ed., Ultramafic and related rocks: John Wiley and Sons, N. Y., p. 352-362.

Himmelberg, G. R., and Loney, R. A., 1973, Petrology of the Vulcan Peak alpine-type peridotite, southwestern Oregon: Geol. Soc. America Bull., v. 84, p. 1585-1600.

Hunter, C. E., 1941, Forsterite olivine deposits of North Carolina and Georgia: North Carolina Dept. Cons. Devel. Bull. 41, 117 p.

Hutchison, C. S., 1975, Ophiolite in Southeast Asia: Geol. Soc. America Bull., v. 86, p. 797-806.

Jackson, E. D., and Thayer, T. P., 1972, Criteria for distinguishing between stratiform, concentric, and alpine type peridotite-gabbro complexes: 24th Int. Geol. Cong. Proceedings, Sec. 2, p. 289-296.

Loney, R. A., Himmelberg, J. R., and Coleman, R. G., 1971, Structure and petrology of the alpine-type peridotite at Burro Mountain, California, U. S. A.: Jour. Pet., v. 12, p. 245-310.

Mercier, J. C., and Nicolas, A., 1975, Textures and fabrics of upper-mantle peridotites as illustrated by xenoliths from basalts: Jour. Pet., v. 16, p. 454-487.

Moores, E. M., 1970, Ultramafics and orogeny, with modes of the U. S. Cordillera and the Tethys: Nature, v. 228, p. 837-840.

Moores, E. M., and MacGregor, I. D., 1972, Types of alpine ultramafic rocks and their implications for fossil plate interactions: Geol. Soc. America Memoir 132, p. 209-223.

Nicolas, A., Boudez, J. L., Boudier, O. F., and Mercier, J., 1971, Textures, structures and fabrics due to solid state flow in some European lherzolites: Tectonophys., v. 12, p. 55-86.

Page, N. J., 1967, Serpentinization considered as a constant volume metasomatic process: A discussion: Am. Mineralogist, v. 52, p. 546-549.

Poldervaart, A., 1950, Correlation of physical properties and chemical composition in the plagioclase, olivine, and orthopyroxene series: Am. Mineralogist, v. 35, p. 1067-1079.

Ragan, D. M., 1963, Emplacement of the Twin Sisters dunite, Washington: American Jour. Sci., v. 261, p. 549-565.

__________, 1969, Olivine recrystallization textures: Mineralogical Mag., v. 37, p. 238-240.

Spry, A., 1969, Metamorphic textures: New York, Pergamon Press, 350 p.

Stevens, R. K., Strong, D. F., and Kean, B. F., 1974, Do some eastern Appalachian ultramafic rocks represent mantle diapirs produced above a subduction zone? Geology, v. 2, p. 175-178.

Swanson, S. E., and Raymond, L. A., 1976, Alteration of the Day Book dunite, Spruce Pine district, western North Carolina (Abs.): Geol. Soc. America Abstracts with Programs, v. 8, p. 282.

Thayer, T. P., 1966, Serpentinization considered as a constant volume metasomatic process: Am. Mineralogist, v. 51, p. 685-710.

__________, 1967, Serpentinization considered as a constant volume metasomatic process: A reply: Am. Mineralogist, v. 52, p. 550-553.

Turner, F. J., 1968, Metamorphic petrology: mineralogical and field aspects: McGraw-Hill, New York, 403 p.

Yoder, H. S., and Sahama, T. G., 1957, Olivine X-ray determinative curve: Am. Mineralogist, v. 42, p. 475-491.

Yurkovich, S. P., 1976, Petrology of the Corundum Hill alpine peridotite, Macon County, North Carolina (Abs.): Geol. Soc. America Abstracts with Programs, v. 8, p. 306.

THE CORUNDUM HILL DUNITE, MACON COUNTY, NORTH CAROLINA

By

Steven P. Yurkovich
Department of Earth Sciences
Western Carolina University
Cullowhee, North Carolina 28723

ABSTRACT

The purpose of this investigation is to determine the geologic history of the Corundum Hill alpine ultramafite. The Corundum Hill complex is composed predominantly of serpentinized dunites (harzbur- gites and olivine orthopyroxenites are observed occasionally). Em- placement of the ultramafite into the adjacent gneisses indicates a lack of structural deformation in the country rock implying a syn-orogenic time of emplacement.

Petrographic analyses established that the ultramafic body went through an early magmatic phase (possibly in the mantle). After crys- tallization, the solidified mass was forcibly injected into the upper crust for cataclases and recrystallization have altered the original mag- matic texture.

Several episodes of metasomatism have affected the ultramafic rocks. First, serpentinization of the dunite occurred with the greatest amount of serpentine found along the shear surfaces within the rocks. This chemical alteration took place late in the emplacement history, probably as the ultramafite was injected into the upper crust. Follow- ing this, the dunite underwent a subsequent thermal event which allowed steatization and amphibolization to occur as a result of influx of hot fluids into the rock fractures.

INTRODUCTION

The ultramafic rocks in western North Carolina are located in the Blue Ridge physiographic province. These deposits (which are com- posed of dunites, peridotites, and troctolites) occupy a linear belt that parallels the axis of the Appalachians extending from the Georgia border northeastward into Virginia (Figure 1). They occur as tabular or lentic- ular bodies with their major axes concordant with the regional foliation of the enclosing rocks. Typically, the ultramafic masses are narrow lenses a hundred meters to about a kilometer in length and several

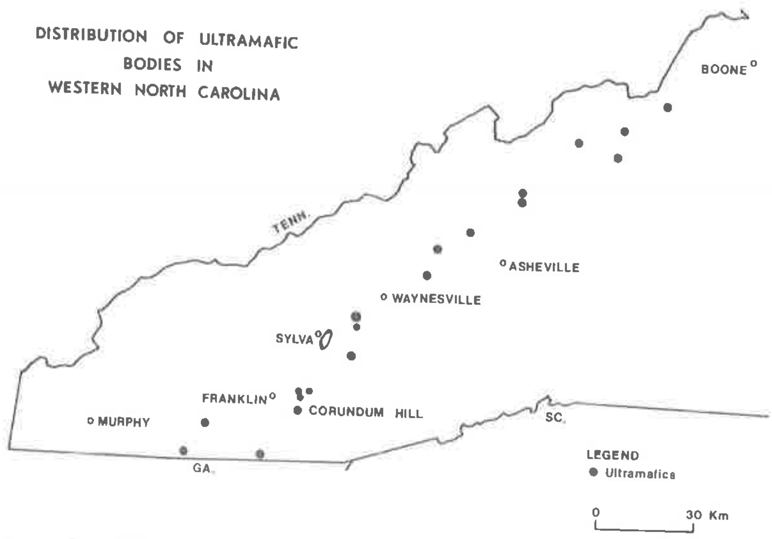


Figure 1. The distribution of ultramafic rocks in western North Carolina. The black circles indicate location and not the size of the bodies.

hundred meters in width. In many cases, the ultramafic rocks are associated with talc schists and/or amphibolites.

The Corundum Hill dunite is located in Macon County, North Carolina, approximately eleven kilometers southeast of Franklin (Figure 1). It is one of the better known ultramafic bodies in the state owning to the quantities of gem quality and abrasive corundum which were mined here through the early part of the century. To date the geologic investigations of this dunite have dealt with the economic mineral potential of the site, particularly the abundances of dunite, corundum, chromite, and vermiculite (Pratt and Lewis, 1905; Hunter and Mattocks, 1936; Hunter, 1941; Hunter et al., 1942; Murdock and Hunter, 1946; Ballard, 1947).

The purpose of this investigation is to establish the petrogenesis of the Corundum Hill ultramafite and within this paper emphasis will be given to the mode of emplacement of the dunite.

Acknowledgment

The author wishes to thank J. R. Carpenter, P. G. White, and J. T. Wilcox for reviewing the manuscript and offering constructive criticisms. This project is supported by the North Carolina Board of Science and Technology grant 631.

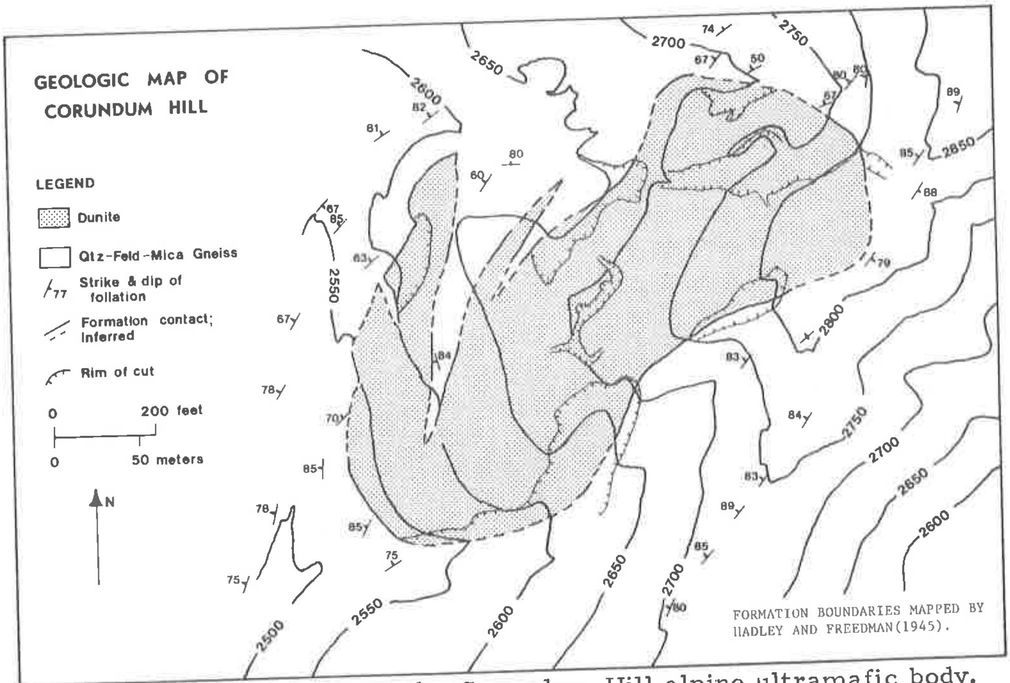


Figure 2. Geologic map of the Corundum Hill alpine ultramafic body.

GEOLOGIC SETTING

Surrounding the Corundum Hill dunite is a muscovite-biotite-plagioclase-quartz gneiss (formerly the Carolina Gneiss). The rocks have well defined layering that may be due partly to relic sedimentary bedding; however, the present texture is probably the result of metamorphic differentiation of mafic and felsic components. Regional strike of the gneissic foliation is N 30-60° E with a dip of 65-85° NW (Figure 2). Locally the dip may be to the southeast. The foliation within the country rock appears to be concordant with the major axis of the dunite body.

These gneisses are the core of an anticlinorium that makes up the southwestern North Carolina area (Hadley and Nelson, 1971). The age of the rocks is uncertain. Fullagar and Odom (1973) report that whole rock rubidium-strontium ages of plutons that cut the gneisses indicate a minimum age of 1.1 b.y. for these rocks.

On the northwest side of the dunite several small exposures of plagioclase-epidote-quartz-hornblende gneiss are seen. Hornblende-plagioclase gneisses were observed in the mine workings on the southeast side of the dunite by Pratt and Lewis (1905). Unfortunately, there are no visible contacts between the gneiss and the dunite. This gneiss could also be of Grenville age or it may be genetically related to the dunite emplacement.

The Corundum Hill ultramafite is a fine grained pale green rock although some grains may reach 20mm in length. The ultramafite is composed predominantly of serpentinized dunite with local occurrences of harzburgites and olivine orthopyroxenites. It is an oval shaped body (400 meters long by 150 meters wide) with its long axis parallel to the regional foliation (Figure 2). On the north side two fingers of dunite extend into the country rock. The contacts of the dunite dip 70-90° to the northwest (Figure 3) suggesting that this is a tabular body, probably a sill.

Field, structural, and petrographic data presented by Yurkovich (1976) show that this dunite can best be described as an "alpine-type" ultramafite. The dunite is separated from the country rock by a talc-amphibole schist which varies in thickness from one to four feet. The talc sheath is in turn separated from the country rock by a vermiculite zone which probably results from hydrothermal alteration of the country rock. There is no evidence within the country rock suggesting the presence of a contact metamorphic aureole, indicating a low temperature of dunite emplacement.

PETROGRAPHY OF THE ULTRAMAFIC BODY

Thin sections of thirty one samples were examined for volume percent of minerals (using standard point counting methods), textures, and microstructures. No oriented thin sections were used.

Mineralogy

The mineralogic composition of the Corundum Hill ultramafite includes the primary minerals olivine, chromite and enstatite with serpentine, talc, chlorite, anthophyllite, cummingtonite (tremolite?), magnetite, and magnesite occurring as secondary phases. Table 1 shows the mineralogy of the three different rock types found in the area.

Within these rocks olivine is the most abundant mineral, having an average grain size of less than 1 mm with some grains reaching 15-20mm in length. The amount of olivine present is inversely related to the amount of serpentine.

Compositions of the olivines were estimated by measuring the position of the (130) olivine x-ray reflection against an internal standard: either the Si (111) reflection or the $KBrO_3$ (012) reflection. Potassium bromate was used in lieu of silicon because in whole rock samples that contained talc, a talc x-ray peak generally interferred with the Si (111) reflection. The samples were run on a Philips diffractometer using CuK_α radiation, a curved crystal monochromator, and a scanning speed of $1/4^o$ 2θ/minute. After determining the two theta value of the (130) olivine reflection (using a minimum of six measurements), the x-ray determinative curve of Yoder and Sahama (1957) was used to estimate the olivine compositions.

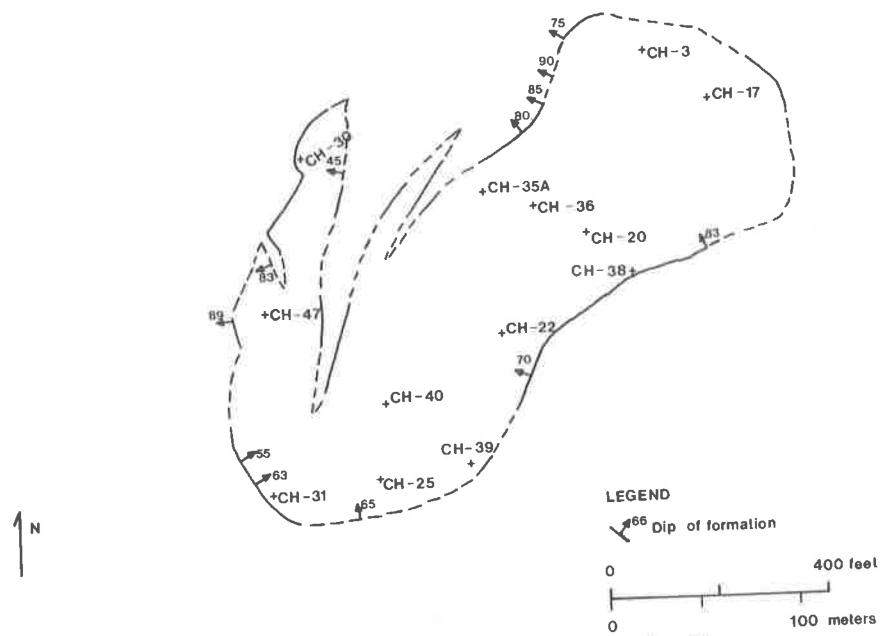


Figure 3. Outline of the dunite showing the direction of dip of the formation contact and the locations of samples used in the x-ray analyses.

Table 1. Variation in the Mineral Assemblages in Volume Percent Among the Rock Types at Corundum Hill.

	Dunite	Harzburgite	Olivine Orthopyroxenite
Olivine	47.1 - 85.4	59.9 - 62.1	30.6
Serpentine	9.6 - 39.6	14.2 - 18.2	3.7
Talc	0 - 14.1	5.2 - 6.1	7.6
Enstatite	0 - 6.9	6.5 - 12.8	54.7
Amphibole	0 - 8.3	10.3 - 12.8	.0
Chlorite	0.2 - 1.3	0.3 - 0.9	2.2
Opaques	0.4 - 3.7	0.8 - 1.6	1.1
Magnesite	.0 - 4.8	trace	.0

Thirteen samples were analysed (Table 2) and the average olivine composition for the dunite was $Fo_{92}Fa_{08}$ (the uncertainty is less than 5 mole percent). An interesting distribution of olivine compositions is observed. The olivines are more magnesian, with the exception of

Table 2. Olivine Compositions Determined by X-ray Diffraction. See Figure 3 for Sample Locations.

Sample Number	Mole Percent Forsterite	Sample Number	Mole Percent Forsterite
CH-3	90.8	CH-35A	94.0
CH-17	91.1	CH-36	89.7
CH-20	89.4	CH-38	95.5
CH-22	89.9	CH-39	93.6
CH-25	91.9	CH-40	91.9
CH-30	89.1	CH-47	94.9
CH-31	92.7		

CH-30, near the dunite margins than within the complex. The data may be explained in one of two ways. First, this distribution is what one would expect during solidification of a magma. Since the margins cool first, the outer olivines are magnesium rich; the inner olivines, which crystallized at lower temperatures would be enriched in iron. However, Carpenter and Phyfer (1975), in microprobe analyses of the Day Book, Newdale, Holcomb Branch, and Balsam Gap deposits, showed that there was no variation in chemistry within or among the ultramafites. Second, nearly all of the olivine compositions are within one standard deviation of the average. If the olivine compositions are constant, as indicated by Carpenter and Phyfer, small errors in experimental technique and/or the method employed might produce the variability observed. If this is correct, a more precise geochemical analysis must be done before this trend can be verified.

Enstatite occurs as interstitial blades between olivine and as pods or veins (up to about 2 cm wide) cutting through the dunite. Many of the veins intersect at nearly right angles. In all cases the orthopyroxene grains are much larger than the olivine and may reach a length of 5 cm. Many of the dunite samples contain enstatite and as the enstatite concentration increases harzburgites and olivine orthopyroxenites are formed. The enstatite is occasionally altered to talc.

Approximately 1 percent of the rock is made up of chromite that may reach a diameter of 1.5mm. The chromite is euhedral to subhedral (with octahedral outlines) and is observed as disseminated grains or in bands. Some of the chromite grains contain non-opaque centers that appear to be olivine inclusions. If so, some of the chromite postdates olivine crystallization. Chromite can be seen in all stages of alteration to a chromium-chlorite.

Serpentine is the major alteration product of olivine. The alteration is concentrated along fracture surfaces and grain margins. After the olivine has completely reacted to form serpentine, ghost images of the original olivine remain outlined by magnetite flakes. X-ray diffraction work suggests that the serpentine is clino-chrysotile.

Two amphiboles are associated with the dunite. Anthophyllite occurs in a needle like growth up to several centimeters in length. A monoclinic amphibole (extinction angle to 17°) is also present as smaller stubby prismatic crystals. It has optical properties similar to either cummingtonite or tremolite. This material may be the dominant amphibole is some of the dunites. The amphiboles form at the expense of olivine as they cut through the olivine grains.

Talc and chlorite form as veinlets throughout the dunite mass. In many parts of the dunite they appear along joint surfaces. Magnesite is observed in the talc veins as disseminated grains (of secondary origin) or less frequently as halos around olivine grains.

Textures and Microstructures

Understanding of the rock structure is important in establishing the evolution of the rock mass. Microscopic examination of the dunite textures suggests at least three different episodes in rock genesis. These are (1) an early development of primary structures, (2) the formation of a flow texture (?) and (3) a later generation of deformational structures.

The primary structures within the dunite include an allotrimorphic granular texture (randomly oriented and interlocking anhedral olivine grains) and euhedral to subhedral chromite grains. The chromite is frequently found in bands, lenses or small pods. Such textures are analogous to those seen in magmatically formed dunites.

The next phase of rock genesis was the generation of pronounced elongation of large olivine grains. These olivine "phenocrysts" are pencil shaped and may be up to 15mm long; however, more commonly they are 3-5mm in length. Within any one thin section the elongated olivine grains are parallel to one another, and in some sections the chromite layers appear to be subparallel to this elongation direction. Since oriented sections were not used, the direction of elongation could not be determined. This olivine orientation may be due to plastic flow of solid dunite as it was injected upwards or to flowage of the mass during a magmatic stage.

Finally, the textures show the development of recrystallization and deformational features. Many thin sections have textures indicating recrystallization of the dunite has taken place. These textures are typified by a fine mosaic (grain size < 0.4 mm) of recrystallized olivine surrounding and in some cases cutting through the large olivine "phenocrysts." The fine mosaic is evidenced by olivine that is unstrained while the "phenocrysts" have undulose extinction, deformation lamellae, and in some cases kink bands. The grain boundaries of the olivine mosaic tend to be straight and meet at high angles. This is typical of annealing recrystallization.

Other mineral grains also show cataclasis. Olivine and chromite have been broken, crushed, and in some instances these fractures are filled with serpentine. Granulation is pronounced, but the broken mineral fragments show no signs of recrystallization.

A distinct lamination appears in the dunite. The laminations (1-2mm in width) appear in hand samples as alternate layers of pale green olivine and dark gray bands of serpentine. These laminations are probably due to shearing in the dunite during emplacement. The shear zones include fractures and crushed olivine and are avenues along which serpentinization has developed. In some shear zones the olivine is completely replaced by serpentine. The shear zones are most prominent near the dunite margins where they are parallel to the formation contacts. Some of the laminated zones have slickensides developed along them.

SERPENTINIZATION

Serpentinization of the dunites began after solidification. The amount of serpentine within the mass is variable (Table 1) with the range for the body being 10-40 volume percent. The serpentine is most abundant along the shear planes where up to 40 percent of the dunite is altered. These shear zones are commonly observed near the dunite margins. However, other dunites (without the shear surfaces) near the border of the formation may have as little as 10 percent serpentine. It appears that serpentinization is controlled by the presence of fractures and/or shear surfaces. This relationship between serpentinization and shear planes indicates that water for serpentinization was introduced into the dunite along the break surfaces from the surrounding country rock and not as a result of autometamorphism of a hydrous serpentine mass.

Serpentinization probably occurred during the latter stages of dunite emplacement. As the dunite was forced toward the surface, it incorporated water from the sediments in the higher crustal levels, and chemical alteration developed along shear surfaces, fractures, and grain margins. It appears that the break surfaces were tight, for little serpentinization is observed in the interior of the rock mass. There is no indication of a volume increase as a result of this process.

Using the experimental data from the system MgO-SiO_2-H_2O the temperature range for serpentinization can be approximated. Johannes (1968) has shown that at pressures below 1Kb, brucite + serpentine form from forsterite + water at 340°C; at a pressure of 3Kbs this reaction occurs at about 390°C. Since brucite is not observed at Corundum Hill, this univariant reaction defines the lower temperature limit of serpentinization. The reaction serpentine = forsterite + talc + water vapor has been determined by Bowen and Tuttle (1949). For pressures below 1Kb and 3Kbs the reaction temperatures are 500°C and

525°C respectively. This reaction then establishes the upper limits of serpentine stability. Addition of Fe to the system (the olivine composition is $Fo_{92}Fa_{08}$) would lower the temperature of the serpentine reactions.

STEATIZATION

A thick talc-amphibole-chlorite sheath surrounds the dunite and also fills joints throughout the rock mass. In places the sheath has a pronounced schistosity. Large blocks of dunite 3-5 meters in dimension are completely enveloped by this talcose material. Veins of talc (occasionally with magnesite) cut through the dunite. Also, in some of the joints chlorite or chlorite-corundum deposits are developed. These observations indicate that steatization, amphibolization and chloritization postdate serpentinization and these processes occurred either during the later stages of emplacement or after emplacement of the dunite.

Within the talc zones anthophyllite is the most common amphibole although a second amphibole may be present. Using the experimental data of Greenwood (1963) the lower temperature limits of the equilibrium pair talc-anthophyllite at pressures of 1Kb, 2Kb, and 2.6Kb are 660°C, 675°C, and 680°C respectively with an upper stability limit at the same pressures of 690°C, 705°C, and 715°C.

Since field studies show that steatization followed serpentinization, the dunite body must have undergone a subsequent heating event after serpentinization. Two hypotheses for talc formation may be applicable to the Corundum Hill dunite.

First, steatization may result from metasomatism between the country rock (gneiss) and the serpentinized dunite. Chidester (1962), in his study of steatization of dunites in Vermont, concluded that Mg migrated from the serpentine to the blackwall (chlorite zone) while Si moved from the country rock into the steatite zone. He suggested that these solutions were probably mobilized from the country rock during regional metamorphism. Since metamorphism at Corundum Hill reached the upper amphibolite facies (Hadley and Nelson, 1971) a similar type of metasomatic reaction may have been generated here. If this hypothesis proves correct the dunite must have been emplaced prior to or contemporaneously with the latest metamorphic event in the region.

Second, steatization may have resulted from hydrothermal alteration related to intrusion of pegmatitic (or granitic) material. Pratt and Lewis (1905) and Cobb (1924) indicate the presence of granitic material within the dunite. No felsic dikes were seen in the dunite by this author; however, the surrounding rocks are cut through by granitic-like material. The hot solutions from this magma reacting with the dunite may have produced the talcose material.

Although no definitive solution can be given to the origin of steatite until geochemical analyses are performed, the following

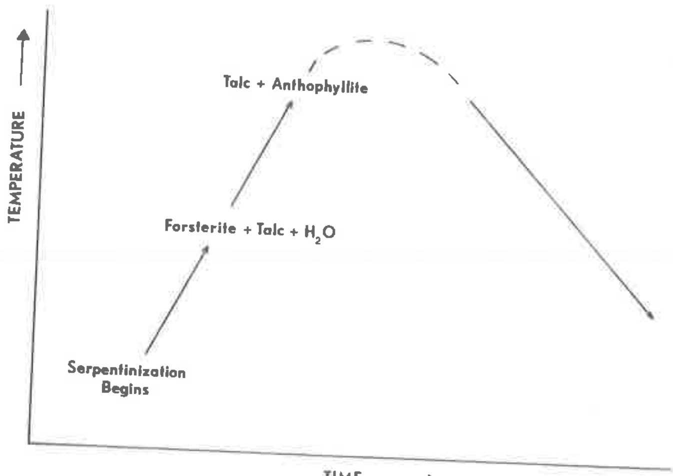


Figure 4. Schematic representation of the chemical alteration of the dunite with respect to temperature and time of metamorphism.

discussion might be pertinent to establishing the origin of the alteration
products. Two models are presented which attempt to explain the re-
lationship between serpentinization and formation of steatites.

If the dunite was emplaced either pre- or synorogenically the
timing of serpentinization with respect to steatization might be depicted
schematically as in Figure 4. During prograde metamorphism, ser-
pentinization would be followed by its dehydration, with the water vapor
dispersed into the country rock. At higher temperatures talc would be
generated and the talc-anthophyllite pair would become stable. As
temperatures declined from peak metamorphism one would expect that
the above mentioned reactions would occur in reverse order with retro-
grade metamorphism (under equilibrium conditions). This should hap-
pen unless (1) the temperature drop is rapid or (2) reaction kinetics
are slow or (3) H_2O was so effectively removed from the system that it
is no longer available. In any case the high temperature phases would
then persist metastably.

A thick sheath of talc-anthophyllite presently surrounds the
dunite. In order for serpentine to form during retrograde metamorphism
water and other ions would have to migrate through this metastable talc
sheath and leave no evidence of this migration (e. g., serpentine is not
found in the talcose layer). In this model the dunite would be serpen-
tinized on two separate occasions.

Another possible solution is that the dunite was emplaced into
the upper crust late in the metamorphism (during cooling) with ser-
pentinization taking place at that time. Later, hydrothermal fluids de-
rived from pegmatitic activity were introduced along fracture surfaces

with subsequent formation of talc and talc-anthophyllite. In this model emplacement must be syn-to late orogenic.

DISCUSSION AND CONCLUSIONS

Several models have been proposed for the emplacement mechan-ism of the alpine-type ultramafic bodies in western North Carolina. Hartley (1973) indicated that the Lake Chatuge sill was emplaced as a gabbroic crystal mush. In a petrofabrics study of the Buck Creek dunite, Sailor and Kuntz (1973) observed brittle deformation features super-imposed upon recrystallization textures. They state "these factors sug-gest that emplacement of the dunite into the crust was achieved when the dunite was relatively cold and incapable of recrystallization, possibly by thrusting." Madison (1968) suggested that the Webster-Addie com-plex was emplaced essentially as an anhydrous solid. Recent work at Webster-Addie by Greenberg (1976) provides evidence that emplace-ment of the body was by diapiric rise of a peridotite mass.

Carpenter and Phyfer (1969) have proposed that some of the alpine ultramafites in the southern Appalachians may have been emplac-ed as a result of intrusion of crystalline serpentine. Astwood, Carpenter, and Sharp (1972) suggest this method for the emplacement of the Dark-ridge and Balsam Gap dunites.

From the evidence presented, the petrogenesis and emplace-ment of the Corundum Hill dunite can best be described by an early magmatic period, followed by crystallization of the magma then tectonic emplacement of the solid into the crust. The magmatic event may have resulted from partial melting and fractionation of mantle material. Coleman (1971) suggests that partial melting of lherzolite at depths of less than 50Km could produce a subalkaline basaltic magma. Differentia-tion of this liquid in the upper levels of the ridge systems could generate gabbros and trondjemites both of which are associated with ophiolites and ocean crust. Coleman indicated that residues developed by partial melting of a primitive mantle are represented by harzburgite-dunite masses upon which the oceanic crust is formed.

This magmatic material moved upward and crystallized in the lower crust or upper mantle. The magmatic stage would account for the development of primary igneous textures, structures, and the varia-tion in olivine compositions (if the x-ray analyses are accurate). In addition, the subparallel alignment and/or elongation of olivine grains might be attributed to the magmatic period. After crystallization the dunite probably possessed a grain size much larger than presently ob-served.

The solidified mass was then reinjected (faulted?) into the upper crust along a zone of weakness (possibly a bedding plane in the gneiss). During this migration cataclasis, plastic deformation (possibly generat-ing flow structure) and recrystallization of the olivine occurred.

According to Ragan (1967) recrystallization of olivine can take place between the incongruent melting temperature of enstatite and the upper stability limit of serpentine. By extrapolating their experimental data, Ave'Lallemant and Carter (1970) indicate that recrystallization of oli- vine may take place at temperatures as low as 500°C under the lower rates of natural deformation.

Stress fields of greater magnitude were set up nearer the mar- gins of the mass than at the center of the complex. These stresses caused greater amounts of cataclastic deformation and fracturing near the dunite margins with the shear zones oriented nearly parallel to the contact. Ragan (1967) states that at temperatures of less than 450°C olivine deforms plastically without recrystallization to produce undulose extinction within the mineral. In this stage of deformation shear planes and undulose extinction were formed.

Forceful injection of the solidified mass was necessary. Some of the ultramafic mass may have been stripped away during injection. On the northwest and southeast sides of the dunite, small exposures of plagioclase-hornblende gneiss are observed. It is possible that these gneisses were derived by metamorphosing gabbroic or dioritic material which was emplaced along with the dunite.

During the later stages of emplacement (or after emplacement) water was added to the dunite and serpentinization occurred along shear surfaces and grain boundaries. At temperatures below 500°C serpen- tinization was occurring contemporaneously with cataclastic deforma- tion. Broken mineral grains now have their fractures filled with ser- pentine. Finally, alteration along fractures produced the talc layer around the dunite. Since the dunite had already solidified and was being emplaced into rocks undergoing regional metamorphism, no contact aureole is found in the surrounding rocks.

Since the foliation with the country rock is concordant with the major axis of the dunite body, emplacement of the ultramafite probably occurred during orogenesis in the Ordovician. Paleomagnetic studies of the Webster-Addie dunite by Madison (1968) yielded a relative Ordo- vician age. Similar ages for other ultramafites have been proposed by Pratt and Lewis (1905), Hadley and Nelson (1971), Hartley (1973) and others. This age coincides with the early folding and regional meta- morphism in the Southern Appalachians. During this time the Southern Appalachians were being folded recumbently and pushed towards the northwest (Hatcher, 1972). The stresses during orogenesis may well have been the driving mechanism that forced the dunite upwards.

REFERENCES

Astwood, P. M., Carpenter, J. R., and Sharp, W. E., 1972, A petro- fabric study of the Dark Ridge and Balsam Gap Dunites, Jack- son County, N. C.: Southeastern Geol., v. 14, p. 183-194.

Ave'Lallemant, H. G., and Carter, N. L., 1970, Syntectonic recrystallization of olivine and modes of flow in the upper mantle: Geol. Soc. Amer. Bull., v. 81, p. 2203-2220.

Ballard, T. J., 1947, Corundum Hill Mine, Macon County, N. C.: Bureau of Mines Report of Investigations 4042, 10 p.

Bowen, N. L., and Tuttle, O. F., 1949, The systems $MgO-SiO_2-H_2O$: Geol. Soc. Amer. Bull., v. 60, p. 439-460.

Carpenter, J. R., and Phyfer, D. W., 1969, Proposed origin for the alpine type ultramafics of the Appalachians (Abstr.): Geol. Soc. Amer., Annual meetings, p. 261-263.

__________, 1975, Olivine compositions from Southern Appalachian ultramafics: Southeastern Geol., v. 16, p. 169-172.

Chidester, A. H., 1962, Petrology and geochemistry of selected talc-bearing ultramafic rocks and adjacent country rocks in North-Central Vermont: U. S. Geol. Survey Prof. Paper 345, 207 p.

Cobb, M. C., 1924, The origin of corundum associated with dunite in Western North Carolina: Ph. D. Thesis, Bryn Mawr College, 43 p.

Coleman, R. G., 1971, Plate tectonic emplacement of upper mantle peridotites along continental edges: Jour. Geophys. Research, v. 76, p. 1212-1222.

Fullagar, P. D., and Odom, A. L., 1973, Geochronology of Precambrian gneisses in the Blue Ridge province of Northwestern North Carolina and adjacent parts of Virginia and Tennessee: Geol. Soc. Amer. Bull., v. 84, p. 3065-3080.

Greenburg, J. K., 1976, The alpine ultramafic problem in the Southern Appalachians: The Webster-Addie Dunite (Abstr.): Geol. Soc. Amer., NE/SE Regional Meetings, p. 185.

Greenwood, H. J., 1963, The synthesis and stability of anthophyllite: Jour. Petrology, v. 4, p. 317-351.

Hadley, J. B., and Freedman, A., 1947, Open File Map in Bureau of Mines Report of Investigations 4042, 10 p.

Hadley, J. B., and Nelson, A. E., 1971, Geologic map of the Knoxville quadrangle, North Carolina, Tennessee, and South Carolina: U. S. Geol. Survey, Misc. Geol. Inv. Map I-654.

Hartley, M. E., III, 1973, Ultramafic and related rocks in the vicinity of Lake Chatuge: Ga. Geol. Survey Bull. 85, 61 p.

Hatcher, R. D., 1972, Developmental model for the Southern Appalachians: Geol. Soc. Amer. Bull., v. 83, p. 2753-2760.

Hunter, C. E., 1941, Forsterite olivine deposits of North Carolina and Georgia: North Carolina Geol. Econ. Survey Bull. 41, 117 p.

Hunter, C. E., and Mattocks, P. W., 1936, Vermiculites of western North Carolina and North Georgia: TVA Div. Geol. Bull., v. 5, p. 1-10.

Hunter, C. E., Murdock, T. G., and McCarthy, G. R., 1942, Chromite deposits of North Carolina: North Carolina Geol. Econ. Survey, Bull. 42, 39 p.

Johannes, W., 1968, Experimental investigations of the reaction forsterite + H_2O = serpentine + brucite: Contr. Mineral. Petrology, v. 19, p. 309-315.

Madison, J. A., 1968, Petrology and geochemistry of the Webster-Addie ultramafic body, Jackson County, N. C.: Ph.D. Thesis, Washington University, 139 p.

Murdock, T. G., and Hunter, C. E., 1946, The vermiculite deposits of North Carolina: North Carolina Geol. Econ. Survey, Bull. 50, 44 p.

Pratt, J. S., and Lewis, J. V., 1905, Corundum and peridotites of western North Carolina: North Carolina Geol. Survey, v. 1, 464 p.

Ragan, D. M., 1967, The Twin Sisters Dunite, Washington: in Ultramafics and Related Rocks, P. J. Wyllie, editor, John Wiley and Sons, New York, 464 p.

Sailor, R. V., and Kuntz, M., 1973, Petrofabric and textural evidence for syntectonic recrystallization of the Buck Creek Dunite, N. C. (Abstr.): Geol. Soc. Amer., Annual Meetings, p. 791-792.

Yoder, H. S., and Sahama, Th. G., 1957, Olivine X-ray determinative curve: Amer. Mineral., v. 42, p. 475-491.

Yurkovich, S. P., 1976, Petrology of the Corundum Hill Alpine Peridotite, Macon County, N. C. (Abstr.): Geol. Soc. Amer., NE/SE Regional Meetings, p. 306.