Southeastern Geology: Volume 25, No. 1
June 1984

Edited by: S. Duncan Heron, Jr.

Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

Heron, Jr., S. (1984). Southeastern Geology, Vol. 25 No. 1, June 1984. Permission to re-print granted by Duncan Heron via Steve Hageman, Professor of Geology, Dept. of Geological & Environmental Sciences, Appalachian State University.

SOUTHEASTERN GEOLOGY

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 25, NO. 1 JUNE, 1984

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

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SOUTHEASTERN GEOLOGY

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May, 1984

DELTAIC DEPOSITS OF THE UPPER MISSISSIPPIAN RAVENCLIFF MEMBER OF THE HINTON FORMATION SOUTHERN WEST VIRGINIA

R. A. SCHALLA *Cities Service Oil and Gas Corporation, 1600 Broadway, Suite 900, Denver, Colorado 80202*

ABSTRACT

The Ravencliff Member is formally proposed as a lithostratigraphic subdivision of the Hinton Formation in southern West Virginia.

The Ravencliff comprises a heterogeneous sequence of sandstones, pebble conglomerates, and interbedded shales, which are shown to be of fluvial and deltaic origin. The Ravencliff Member is restricted to those strata which lie above the top of the Little Stone Gap Member (Avis Limestone) and below the "Hot Shale" marker bed.

Analysis of subsurface data, principally geophysical and driller's logs, permits paleoenvironmental reconstruction of a Ravencliff deltaic complex in McDowell County, West Virginia. Subsurface maps, stratigraphic cross-sections, and geophysical log (gamma ray - density) character were the principal criteria used in the delineation of three essentially contemporaneous environments of deposition within the Ravencliff of this area. These are: 1) delta-plain and distributary channel; 2) delta front and shoreline; 3) delta platform and prodelta. Each of these lithofacies represents a distinct environment within the southward prograding deltaic complex.

INTRODUCTION

This project was undertaken in an effort to refine the lithostratigraphy of a portion of the Upper Mississippian Chester (Mauch Chunk) Series in southern West Virginia. Hydrocarbon production has been established from sandstones of the Chester Series in this area and in many cases these sandstones have proven to be prolific gas producers. Hydrocarbon accumulations are commonly the result of stratigraphic traps associated with facies variations within these strata. For this reason, an understanding of the lithofacies relationships within the Chester Series is important to petroleum exploration in this region. The subsurface maps and stratigraphic cross-sections presented in this report provide a paleodepositional model for the Ravencliff Member of the Hinton Formation.

The study area covers a total of approximately 400 square miles and is located within central and western McDowell, and southwestern Wyoming Counties, West Virginia (figure 1).

PREVIOUS WORK

Campbell (1893), and Campbell and Mendenhall (1896), were the first workers to establish a stratigraphic section for the Upper Mississippian strata dealt with in this study. Later detailed studies and mapping by Reger (1926), Wilpolt and Marden (1949; 1959), Harris and Miller (1964), and others resulted in revisions of Campbell and Mendenhall's (1896) original units, and established the presently accepted stratigraphic section for southern West Virginia and adjacent areas (figure 2). Although a brief discussion of the results of this study has already been presented (Schalla, 1980), no detailed subsurface study of the Upper Mississippian strata of this area has yet been published.

STRATIGRAPHY

General Statement

The heterogeneous sequence of sandstones, quartz pebbie conglomerates, and

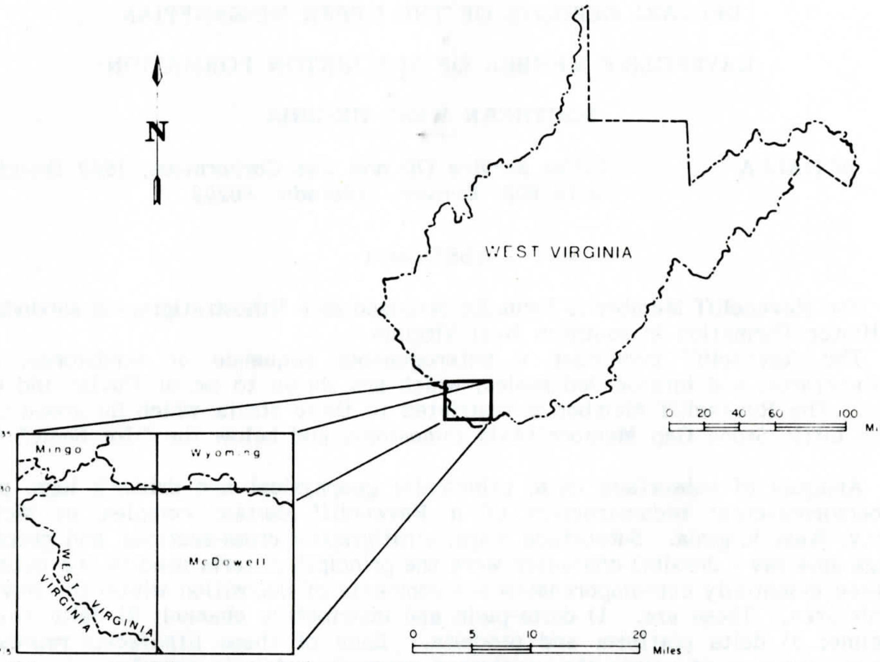

Figure 1. Location map showing study area.

	REGER (1926)		WILPOLT AND MARDEN (1959)			THIS STUDY		
MAUCH CHUNK SERIES	BLUESTONE GROUP	NUMEROUS FORMATIONS	PENNINGTON GROUP	BLUESTONE FORMATION	UPPER MEMBER RED MEMBER GRAY SHALE MEMBER	PENNINGTON GROUP	BLUESTONE FORMATION	
	PRINCETON CONGLOMERATE			PRINCETON SANDSTONE	(CONSIDERED TO BE EQUIVALENT TO RAVENCLIFF IN SUBSURFACE)		PRINCETON SANDSTONE	
	HINTON GROUP*	L. PLUTO SHALE FALLS MILLS SS. TALLERY SS. LOW GAP SS. AVIS SANDSTONE AVIS LIMESTONE NUMEROUS FORMATIONS STONY GAP SS.		HINTON FORMATION	UPPER RED MEMBER LIMESTONE MEMBER MIDDLE RED MEMBER STONY GAP SANDSTONE MEMBER		HINTON FORMATION	RAVENCLIFF MEMBER LITTLE STONE GAP MEMBER
	BLUEFIELD GROUP	NUMEROUS FORMATIONS	BLUEFIELD FORMATION			BLUEFIELD FORMATION		

* REGER'S HINTON GROUP CONTAINED 44 NAMED SUBDIVISIONS MANY OF WHICH ARE NOT SHOWN ON THIS CHART

Figure 2. Upper Mississippian correlation chart showing previous stratigraphic nomenclature in southern West Virginia and the relationship of the Ravencliff Member.

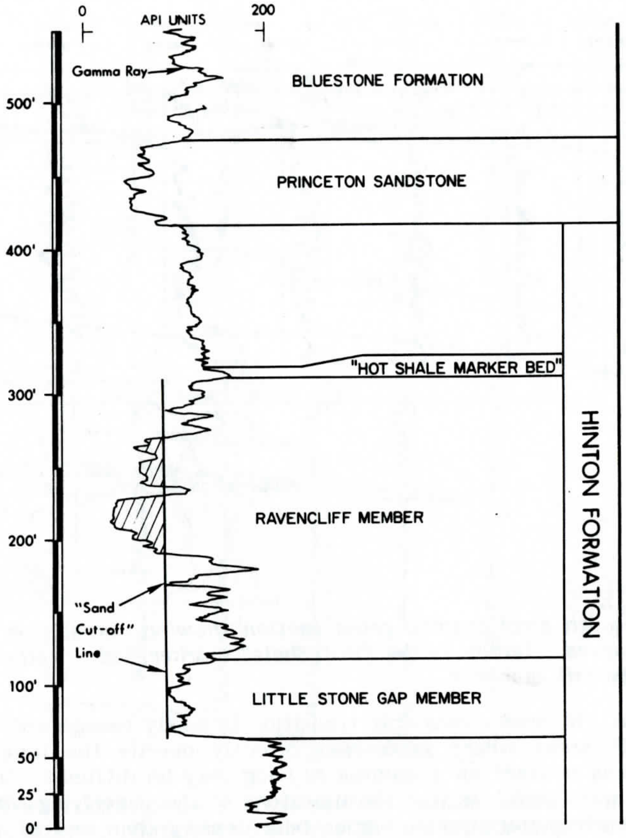

Figure 3. Gamma ray log showing typical character of units discussed in this report, as well as the "Sand Cut-off" line used in mapping.

interbedded variegated shales which directly overlies the Little Stone Gap Member of the Hinton Formation in central McDowell County, West Virginia are, herein, formally designated the Ravencliff Member. The name Ravencliff or Ravenscliff is a drillers' term having widespread acceptance in the petroleum industry in southern West Virginia and adjacent parts of Kentucky and Virginia. The origin of this name is unknown to the writer; however, it appears to be generally accepted by the West Virginia Geological Survey as an informal designation for one of the upper sandstones of the Chester (Mauch Chunk) Series (Haught, 1959).

The Ravencliff Member, as defined in this report, comprises a distinctive group of lithologies, which can be identified easily in the subsurface through the use of gamma ray and other geophysical logs. Correlation of subsurface data with descriptions of surface exposures in Mercer County, West Virginia, suggests that the Ravencliff is laterally equivalent to a portion of the upper Hinton Group of Reger (1926). Because many of the lithologic units delineated by Reger (1926) in outcrop are of such limited thickness and areal extent, their value in subsurface applications is very limited. Additionally, Harris and Miller's (1958) demotion of Reger's Hinton Group to formational status indicates that a reevaluation of the unit boundaries within these strata is necessary. The Ravencliff Member is, therefore, proposed in an effort to refine stratigraphic relationships within the upper Hinton Formation of the Pennington Group.

Contacts

An abrupt upward transition from gray, argillaceous, fossiliferous limestone, to variegated shales, sandstones, and conglomerates marks the lower contact of the

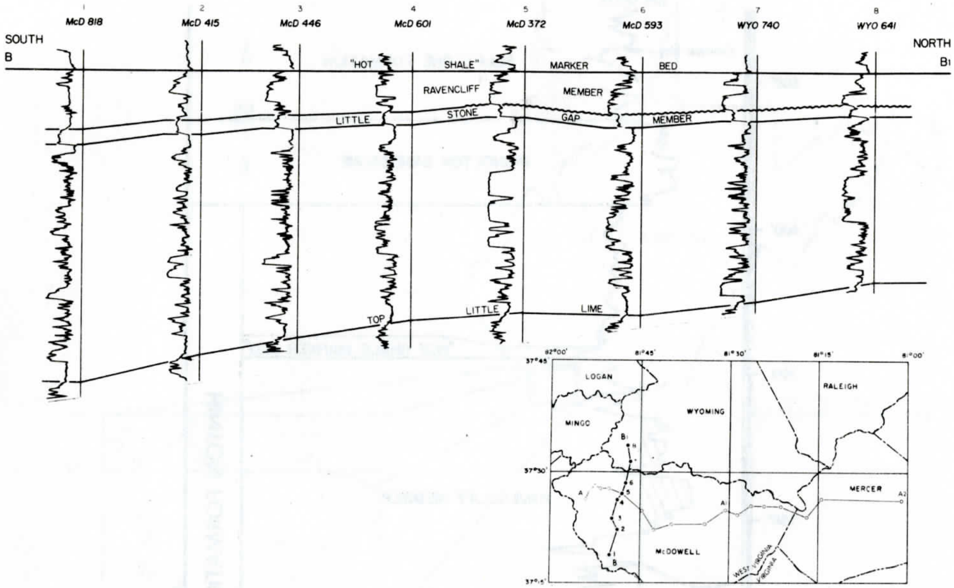

Figure 4. North-south stratigraphic cross section showing correlation of eight wells with gamma ray curves. Datum is the "Hot Shale" marker bed. Wells are designated by West Virginia permit numbers.

Ravencliff Member. In most areas this transition is easily recognized on gamma ray logs (figure 3). In areas where sandstones directly overlie the limestone, precise determination of this contact on a gamma ray log may be difficult. In such cases a bulk density log is most useful, in that the limestone of the underlying Little Stone Gap Member exhibits a characteristically higher bulk density than any of the lithologies composing the Ravencliff Member. If a density log is not available, analysis of well cuttings generally allows determination of this contact with five to ten feet accuracy.

Although the contact is generally conformable, the Ravencliff Member may overlie the Little Stone Gap Member (Avis Limestone of Reger, 1926) with local disconformity as a result of erosion by Ravencliff channels. Evidence of this is present in western McDowell and southwestern Wyoming Counties, where fluvially deposited sandstones and siltstones directly overlie an abnormally thin Little Stone Gap Member (figure 4).

The upper limit of the Ravencliff Member is defined as the base of a distinctive shale unit, whose lowermost strata commonly exhibit anomalously high radioactivity. These strata appear as a "hot streak" on gamma ray logs and are, therefore, of considerable utility in subsurface correlation. These highly radioactive shales have been traced in the subsurface throughout the study area and have proven to be the only consistent marker bed within the upper Hinton Formation. Additionally, these strata are thought to represent a depositional unit which is genetically distinct from the subjacent Ravencliff Member. These shale strata are nominally referred to in this report as the "Hot Shale" marker bed (figure 3). The "Hot Shale" marker bed conformably overlies the Ravencliff Member in all areas studied.

In western McDowell County, where the Ravencliff Member interfingers with laterally equivalent strata, the upper contact of the Ravencliff is indeterminate and the "Hot Shale" marker bed serves only as a guide to gross lateral correlations.

Age and Correlation

Reger (1926) analysed numerous collections of invertebrate fossils from shales and limestones of the "Hinton Group" to the east of the study area. These collections indicated a Late Mississippian (Chesterian) age for these strata in Summers, Mercer, and Monroe Counties, West Virginia. Later workers (Butts, 1940; Cooper, 1944;

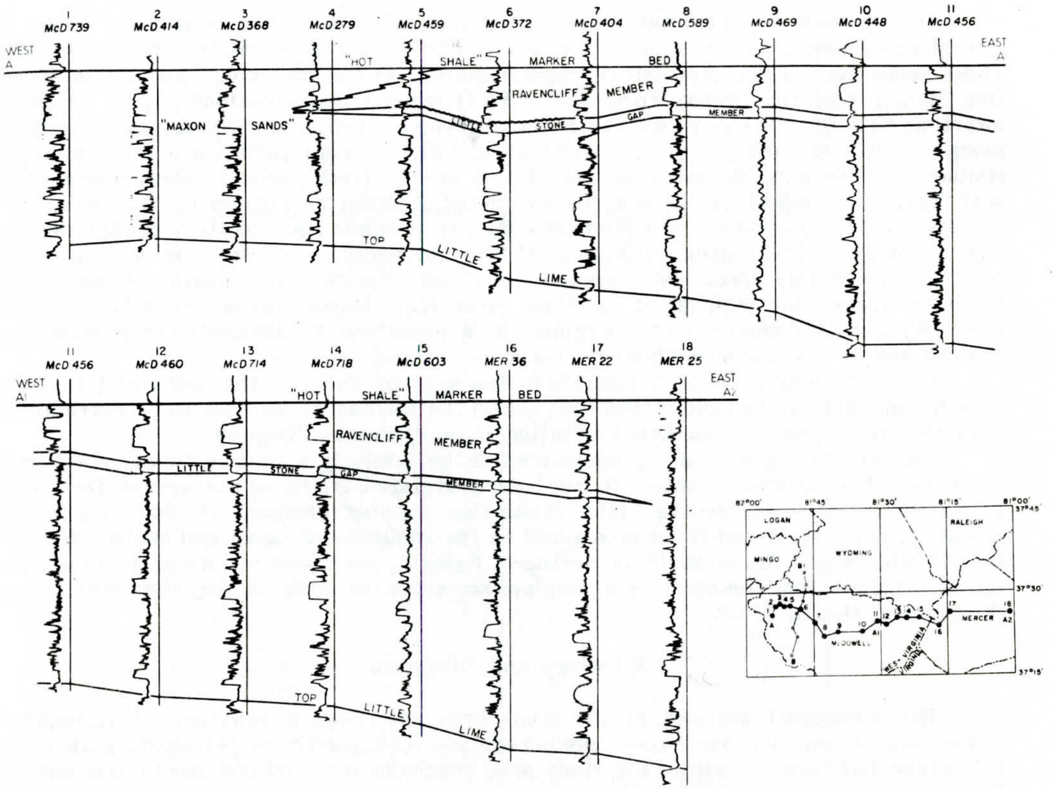

Figure 5. East-west stratigraphic cross section showing correlation of eighteen wells with gamma ray curves. Datum is the "Hot Shale" marker bed. Wells are designated by West Virginia permit numbers.

Ettensohn and Peppers, 1979; and others) have extended Reger's (1926) initial age assignment for the "Hinton Group", to include laterally equivalent strata to the south and west of the study area. Palynological analysis (W. C. Meyers, personal communication) of a core taken by Cities Service Company, in southern Nicholas County, West Virginia, indicates a Late Mississippian age for strata of the upper "Hinton Group" in this area as well. No paleontologic or palynologic studies have been made of the Ravencliff Member within the study area.

Wilpolt and Marden (1959, p. 602) considered the "Ravencliff sand" in southern West Virginia to be equivalent to Reger's (1926) Princeton Sandstone, in Summers and Mercer Counties (figure 2). This correlation is not supported by the findings of this study. For example, in north central Mercer County, United Fuel Gas Company's Parker Wilson and others #9341 (MER-17) test well penetrated the top of the Ravencliff Member at a depth of approximately 700 feet. According to Reger (1926) this well is situated next to surficial exposures of the Princeton Sandstone.

The Ravencliff Member has been traced in the subsurface to the east of the study area where it is considered to be equivalent to part of the upper Hinton Group of Reger (1926), specifically, the interval from the top of the Avis Limestone through the Falls Mills Sandstone, possibly including all or part of the Lower Pluto Shale (figure 2). This correlation appears reasonable, based upon the gross lithologic similarities of Reger's outcrop descriptions and available geophysical and driller's logs, and the generally comparable thicknesses of these intervals. In western Mercer County, subsurface correlations indicate a total thickness for the Ravencliff Member of between 410 and 465 feet. This compares well with the 430 foot maximum thickness recorded by Reger (1926) in Mercer County, for the interval between the top of the Avis Limestone and the base of the Pluto Limestone.

To the west of the study area, the Ravencliff Member is considered to be a lateral equivalent of what is nominally referred to in this report as the "Maxon sands". These sands occupy the same stratigraphic interval as the Ravencliff and Little Stone Gap Members of the Hinton Formation. Stratigraphic cross-sections (figure 5) show interfingering of "Maxon sands" with the Ravencliff and Little Stone Gap Members along the western margin of the study area. The "Maxon sands" are a genetically similar, but distinct, depositional unit which are at least partially contemporaneous with Ravencliff deposition. The term Ravencliff is often misapplied to "Maxon sands" in Pike County, Kentucky, and Buchanan County, Virginia, due to their similar depths and lithology. This misidentification of the Ravencliff confuses the stratigraphic relationships in this area. Because the Ravencliff Member is a separate depositional event in space, but equivalent in time with the "Maxon sands" of Pike County, Kentucky and Buchanan County, Virginia, it is important to differentiate it from the "Maxon sands" and other equivalent strata to the west.

Detailed analysis of both subsurface and outcrop data, to the south of the study area by Englund and DeLaney (1966) has shown the Ravencliff Member to be correlative with the lower part of the Lee Formation in southwestern Virginia.

The Ravencliff Member is considered to be laterally extensive to the north and northeast of McDowell County. It has been correlated in the subsurface as far north as Raleigh County, where the Little Stone Gap Member pinches out, due to a facies change, and is considered to be equivalent to the sequence of sands and shales referred to by drillers as "Ravencliff" in northern Fayette, and western Nicholas Counties. Gross lithologic and paleontologic similarities serve as a basis for this correlation (Kamm and Heald, 1983).

Lithology and Diagenesis

The Ravencliff Member of the study area comprises a sequence of variegated shales, sandstones, and occasional conglomerates. The paucity of petrologic data from the Ravencliff Member within the study area precludes detailed analysis of this unit in McDowell County. The lithologic descriptions contained in this report were derived largely from core and outcrop descriptions of strata considered to be equivalent to the Ravencliff to the east and north. These data support the limited petrologic information available from drill cuttings from within the study area.

Kamm and Heald (1983) reported that sandstones of the Ravencliff Member are composed of approximately 90 percent quartz, in the form of both detrital grains and secondary overgrowths. The remaining 10 percent includes authigenic carbonate, detrital phyllosilicates, authigenic kaolinite, and small amounts of feldspar, heavy minerals and opaques. The sand grains are generally fine to medium-grained, sub-angular to sub-round, and moderately well-sorted. Quartzose pebbles up to 2.0 centimeters in diameter are commonly scattered throughout the sandstones. Conglomeratic zones up to several feet in thickness may be interbedded.

Shales, which are commonly intercalated in the sandstone strata of the Ravencliff, range in color from dark gray, to various shades of red and green (Reger, 1926). These shales are generally silty and may grade upward into very fine-grained sandstones.

The diagenetic history of sandstones within the Ravencliff Member has been the subject of investigation of several recent studies (Pawlowski and Heald, 1978; Kamm and Heald, 1983). These workers indicated that post-depositional compaction attending pressure solution is among the most important processes involved in primary porosity reduction in these sands. Quartz cement is abundant throughout these sandstones and stylolite development is common. Kaolinite and other clay minerals, which may be the result of feldspar decomposition, occur in localized patches. Authigenic pyrite is present and occurs most extensively within shale beds; it is also localized along stylolitic seams within sandstones.

Depositional Environment

The principle tectonic feature influencing sedimentation in the Appalachian Basin during late Chesterian time was an elongate highland, situated to the east of the study

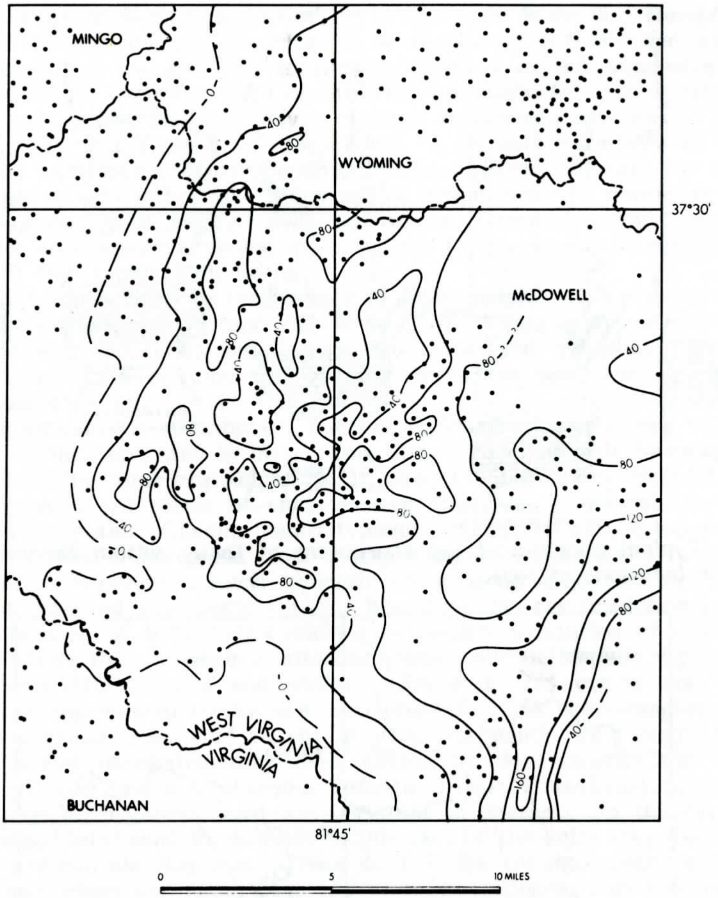

Figure 6. Net sandstone isolith map of the Ravencliff Member. Shows general north-south orientation of sandstone thicks and abrupt northwest-southeast termination of sand deposition. Black dots denote subsurface control points. The contour interval is 40 feet.

area (DeWitt and McGrew, 1979). Fluvial-deltaic deposition is thought to have dominated a broad coastal plain which separated the highlands in the east from marine environments to the west (Donaldson and Shumaker, 1978). Throughout late Chesterian time the rate of sedimentation in the West Virginia part of the Appalachian Basin exceeded basin subsidence, resulting in an infilling of the geosyncline and progradation onto adjacent shelf areas (DeWitt and McGrew, 1979).

The Ravencliff Member of McDowell County, is interpreted as representing a major, southward prograding, deltaic complex. Analysis of subsurface data, principally geophysical and driller's logs permits paleoenvironmental reconstruction of this deltaic complex, and delineation of three distinct environments of deposition within it. The paleoenvironmental interpretations presented in this report are in general agreement with the regional interpretations of Donaldson and Shumaker (1978), and are supported by the findings of numerous recent workers in and near the study area (Englund and DeLaney, 1966; Ellsworth, 1976; Whisonant and Scolaro, 1979; DeWitt and McGrew, 1979; Ettensohn and Peppers, 1979; Ettensohn, 1980).

Data collected from approximately 220 geophysical logs served as the basis for the Net Sandstone Isolith Map of the Ravencliff Member presented in this report (figure 6). The total sand thickness encountered in each well was estimated by establishing a "sand cut-off line" on each gamma ray log. The cut-off line was determined by aligning a straight edge parallel to the API units scale, at a position passing through

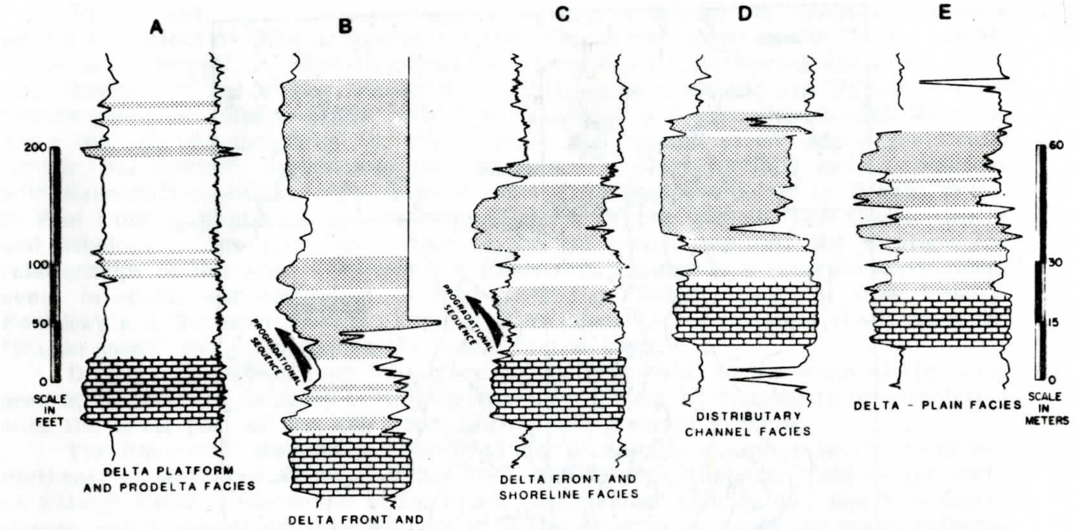

Figure 7. Typical geophysical log characters of facies within Ravencliff delta; see figure 8 for locations of logs.

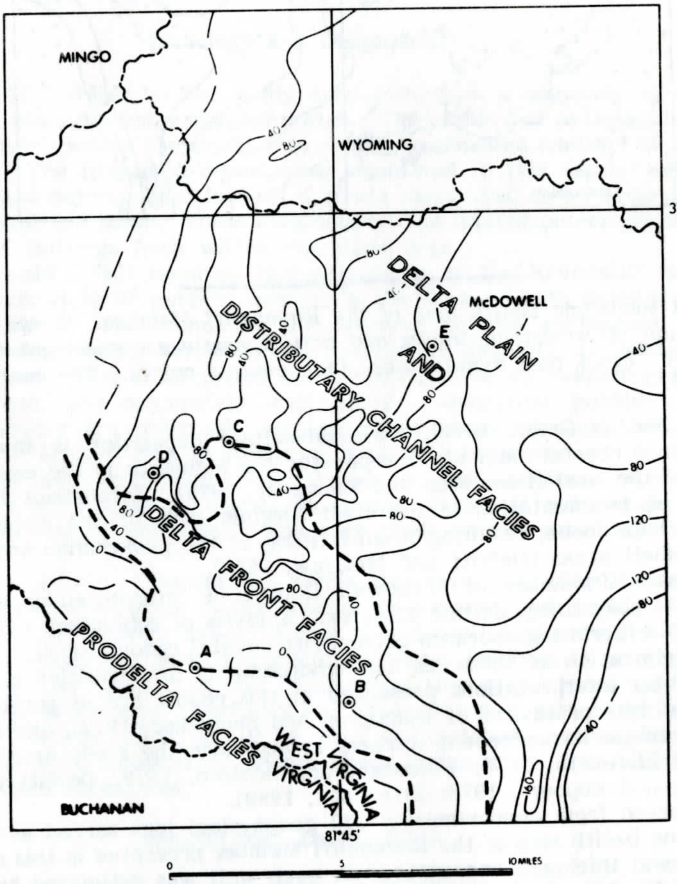

Figure 8. Net sandstone isolith map showing approximate facies boundaries. *Letters (A–E) refer to geophysical logs illustrated in figure 7. The contour interval is 40 feet.*

the point of maximum gamma ray deflection in the Little Stone Gap Member. This line was projected upward, through the interval of the Ravencliff Member, and the total thickness of sand was estimated by totaling the footage of the units whose deflection exceeded this cut-off (figure 3). Allowance was then made for thin interbeds of shale which may have been included. This method is considered superior to relying on a specific API unit cut-off, as it is independent of possible variations in logging tool calibration and sensitivity. Environmental interpretations necessary for mapping and facies boundary determinations are the result of geophysical log character analysis, as described by Pirson (1970), Selley (1978), and others. Gamma ray logs were relied upon extensively for environmental determinations, due to the dearth of self-potential logs in this part of the Appalachian Basin.

The three essentially contemporaneous environments of Ravencliff deposition, which have been delineated by this study are: 1) delta-plain and distributary channel; 2) delta front and shoreline; and 3) delta platform and prodelta. Each of these lithofacies represents a distinct environment within the southward trending progradational deltaic complex.

The delta-plain and distributary channel facies is characterized by north-south trending sand thicks, separated by intervening thins. Total sand thicknesses range from about 20 feet in the inter-channel areas, to over 100 feet along channel axes. This facies was probably the result of anastomosing distributary channels crossing a low gradient subaerial plain. Channel sands typically exhibit an abrupt basal contact and fining upward cycles, which grade from coarse (conglomeratic?) sand to silt (figures 7 and 8). Thick, uninterrupted sands typically occur along major channel axes. These may represent repeated incomplete cycles. Where gamma ray resolution is insufficient to allow delineation of individual cycles, bulk density indication of repeated porous zones may allude to their presence. Interdistributary and delta-plain deposits generally occur as interstratified sands and shales. Fining-upward cycles may be present, however, coarsening-upward cycles are also present. The coarsening-upward cycles, which probably grade from mud to silt or very fine sand, are thought to represent crevasse splays and interdistributary lakes, similar to those described by McBride and others (1975, p. 499) in the Cretaceous Difunta Group of northeastern Mexico.

Sand bodies with general east-west orientation characterize the delta front and shoreline facies. Total sand thicknesses in this part of the delta may exceed 100 feet, but typically are considerably less. These sand bodies are interpreted as representing distributary mouth-bars and sublittoral and shoreline deposits. Sands deposited in this facies are characterized by gradational lower contacts and upward coarsening cycles, grading from mud to fine- to medium-grained sand. In the Ravencliff delta these coarsening-upward cycles appear to have often been obscured by reworking of the delta front sediments by distributary channels, as progradation progressed. Down-dip these sands are an indication of maximum wave base during delta progradation.

The delta platform and prodelta facies is characterized by thick shale sequences with occasional thin sand or silt interbeds. These interbeds seldom exceed 1 to 2 feet in thickness and are interpreted as the result of hyperpycnal underflows, which originated along the delta front. Upward-fining is apparent in some of these sand units, but most exhibit a uniform grain size character, which is thought to be the result of intense bioturbation.

GEOLOGICAL HISTORY AND PALEOGEOGRAPHY

The sandstones which underlie the Little Stone Gap Member in the western part of the study area, are considered to be correlative with the "Maxon sands" farther to the west (figure 5; logs 6-8). These sandstones indicate that late Chesterian deltaic sedimentation was initiated in western McDowell County before Little Stone Gap time. This period of delta construction will be referred to as the "Maxon delta", for the purpose of differentiating it from later delta construction phases. During the initial period of "Maxon delta" construction, the depo-center appears to have been located, in western McDowell County (figure 9a). Prodelta muds which overlie the initial "Maxon sands" and underlie the Little Stone Gap Member in the western part of the study area (figure 5; logs 6-8), are evidence of westward progradation (figure 9b). As the depo-center of the "Maxon delta" shifted westward, an extensive carbonate bank developed

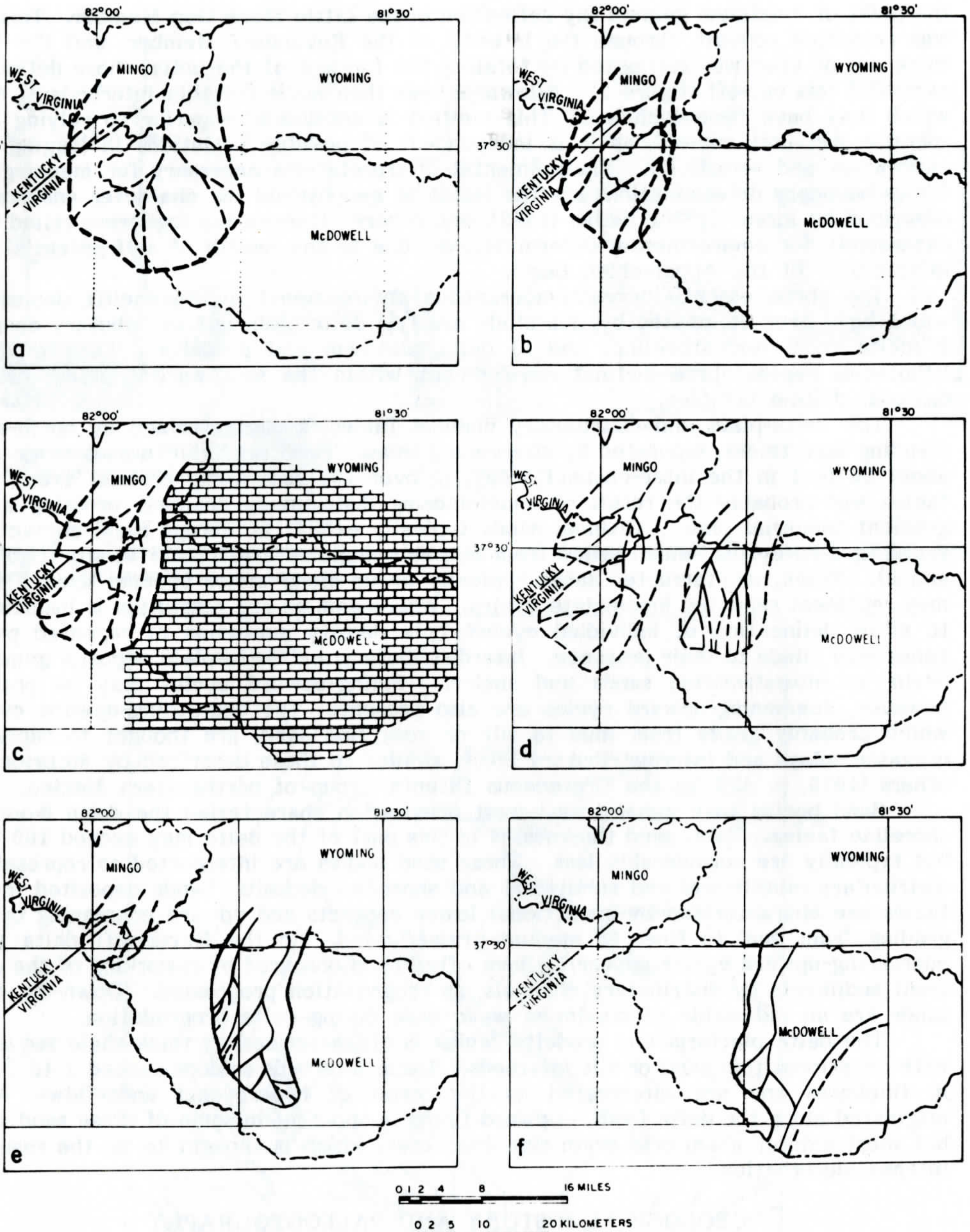

Figure 9. Paleogeographic reconstructions: a) possible maximum extent of initial "Maxon delta"; b) westward progradation of "Maxon delta"; c) continued westward progradation of "Maxon delta" and development of Little Stone Gap carbonate bank; d) initial development of Ravencliff delta; e) southeasterly progradation of Ravencliff delta and destruction of "Maxon delta"; f) final eastward shift in Ravencliff depocenter and destruction of earliest Ravencliff lobe.

to the east. This period of carbonate deposition, which is represented by the Little Stone Gap Member, was probably the result of a decrease in clastic input from the delta to the west. The Little Stone Gap carbonates were eventually deposited over the most easterly part of the "Maxon delta" (figure 9c).

The Little Stone Gap carbonate bank was initially inundated by Ravencliff sediments in west central McDowell County, where channel sands directly overlie, or erode into, the subjacent limestone (figure 4). To the east, in central McDowell and western Mercer Counties, the initiation of Ravencliff sedimentation is represented by dark gray shales, which directly overlie the Little Stone Gap Member (figure 5). These shales represent prodelta muds. Whether the initiation of Ravencliff sedimentation was the result of avulsion of the river system which was feeding the "Maxon delta", or was the result of a shifting in some other river system is unknown. The interfingering of the Ravencliff Member with "Maxon sands" in western McDowell County, is evidence that both of these deltaic complexes coexisted, at least during the early stages of Ravencliff sedimentation (figure 9d).

Sometime after the onset of Ravencliff sedimentation, the "Maxon delta" was abandoned and the depo-center of the Ravencliff delta shifted progressively eastward. This is evidenced by an increasingly thick shale interval between the Ravencliff channel sands and the subjacent Little Stone Gap Member (figure 5; logs 11-15), in eastern McDowell County (figures 9e and 9f).

The laterally extensive "Hot Shale" marker bed, which denotes the end of Ravencliff deposition, is considered to be non-deltaic in origin. Its widespread occurrence in southern West Virginia suggests that it may represent a period of localized marine transgression.

CONCLUSIONS

Three conclusions may be drawn from this study of the Ravencliff Member in southern West Virginia.

(1) The Ravencliff Member in McDowell and adjoining areas of Wyoming and Mingo Counties, West Virginia, is of fluvial-deltaic origin and can be differentiated in the subsurface from other similar, but distinct depositional events.

(2) Facies relationships within the Ravencliff delta can be documented through the use of subsurface maps and geophysical log character.

(3) Delta progradation proceeded in a south-southwesterly direction, with progressive shifting of lobe construction to the east-southeast.

ACKNOWLEDGEMENTS

I would like to thank Cities Services Oil and Gas Corporation for allowing me to publish this study. I am also indebted to Jon Huffman and Mujeeb Cheema for their assistance and encouragement, to Carla Heinrich and Ava Waggoner for drafting the figures, and to Julie Simmons for typing the manuscript. The critical review of the manuscript by W. D. Rizer and John M. Dennison is much appreciated.

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GEOCHEMISTRY AND PETROGENESIS OF TRONDHJEMITE DIKES, BLUE RIDGE, NORTH CAROLINA-GEORGIA

LAURA F. WOOD* *Department of Geology, Vanderbilt University, Nashville, Tennessee 37235*

CALVIN F. MILLER *Department of Geology, Vanderbilt University, Nashville, Tennessee 37235*

ABSTRACT

Trondhjemite dikes are common in the southeastern Blue Ridge of Georgia and North Carolina. These fine-grained dikes are characterized by the presence of abundant plagioclase, some occurring as euhedral, oscillatory-zoned phenocrysts, quartz, and minor biotite, K-feldspar, and muscovite. They are mineralogically and chemically similar to the Whiteside pluton (Miller and Kish, 1980), but textural features indicate that, unlike the Whiteside, they are post-metamorphic and generally post-tectonic. The dikes are characterized chemically by high SiO2 ($\gtrsim$ 70 wt%) and low K20 (1-2 wt%), low mafics (FeO+MgO $\lesssim$ 2 wt%) but relatively high Mg/(Fe+Mg) (~ 0.5 mol), high Sr (500-1000 ppm), very low Rb (10-40 ppm), and rather low Ba (300-600 ppm). REE patterns show low total REE, extreme depletion in HREE (La 20 x chondrite, Lu ~ chondritic), and a small positive Eu anomaly. Initial 87Sr/86Sr ratios are $\lesssim$ 0.704. Chemical data constrain source material of the trondhjemites as follows: (1) high Sr and absence of a negative Eu anomaly indicate a feldspar-poor source; (2) low Rb and 87Sr/86Sr suggest a mafic to ultramafic source, but (3) high SiO2 seemingly eliminates ultramafic material; (4) strong depletion in HREE indicates residual garnet and/or hornblende; either of these minerals could also account for the positive Eu anomaly. The best candidates for source material for the trondhjemite dikes are quartz eclogite and plagioclase-poor amphibolite. Using published partition coefficients and known LIL concentrations for typical basalt compositions, our melting models suggest that eclogite of low K tholeiitic composition is the most reasonable source for the trondhjemites. Generation of these peraluminous dikes from a mafic, metaluminous source requires a melting reaction such as (Mg, Fe, Ca)gt → cpx + liquid. Interpretation of plagioclase-quartz textural relations in light of experimental phase equilibria suggests that crystallization commenced at a minimum depth of about 30-40 km.

INTRODUCTION

Trondhjemites are defined as leucotonalites with sodic plagioclase (An < 30) (cf. Barker, 1979 and Streckeisen, 1976). Biotite and, less commonly, hornblende, are the typical mafic phases. In terms of major elements, they are characterized by high SiO2 and low FeO*+MgO, K_2O, and K/Na ratio.

Trondhjemites occur in various igneous suites of all ages. They are significant constituents of Precambrian gneiss and greenstone terranes and are found along continental margins ranging in age from the Proterozoic to Cenozoic. They have been described in sub-volcanic calc-alkaline suites of island arcs and as members of ophiolite complexes (Barker, 1979).

Small trondhjemite dikes are widespread in the Blue Ridge. Almost all are located within the Hayesville-Fries thrust sheet (Hatcher, 1978), a slab of early Paleozoic metaclastics and small mafic-ultramafic bodies which is thought to have been thrust over a Grenville-age basement during the Ordovician (Hatcher, 1978). Most of the trondhjemite dikes have orientations which parallel the structural trend of the Blue Ridge, though some are perpendicular to the regional trend (Yurkovich, 1982); the dikes may either cross-cut or parallel the local foliation. These small intrusions are

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*Present address: Department of Geology, University of North Carolina, Chapel Hill, North Carolina 27514.

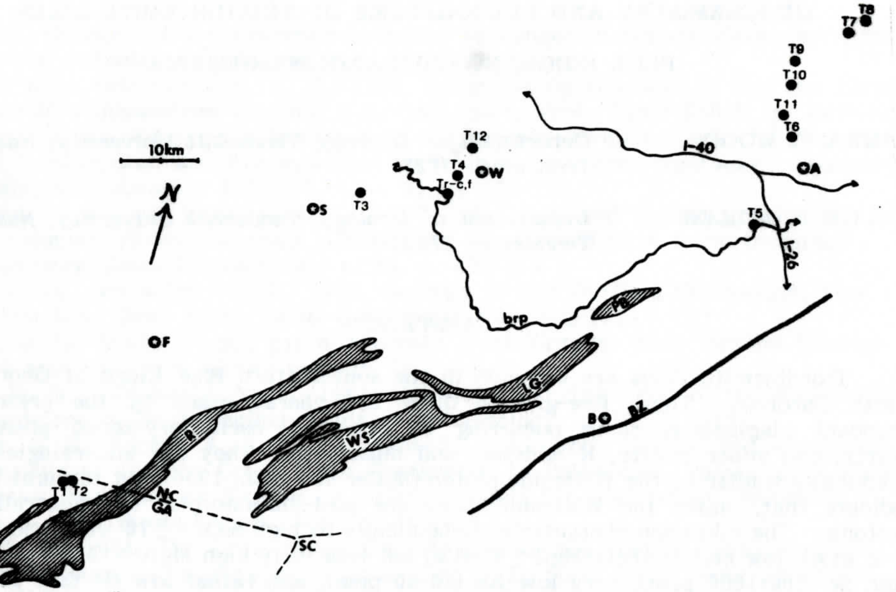

Figure 1. Map showing collection localities (indicated by dots and sample numbers) of trondhjemite dikes in northern Georgia and North Carolina. S = Sylva; W = Waynesville; F = Franklin; A = Asheville; brp = Blue Ridge Parkway; BZ = Brevard Zone; WS = Whiteside Pluton; PB = Pink Beds Pluton; R = Rabun Pluton; LG = Looking Glass Pluton.

associated with larger mid-Paleozoic trondhjemite-granodiorite plutons which occur in a belt extending from central Alabama through central Virginia (Greenberg and Kish, 1976; Miller and others, 1983a).

This paper presents a study of thirteen dikes located in northern Georgia and North Carolina (Figure 1)[1]. These dikes were petrographically and chemically analyzed in order to constrain their origin and arrive at a possible petrogenetic model.

METHODS

Major and trace element abundances in 16 samples were determined by standard X-ray fluorescence methods by the authors at Vanderbilt University and by X-ray Assay Laboratories. A single REE pattern was obtained by neutron activation analysis by Allen Glazner at UCLA. Phenocryst compositions were determined by microprobe at UCLA for two samples.

PETROGRAPHY

The trondhjemite dikes are fine-grained, porphyritic and weakly or non-foliated. The thirteen dikes sampled in this study have maximum widths of 2 meters. Intrusive contacts with the gneissic country rock are sharp and generally cross-cutting. The fabric of the trondhjemite dikes suggests that they were emplaced post-tectonically and post-metamorphically as there is little intergranular shearing, foliation is weak or absent, and delicate igneous textures are preserved (Figure 2).

Euhedral (0.5-2mm) phenocrysts of plagioclase are common; many are oscillatorily zoned (Figure 2) and a number of them contain quartz inclusions (Figure 3). Quartz also occurs as sparse phenocrysts in a few samples. Biotite is ubiquitous and muscovite common as phenocryst phases (see Figure 2). In addition to plagioclase, quartz, and micas, the groundmass and microphenocryst assemblage includes K-feldspar and small amounts of the accessories sphene, epidote, apatite, opaque minerals, and zircon. The groundmass, with grain sizes ranging from .05-.3 mm, typically comprises at least 80% of the rocks. Plagioclase phenocrysts are oligoclase (An 13-28 in samples analyzed by

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[1] Exact localities are available, upon request, from the authors.

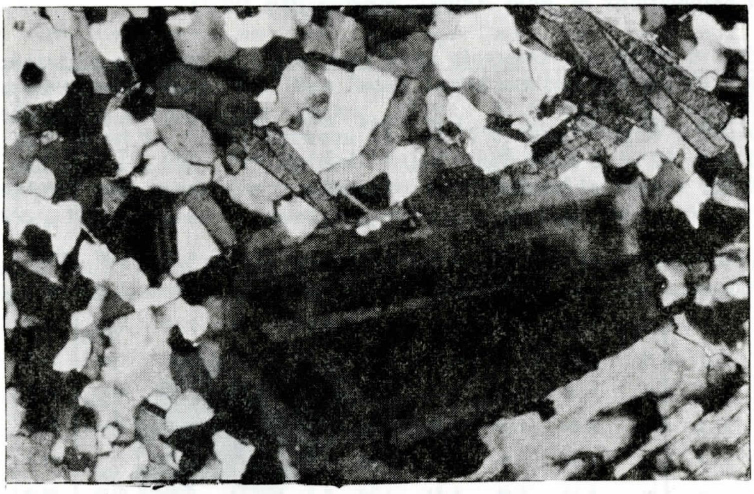

Figure 2. Photomicrograph showing phenocrysts of euhedral oscillatory-zoned plagioclase and primary biotite and muscovite. Field of view approx. 2 mm.

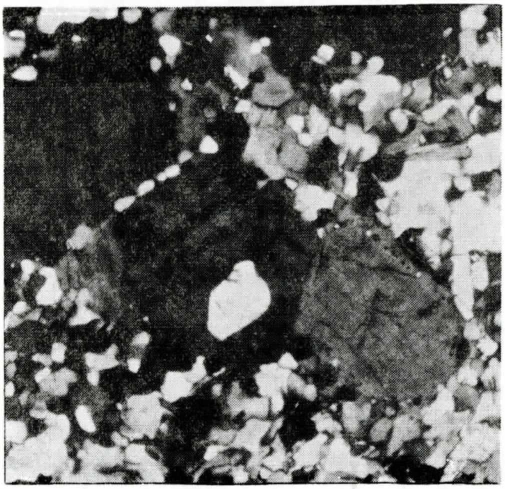

Figure 3. Photomicrograph showing subhedral quartz inclusion in plagioclase phenocryst; field of view approx. 3 mm.

microprobe). Oscillatory zoning, though obvious optically, is weak and apparently reverse overall (for example, core = An 22, rim = An 27). Biotite is relatively Al-rich (~ 1.45 Al/11 oxygen anhydrous formula), but it is somewhat poorer in Al than most biotites from muscovite-bearing rocks (cf. Miller and others, 1981). The ratio Fe/(Mg + Fe, mol) is rather low for such felsic rocks (~ 0.5-0.6, FeO ~ 22 wt%, MgO ~ 10%). Muscovite is celadonitic (Fe_2O_3 + MgO + TiO_2 >4 wt.%, Al < 2.6 per 11 oxygen anhydrous formula), as is common in primary igneous muscovites (Miller and others, 1981). Epidote, with 11-14 wt% Fe_2O_3, can be represented almost entirely as an epidote (~ 75%)--clinozoisite (~ 25%) solid solution.

Plagioclase and quartz are the most abundant leucocratic minerals (60-70% and 25-30% respectively)[2] while muscovite and K-spar are minor (< 10% total). Biotite (< 10%) is the only mafic silicate phase. Modes fall within the trondhjemite field of Streckeisen (1976).

[2] Modes were estimated from inspection of thin sections and stained slabs.

Table 1. Compositions of Blue Ridge Trondhjemite Dikes.

sample no.	SiO_2	Al_2O_3	Fe_2O_3*	MgO	CaO	Na_2O	K_2O	MnO	TiO_2	P_2O_5	L.O.I.	Total	Rb	Sr	Ba
T1	71.7	15.4	1.09	0.42	2.00	5.64	1.72	0.01	0.11	0.03	1.00	99.3	35	600	590
T2	72.2	15.9	1.09	0.40	2.12	6.50	0.65	0.01	0.10	0.06	0.93	100.1	20	662	440
T3	72.0	16.3	0.91	0.57	2.84	5.61	0.97	0.01	0.14	0.05	0.62	100.2	20	685	320
T4a	69.2	17.5	1.17	0.59	3.11	6.37	1.11	0.01	0.15	0.06	0.16	99.6	10	1147	340
T6b	72.5	15.5	1.57	0.49	2.09	5.29	1.20	0.01	0.13	0.05	0.62	99.6	10	747	400
T6c	75.2	14.9	0.90	0.32	1.72	5.06	1.18	0.00	0.06	0.03	0.62	100.2	32	558	380
T7	69.8	17.6	0.95	0.56	2.86	6.02	1.32	0.01	0.14	0.03	0.85	100.4	24	751	490
T8	71.6	16.1	1.45	0.56	2.49	4.93	1.50	0.01	0.14	0.04	0.62	99.7	30	693	390
T9a	70.3	16.7	0.76	0.28	1.10	6.11	3.21	0.00	0.06	0.03	1.00	99.7	70	385	930
T10e	71.8	15.6	1.30	0.49	2.01	6.01	1.88	0.01	0.12	0.03	0.47	99.9	25	756	380
T10f	67.9	17.5	1.60	0.84	2.99	6.29	1.40	0.02	0.27	0.08	0.47	99.5	28	1045	340
T10x	68.5	17.4	1.75	0.57	2.96	6.26	1.43	0.02	0.27	0.06	----	99.3	27	974	310
T11	72.8	15.4	2.06	0.43	2.06	5.83	1.43	0.01	0.13	0.04	0.39	99.9	22	565	410
T12	71.9	16.1	0.46	0.36	2.34	5.57	1.08	0.00	0.07	0.04	1.16	99.3	20	708	290
Tr-c	71.5	16.3	1.36	0.40	2.56	5.52	0.95	0.02	0.17	0.05	----	98.8	17	549	420
Tr-f	72.9	15.1	1.10	0.46	2.32	5.58	1.02	0.01	0.13	0.05	0.31	99.1	17	486	410
T-en	70.7	16.3	1.78	0.49	2.65	5.28	1.60	0.01	0.19	0.02	----	99.1	27	815	429

*total Fe as Fe_2O_3

L.O.I.= Ignition Loss

(Major element oxides as weight %; Rb, Sr and Ba as ppm).

REE of Tr-c = La:8.1, Ce:16.9, Sm:0.74, Eu:0.26, Tb:0.11, Lu:0.026, ppm

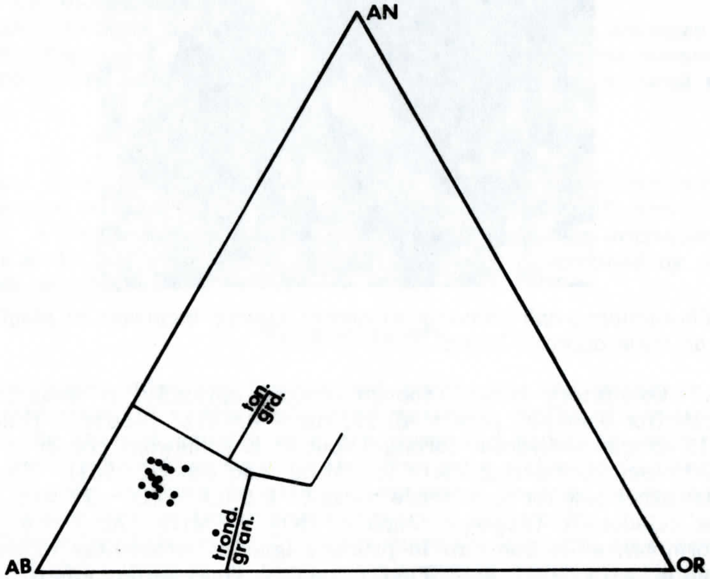

Figure 4. Qz-Ab-Or ternary showing trondhjemite (trond.), granodiorite (grd.), granite (gran.), and tonalite (ton.) fields as defined by Barker (1979; modified from O'Connor, 1965). Dikes analyzed in this study (dots) fall within the trondhjemite field.

CHEMICAL DATA

Representative analyses of major and trace elements are shown in Table 1. The elemental oxide abundances of the Blue Ridge trondhjemites fall within the limits of typical trondhjemite compositions (Barker, 1979). SiO2 concentrations fall mostly between 70 and 72 wt.%; Al_2O_3 between 15 and 17%; (FeO*+MgO) 1-2%; CaO 2-3%; Na20 5-6%, and K20 1-1.5%. Fe*/Mg(mol) ratios range from 1 to 2. The Blue Ridge trondhjemites plot well within the trondhjemite field defined by Barker (1979; modified from O'Connor, 1965), which is based on normative proportions of An, Ab, and Or (Figure 4).

The trondhjemites are low in Rb (~ 10-40 ppm), high in Sr (500-1000 ppm), and have high K/Rb (400-500) and low Rb/Sr (0.01-0.05). Barium concentration is low (300-600 ppm). Rare earth elements (REE) are very low, with heavy REE being extremely depleted (~ chondritic). The concentrations of the large-ion lithophile elements (LIL) K, Rb, Sr, Ba, and the REE are shown in Figure 5 relative to Kay's (1980) normalization factors.

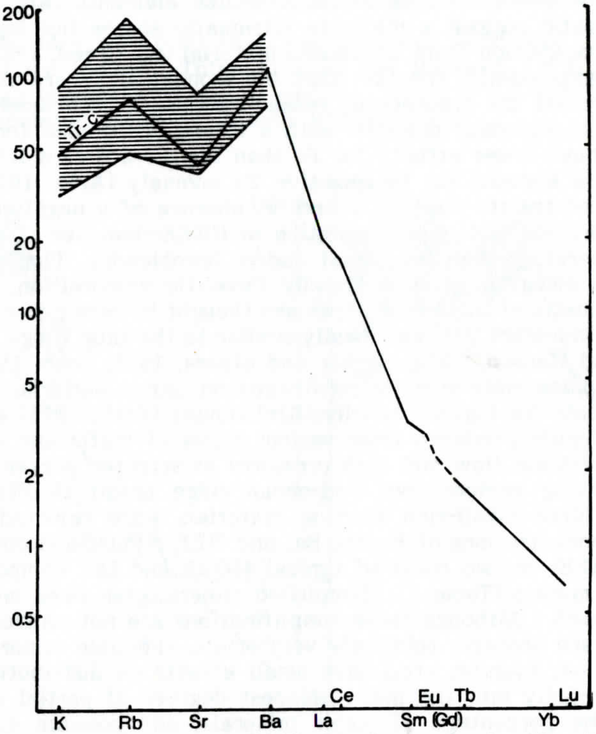

Figure 5. Field of LIL (K, Rb, Sr and REE) abundances in the Blue Ridge trondhjemite dikes relative to Kay's (1980) normalization factors (chondritic for Ba, Sr, and REE; Rb and K are 0.2 x MORB). One sample with a composition affected by crustal fractionation (high K, Rb, Ba, low Sr) omitted. Pattern of sample Tr-c illustrated. REE patterns for dikes reported by Sando and others (1983) and Miller and others (1983a, b) are very similar, though not quite so HREE-depleted (LREE 8-22 x chondrite, HREE 1.5-3).

Kish and others (1975) report a low initial 87Sr/86Sr for one trondhjemite dike of 0.7039. Sando and others (1983) and Miller and others (1983a, b) have obtained initial 87Sr/86Sr values ranging from 0.7034 to .7043 and one slightly positive ϵ_{Nd} value.

PETROGENESIS

Textural and chemical features indicate that the compositions of the trondhjemites are close to those of liquids and that they have not been affected strongly by fractionation taking place at crustal pressures. The low percentage of phenocrysts indicates that they are not cumulates and that liquid was by far the dominant phase. Furthermore, if fractionation had occurred, the fractionating assemblage would have been plagioclase-rich since plagioclase is the dominant phenocryst. However, because plagioclase has a strong affinity for Eu and Sr, extensive fractionation of plagioclase induces a negative Eu anomaly and depletes Sr in residual liquids; the small positive Eu anomaly and high Sr in these rocks argue strongly against such a process. We thus take the rock compositions to be reasonable approximations of liquid compositions.

The very low concentrations of incompatible elements (Rb, REE, Ba) and low initial 87Sr/86Sr ratio suggest a mafic to ultramafic source but high SiO2 seemingly eliminates direct derivation from ultramafic material (cf. Green, 1980, but see Kushiro, 1972, for an opposing view). The fact that the trondhjemites are strongly depleted in HREE further suggests the presence of residual garnet or hornblende because they are the only potentially abundant minerals with a strong preference for HREE over light REE; since they have lower affinity for Eu than for the adjacent REE, either of these minerals could also account for the positive Eu anomaly (Arth, 1976).

The above constraints--high SiO2 and Sr, absence of a negative Eu anomaly, low Rb and initial 87Sr/86Sr and strong depletion in HREE--indicate a feldspar-poor mafic source whose mineralogy includes garnet and/or hornblende. Plagioclase-poor amphibolite and quartz eclogite, given a broadly tholeiitic composition, satisfy the above criteria. Partial melts of these rock types are thought to have generated trondhjemites in other terranes which are compositionally similar to the Blue Ridge trondhjemites (for example, Arth and Hanson, 1972; Barker and others, 1976; Arth, 1979; Barker, 1979).

In order to place more specific constraints on source material, we used published partition coefficients for high-silica (rhyolitic) liquids (Arth, 1976) and calculated the compositions of liquids produced from various types of mafic and ultramafic sources under varying conditions (low and high pressure) at selected percentages of melting.[3] We chose as starting compositions mid-ocean ridge basalt (MORB) and island arc tholeiite (IAT). More alkali-rich starting materials were rejected because of their unsuitably high concentrations of K, Rb, Ba, and REE. Published concentrations of K, Rb, Sr, Ba and REEs in two rocks of typical MORB and IAT composition (Kay, 1980) were used in our models (Table 2). Simplified mineralogies were modelled and modal melting was assumed. Although these simplifications are not entirely realistic, errors inherent in them are probably relatively very small. Possible minor minerals (quartz, spinel, K-spar, mica, kyanite, etc.) have small effects on distribution coefficients or would disappear rapidly into the melt. Modest degrees of partial melting would not affect strongly the percentage of major minerals, and possible variability in major mineral modes is limited.

Simplified amphibolites (70% hornblende + 30% plagioclase) and eclogites (55% clinopyroxene + 45% garnet) were modelled for fractions of melting at intervals from 0-100%.[4] The calculated normalized concentrations of LILs in the hypothetical melts

[3] We used the modal melting equation from Hanson (1978):

$$C_L = C_o/[D_o + F(1-D_o)]$$

where C_o = wt. concn. of trace el. in parent,
C_L = wt. concn. of trace el. in melt,
D_o = bulk distribution coefficient of trace element,
F = weight fraction of melt relative to original solid.

[4] Peridotites and gabbros were also modelled as sources. Matches with trondhjemite compositions were generally entirely unsatisfactory as suggested in the constraints above. The only possible match was for chondritic to slightly enriched garnet periodotite with very small (< 5%) fractions of melting; we consider this an implausible source material on the basis of trondhjemite SiO_2 (> 70 wt.%), MgO (< 1%), and Fe/Mg ($\geq$ 1), all seemingly incompatible with direct derivation from an olivine-rich source.

Table 2. Normalized LIL Concentrations in Model Source Materials.

	MORB[a]	IAT[b]	Normalizing Values (ppm)[c]
Ce	10.0	3.8	0.98
Sm	12.6	5.2	0.23
Eu*[d]	12.2	5.3	—
Gd*[d]	12.0	5.3	—
Yb	10.4	6.1	0.25
K	5	23.1	176
Rb	5	20.9	0.22
Sr	10	14.6	14.0
Ba	4.8	28	3.8

a Ocean ridge basalt D1 (Kay and Hubbard, 1978; Kay, 1980)
b Island arc tholeiite Tofua 17 (Kay and Hubbard, 1978; Kay, 1980)
c Leedey chondrite, except K and Rb, which = 0.2 x MORB (Kay, 1980)
d Extrapolated assuming smooth REE patterns

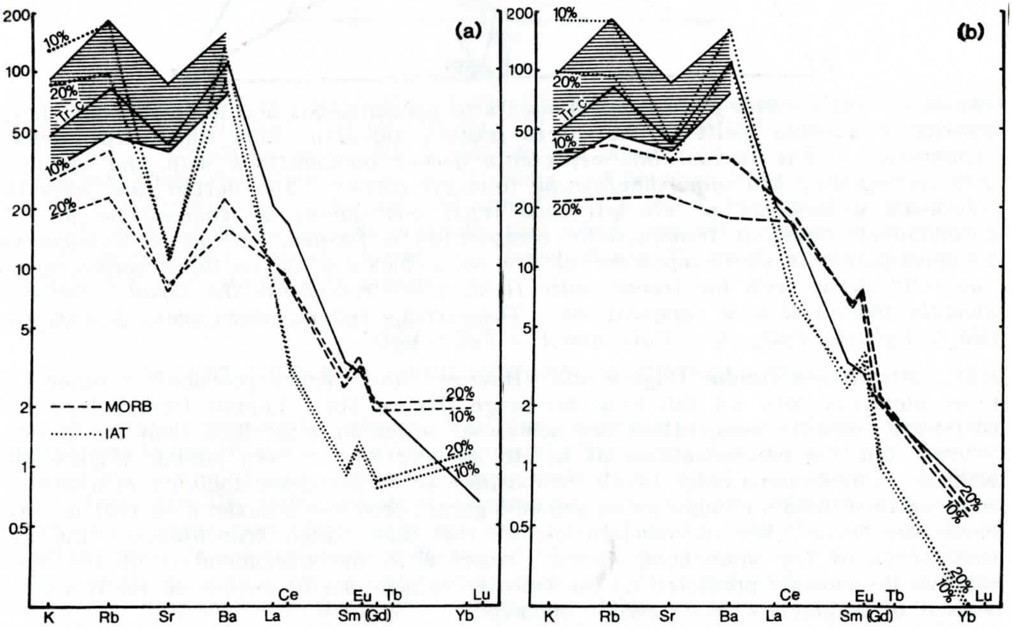

Figure 6. Calculated concentrations (normalized as in Figure 5) of LIL elements in hypothetical melts of mid-ocean ridge basalt (MORB) and island arc tholeiite (IAT) produced after 10% and 20% melting of the original material. (a) amphibolite residue (70% hbl, 30% plag). (b) eclogite residue (55% cpx, 45% garnet).

at fractions = 10 and 20%, the most reasonable values, are plotted in Figure 6.

An amphibolite of mid-ocean ridge basalt composition yields a liquid that is lower in K, Rb, Sr and Ba than the trondhjemites (Figure 6a). The REE pattern of this hypothetical liquid is more depleted in the middle REE and not as strongly depleted in HREE as that of the trondhjemites (Figure 6a). An amphibolite of island arc tholeiite composition produces a liquid with K, Rb, and Ba concentrations similar to the trondhjemites but much lower Sr (Figure 6a). The REE concentrations are all too low, and, as with the MORB, the shape of the pattern does not match that of the trondhjemites.

The concentrations of REE in liquids derived from eclogite of MORB composition are similar to concentrations in the trondhjemites and the shapes of model and observed

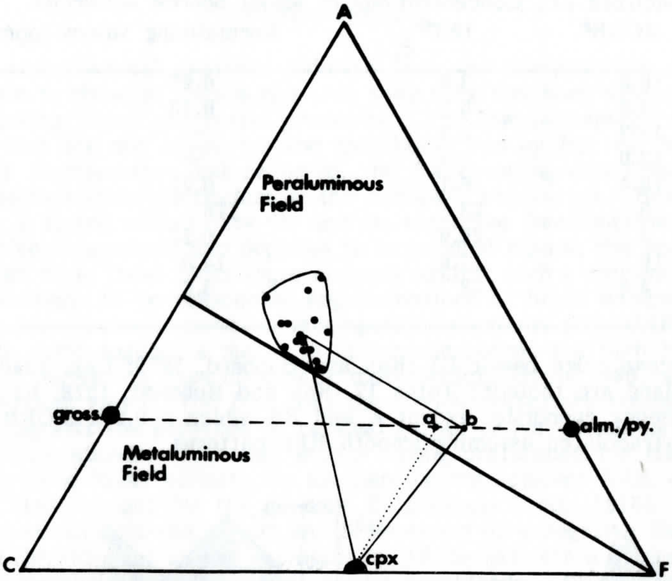

Figure 7. ACF ternary diagram, divided into peraluminous and metaluminous fields, showing a possible melting reaction to produce liquid of Blue Ridge trondhjemite composition. The dashed line represents garnet compositions with end members grossular (gross.) and almandine/pyrope (alm/py) garnet. The dotted line joins the subsolidus assemblage of clinopyroxene (cpx) and garnet of composition a. As peraluminous liquid of trondhjemitic composition is formed, the garnet composition becomes poorer in Ca (composition b). The solid lines are the tie lines connecting the two solid phases with the trondhjemite field, which represents the liquid. The dots indicate individual dike compositions. The ternary end members are: $A = Al_2O_3-(Na_2O+K_2O)+5/3P_2O_5$; $C = CaO$; *and* $F = FeO^*+MgO$.

REE patterns are similar (Figure 6b). However, the pattern produced in other LIL trace elements does not fall into this range (Figure 6b). Liquids from eclogite of island-arc tholeiite composition are somewhat lower in total REE than the trondhjemites, but the concentrations of K, Rb, Sr and Ba are very similar (Figure 6b). Whether of mid-ocean ridge basalt composition or of island-arc tholeiite composition, an eclogite with 55% clinopyroxene and 45% garnet produces a model melt that is much closer in trace element composition to the Blue Ridge trondhjemites than an amphibolite of the same composition. Absence in the trondhjemites of the large positive Eu anomaly predicted by the eclogite models can be accounted for if a small amount of plagioclase fractionation occurred.

With eclogite of either island-arc tholeiite or mid-ocean ridge basalt composition as the probable source for these trondhjemites, there remains the problem of a metaluminous source generating a peraluminous liquid. A possible melting reaction which could yield peraluminous liquid is shown on the ternary ACF (Figure 7). On this ACF diagram, we plotted the approximate compositions of eclogitic garnet and clinopyroxene (the major solid phases in eclogite) and the Blue Ridge trondhjemites, which represent the supposed coexisting liquid phase. A possible eclogite melting reaction which could produce a peraluminous liquid is as follows: (Mg, Fe, Ca)gt → liq + cpx + Ca-poorer gt. In other words, the garnet/clinopyroxene ratio decreases as the liquid forms and preferentially concentrates Al from the garnet breakdown in the liquid.

The generation of peraluminous melts from quartz eclogite was demonstrated experimentally by Stern and Wyllie (1978), who used natural rock as starting material, and later by Sekine and others (1981), using synthetic compositions with relative amounts of clinopyroxene and garnet similar to those of the modelled eclogite of this study. The lower temperature partial melts produced in both experimental investigations were similar to the Blue Ridge trondhjemites with respect to relative

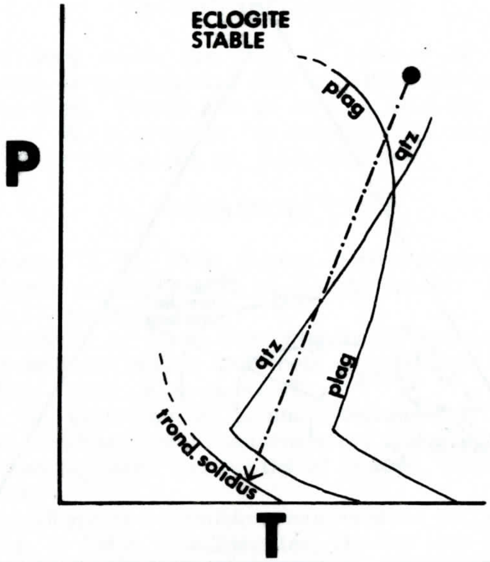

Figure 8. Schematic representation of the stability fields of quartz, plagioclase, and felsic liquid with limited H_2O (cf. Wyllie, 1977). Dashed-dotted line defines possible movement of trondhjemite magma to explain the occurrence of quartz inclusions in plagioclase phenocrysts (see Figure 3). At some high pressure and temperature, quartz was on the liquidus and plagioclase was absent. As P and T dropped, both phases co-precipitated until quartz became unstable; that which was not included in plagioclase grains reacted back into the liquid. At lower P and T, quartz reappeared on the liquidus and precipitated with plagioclase.

proportions of Al_2O_3, CaO and FeO + MgO.

ASCENT OF MAGMA AND HISTORY OF CRYSTALLIZATION

As stated previously, quartz inclusions commonly occur within plagioclase phenocrysts in the trondhjemite dikes (Figure 3). This phenomenon may have occurred because the ascending magma left the quartz stability field and entered the plagioclase field. These stability fields are shown schematically in Figure 8. When crystallization began, we suggest that plagioclase was unstable and quartz was the liquidus phase. When the pressure and temperature decreased, plagioclase joined quartz on the liquidus. Later, quartz became unstable and reacted back into the liquid, except for that which had become trapped inside the plagioclase grains. It did not begin to crystallize again until the magma approached its ultimate level of emplacement. Biotite was also on the liquidus at a very early stage, suggesting reasonably high P_{H_2O}.

Minimum depth of initial crystallization can be crudely constrained by plotting the trondhjemites in the Qz-An-Ab-Or-H2O system. Figure 9 is a two-dimensional representation of that system with An contours on the plagioclase-quartz cotectic shown approximately (Winkler, 1976).[5] However, with rising pressure the cotectic rises (An increases) and moves toward the albite corner. The trondhjemite field lies above the cotectic (is richer in An) when the system is at P_{H_2O}=7kb. Extrapolating from the rate of change with pressure at lower pressures, the trondhjemite field would be coincident with the cotectic at very roughly 10kb (cf. Winkler, 1976). Thus, liquid of Blue Ridge trondhjemite composition will crystallize plagioclase and quartz at $P_{H_2O} \sim$ 10kb, suggesting initiation of crystallization at great depth.

CONCLUSIONS

The trondhjemite dikes of this study represent post-tectonic intrusions which were emplaced as crystal-poor magmas whose phenocryst assemblage appears to record

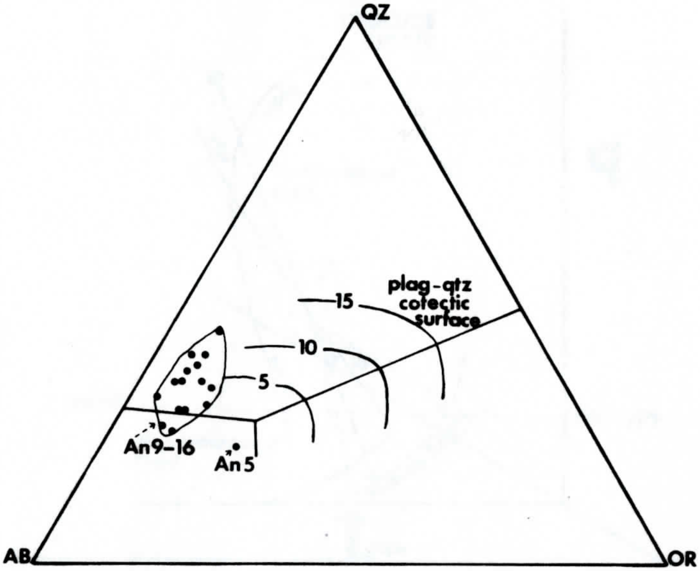

Figure 9. Two-dimensional representation of the system Qz-An-Ab-Or at P_{H_2O}=7kb with the plagioclase-quartz cotectic surface contoured at 5, 10 and 15% An (curved lines) (from Winkler, 1976). The Blue Ridge trondhjemite compositional field, which is outlined (dots represent individual dike compositions), lies above the 7kb cotectic and thus will intersect the cotectic only at somewhat higher pressures (roughly 10kb by extrapolation of Winkler's data). We suggest that the trondhjemite magmas commenced crystallization at or above this pressure, that is, at depths $\gtrsim$ 35km.

ascent from depths of at least 30-40 km. We propose eclogite, equivalent in composition to low K tholeiite, as the probable source. The data do not discriminate clearly among possible compositions of the tholeiitic source. With respect to REEs, mid-ocean ridge basalt serves slightly better than island-arc tholeiite, but K, Rb, Sr and Ba concentrations in model liquids from island-arc tholeiite are closer to those of the Blue Ridge trondhjemites.

Almost all of the trondhjemite dikes and associated low K plutons in the Blue Ridge of northern Georgia and North Carolina are located within the Hayesville-Fries thrust sheet, which was probably in place when the trondhjemitic magmas intruded. It is therefore puzzling that the data suggest a deep mafic crust beneath this portion of the Blue Ridge about 400 m.y. ago. Low K, Rb, and $^{87}Sr/^{86}Sr$ ratio place severe limitations on extent of contamination as the trondhjemitic magmas rose through the tectonically thickened continental crust before final emplacement, for significant crustal contamination would rapidly increase these values. Our data thus suggest that depleted mafic material was present at the base of, or beneath, the crust in the mid-Paleozoic, that it partially melted to yield trondhjemitic magmas, and that these magmas were able to ascend through the crust with little contamination.

Added note: Nd, O, Pb and additional Sr isotope data acquired since the completion of this study by T. W. Sando, G. C. Solomon, and P. D. Fullagar are consistent with the depleted mafic source postulated in this paper (cf. Miller and others, 1983b; Sando and others, 1983).

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[5] Although the trondhjemite magmas were probably not water-saturated at high P, changes in the Qz-An-Ab-Or phase boundaries will be similar at P_{H_2O} < P total, and thus conclusions drawn from the water-saturated diagrams should be at least qualitatively correct.

ACKNOWLEDGEMENTS

We appreciate very much the help of Leonard S. Wiener and Stephen P. Yurkovich, who showed us trondhjemite dike localities and provided useful comments and discussions in the field. Thanks also go to S. A. Kish for useful discussions of this project and to John J. W. Rogers for his helpful suggestions and comments. This project was supported by NSF grant no. EAR-8109055.

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FACTORS AFFECTING PEAT DEPOSITION IN A HOLOCENE RIVER-FLOODPLAIN SWAMP NEAR WILMINGTON, NORTH CAROLINA

KATHI R. KRONENFELD *Department of Geology, University of North Carolina,*
*Chapel Hill, North Carolina 27514**

ABSTRACT

A small peat deposit is located in the Holocene sediment on a peninsula between the Cape Fear and Northeast Cape Fear Rivers near Wilmington, North Carolina. Two distinct sedimentary environments can be recognized in the study area: 1) the river-edge environment, and 2) the swamp-forest environment. Sediment in the river-edge environment is dominantly silty clay deposited by the rivers, with little organic material. Vegetation is a reed-sedge assemblage, dominated by *Typha* (cattail), *Mariscus* (sawgrass), and *Taxodium* (cypress). The forest environment is found in the central portion of the peninsula. The sediment is composed almost entirely of organic material. A small amount of mineral sediment is supplied by a Pleistocene beach sand which underlies the study area. The vegetation is dominated by *Acer rubrum* (red maple), *Nyssa sylvatica* (black gum), and *Liquidambar styraciflua* (sweet gum).

Sediment supply is a major factor controlling peat formation in the study area. In the river-edge environment, organic material is diluted by river sediment, whereas the swamp-forest environment has a high organic content because the river sediment is lacking. The silty clay from the rivers acts as an aquiclude for groundwater flowing through the Pleistocene sand unit. This restriction in the water flow creates conditions favorable to the preservation of the organic material as peat.

Radiocarbon dating of a basal peat yielded an age of 2840 ± 50 years BP. The rate of Holocene sea level rise has probably controlled the rate of peat formation.

INTRODUCTION

Studies of the Holocene sediment of the Atlantic Coastal Plain have traditionally concentrated on the barrier island system and the major estuaries. In the recent years, a considerable amount of work has been done in the large pocosin swamps and Carolina bays (Whitehead, 1972; Staub and Cohen, 1979; and Ingram and Otte, 1980). In contrast, the river floodplains have received relatively little attention.

Small peat deposits are present in the river floodplains. Little detailed work has been done on these deposits, however, partially because their small size and relatively low organic content has made them economically unattractive. One such river floodplain deposit is examined in this study. The objectives of this research were:

1) Description of the Holocene sediments associated with the lower Cape Fear and Northeast Cape Fear Rivers near Wilmington, North Carolina.
2) Determination of the relationship between sediment distribution patterns and the occurrence of peat and organic sediments in the area.
3) Interpretation of the Holocene sedimentary history of the area, with particular emphasis on factors controlling the formation of peat and its distribution patterns.

GEOLOGIC AND GEOGRAPHIC SETTING

The study area is located in New Hanover and Brunswick counties, immediately west of Wilmington, North Carolina (Fig. 1). It includes the northern half of Eagle Island and the southern part of the peninsula between the Cape Fear and Northeast Cape Fear Rivers, and is detailed on the Wilmington and Castle Hayne 15' U.S. Geological Survey topographic maps.

**Current Address: Department of Geological Sciences, University of Southern California, University Park, Los Angeles, California 90089.*

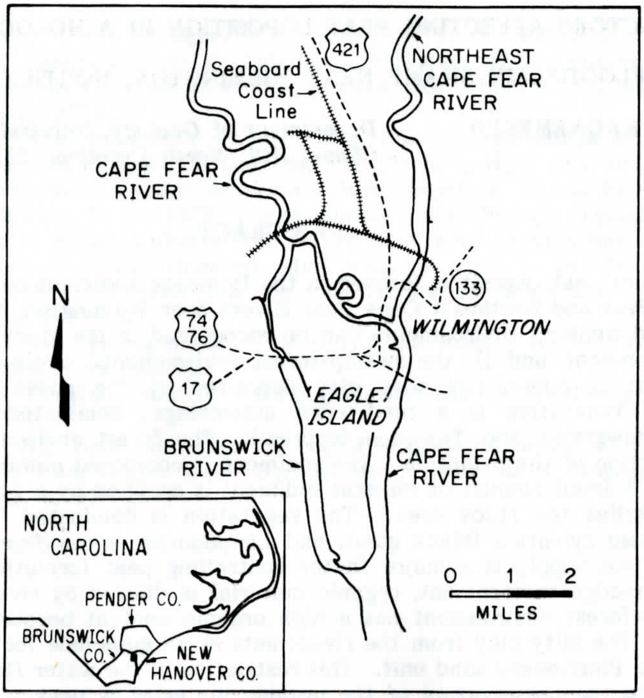

Figure 1. Location map of the Wilmington area, North Carolina.

The maximum elevation within the study area is slightly over 25 ft (7.6m), suggesting that it is located on the western edge of the Pamlico terrace. This is the youngest of three late Pleistocene marine terraces found on the Lower Coastal Plain of eastern North Carolina (Stuckey, 1965). The Suffold Scarp, which normally forms the western boundary of the Pamlico terrace, cannot be delineated in the study area.

A late Pleistocene beach sand is a prominent unit within the study area. It was deposited during the Peorian high stand of the sea, and was later reworked by the wind into dunes (Bain, 1970). The sediment is now composed of homogenous, well sorted, medium to fine sand, and it forms a broad ridge along the center of the peninsula.

When sea level dropped during the Wisconsin glacial stage the surficial sands were dissected by streams. Later, the Holocene sea-level rise led to the deposition of sediment in these dissected stream valleys. Within the study area, the surface of this Holocene sediment does not rise more than three to five feet above mean sea level.

The Cape Fear River is one of the major rivers draining the deeply weathered crystalline rocks of the North Carolina Piedmont. Its drainage basin extends westward beyond Greensboro, North Carolina. The Northeast Cape Fear River is much smaller, and is confined to the lower Coastal Plain. The Brunswick River is a former channel of the Cape Fear River that has been dredged and widened.

METHODS

Sample Collection

Samples were collected at 54 locations within the study area (Fig. 2). Nineteen sites were cored with a Davis peat sampler and the remaining 35 sites with a Macauley peat sampler. All sites were sampled at 1 ft vertical intervals, from the surface to the total depth penetrated. Neither type of corer could penetrate thick sand layers or cohesive clays. Each 1 ft sample was homogenized by hand, then placed in a water-tight plastic bag for transportation back to the laboratory. Data recorded at each site included depth to the water table, dominant vegetation, a physical description of each sample collected, and a description of the material underlying the deepest sample

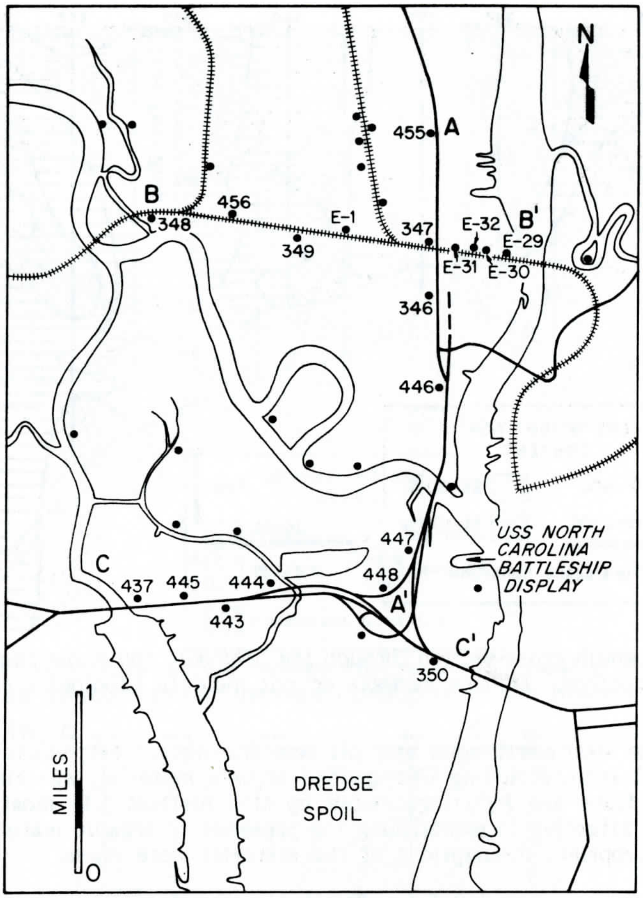

Figure 2. Map of the study area showing sample sites. Numbered sites appear in the cross-sections used in this study.

collected, when that could be determined.

Moisture and Ash Determinations

In the laboratory, all samples were tested to determine percentages of moisture and dry-weight ash, according to the following procedures: About 100 grams of each sample was weighed, dried to constant weight at 105°C, and reweighed to yield percent moisture. The dried samples were then combusted at 550°C until the organics were burned off; percent dry-weight ash was then calculated. The percent dry-weight ash is approximately the reciprocal of the organic content.

The following classification system based on the percent dry-weight ash content is used in this report:

1) peat	<14% dry-weight ash
2) mineral-bearing peat	14-25% dry-weight ash
3) very peaty sediment	26-50% dry-weight ash
4) peaty sediment	51-75% dry-weight ash
5) sediment	76-100% dry-weight ash

Size Analysis

Most samples were examined under a binocular microscope in a manner similar to that described by Ingram (1965). This method involves visual estimation of the relative

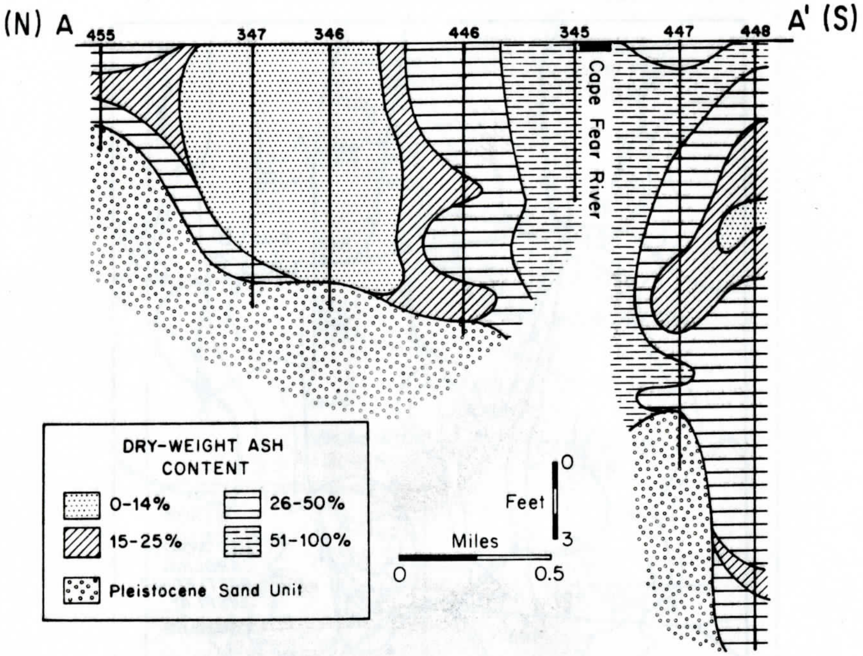

Figure 3. North-south cross-section through the peninsula and Eagle Island, showing the dry-weight ash content. (Figure symbols do not indicate lithologies of the Holocene sediments.)

abundance of the Wentworth sand and silt size fractions. All sediment smaller than coarse silt (0.031 mm), including fine-grained organic material, was classified as clay, since these fractions are indistinguishable by this method. In most samples visual examination was effective in establishing the presence of organic material, but not the amount, and appropriate descriptions of the material were made.

RESULTS

Organic Material

Cross-sections through the study area show that the peat with the lowest ash content (that is, sediment composed almost entirely of organic material) is found in the central portion of the peninsula. This organic-rich Holocene sediment overlies the Pleistocene sand unit in the north, as can be seen in cross-section A-A' (Fig. 3). Peat abruptly overlies the sand in some places, while in others there is a more gradual transition from peat to sand. Little of the Holocene sediment found in the north-central portion of the study area contains less than 50% organic material.

The organic content gradually decreases towards the rivers. Sediment containing less than 50% organic material is typically found next to the rivers. Cross-section B-B', which runs east-west across the peninsula (Fig. 4), shows that the sediment approaching both rivers contains a layer of very peaty clay at middle depths. This very peaty sediment interfingers with the peat in the central portion of the peninsula. A layer of clayey peat extends westward from the peat deposit, underlain by high-ash sediment and overlain by the very peaty clay.

The east-west cross-section C-C' across Eagle Island (Fig. 5) shows a pattern similar to B-B' in its western half. Two lenses of very peaty clay interfinger with high-ash sediment near the Brunswick River. The eastern portion of cross-section C-C', however, is quite different from that of the peninsula. A 4 ft (1.2 m) thick layer of peat is found about 3 ft (0.9 m) below the surface in the northeast corner of Eagle Island (cores 350 and 448). The peat is surrounded by very peaty sediment, and the whole package is underlain by high-ash sediment.

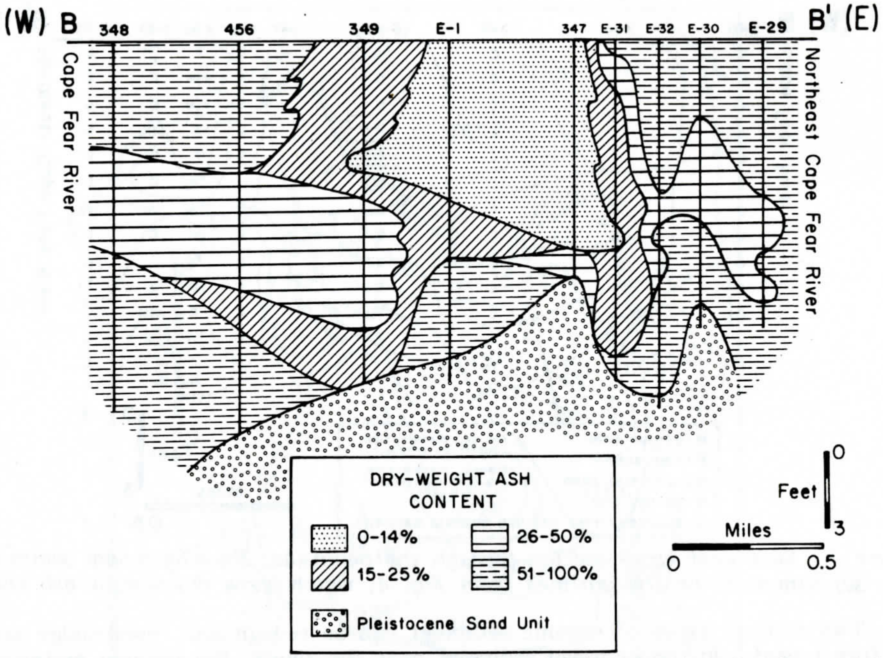

Figure 4. East-west cross-section through the peninsula, showing the dry-weight ash content. (Figure symbols do not indicate lithologies of the Holocene sediments.)

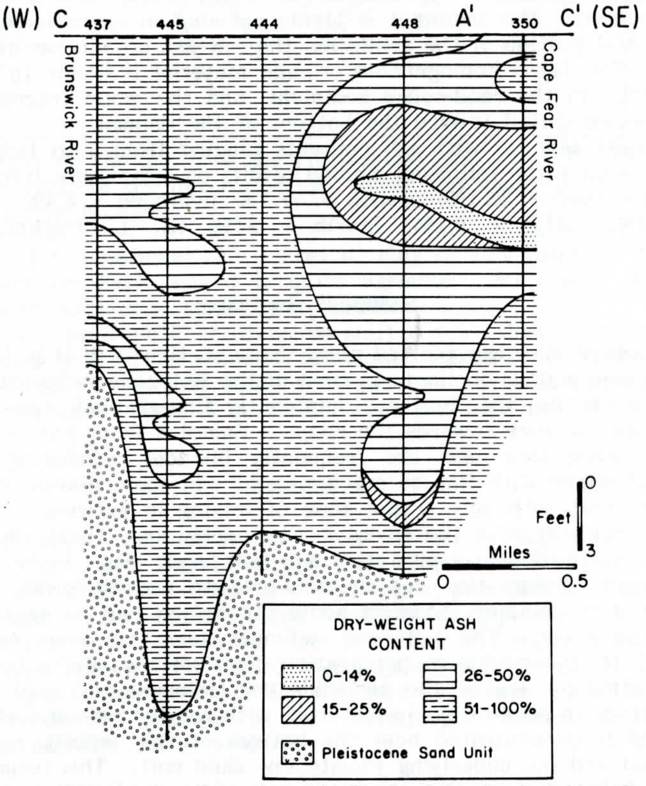

Figure 5. East-west cross-section through Eagle Island, showing the dry-weight ash content. (Figure symbols do not indicate lithologies of the Holocene sediments.)

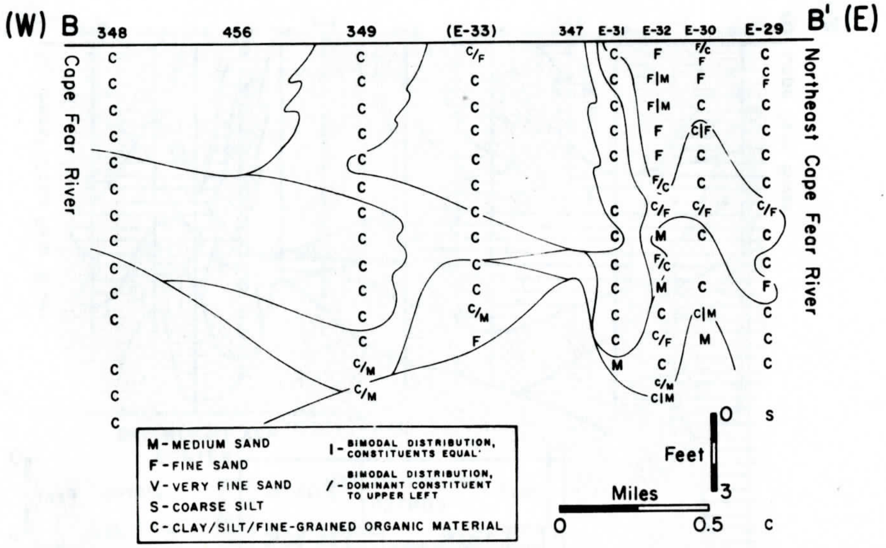

Figure 6. East-west cross-section through the peninsula, showing modal sediment size data superimposed on the outlines from Fig. 4, which show dry-weight ash content.

Two distinct types of organic sediment can be recognized: reed-sedge sediment and forest peat. In the high-ash sediment near the rivers, the organic material is of the reed-sedge type and is fibric to hemic, with many fragments of grass or reed. Additionally, wood, probably cypress (L. J. Otte, 1981, oral communication), is scattered throughout. The sediment is highly variable in appearance.

In the central portion of the peninsula, the low-ash organic sediment is composed of forest peat. This fairly homogeneous organic material is hemic to sapric. Wood is less common than in the reed-sedge sediment, and the wood fragments present are slightly more concentrated toward the bottom of the deposit.

For five peat samples with ash contents ranging from 3 to 15%, proximate and ultimate analyses gave the following median results on a moisture free basis: volitiles - 62.0%; fixed carbon - 32.1%; carbon - 57.1%; hydrogen - 4.4%; oxygen - 30.0%; nitrogen - 1.3%; sulfur - 0.2%; Btu/lb - 9400 (R. L. Ingram, 1983, written communication).

Sediment Analysis

Modal sediment size was plotted along cross-section B-B' (Fig. 6). Silt, clay, and fine-grained organic matter dominate in most of the Holocene sediment. In the central portion of the peninsula, the fines are dominantly fine-grained organic material; this material gives way to clay towards the rivers. Samples from just above the contact with the Pleistocene sand unit are frequently bimodal, consisting of fine-grained organic material mixed with fine to medium sand. In some places, peat overlies the Pleistocene sand unit with only a thin zone of mixing in between.

Holocene sediment near the rivers is dominantly grey to brown silty clay, with occasional thin layers of nearly pure clay and fine sand. Silty layers are present but are less common. Bioturbation due to plant growth and burrowing animals such as fiddler crabs is very common, however some lamination can be seen, particularly in very clay-rich sediment. The Holocene sediment near the rivers extends to depths greater than 27 ft, the maximum penetration depth of the corers used.

In the central portion of the peninsula the Holocene sediment is dominated by dark brown to blackish brown organic material, with varying amounts of fine to medium sand. The sand is concentrated near the bottom of the deposit, near the boundary between the peat and the underlying Pleistocene sand unit. This boundary is sharp; it shows no sign of disturbance other than that caused by the initial growth of the plants that formed the peat. No sedimentary structures are present in the Holocene

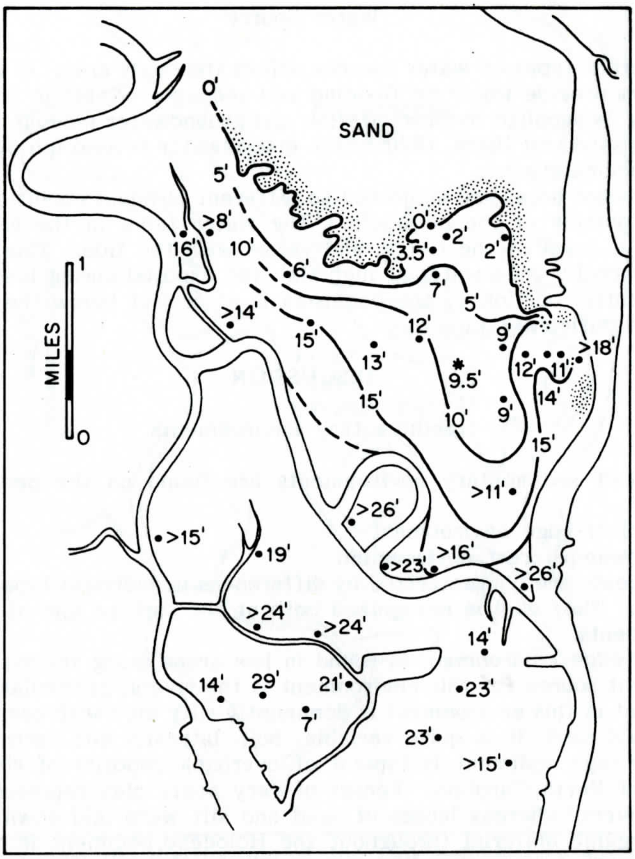

Figure 7. Map showing the depth to the Pleistocene sand unit. The asterisk indicates the site for the carbon-14 age determination, which was made on a peat sample from a depth of 9.5 ft (2.9 m).

sediments, except for occasional thin lenses of clay or fine sand.

The sediment on Eagle Island is quite variable. Silty clay is the dominant sediment type, with occasional small sandy areas.

The channel of the Cape Fear River has shifted more than once in the study area. The Brunswick River is a former channel, abandoned relatively recently. At least one other former channel, now totally filled with sediment, crosses Eagle Island (Fig. 7); sites with more than about 20 ft (6 m) of Holocene sediment are probably part of this old channel.

Vegetation

Two distinct plant communities grow on the peninsula. Near the rivers, the dominant plant species are *Typha* (cattail) and *Mariscus* (sawgrass). Numerous less abundant species of reeds and sedges are also found, particularly *Pontederia cordata* (pickerel weed), *Scirpus* (threesquare), and *Sagitaria* (duck potato). *Taxodium* (cypress) is the only common tree species. In the central portion of the peninsula the vegetation is dominated by *Acer rubrum* (red maple), *Nyssa sylvatica* (black gum), and *Liquidambar styraciflua* (sweet gum). *Smilax laurifolia*, *Magnolia virginiana* (sweet bay), and species of *Woodwardia* (ferns) are common in the understory. Species of *Pinus* (pine) are occasionally seen, but are not common.

Water Source

Two different types of water sources affect the study area. In the river-edge the areas, the rivers provide water by flooding and seepage. Water in the central portion of the peninsula is supplied by precipitation and groundwater moving southward through the Pleistocene sand unit (Bain, 1970). The groundwater is acid (ph 5.3) in areas where peat is now accumulating.

The river-edge areas are subjected to different physical conditions than are those in the central portion of the peninsula. The water table in the river-edge areas is controlled by the level of the rivers, and varies with the tide. Thus, the marsh along the rivers is generally underwater at high tide and exposed during low tide. The water table in the central portion of the peninsula is at or just below the surface and does not fluctuate significantly.

DISCUSSION

Sedimentary Environments

Two distinct sedimentary environments are found on the peninsula within the study area:

(1) the river-edge environment
(2) the swamp-forest environment

These environments are characterized by differences in sediment type, organic content, and vegetation. They can be recognized both at the surface and at depth within the Holocene sediments.

The *river-edge environment* is found in low areas along the rivers (Fig. 8). The primary sediment source for this environment is the rivers, particularly the Cape Fear River. Sediment in this environment is dominantly silty clay with occasional thin layers of clay, silt, and sand; it is quite variable, both laterally and vertically.

The river-edge sediment is typical of overbank deposits of rivers in the lower Coastal Plain of North Carolina. Lenses of very peaty clay represent periods of low water in the rivers, whereas lenses of sand and silt were laid down during high-river stages. The organic material throughout the Holocene sediment is of the reed-sedge type, suggesting that the present day plant community has remained stable through recent time.

The river-edge environment is widest along the Cape Fear River, indicating that this river supplies significantly more sediment than does the Northeast Cape Fear River. This is due primarily to the difference in size of the two rivers.

The *swamp-forest environment* is found in the center of the southern part of the peninsula (Fig. 8). The sediment is composed almost entirely of organic material of the forest peat type. Wood tends to be concentrated towards the bottom of the deposit due to the fact that cypress does not grow well on a peat substrate (C. J. Richardson, 1982, oral communication). Other types of wood tend to decompose more completely when buried, thus enhancing the cypress component.

The primary source of mineral sediment for this area is the underlying Pleistocene sand unit. Sand is therefore concentrated towards the bottom of the deposit, near the boundary between the peat and the underlying sand. Some fine sand is dispersed throughout the organic-rich sediment, however, and was probably blown into the area by the wind from an area of Pleistocene sand that is exposed to the north.

In the central and northern parts of the peninsula, peat lies directly on top of the Pleistocene sand unit. Further north, this peat layer pinches out. As the peat deposit has accumulated, it has spread northward over the broad sand ridge forming the spine of the peninsula.

The forest peat reaches its maximum thickness of about 10 to 12 ft near the southern tip of the swamp-forest area. This is considerably shallower than the deepest river-edge sediment (Fig. 7).

Eagle Island

Much of Eagle Island has been disturbed by human activity, particularly by the

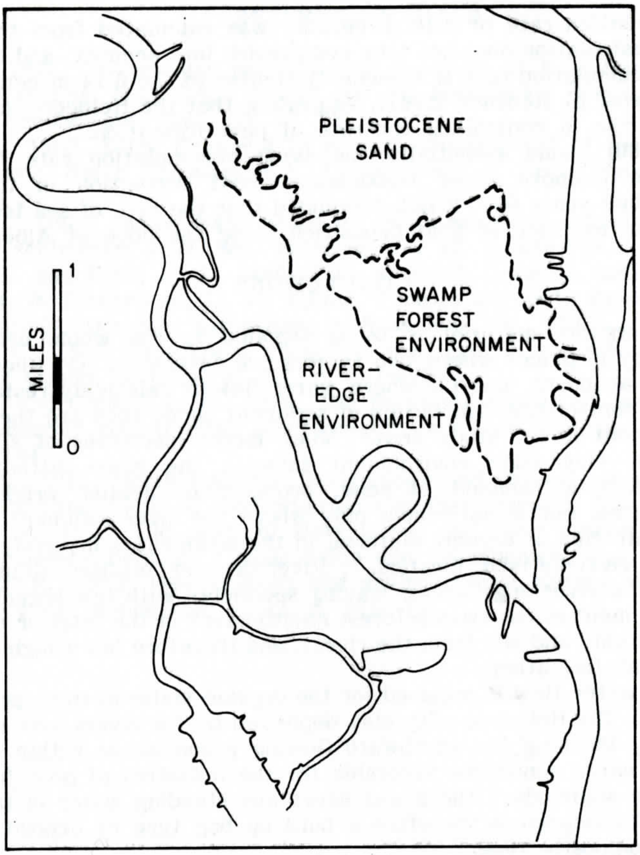

Figure 8. Map showing the distribution of the two sedimentary environments on the peninsula.

dumping of dredge spoil onto its southern end. Highway and building construction, as well as site preparation for the U.S.S. North Carolina battleship display, have profoundly affected the surface of large areas of the northern half of the island. Relatively undisturbed areas are now dominated by the river-edge environment. Some areas in the interior of the island have a plant community similar, but not identical, to the swamp-forest vegetation. Cypress is much more common, and is dominant in a few sandy areas.

Sediment found in samples collected on Eagle Island is dominated by silty clay from the rivers. Much variability is evident, however. The lens of peat found in the northeast corner of the island is similar, but not identical, to that found in the swamp-forest environment. The principal difference between this peat and the swamp-forest type is that the Eagle Island peat does not sit directly upon the Pleistocene sand; instead, it is underlain by a layer of very peaty clay that is underlain by sand. This intermediate very peaty clay layer is similar to the river sediment, and indicates that the area was probably part of the river-edge environment before peat formation began there. The vertical variation in depositional environments was probably caused by the migration of the main channel of the Cape Fear River.

AGE OF PEAT DEPOSIT

The thickest swamp forest peat observed reached 9 1/2 ft (2.9 m). At this site, the peat is underlain by the Pleistocene sand unit (Fig. 7). A sample was collected from the base of the peat deposit, and carbon-14 analysis determined its age to be 2840 ± 50 years BP.

An accumulation rate of 0.10 m/century was calculated from the thickness and age data. This calculation does not take compaction into account, and so is a minimum figure. This accumulation rate is reasonably similar to the 0.14 m/century rate of sea level rise calculated by Redfield (1967), suggesting that the Holocene sea level rise has played a major role in controlling the rate of peat formation.

Erlich (1980) found essentially the same accumulation rate for the organic sediment of the Roanoke River floodplain. Peat deposition in that area began approximately 5000 years BP. Erlich concluded that the rate of sea level rise was the factor controlling the rate of peat deposition along the edge of Albemarle Sound.

CONCLUSIONS

Abundant organic material is being supplied to the whole area, but peat is accumulating only in places where two factors are favorable: (1) where river-derived clastics are not abundant, and (2) where water-flow is relatively restricted.

The rivers control the distribution of sediment, since they are the primary source of mineral sediment in the study area. Much larger quantities of silt and clay are deposited in the river-edge environment than in the more distant swamp-forest environment. A small amount of sand, probably of aeolian origin, is deposited throughout the area, but is noticeable only where the river sediment is absent.

The concentration of organic material in the sediment is inversely proportional to the amount of river-derived clastics. River-derived clastics dilute the organic contribution to the river-edge areas, leaving sediments with low organic contents. In contrast, the sediment in the swamp-forest environment in the interior of the peninsula is less diluted by clay and silt from the rivers, and therefore has a high organic content that leads to peat formation.

Restricted water flow is required for the organic material to be preserved as peat (Waksman, 1942). The Holocene silty clay deposited by the rivers acts as an aquiclude, at least partially blocking the southward-flowing groundwater within the Pleistocene sand unit; this created conditions favorable for the initiation of peat formation in the area north of the aquiclude. There was never any standing water in the area of peat accumulation. The deposit is therefore a built-up bog type of deposit, rather than a filled basin or composite type.

The elevation of the water table is limited by the level of the surface on the river-edge sediment. This is controlled by the level of the water in the rivers. Rising sea level during the Holocene has caused the river level in the study area to rise. The rate of peat formation is therefore ultimately limited by the rate of sea level rise, as was suggested by the carbon-14 date obtained for the base of the deposit.

Changes in the course of the Cape Fear River through time have shifted the locations of swamp-forest environmental conditions. The tongue of peat on Eagle Island formed when the river occupied a different channel, perhaps the Brunswick River channel. When the river shifted northward to its present course, the flow of groundwater from the Pleistocene sand unit was cut off. The northern half of Eagle Island then began receiving sediment from the river, and peat formation ceased.

ACKNOWLEDGEMENTS

I would like to thank Roy L. Ingram for his guidance in this study, also Walter H. Wheeler and John J. W. Rogers for their support and advice. Special thanks go to Lee J. Otte who provided much assistance and encouragement during the early stages of this research. I wish to express my appreciation to the reviewers of this article for their many valuable suggestions.

This study was made possible in part by funding from Sigma Xi, the United States Department of Energy (Grant DE-AC18-79FC14693 to Roy L. Ingram), and the North Carolina Geological Survey.

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PETROLOGY OF THE HUNTERSVILLE CHERT

S. J. SHERRARD — *Chevron U.S.A., P. O. Box 6056, New Orleans, Louisiana 70174*

M. T. HEALD — *Department of Geology and Geography, West Virginia University, Morgantown, West Virginia 26506*

ABSTRACT

The Huntersville Chert of Devonian age is an important gas producer in West Virginia and Pennsylvania. In five cores from north-central West Virginia, western Maryland and central Pennsylvania, 7 lithotypes were recognized: clean chert, chert with dark organic material, spicular chert, dolomitic chert, glauconitic chert, silty argillaceous chert and dolomitic silty argillaceous chert. The chert is apparently of primary origin with siliceous organisms playing a major part in the accumulation of the silica.

Pores occur in the canals of spicules, within chalcedony, and along incompletely filled fractures and voids after removal of calcite. Fractures developed in 7 stages and resulted from shrinkage of early silica, slumping, differential compaction and tectonic activity. Dark organic material migrated repeatedly along the tectonic fractures. The best porosity and permeability for hydrocarbon accumulation is in the purer brittle chert which fractured readily.

INTRODUCTION

The Huntersville is the chert facies of the Devonian Onesquethaw Stage and underlies most of West Virginia, southwestern Pennsylvania and western Maryland. The location of the wells from which samples were studied is shown in Figure 1. Cores were obtained from 5 wells and cuttings from 10 other wells. Most of the wells are in gas fields that have produced from the Huntersville Chert and are situated at or near the crests of anticlines. Much of the study was concentrated in the northern part of the Huntersville because of the greater availability of samples and well logs. The investigation was based primarily on thin sections supplemented by X-ray mineral analysis, elemental analysis and pyrolysis.

Most previous work on the Huntersville Chert has been concerned with stratigraphic and structural relationships. Dennison (1961) included the Huntersville in his Onesquethaw Stage which is the interval between the top of the Tioga metabentonite and the top of the Deerpark Stage. On the basis of a study of the Huntersville in northern West Virginia, Amstadt (1980) subdivided the unit into several lithofacies. The emphasis of Skoff's (1980) investigation was on the sedimentation of the Onesquethaw Stage.

STRATIGRAPHY

The Huntersville Chert is a member of the Onesquethaw Stage and has as lateral facies equivalents the Onondaga Limestone and Needmore Shale (Dennison, 1961). Where present, the Tioga metabentonite marks the top of this stage. The Onesquethaw Stage is early middle Devonian and followed the Deerpark Stage, represented by the Oriskany Sandstone, and preceded the Cazenovia Stage, represented by the Marcellus Shale and the Millboro Shale. Across West Virginia the Onesquethaw generally thickens from east to west with the thickest Huntersville Chert sequences in the north-central area of the state.

Amstadt (1980) and Skoff (1980) divided the Onesquethaw Stage into five basic stratigraphic units mainly on the basis of gamma-ray logs:

Onondaga Limestone
Upper unit }
Middle unit } Huntersville Chert
Lower unit }
Needmore Shale

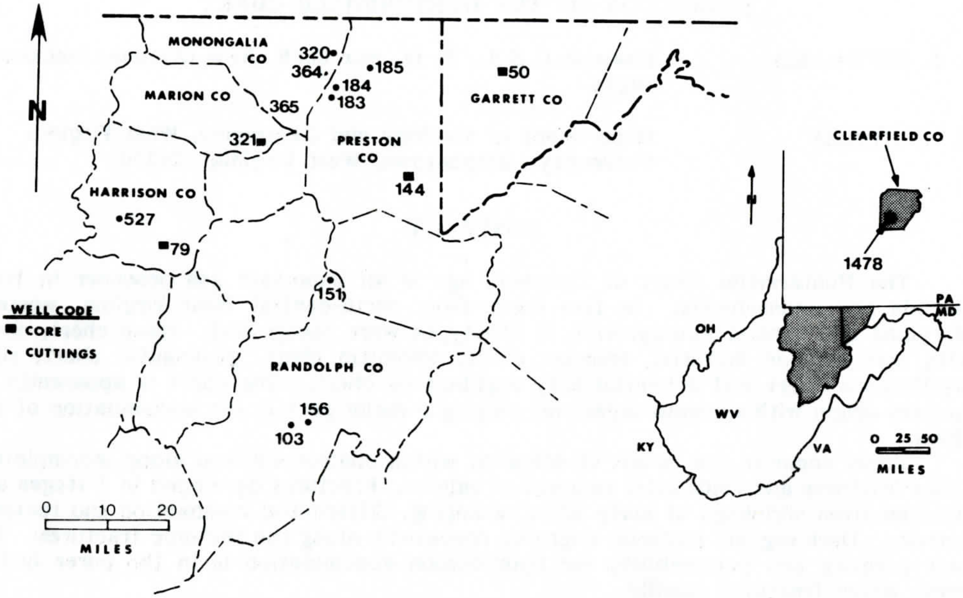

Figure 1. Map showing location of wells studied. Well numbers are county permit numbers.

The upper unit of the Huntersville is composed of the purest chert and increases in thickness from 20 feet in eastern Garrett County, Maryland to 120 feet in Lewis County, West Virginia (Skoff, 1980). The middle unit of the chert is silty and argillaceous especially near the top and has a relatively constant thickness across most of the study area but thins considerably in Garrett County, Maryland. The lower unit of the Huntersville consists of clean chert in western West Virginia and grades into shaly chert to the east. The chert becomes increasingly argillaceous at depth and forms a gradational contact with the Needmore Shale.

CHERT LITHOTYPES

The Huntersville chert varies in lithology because of differences in content of organic material, dolomite, glauconite, silt and argillaceous material. The chert has been classified into the following lithotypes: clean chert, chert with dark organic material, spicular chert, dolomitic chert, glauconitic chert, silty argillaceous chert and dolomitic silty argillaceous chert.

Clean Chert: The purest chert is white to light gray in hand specimen and appears white in thin section. With small amounts of organic material the chert becomes medium dark gray and has a tan appearance in thin section. Even the cleanest chert in the Huntersville has at least small amounts of dolomite and silt-size quartz detritus.

Chert With Dark Organic Material: This lithotype has fine specks and blebs of dark organic material dispersed through the chert. In thin section with transmitted light the chert is light brown to dark brown depending on the amount of organic material. In reflected light it is milky white or brownish where organic material is more concentrated.

Spicular Chert: This lithotype is characterized by an abundance of rod-shaped structures (Pl. 1A) that are probably spicules of the silicic sponges, Demospongia and Hexactinellida, Steven A. Warshauer, personal communication). The spicules are most commonly monaxons with some tetraxons. They are easily seen where dark organic material surrounds the spicules and fills their central canals. The spicules are difficult to detect where the dark organic material was lost during recrystallization.

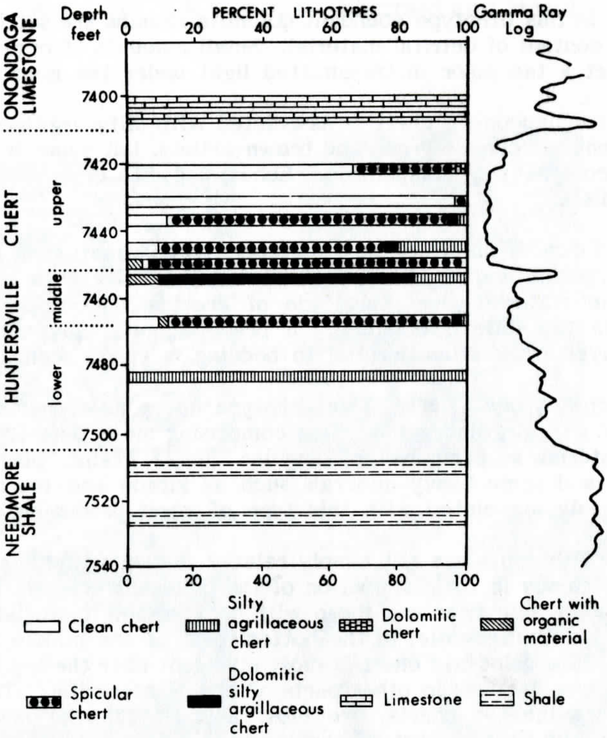

Figure 2. Lithotypes in the George Well (Gar. 50) Garrett County, Maryland.

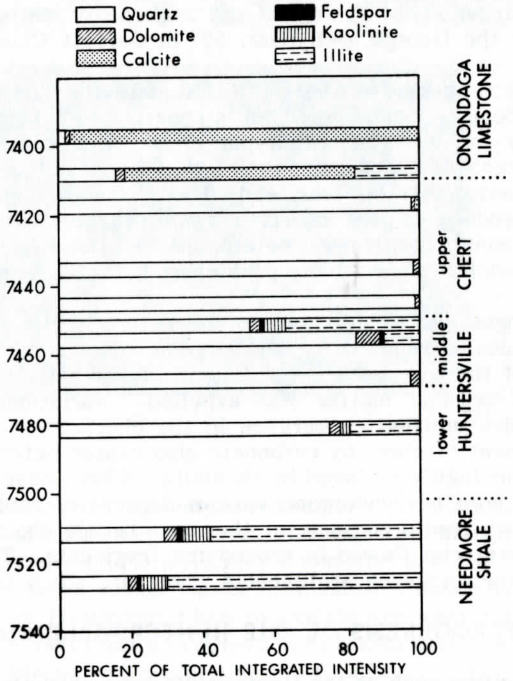

Figure 3. Mineralogy based on X-ray diffraction analysis of core samples from the George Well (Gar. 50) Garrett County, Maryland.

Dolomitic Chert: In this lithotype abundant dolomite rhombs are set in a chert matrix with only a minor content of detrital material. Small amounts of organic material give the dolomitic chert a tan color in transmitted light under the microscope.

Glauconitic Chert: Glauconitic chert is associated with silty argillaceous chert. The glauconite is present as rounded green and brown pellets, but some is highly ellipitical from flattening accompanying compaction. Euhedral pyrite crystals occur in many of the glauconite pellets.

Silty Argillaceous Chert: This lithotype has a chert matrix containing greater than 33% silt-sized detrital grains and argillaceous material, generally with small amounts of dolomite. Organic material gives this type of chert a dark appearance under the microscope. When the slide is composed of predominantly clay minerals, illite and kaolinite, a preferred orientation parallel to bedding is easily seen.

Dolomitic-Silty Argillaceous Chert: This lithotype has a chert matrix with dolomite rhombohedrons and silty argillaceous detritus comprising more than 33% of the sample. The silt-sized material is composed of angular quartz grains, plagioclase feldspar, glauconite pellets and some heavy minerals such as zircon and tourmaline. Organic material is commonly associated with this type of chert producing a brown color in transmitted light.

Although the lithotypes are not simply related to stratigraphic position in detail, there is some consistency in the distribution of the different types in the section. The purer cherts, including spiculites and those with dark organic material, are common in the upper unit of the Huntersville, in the bottom part of the middle unit and the top of the lower unit. The dolomitic chert is most abundant near the top of the upper unit and in some argillaceous zones in other parts of the Huntersville. The more detrital types, silty and argillaceous cherts, are most abundant in the middle unit of the Huntersville and in the bottom part of the lower unit. The glauconitic chert occurs in patches in local silty argillaceous zones in the upper unit of the Huntersville. Examples of the distribution of lithotypes and mineralogy in the section are shown in Figures 2 and 3 for the George Well (Gar. 50) in Garrett County, Maryland.

Heterogeneity In The Huntersville Chert: The Huntersville Chert characteristically has a mottled, heterogeneous appearance which is apparent in both thin sections and cores (Pl. 1B). Generally the different lithotypes grade into each other with no definite boundaries. The heterogeneity is due in part to depositional conditions. At times of greater current activity, clay minerals and silt-size quartz and feldspar were brought into the basin to produce impure chert. Organic structures which appeared to be sponge masses also contributed to the heterogeneity. Detritus washed in around these masses resulted in patches of relatively pure chert being surrounded by silty chert (Pl. 1C).

Diagnetic changes also contributed to the heterogeneity of the chert. Some of the chert with spicules surrounded by dark organic matter grades into clear chert in which the outlines of the spicules are very vague. Apparently this chert recrystallized and in the process, organic matter was expelled. Variations in degree of recrystallization produced a mottled appearance in the chert.

Local replacement of chert by carbonate also caused heterogeneity. Some of the argillaceous chert was highly replaced by dolomite. Chert near limestone facies as in the Preston County core has undergone various degrees of replacement by calcite.

Early brecciation fractured some of the more brittle chert and relatively ductile argillaceous cherty material flowed in around the fragments. The boundaries between the fragments and the enclosing argillaceous chert are sharp (Pl. 1D).

PETROGENESIS OF THE HUNTERSVILLE CHERT

Thick chert sections such as the Huntersville have been attributed to replacement of limestone or to primary accumulation through either inorganic or organic processes. Although chert nodules formed by replacement in limestones associated with the Huntersville, evidence of large scale replacement is lacking. Shells of silicified

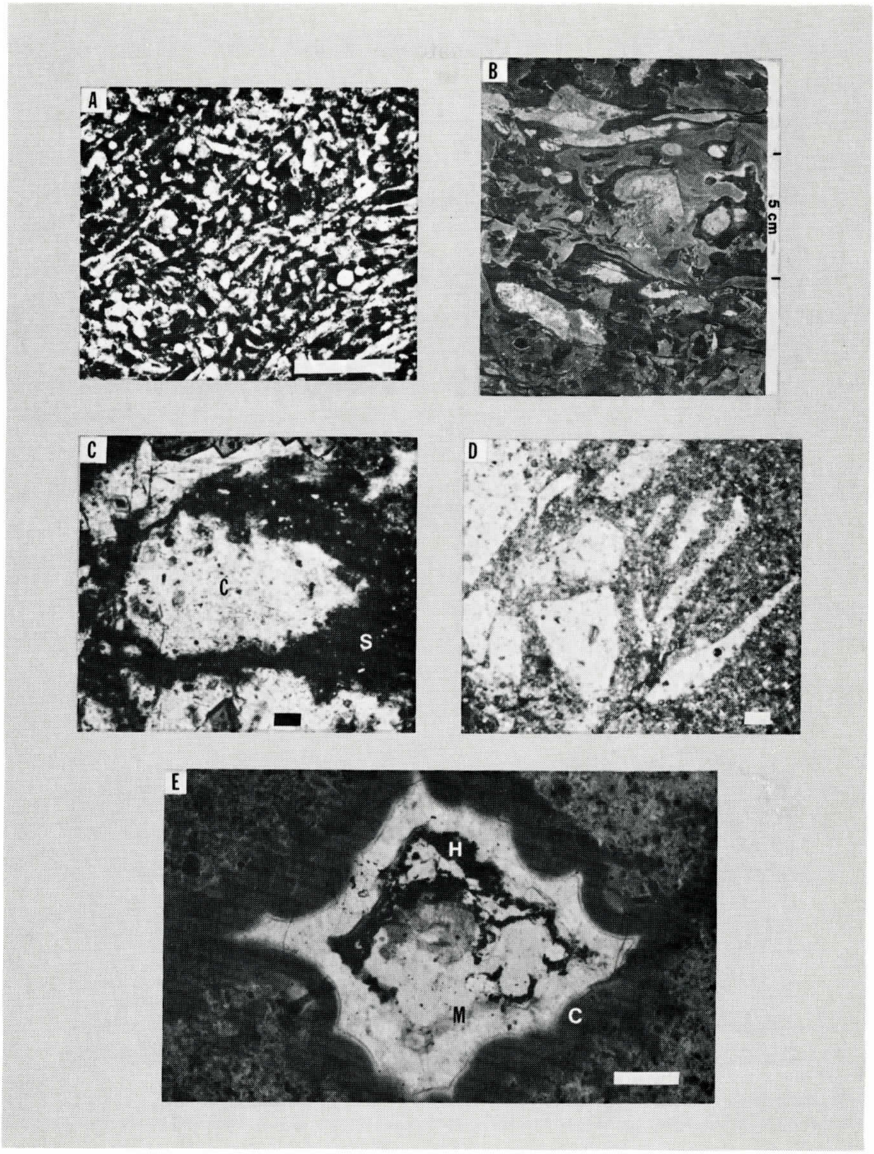

Plate 1 Samples from wells. Numbers refer to county permit numbers. A) Photomicrograph of chert rich in sponge spicules. Clearfield Co. 1478, 7445.4 feet. B) Photograph of core slab. Marion Co. 321, 7382.5 feet. C, D, and E are photomicrographs. C) Light colored chert (C) which may represent sponge masses surrounded by dark silty chert (S). Preston Co. 144, 5622 feet. D) Brecciated light colored chert enclosed in argillaceous chert. Clearfield Co. 1478, 7441.9 feet. E) Vug lined with chalcedony (C) and filled with microcrystalline quartz (M) and fossil hydrocarbon (H). Clearfield Co. 1478, 7443.9 feet. All photomicrographs are in plain light. Bar lengths are 0.2 mm.

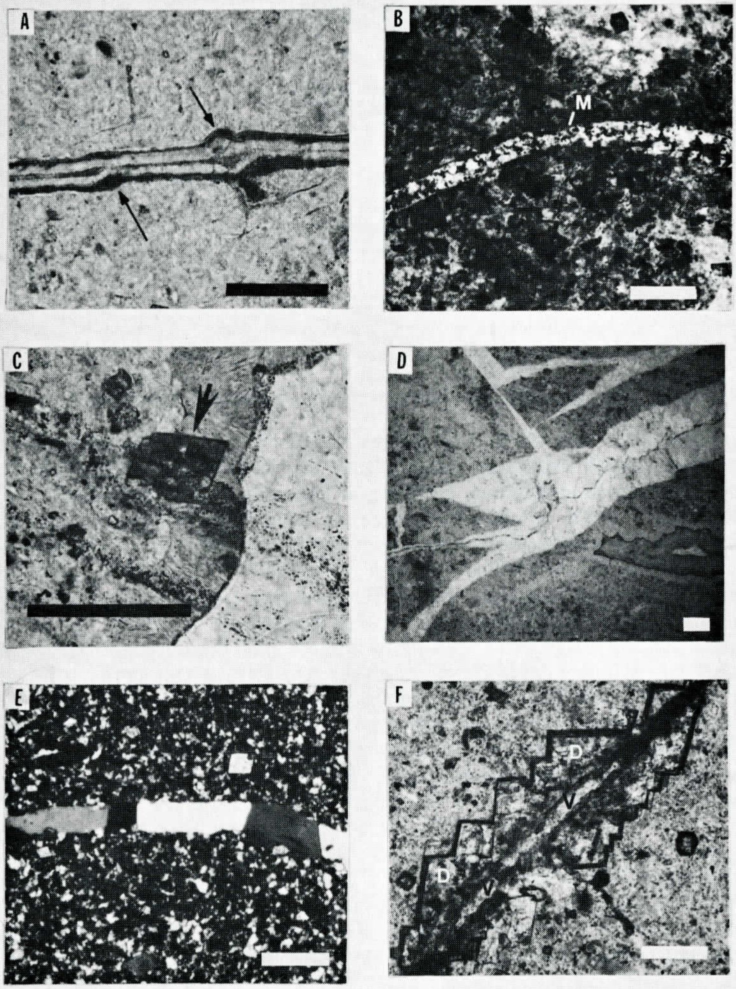

Plate 2 Photomicrographs of well samples. Numbers refer to county permit numbers. A) Voids (arrows) at kinks resulting from lateral movement along a chalcedony vein. Plain light. Garret Co. 50, 7430.9 feet. B) Fossil mold (M) filled with microcrystalline quartz and fossil hydrocarbon (black). Plain light. Preston Co. 144, 5623.5 feet. C) Mold (arrow) resulting from leaching of a calcite rhomb. Plain light. Garret Co. 50, 7467.2 feet. D) Coarse mosaic quartz filling fractures which formed during late brecciation of the chert. Plain light. Garret Co. 50, 7430.9 feet. E) Long quartz crystals that formed in late fracture. Crossed nicols. Garret Co. 50, 7430.9 feet. F) Dolomite (D) replaced quartz vein (V) and adjacent chert matrix. Plain light. Clearfield Co. 1478, 7443.9 feet. Bar lengths are 0.2 mm.

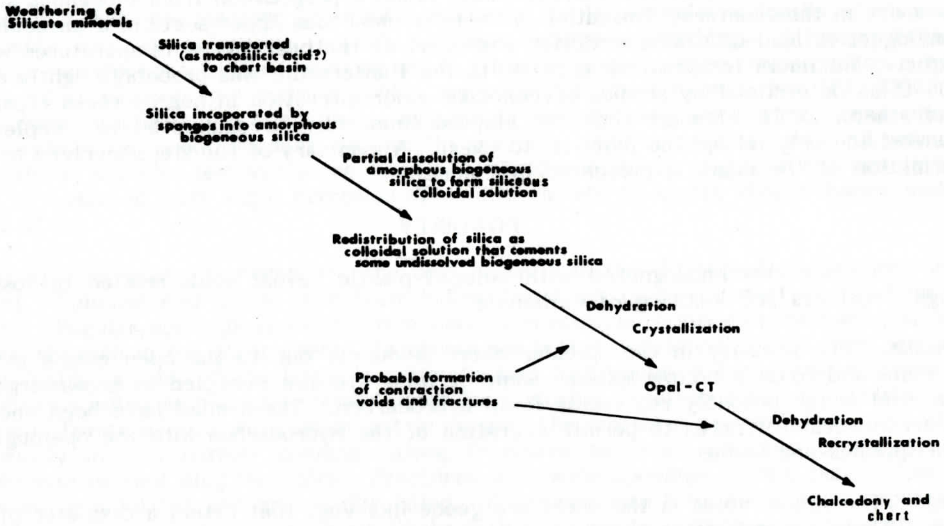

Figure 4. Postulated sequence in formation of chert in the Huntersville.

calcareous organisms are well preserved in the chert but they are rare. If limestone had been extensively replaced, one would have expected these shells to be much more common. Dennison (1961) favored a primary inorganic origin for the chert and cited as evidence structures attributed to slump of plastic silica gel. The present writers favor a primary origin but feel that biogenic activity may have played a major role in the formation of the chert. Siliceous sponges were present and locally produced chert made up almost entirely of spicules. Irregular patches in some of the cores have been interpreted by Alan C. Donaldson (personal communication) as sponge masses which were probably Demspongia and Hexactinellida. Although spicules were not conspicuous in some thin sections, gradations from recognizable spicules to massive chert were common. This suggests that siliceous organisms had been more abundant but their forms were obliterated during the transformation of the silica to its present crystalline form. Chert formation was evidently favored in the deeper parts of the basin where siliceous organisms flourished. The development of troughs for accumulation and arches to restrict circulation was believed to be controlled by basement structures (Skoff, 1980). The original source of the silica is uncertain.

Volcanic activity, indicated by the Tioga metabentonite, was occurring during Huntersville time and may have been a source for some of the silica, but no evidence of pyroclastics was found in the chert. Detritus in the chert is quartzose rather than feldspathic as would be expected with volcanic sources. Probably the chief source of silica was from the terrestrial weathering of silicate minerals. Because the chert basin was at low latitudes in Huntersville time, weathering of silicate minerals was most likely favored by the tropical to subtropical climate that existed at that time. The released silica was then transported to the chert basin where it was incorporated by sponges into amorphous siliceous spicules. Modern diatoms have been shown to be very effective in removing silica from sea water, even at low concentrations, where the only source of silica was from ocean and river water and no volcanics were involved (Calvert, 1966).

The remaining sequence of events in the formation of the Huntersville Chert is probably similar to that proposed by Oehler (1976). After accumulation of the spicules, the amorphous biogenous silica was partially dissolved and redistributed as a siliceous colloidal solution that cemented some of the undissolved biogeneous silica. The dissolution and redistribution started early before compaction of the spicules as evidenced by loosely packed spicules cemented by chalcedonic quartz. The silica mass was then subject to dehydration and crystallization to form opal-CT, a poorly ordered form of cristobalite and tridymite. Contraction voids and syneresis cracks developed during dehydration. Additional dehydration and recrystallization produced the final form of the chert. Ernst and Calvert (1969) found a progression from biogenous silica to chert in the Monterey Formation with a diatomaceous upper section, a porcelanite (analogous to opal-CT) middle section and chert at the base where temperatures were higher. Maximum temperature attained in the Huntersville was probably slightly over 200°C as determined by studies of conodont color alteration in nearby rocks (Epstein and others, 1977). Enough time has elapsed since silica accumulated to completely convert the original opaline material to chert. A summary of the steps involved in the formation of the chert is presented in Figure 4.

POROSITY

Thin sections impregnated with colored plastic reveal voids related to fossils, vugs, fractures and leaching of carbonate.

Fossils: The porosity in the spicular chert is mainly due to the open canals in the spicules and voids in hollow spines. Some of these are now occupied by brown organic material which probably represents fossil hydrocarbon. There must have been enough interconnection of pores to permit migration of the hydrocarbon into the openings in the spicules and spines.

Vugs: The larger voids in the chert are geode-like vugs that attain a diameter of 10 mm. These formed in the relatively pure chert apparently as a result of contraction in the transformation of amorphous silica to crystalline form. Micropores occur in the botryoidal chalcedony that lines the vugs. The pores are present in bands parallel to the botryoidal surfaces and apparently resulted from interruptions in growth of the chalcedony. Some of the pores are occupied by brown organic material but others remained open as indicated by the presence of colored plastic in thin section. Microcrystalline quartz formed in the vugs after chalcedony; however, relatively large voids were left where quartz filling was incomplete. Some of the spaces between the quartz crystals are occupied by fossil hydrocarbon (P1. 1E).

Submicroscopic Pores: Several lines of evidence indicate that a notable part of the porosity in the chert exists in voids too small to be observed in thin section. Porosity determined by core analysis and by comparing bulk density to matrix density is several percent higher than that measured in thin section. Tiny voids between crystals is suggested where the chert has low index, is abnormally dark in transmitted light and has a white frosted appearance in reflected light. The presence of pores between fibers in the chalcedony is also suggested by the golden brown color which occurs where hydrocarbon was migrating through the chalcedony to vugs and openings in fractures. The chalcedony is white where veins truncated the chalcedony and acted as a seal to prevent hydrocarbon migration. Patches of clay in the chert are more porous than shale because the surrounding rigid chert acted as a support and prevented the clay from being highly compacted.

Local variation in porosity can be determined by observing salt crusts on the core. The cores typically contain some salt from saline formation waters. When the core is wet and left to dry capillary action in porous areas brings saline water to the surface where salt forms on evaporation. Solid salt crusts form on argillaceous zones and minor salt develops on impure chert, but tight chert patches are free of salt. Even small scale variations in porosity are clearly indicated by salt distribution.

Fractures: Fractures produced the best effective porosity in the chert. The fractures are most highly developed in the brittle purer chert and tend to die out in the more argillaceous chert. Organic material migrated into the fractures from the relatively organic-rich argillaceous chert. Some fractures are filled with chalcedony but some pores were left within the chalcedony. Where lateral movement occurred along the fractures after chalcedony filling, voids developed at kinks at the border of the fractures (P1. 2A). Good permeable channels result where mosaic quartz only partially fills the fracture openings.

Leaching: A few molds of small fossils indicate that some secondary porosity developed from the leaching of calcite. Some of the large shells of fossils which were probably initially calcareous appear to have undergone leaching. Microcrystalline quartz has grown into the resulting voids although some cavities remained open and were filled with organic material which apparently is fossil hydrocarbon (Pl. 2B). In the cores studied these shells were uncommon but if there were some areas where these shells had been abundant, noteworthy porosity could be present. A small amount of calcite was also removed along fractures. Evidence for all stages in the dissolution can be observed from slight corrosion of calcite rhombs to empty rhomb-shaped molds (Pl. 2C).

Factors Favorable For Porosity: The best porosity in the Huntersville is in the purer chert. Spicules and spines with their hollow tubes are most abundant in this type of chert. Vugular openings resulting from early contraction of siliceous masses also are more common in the purer chert. Fractures are better developed in this chert because it is highly brittle. Some large openings formed in brecciated zones where irregular fragments acted as props to prevent voids from closing. Scattered quartz crystals that grew in fractures also were important in keeping the fractures propped open. Porosity is easily lost in narrow openings along fractures as small amounts of quartz or carbonate cement plug the voids. Fractures with wide openings permit easy inflow of cementing solutions and are readily filled. Fractures with intermediate size openings (0.03 mm to 2 mm wide) are most likely to have porosity preserved as they are not as readily filled as the small fractures and cementing solutions do not enter as easily as in the case of fractures with wide openings.

Porosity is preserved in patches of argillaceous chert when surrounding rigid chert forms a framework to prevent compaction of the clay. Patchy distribution of clay may result from slumpage of sequences of shaly beds and siliceous beds. Also, patches of clay may result from deposition of argillaceous material around organic structures such as siliceous sponges. Additionally, ductile argillaceous material may flow in around early brecciated fragments of chert. Argillaceous materials distributed in layers were unfavorable for porosity as voids were lost through normal compaction. Furthermore, laminations of clay promote development of stylolite seams with resulting silica solute which leads to cementation of voids in the chert.

The small amount of secondary porosity produced by calcite leaching may have been caused by the more acid conditions related to the maturation of organics in the chert. Flow of acidic fluids along fractures would account for the observed leaching of calcite in some of the fracture zones.

SEQUENCE OF FRACTURING AND OPEN SPACE FILLING

Vugs: The earliest openings to form in the chert were vugs apparently resulting from contraction attending conversion of amorphous silica to crystallized forms. The vugs tend to be elongate with maximum openings of 10 mm and are best developed in the purer chert. Commonly two or three generations of chalcedony (fibrous quartz) were deposited in botryoidal forms. Fibers radiate out from multiple growth centers which occur along the margins of the vugs in the first generation. New growth centers formed in the later generations of chalcedony so that the quartz fibers are not continuous from one generation to another. In thin sections most of the chalcedony is tan although some is white. After chalcedony had crystallized, coarse quartz formed in the central part of some of the vugs where space was available.

Tan Chalcedony Veins: Tan chalcedony veins commonly link the vugs. Some veins have one generation of chalcedony and others have two generations. In the first generation, quartz fibers grew perpendicularly to the walls of the veins whereas in the second generation, fibers generally formed an oblique angle with the surface of the earlier chalcedony. The veins formed at about the same time as mineral deposition in the vugs as the fibrous quartz in the veins is continuous with the fibrous quartz in the vugs. The openings for these veins may have resulted from contraction of the silica mass or from movements along lines of weakness centered on the vugs.

Early Brecciation: In the early stages of diagenesis the pure chert became rigid before the impure argillaceous lithotypes became well lithified so that during slumping and/or differential compaction, the pure chert brecciated and the more ductile argillaceous material tended to flow in around the resulting fragments. Some spaces were left between the fragments and these later became filled with coarse secondary quartz.

White Chalcedony Veins: Tectonic activity caused well-defined fractures which locally became filled with white chalcedony. The fibers in these veins are well aligned perpendicular to walls of the veins and are free of organic material. These veins clearly cut across the earlier formed tan chalcedony.

Fine Mosaic Quartz Veins: Fine mosaic quartz formed narrow veins generally less than 0.03 mm wide after the period of chalcedony development. Dark organic material apparently representing fossil hydrocarbon occurs with quartz in most of these veins. The veins are free of organic matter where they cross earlier pure chalcedony veins. Apparently, the earlier chalcedony acted as favorable seed crystals so quartz grew rapidly at the intersection before hydrocarbon bleeding from the chert matrix entered the vein. Commonly the mosaic quartz veins narrow down to hairline fractures with only organic material. Separate hairline fractures with organic material and no mineral matter also formed at this time. In the later development of the fine mosaic quartz, less hydrocarbon was deposited so that nearly clear quartz formed.

Coarse Mosaic Quartz Veins: The next veins to form had small crystals at the borders grading into very coarse mosaic quartz at the center of the veins. Somewhat of a comb structure formed at the juncture of the quartz that had grown out from the two walls of the veins. Dark organic material is common along this juncture. Coarse mosaic quartz also surrounds fragments resulting from tectonic brecciation of the chert (Pl. 2D).

Vein With Large Quartz Crystals: The last quartz veins to form have large single crystals (Pl. 2E). In the smaller veins the quartz not only grew completely across the veins but also commonly extended along the veins for a distance equal to six times the width of the veins. Conditions must have been particularly favorable for slow quartz growth at this stage in contrast to earlier conditions favoring chalcedony and fine mosaic quartz. In a few places veins were wide and interlocking crystals formed in the veins. Crystals at the margins of the veins were as large as those at the center and thus these veins differed from those of mosaic quartz where grain size markedly increased from the margins to the centers of the veins.

Carbonate Veins: The last veins to form are composed mainly of coarse carbonate. Calcite is the chief carbonate with smaller amounts of dolomite which formed mostly by replacing calcite. These veins are more common in the carbonate bearing chert such as in the upper part of the Preston County well. In some veins quartz is intergrown with the carbonate or is the chief filling along short stretches of the veins. Quartz filled the veins where they crossed early quartz veins apparently because the earlier quartz acted as seed crystals and enabled the quartz to grow at the intersection before precipitation of carbonate. The quartz is dark from organic material which filled voids that occurred in the quartz at the intersections. Some of the carbonate veins narrow down and grade into hairline fractures filled only with dark organic matter.

MINERAL REPLACEMENT

Quartz in the form of chert, chalcedony and coarse crystals was partially replaced by calcite and dolomite from early to late stages of diagenesis. Replacement of the chert matrix by calcite was particularly common in the upper part of the Huntersville because of the availability of carbonate from the overlying Onondaga limestone and calcitic shales. In the lower parts of the interval, calcite was more limited but rhombohedrons growing beyond the walls of calcite veins commonly replaced part of the adjourning chert matrix.

Replacement by dolomite appears to have occurred in two stages. First stage dolomite is cut by early hairline fractures which are lined with organic material. This dolomite occurs as small rhombohedrons in all chert lithotypes but is more abundant in the chert that was originally silty and argillaceous. In the purer chert early replacement dolomite is more common in milky white chert which apparently had numerous tiny pores. These pores facilitated the movement of fluids and replacement of chert. The second stage of replacement produced dolomite that was more massive and developed after formation of carbonate veins. This later dolomite appears to have most easily replaced the macrocrystalline quartz found in veins and the center of vugs. In addition to replacing the macrocrystalline quartz, the dolomite extended into the surrounding chalcedony and chert matrix (Pl. 2F). Some of the dolomite may have been forming at the same time as the hydrocarbon development inasmuch as transformation of clays with release of magnesium occurs in the same temperature range as the generation of oil (Heraux and others, 1979; Rimstidt, 1983). Traces of this oil may be represented by the fossil hydrocarbon in the vugs and fractures.

Although pyrite occurs only in trace amounts, it had a strong tendency to replace chert matrix and macrocrystalline vein quartz. The pyrite is euhedral and formed after carbonate veins but before late dolomite. Some of the pyrite crystals which replaced vein quartz are completely surrounded by late replacement dolomite that left the pyrite intact.

MIGRATION OF ORGANIC MATERIAL

Dark brown organic material occurs within spicule canals and is intimately associated with the fine quartz in the chert. Some of this organic material presumably formed syngenetically and some may have been introduced early in the history of the chert. Migration of the organic material, probably as a hydrocarbon, was promoted by several processes. Some of the interstitial organics were apparently expelled where extensive recrystallization of the chert occurred. Also entrapped organics were released as silica dissolved along stylolite seams which are more common in the argillaceous chert. Stylolitic solution was early and continued during and after the periods of fracturing. Hydrocarbons were moving actively during the time of fracturing. Some of the fractures acted as channelways for hydrocarbon which bled out of the chert matrix. Hairline fractures originating in clean chert but tapping organic rich chert are lined with fossil hydrocarbon. Hydrocarbon migrating into chalcedony produced a golden brown color. This chalcedony is white where truncating veins had sealed off the chalcedony preventing hydrocarbon migration.

A dark organic filling occurs in the geode-like vugs in the chert which remained open after quartz had grown in the vugs. Fractures with coarse mosaic quartz were important avenues for movement of hydrocarbons as organic material is commonly included in the quartz or fills residual openings along the fractures. Late migration of hydrocarbon is indicated by the dark organic material in the youngest carbonate veins. Golden brown organic material occurs in some of the molds resulting from late carbonate leaching.

A summary of the relative time of fracturing, mineral replacement and migration of organic material is given in Figure 5.

CONCLUSIONS

The Huntersville Chert appears to be of primary origin mainly through biogenic activity. The chert has a heterogeneous patchy appearance from organic structures and variations in mineralogy. Locally the chert is silty, argillaceous and/or dolomitic. Dark organic material which may be an important *in situ* source of some of the hydrocarbon is common in much of the chert.

Pores occur in canals of spicules and apparently within the crypto-crystalline quartz aggregates in the low index chert. Voids are present in vugs and cracks resulting from shrinkage of the silica. Leaching of calcite occurred on a small scale. Fractures developed during several stages. Where mineralization was incomplete along these openings significant secondary porosity remained. Organic matter along the fractures indicated that they were important in the migration of hydrocarbon. The fractures facilitated the bleeding off of hydrocarbon from organic matter in the matrix of the

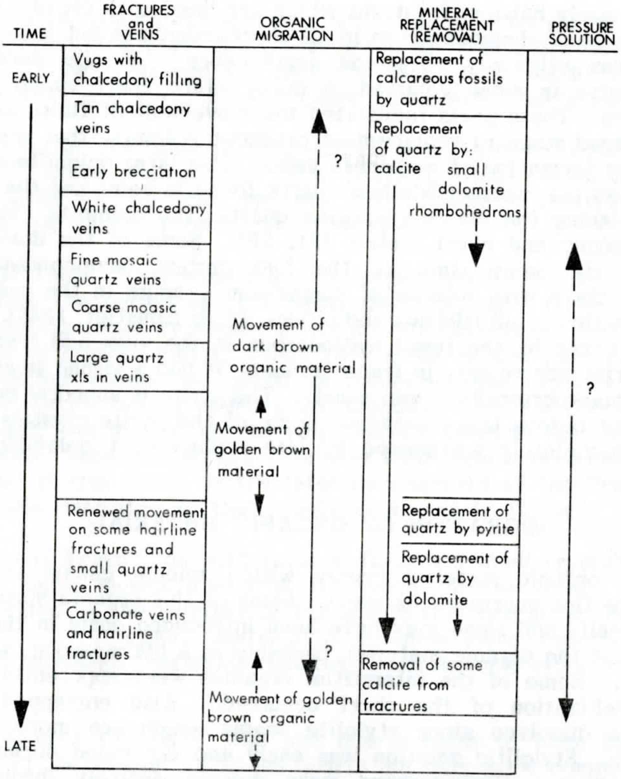

Figure 5. Chronology of fracturing, vein formation, mineral replacement and migration of organic material.

chert and permitted entry of hydrocarbons from nearby beds. The most favorable areas for secondary porosity are in the purer cherts which were more brittle and fractured readily.

ACKNOWLEDGMENTS

Grateful appreciation is extended to Consolidated Gas Supply Corporation for financial support for this study. The writers express their gratitude to John M. Dennison and J. Fred Read for critical review of the manuscript.

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MOLYBDENUM OCCURRENCE IN THE SALISBURY PLUTON, ROWAN COUNTY, NORTH CAROLINA

BRIAN L. ENGLISH *Department of Geology, University of North Carolina, Chapel Hill, North Carolina 27514*

ABSTRACT

Molybdenite occurs in three approximately 400 m.y.-old granitic plutons of the Charlotte Belt. A new occurrence associated with a possible trace element fractionation center located in the northeast portion of the Salisbury pluton, North Carolina, is reported here. The molybdenite, which constitutes less than 0.025 percent of the sampled rock, occurs as disseminated scales in the adamellite and as localized aggregates in a shear zone. The low initial $^{87}Sr/^{86}Sr$ suggests the magma originated in the upper mantle/lower crust. Fractionation of the incompatible elements concentrated the molybdenum in a late stage magma.

The Salisbury pluton, Newell prospect, and Woodleaf quarry molybdenite occurrences suggest molybdenum mineralization may be present in other granitic intrusions of the 400 m.y.-old, synmetamorphic plutonic group.

INTRODUCTION

The Salisbury pluton is one of the most studied granitic plutons in North Carolina. Numerous dimension stone quarries afford high quality exposures and samples for petrographic and geochemical analysis. Previous studies include modal analysis (Chayes, 1950-1951), mapping and petrography (Phillips, 1967), geochemistry (Lemmon, 1969), Rb-Sr isotope geochemistry (Fullagar, 1971), and a summary paper (Fullagar and others, 1971). Butler and Ragland (1969) show the pluton to be critical in interpreting the geological history of plutonic activity in the southeastern Piedmont. The molybdenum occurrence reported here was noted in the course of my research on zircon morphology and U-Pb geochronology of the Salisbury pluton.

GEOLOGIC SETTING

The Salisbury pluton, located in Rowan County, North Carolina (Figure 1), lies within the Charlotte belt of the Piedmont province. The Charlotte belt consists of regionally metamorphosed sedimentary and volcanic rocks with northeast-striking schistosity (Phillips, 1967), intruded by numerous plutons ranging in composition from felsic to ultramafic. The Charlotte belt is a zone of moderate metamorphic grade between two belts of lower-grade rocks (King, 1955). The Salisbury pluton, which forms a topographic high throughout its outcrop area, has a generally concordant relationship with the surrounding metasedimentary rocks. The pluton measures approximately 15 km by 3 km with a total outcrop area of 47 km^2 (Phillips, 1967). In most areas, the thin overburden is a major factor contributing to the presence of numerous quarry operations in the area (Councill, 1954).

PETROLOGY AND GEOCHEMISTRY

The Salisbury pluton consists of pink and white albite adamellite (Figure 2), with quartz, plagioclase, and potassium feldspar constituting 93-98 percent of the rock by volume (Phillips, 1967). Lemmon (1969) reported the pink coloration is imparted by the plagioclase feldspar which has a higher Fe content than the plagioclase in the white adamellite. Contacts between the pink and white varieties are gradational.

Phillips (1967) reported three main textural varieties present in the pluton: relict hypidiomorphic granular; mortar; and mylonitic. Areas in the northeast and southeast portions of the pluton contain rocks with a mortar texture. A texture intermediate between relict hypidiomorphic granular and mylonitic predominates throughout the remainder of the pluton. Phillips attributed these textural varieties to varying degrees

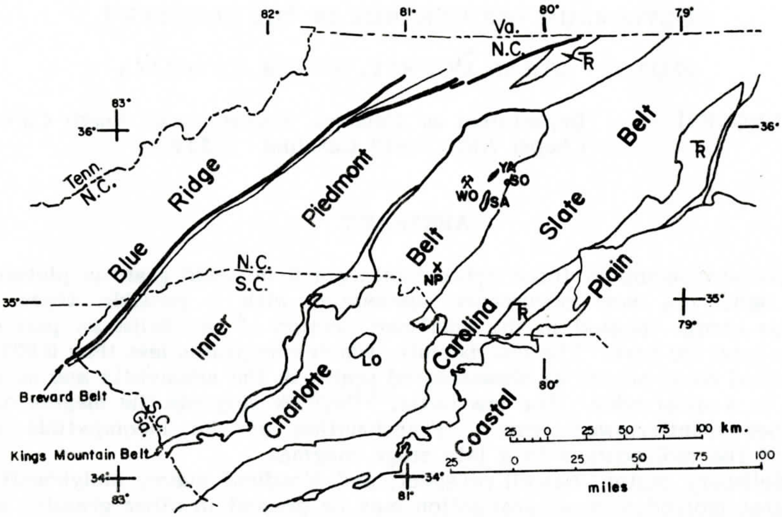

Figure 1. Geologic setting and location of the Salisbury (SA), Lowrys (LO), Southmont (SO), and Yadkin (YA) plutons, and Newell prospect (NP), and the Woodleaf Quarry (WO). Map modified from Fullagar (1971).

of deformation and metamorphism.

The following petrographic observations are modified from Phillips (1967). Two generations of quartz are present in the Salisbury pluton. The first (type I) occurs in large, often fractured, grains (3 mm-6 mm) with highly undulatory extinction. The second generation (type II) exists as an interlocking mosaic of fine grains (average 0.1 mm) bordering the larger minerals, or as linear bands in samples with mylonitic texture. Plagioclase is present as unzoned laths that frequently are bent or broken. Anhedral K-feldspar occurs as flame perthite or as patch perthite. Accessory minerals can be divided into two groups: (I) apatite, magnetite, monazite, rutile, sphene, zircon, first generation ore minerals, some biotite, and some chlorite, which formed before granulation; and (II) epidote, fluorite, garnet, staurolite (?), second generation ore minerals, some biotite, and some chlorite, which formed at the same time or after type II quartz and therefore are considered to be metamorphic minerals.

Fullagar (1971) classified the Salisbury pluton as being synmetamorphic with an Rb-Sr whole rock age of 406 +/- 6 m.y. (recalculated using) λRb = 1.42 x 10^{-11}/yr; Steiger and Jäger, 1977) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7031 +/- 0.0010. The low initial $^{87}Sr/^{86}Sr$ indicates that the Salisbury magma fractionated from the upper mantle or lower crust and was not contaminated by ancient sialic crustal components (Fullagar and others, 1971). Butler and Ragland (1969) report Salisbury pluton samples plot at the extreme end of a metavolcanic differentiation trend. However, the metavolcanics are now known to be significantly older than the Salisbury pluton (Butler and Fullagar, 1978).

Lemmon (1969) and Fullagar and others (1971) discuss the crystallization history of the Salisbury pluton, based on major and trace element analyses. Major element analyses indicate the pluton underwent "eutectic" crystallization with crystallization starting around centers in the south-central and northern parts of the pluton and proceeding outward. Iron and trace element analyses contradict the "eutectic" crystallization model, indicating the pluton underwent fractional crystallization from the margins inward. Increased concentrations of the incompatible elements Nb, Th, U, Y, and Zr are present in the south-central and northern portions of the pluton. Fullagar and others (1971) suggests this conflict could be resolved if the pluton underwent late-stage autometasomatism.

MINERALIZATION

Two distinct types of mineralization are observed in the Salisbury pluton. Gold-

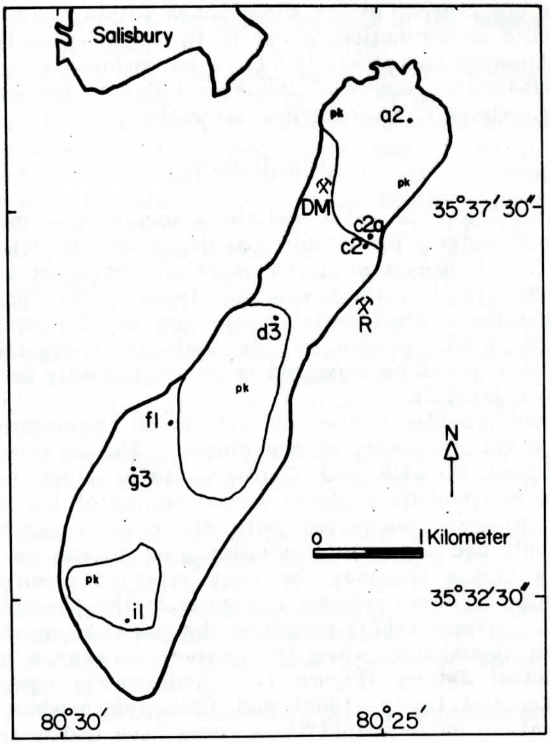

Figure 2. Map of the Salisbury pluton showing collection localities, the Reimer (East Shaft) (R), and Dunn Mountain (DM) mines, and the distribution of pink adamellite (pk). Base map from Phillips (1967).

bearing quartz veins at the Dunn Mountain and Reimer mines (Figure 2) are described by Kerr and Hanna (1893) and Nitze and Hanna (1896). The gold mineralization is strongly localized in shear zones and apparently younger than 368 m.y. (Butler and Fullagar, 1978). The second type of mineralization, molybdenite disseminations, is described here.

Molybdenite was noted in the non-magnetic, heavy mineral (density greater than 3.35) separate for samples collected at the Crystal Pink Quarry, located off SR 2132, in the northeast portion of the pluton (Locality A-2; Figure 2). Mineral separation using a Wilfley table, heavy liquids, and a Frantz magnetic separator indicate that molybdenite constitutes up to 0.025 percent of the samples.

Sample A-2 consists of light pink adamellite which displays mortar texture in thin section. The opaque minerals, in decreasing order of abundance, are magnetite, pyrite, molybdenite and arsenopyrite. Two generations of disseminated sulfide minerals are present. The first, associated with the type I quartz, consists of pitted and embayed pyrite grains (up to 0.8 mm across) and individual scales of molybdenite (up to 2.0 mm across). The second generation sulfide minerals, associated with the type II quartz, consist of euhedral pyrite grains (up to 0.3 mm across), subhedral arsenopyrite (up to 0.4 mm across), and molybdenite scales (up to 0.2 mm across). There is no evidence of extensive hydrothermal alteration.

Sample A-2-S consists of a mylonitized shear zone, less than 4 cm wide, that shows extensive alteration. The primary minerals are albite, quartz, epidote, muscovite, and group II chlorite; K-feldspar is absent. Hematite, limonite and manganese oxide staining is pervasive and some albite grains have weathered to kaolinite. Sulfide minerals consist of pyrite and molybdenite. Pyrite occurs as disseminated grains (up to 0.5 mm across) and as very fine-grained delicate rinds on some quartz grains. Molybdenite occurs disseminated as fine-grained aggregates. Anecdotal evidence from the quarrymen suggest molybdenite aggregates several centimeters across are present in the shear zones within the Central Pink Quarry.

The molybdenite disseminations occur in the northern portion of the pluton, the area observed by Fullagar and others (1971) to be enriched in the incompatible trace elements. A reconnaissance search of this area failed to find evidence of stockwork, sulfide veinlets, or extensive hydrothermal alteration.

DISCUSSION

The low initial $^{87}Sr/^{86}Sr$ ratio and the association of molybdenum with other incompatible elements suggest fractionation as the genetic model. Moorbath and others (1967) and Clark (1971) discuss an upper mantle/lower crust source for the metals associated with porphyry deposits having low initial $^{87}Sr/^{86}Sr$ ratios. Assuming a similar upper mantle/lower crust metal source for the Salisbury pluton, the incompatible molybdenum would remain in the melt until crystallization was almost complete. Molybdenite would be expected to occur primarily as flakes with scattered patches of aggregate crystals.

This occurrence provides further constraints on the nature of the metasomatic event that affected the chemistry of the pluton. The absence of typical porphyry deposit characteristics (Lowell and Guilbert, 1970; Hollister and others, 1974) associated with the molybdenite suggests second boiling of the magma did not occur. Further, the ore minerals associated with the type I quartz indicate that the incompatible elements had begun to crystallize prior to the metasomatic event.

The Salisbury pluton becomes the third reported occurrence of molybdenite associated with metamorphosed granitic intrusions in the Charlotte belt of the North Carolina Piedmont. Privett (1974) mentions the presence of molybdenite associated with zeolite-bearing assemblages along the contact between a diorite and a granitic stock in the Woodleaf Quarry (Figure 1). Radiometric ages are unavailable but petrography (Watson and Laney, 1906) and field relationships (J.R. Butler, 1983, personal communication) indicate that these rocks have undergone at least one period of metamorphism. Worthington and Lutz (1975) describe the Newell prospect (Figure 1) as a porphyry type deposit. They report the host rock to be a quartz monzonite and the K-Ar biotite age of the primary mineralization to be 417 +/- 15 m.y.

Speer (1978) documented a series of molybdenite occurrences associated with the 300 m.y.-old post-metamorphic granitic intrusions of the Piedmont Province. Based on data presented here, I speculate that a similar series of occurrences may exist associated with the 400 m.y.-old synmetamorphic granitic intrusions of the Charlotte belt. Geochemical analysis for molybdenum should be conducted on the Lowrys, Southmont and Yadkin plutons to test this hypothesis.

ACKNOWLEDGEMENTS

This author wishes to express his gratitude to the L. S. Starrett Company, operators of the Crystal Pink Quarry, for their permission to conduct field work at the quarry site. Special thanks are due W. E. Lefler, the quarry foreman, for his professional assistance and pleasant company on a cold and rainy day.

This author appreciates the thorough and incisive reviews by J. R. Butler and P. D. Fullagar. The author is also grateful to P. G. Feiss for many informative discussions during the course of this research.

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THE TURNERSBURG INTRUSIVE: PETROGENESIS OF A METAMORPHOSED ALPINE ULTRAMAFITE IN THE EASTERN INNER PIEDMONT IREDELL COUNTY, NORTH CAROLINA

DONALD R. PRIVETT *One Circle Street, Great Falls, South Carolina 29055*

ABSTRACT

A metamorphosed alpine ultramafic intrusion crops out in the eastern Inner Piedmont belt in north central Iredell County, approximately 13 km northwest of Statesville, North Carolina.

This rock body is approximately 5.7 km long by 1.6 km wide and is concordantly intrusive into a complex of amphibolites, diorites, hornblende gneisses and granite gneisses. Alignment of layers within the ultramafite is N. 35° W.

The ultramafic rock is composed of massive talc, anthophyllite-talc-actinolite, actinolite-anthophyllite, talc-pyrite and hornblende-talc assemblages. The most common lithology is a white to light green massive soapstone which in some outcrops contains 5-10% idioblastic pyrite cubes (5-8 mm) in a matrix of fine grained xenoblastic talc (1.0 mm) and accessory actinolite needles (2.0 mm).

Unusual zoned speroidal structures (0.4 to 1.0 meter) are present in one outcrop. They are encased in vermiculite and have a 2.5 cm thick outer rim of acicular talc and long slender green actinolite crystals that inwardly change abruptly to massive soapstone, then to a central core composed of coarse grained granular to acicular talc and actinolite. The centers weather more rapidly than the margins thus forming a central depression.

Mineralogy, textures, and chemistry suggest that a pyroxenite or peridotite was emplaced and subsequently metamorphosed and metasomatized twice; first to the upper amphibolite facies followed by retrogression to the greenschist facies.

INTRODUCTION

An elongate, elliptical alpine ultramafic intrusion crops out in the diorites, hornblende and granite gneisses, and mica schists of the eastern Inner Piedmont belt of north central Iredell County, 13 km northeast of Statesville and 1.6 km southeast of Turnersburg, North Carolina (Figure 1), and 71 km north of Charlotte, North Carolina.

The purpose of this study is to present mineralogical and petrographic data in support of a proposed geologic history for this altered ultramafite.

GEOLOGIC SETTING

In the southern crystalline Appalachians, the Inner Piedmont (Figure 1) of North Carolina is an 80-100 km wide; fault-bound allochthonous metamorphic and igneous complex (Rankin, 1975; Cook and others, 1979) it is separated from the Blue Ridge on the northwest by the Brevard and Yadkin faults (Espenshade, 1975) and from the Kings Mountain belt and Charlotte belt on the east by the Kings Mountain Shear Zone (Horton, 1981) and the Eufola fault (?) (Milton, 1981). Since pioneering work of Overstreet and Griffiths (1955) and Griffin (1971), our knowledge has been greatly increased as a result of work by Rankin (1975), Espenshade, and others (1975), Hatcher (1978), Harper and Fullagar (1978), Horton (1981), Clarke (1981), Goldsmith (1981), and Milton (1981).

Folded and refolded westward-transported rootless nappes of layered gneiss, amphibolites, mica schist and sillimanite-mica schist are intruded by concordant granitoid to intermediate metamorphosed intrusives and the scattered ultramafics.

The rocks of the belt appear to be metamorphically zoned with lower grade (kyanite to garnet) flanks and a higher grade (sillimanite) central area. (Griffin, 1971;

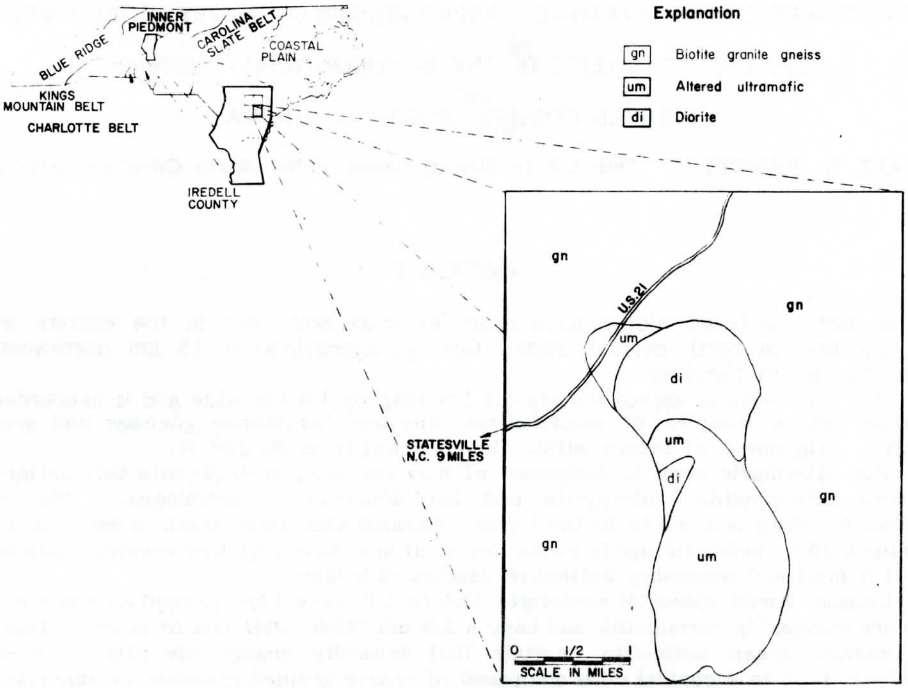

Figure 1. Location and generalized geologic map of the southeastern quarter of the Harmony 7 1/2 quadrangle showing the Turnersburg ultramafic intrusion and host rocks.

Goldsmith, 1981). Zonation is complex and varies along strike and across the belt. The metamorphic thermal maxium occurred 450 - 410 m.y. ago (Butler, 1972).

In their comprehensive review Misra and Keller (1978) grouped the Appalachian ultramafic and mafic rocks into two belts; a well-defined belt along the axis of the Blue Ridge and a poorly defined more irregular belt in the Piedmont. They also summarized the problems of origin of the Piedmont ultramafics concluding that we need more data on the distribution in order to evaluate plate tectonic models and to determine if the ultramafics were emplaced: (1) into a ensialic basin in two stages, initially in the lower crust then into the upper crust during subduction or (2) in an ensimatic back-arc basin later closed by subduction; the ultramafics rising as diapirs.

The petrotectonic history of the Inner Piedmont is still far from being understood. The history and distribution of the ultramafics are vital to the decipherment of the complex history of the Piedmont. Few of the ultramafics have been studied in detail. Recently Bryan and Griffin (1981) described three small intrusions in the South Carolina Inner Piedmont.

General Description

The Turnersburg ultramafic complex has never been studied in detail although it was noted by Hunter and others (1942), who described chromite associated with a "serpentinized dunite" located on the south side of the South Fork of the Yadkin River about 2.5 km south of Turnersburg. LeGrand (1954) referred to scattered occurrences of soapstone and coarsely crystalline enstatite rock east of Turnersburg. Larrabee's map (1966) showed an ultramafic rock complex located south of Turnersburg. The author searched that area but was unable to find any outcrops of ultramafic rocks.

Outcrops of the intrusion are poor and are restricted to roadcuts, residual boulders and to exposures adjacent to the South Yadkin River. The ultramafic rocks are intruded into a biotite granite gneiss unit which includes amphibolite, mica schist, diorite and the ultramafic. The upper limit of metamorphism was determined to be kyanite grade in the surrounding pelites; the mineral assemblages in the ultramafic are

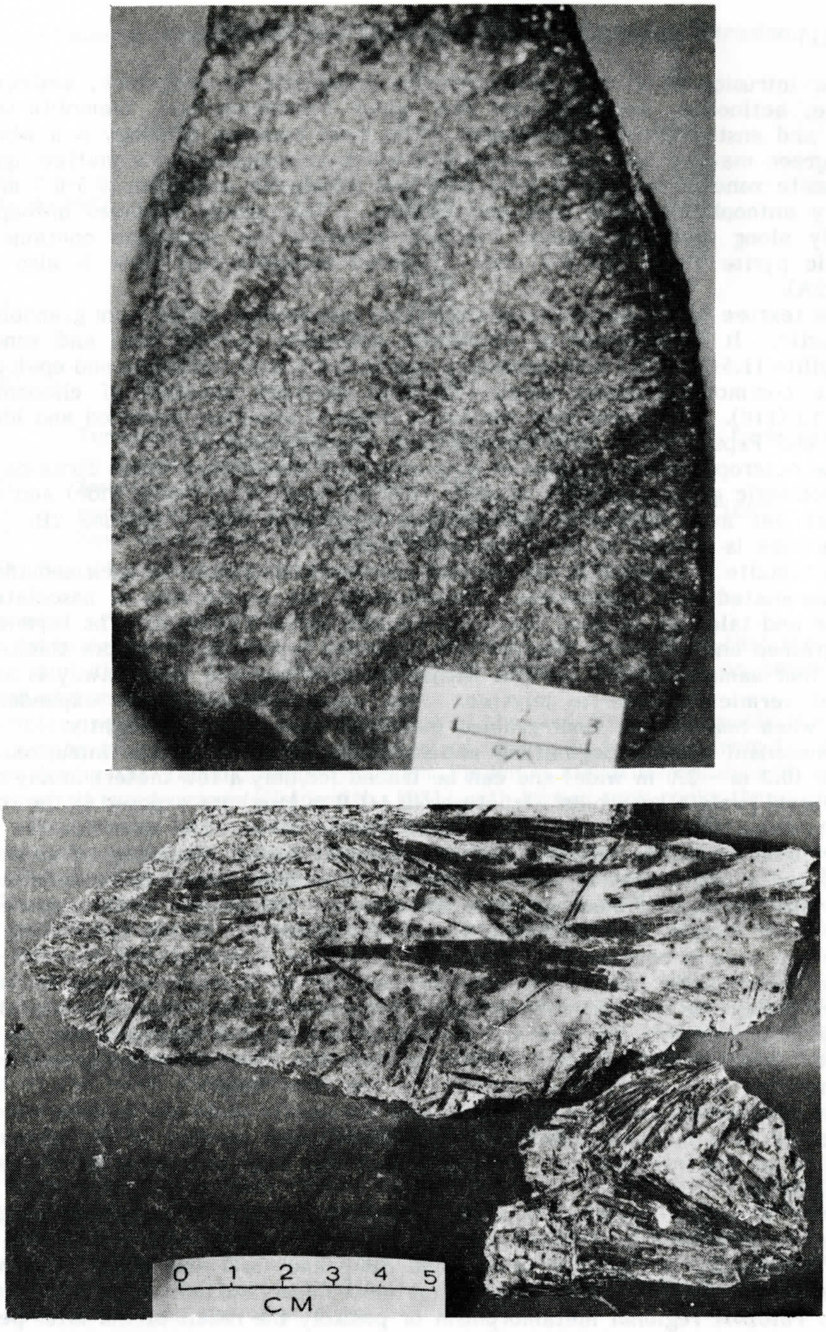

Figure 2. Lithologies and structures in the Turnersburg ultramafic. A. Layered actinolite-talc rock. B. Porphyroblastic actinolite-talc rock.

stable throughout greenschist and amphibolites facies (Evans, 1977). Replacement reactions in the ultramafic can be used to determine relative metamorphic events.

The oval-shaped intrusion trends north-south and is estimated to be 5.7 km long by 1.6 km wide. The strike of the talc-tremolite layers within the ultramafic is N 35^{o} W; and the dip is nearly vertical. If the general map area is correct, the 7.7 km^{2} area of the Turnersburg makes it one of the largest in the Southern Appalachians.

LITHOLOGY AND PETROGRAPHY

The intrusion is a very heterogeneous composite of soapstone, anthophyllite-actinolite, actinolite, actinolite-talc and minor hornblende-talc, tremolite-talc-vermiculite and enstatite-rich assemblages. The most common lithology is a white-grey to pale green massive talc rock (soapstone) which is composed of a matted aggregate of decussate xenoblastic talc (0.2-0.8 mm) and idioblastic actinolite (0.5-0.7 mm) plus accessory anthophyllite (1.0 mm) and epidote. Talc partly replaces anthophyllite, especially along cleavage planes. In one exposure the soapstone contains 5-10% idioblastic pyrite cubes (0.5-0.2 cm). Layered tremolite-talc rock is also present (Figure 2A).

The texture of the actinolite-anthophyllite assemblage varies from granoblastic to lepidoblastic. It is composed of idioblastic actinolite (0.7-3.0 mm) and xenoblastic anthophyllite (1.5-3.0 mm) with accessory talc (0.2-0.3 mm), chromite and epidote. The actinolite commonly contains small (0.8 mm) exsolved lamellae of clinoamphibole parallel to (110). These are very similar to exolved lamellae described and identified by Ross and Papike (1968) as a calcium-poor clinoamphibole.

One outcrop exposes an unusual actinolite-talc assemblage which contains 30-40% large idioblastic green actinolite crystals (2-8 cm long by 2-4 mm wide) and clusters which are set in a matrix of fine-grained white-gray talc. (Figure 2B). Blocky enstatite rock is present in talc-tremolite rocks.

Vermiculite is found near the contact between the ultramafic rock and the gneiss and is associated with pegmatites. Occasionally the vermiculite is associated with tremolite and talc, but in most cases vermiculite alone is present. The vermiculite is coarse-grained and occurs in clusters (1-2 cm), and layers up to 10 cm thick. X-ray data of four samples at the contact with a spheroid and one meter away is an interstratified vermiculite-chlorite mixture. The coarse-grained flakes expanded significantly when heated; the finer-grained materials expanded only slightly.

Concordant granite pegmatites and hornblendite dikes cut the intrusion. These are small (0.3 m - 3.0 m wide) and can be traced for only a few meters in any outcrop.

Unusual prolate spheroidal structures (0.4-1.0 m long) are exposed in the road cuts of County Road 2152 just south of the South Yadkin River. The spheroids are encased in a vermiculite-chlorite mixture and are internally zoned, consisting of an outer rim (2-3 cm) and a larger central core of granular to acicular talc. Some spheroids have an outermost selvage of black granular hornblende. The centers weather more readily than the outer margin, thus producing a central depression. Petrographic study reveals that the rim area is composed of idioblastic actinolite porphyroblasts in a fine-grained matrix of xenoblastic talc. The interior consists of matted xenoblastic talc plus accessory phlogopite and chromite (0.04-0.2 cm) and enstatite. Enstatite was identified optically; it is concentrated near the centers of the altered spheroids and appears to be relict. Scotford and Williams (1983) reported that foliation wraps around enstatite suggesting that it is relict in ultramafics of the Blue Ridge.

The relationship between the diorite and ultramafic could not be determined. The diorite could be a metamorphosed silicic differentiation product of the ultramafic or it could be an old intrusion into the gneiss that was used as access for the ascent of the ultramafic.

When a single chemical analysis (Table 1) of the Turnersburg intrusion is compared to published data from unaltered pyroxenites, the total iron (Fe_2O_3 + FeO) in the Turnersburg intrusion is lower, and SiO_2 and CaO are higher; the result of metasomatism. These changes suggest metamorphism and metasomatism associated with the Palezoic regional metamorphism or possibly the result of the later pegmatite intrusives. The small percentage of Al_2O_3 also suggest an original pyroxenite or peridotite. Al_2O_3 is rather inert during metasomatism; therefore, if Al_2O_3 is low in the altered rock, the percentage in the original rock was likely low.

CONCLUSIONS

The Turnersburg ultramafic is the end product of a long and complex history. Additional geochemical and geophysical data will be required to better characterize the nature of the protolith and the three dimensional character of the intrusion in order

Table I. Chemical Analysis of Talc Rock from the Turnersburg Intrusion.(1)

	I	II	III
SiO_2	56.4	50.5	54.6
TiO_2	0.05	0.53	0.17
Al_2O_3	2.1	4.10	2.19
Fe_2O_3 & FeO	6.5	10.48	9.53
MnO	0.2	0.13	0.15
MgO	22.7	21.71	31.08
CaO	9.1	12.00	1.6
Na_2O	0.03	0.63	0.14
K_2O	0.08	0.21	0.08
L01*	1.94		
TOTAL	99.1	99.29	99.54

I Turnersburg intrusion. R. N. Sales, Analyst, talc rock from exposure on County Road 2152.
II Average pyroxenite, Nockolds, 1954, p. 1022.
III Pyroxenite with rhombic pyroxene, Nockolds, 1954, p. 1022
* Loss on ignition, probably H_2O

(1)Analysis by standard wet chemical methods for silicates.

to determine a more complete history.

A tentative model presented herein is based on the assumption that the enstatite is relict and the low Al_2O_3 content suggests that a pyroxenite or peridotite was emplaced and metamorphosed.

Emplacement and initial metamorphism of the protolith was possibly the result of the subduction of a back arc basin; thrusting of nappes to the west from the root zone occurred during the Taconic orogeny. Inner Piedmont structure consists of a series of stacked and thrusted nappes (Goldsmith, 1981) with higher rank rocks thrust over lower grade. The ultramafic may have been brought up on a thrust. The Kings Mountain shear zone may represent a major tectonic boundary (Nelson, 1981; Horton and Butler, 1981). The shear zone truncates Inner Piedmont nappes detaching them from their root zone (Horton, 1981). Exposures are very poor and faults striking parallel to foliation of the gneiss may not have been detected. The ultramafic is only 10 km west of the Eufola fault (Milton, 1981).

Initial metamorphism produced the talc and anthophyllite and left relict enstatite and chromite; subsequent metamorphism during the Acadian (?) orogeny was responsible for lower grade greenschist metamorphism and pegmatite intrusions producing the actinolite, epidote, vermiculite and chlorite when the Kings Mountain belt was possibly sutured to the Inner Piedmont.

ACKNOWLEDGEMENTS

I appreciate the critical review of this paper by Robert Butler, Department of Geology, University of North Carolina, Chapel Hill, N. C.; N. Telford, Ebasco Corporation, Liberty, North Carolina; R. D. Hatcher, Department of Geology, University of South Carolina, Columbia, S. C., and James L. Carew, Department of Geology, College of Charleston, Charleston, S. C., and Kim Griffin of H. H. Abernathy, Jr.'s office who typed the final copy.

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