STATE OF ARKANSAS
ARKANSAS GEOLOGICAL COMMISSION
Norman F. Williams, State Geologist

MISCELLANEOUS PUBLICATION 18-B

CONTRIBUTIONS
to the
GEOLOGY of ARKANSAS
VOLUME II

Edited By
John David McFarland, III
William V. Bush

Little Rock, Arkansas
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PREFACE

The preparation of this volume was made possible by the work of several people. J. M. Howard reviewed and assisted in editing C. Milton's miserite paper. B. R. Haley and C. G. Stone gave freely of their time in reviewing, proof reading, and editorial suggestions. L. P. Kelone and Susan Young drafted various figures and prepared the cover. Adrian Hunter and Sheila Curd printed, collated, and bound this volume. Our thanks goes to all of these people and the rest of the staff of the Arkansas Geological Commission for their assistance. A very special thanks goes to Loretta Chase who typed and composed the manuscripts and revisions.

John David McFarland, III
William V. Bush
Editors

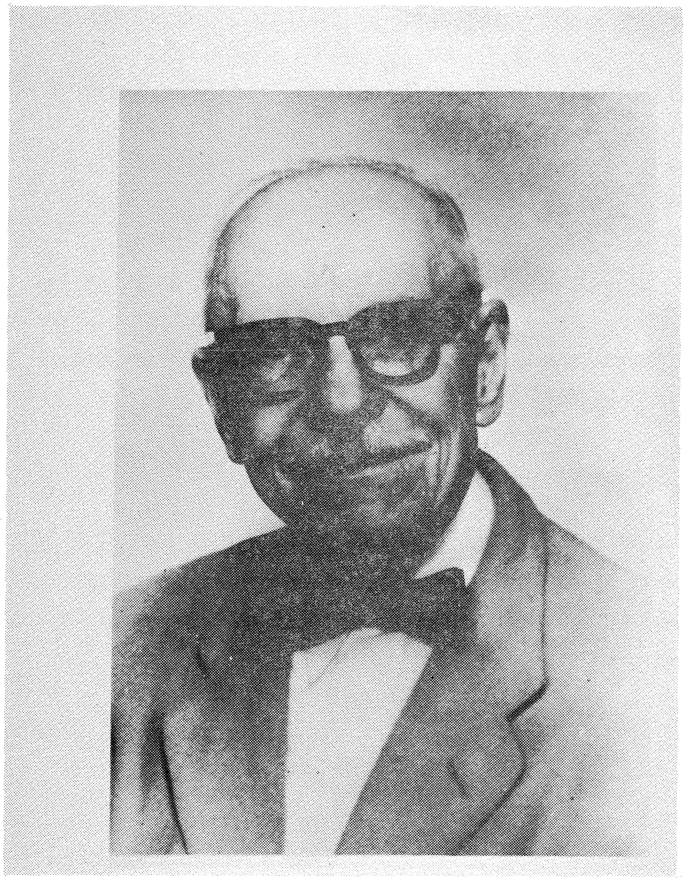
COMMISSIONERS

TABLE OF CONTENTS

J. Francis Williams
1862 – 1891

DEDICATION

It is both fitting and proper that this volume be dedicated to three outstanding geologists, J. Francis Williams, Waldemar T. Schaller, and Hugh D. Miser, who through their initial work on the geology and mineralogy of Arkansas, laid the foundation on which the present geologic framework is being constructed. Their scientific dedication, enthusiasm, careful attention to detail, and ability to put together "the big picture" is a lesson to all concerned with furthering geologic knowledge. Not only should these men be recognized for their scientific endeavors, but also for the goodwill and high esteem of the state of Arkansas which they spread wherever they traveled. Arkansas is proud to have these "pioneers" of modern geologic thought counted in her ranks. The following biographical sketches were written by Dr. Charles Milton, a personal friend of two of these individuals.



Waldemar T. Schaller
1882 – 1967

Hugh Dinsmore Miser
1884 – 1969

John Francis Williams
1862 – 1891

In his brief 29 years of life John Francis Williams achieved undying fame for his pioneer contributions to Arkansas mineralogy and petrology, culminating in the monumental work on The Igneous Rocks of Arkansas. Three years before his death, he was one of the thirteen founders who in Ithaca, New York, organized the Geological Society of America.

He was born in Salem, some 40 miles northeast of Troy in upper New York; and attended schools there and in New Hampshire; he received a degree in civil engineering from the Rensselaer Polytechnic Institute at Troy. In 1881 he traveled in Norway and Sweden, and studied mineralogy and petrography at the University of Gottingen in Germany. In 1883 he did railroad engineering work in Vermont, later becoming assistant in chemistry and natural science at his alma mater. There under the guidance of Professor Henry B. Nason he studied the slates of the region, publishing his first report in 1884; the following year he received the B.S. degree from Rensselaer. In 1885 he traveled in Italy and Sicily; and his doctoral dissertation dealt with volcanic rocks in Italy, published (in German) in 1887. That year he returned to the United States, becoming director of the technical museum of the Pratt Institute of Brooklyn, rich in mineral and rock collections. In 1889 he became honorary fellow and then instructor ("docent") at Clark University in Worcester, Massachusetts. There he was asked by Professor J. C. Branner to study the igneous rocks of Arkansas. The following year he (Williams, 1890) published a study of two Arkansas minerals, manganopectolite and eudialyte. He had worked himself to the limit, and beyond it, weakened by malaria contracted in the swamps of Arkansas. According to contemporary accounts, his last efforts were devoted to completion of The Igneous Rocks of Arkansas, one of the great classics of American geology. Professor J. F. Kemp concluded his memorial (Kemp, 1891) from which these notes are taken, with the words, "It has never been the lot of the writer to know intimately a more generous, frank, and lovable man than J. Francis Williams, and it is impossible to speak of him without the deepest emotion. His character was such as to indescribably endear him to his friends, and his abilities and preparation for his work were of the highest order. His results were such as to secure him in all the future one of the most honorable places in the records of American geological science."

Waldemar Theodore Schaller
1882 – 1967

Waldemar T. Schaller, dean of U. S. Geological Survey mineralogists, was born in Oakland, California. In 1903, with a B. S. degree from the University of California, where he studied geology and mineralogy under Professors Lawson and Eakle, and analytical chemistry under Professor Blasdale, he joined the Geological Survey as Assistant Chemist. In 1912 he traveled to Europe, where he received the Ph.D. degree from the University of Munich. In his long career with the Geological Survey, he did distinguished research in many fields, among them pegmatites and saline deposits. He received many major honors from geological societies here and abroad; he published nearly 300 papers, describing over 40 new mineral species.

He was friendly and generous, always ready to give his time and talents to the problems of others. Through Clarence S. Ross, he became interested in Arkansas mineralogy, and so advanced the pioneer studies of J. Francis Williams on miserite.

Hugh D. Miser
1884 – 1969

Hugh Dinsmore Miser was born at Pea Ridge, Arkansas, and educated in the schools of that state; at the University of Arkansas he took his degree in geology. His first employment was with the U. S. Geological Survey, and after brief interruptions to serve the states of Arkansas (as Acting State Geologist) and Tennessee (as State Geologist) he remained with the U. S. Geological Survey. There hundreds of young geologists found inspiration under his leadership. The geological maps of Oklahoma and Arkansas, and over eighty geological reports from diverse localities in the United States, were but one phase of his varied activities; he was not only a leading authority on the geology of central North America, but was recognized internationally as one of the world's great petroleum geologists. He was active in very many scientific organizations, and in 1955 received the Department of the Interior's highest honor, the Distinguished Service Award, in recognition of his outstanding career not only as a scientist and administrator, but as a citizen and leader. "His warmth of personality, his love for his fellowmen, and his life dedicated to geology, won the respect and affection of all who knew him. When he passed away, we lost a wonderful friend, and we shall miss him very much." (Cohee, 1969)

Editors note: H. D. Miser worked for the Arkansas Geological Survey for one month before going on to the U.S.G.S.. He received a salary of $1.00 per day from Arkansas, but was raised to $2.00 per day when he went with the federal survey.

SILURIAN STRATIGRAPHY OF THE ARKANSAS OZARKS

by

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ABSTRACT

Rocks of Silurian age in the Arkansas Ozarks have been assigned to four lithostratigraphic units. In ascending order, these are: the Brassfield Limestone, the Cason Shale, and the St. Clair and Lafferty Limestones. Only the upper part of the Cason, the so-called "button-shale" of authors, is Silurian. The lower part of the shale section traditionally assigned to the Cason is stratigraphically below the Brassfield and is Ordovician in age.

Ozark Silurian strata have a composite thickness of about 250 feet, but only a little over 100 feet is preserved at any locality. Erosion during and after their deposition has given these strata an irregular distribution. They are restricted to three general outcrop regions: an eastern region in Independence and Izard Counties near Batesville; a central region in Stone County in the vicinity of Blanchard Spring; and a western region centered around Gilbert in Searcy County.

The Brassfield appears to rest unconformably on older strata. It is dominated by poorly washed biosparite (wackestone-packstone) with subordinate biomicrite (wackestone) and biosparite (grainstone). Allochems are comprised almost entirely of skeletal remains. Crinozoans are most abundant, but significant amounts of other common Paleozoic fossil groups are also present. The limestone represents deposition in a semi-protected, subtital environment.

The Cason "button shale" rests unconformably on the Brassfield and older strata. It is a phosphate pebble-bearing, silty, dolomitic clay-shale characterized by oncolite-like algal growths that were originally spherical in shape but have been flattened parallel to bedding by post-depositional compaction. The unit becomes calcareous upward and grades conformably into the base of the overlying St. Clair Limestone.

The St. Clair contains textures much like those found in the Brassfield, and the two limestones are difficult to separate where the "button shale" is absent. The base of the St. Clair is characteristically an ostracode-rich biomicrite, above which occurs crinozoan-rich biosparite and poorly washed biosparite. The St. Clair grades conformably into the Lafferty, which is an ostracode biomicrite that is lithically similar to the base of the St. Clair.

The Cason "button shale" is the detrital base of a transgressive-regressive cycle that includes the St. Clair and Lafferty Limestones. The middle, coarse-grained St. Clair represents a portion of a crinozoan barrier sand that sheltered on its leeside a quiet-water lagoonal environment in which the basal St. Clair biomicrite (transgressive phase) and Lafferty (regressive phase) accumulated.

INTRODUCTION

Rocks of Silurian age in the Arkansas Ozarks have been assigned to four lithostratigraphic units. From oldest to youngest, these are: the Brassfield Limestone, the Cason Shale, and the St. Clair and Lafferty Limestones. Confusion about the internal lithic succession of the Cason interval, and the age assignment of its different units, has been outlined earlier (Amsden, 1968; Craig, 1969, 1975; Wise and Caplan, 1979) and will be commented on below. Only the uppermost Cason, the so-called "button shale" of authors, is Silurian. It will be the only portion of the Cason discussed here.

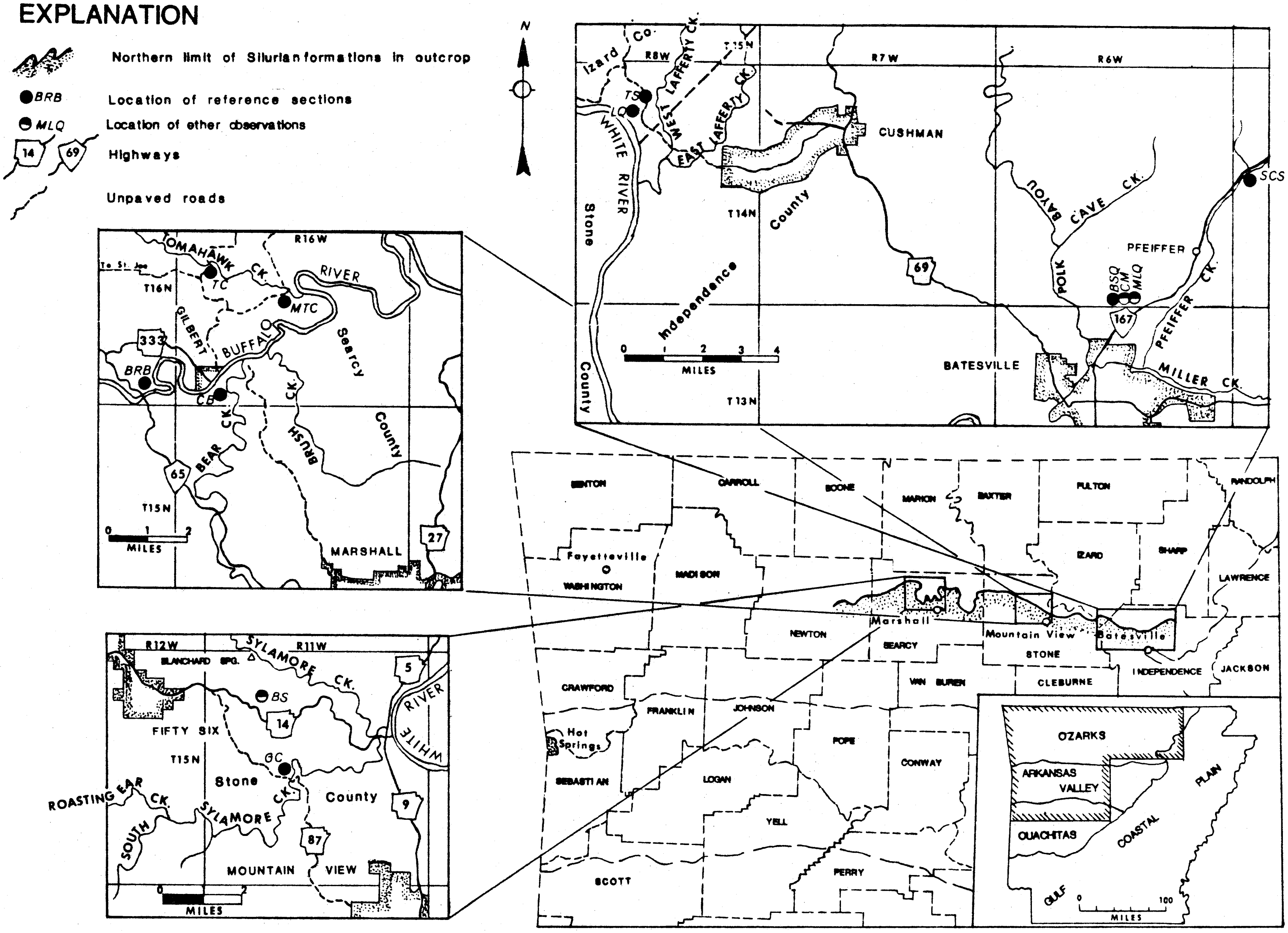
These Silurian strata crop out in a discontinuous belt that extends from their easternmost exposures in the Batesville district, Independence County, west to the vicinity of Snowball in Searcy County (Haley, et al., 1976). They are unconformably underlain by phosphatic sandstone and shale traditionally assigned to the Cason Shale, or where these beds are absent, by the Fernvale Limestone. Both of these units are Upper Ordovician. They are unconformably overlain almost everywhere by the Boone Formation, but in a few areas of limited extent the Penters Chert (Lower Devonian) or the Sylamore Member (Upper Devonian) of the Chattanooga Shale is the unconformable superjacent unit.

Within this east-west tract, occurrences of Silurian rocks are restricted to three general areas, between which they have been removed entirely by erosion prior to the deposition of the Boone Formation (Fig. 1). The easternmost of these areas is the aforementioned Batesville district, in which the Silurian crops out from St. Clair Spring, five miles north of Batesville, west to the vicinity of Guion on the White River in Izard County. The Brassfield is known from only a few localities in the district, but the Cason "button shale" and St. Clair and Lafferty Limestones are widespread. The middle region of Silurian outcrop is in the vicinity of Blanchard Spring in Stone County. No Brassfield is known from this area, and the St. Clair and Lafferty occur only at isolated localities within Blanchard Spring Recreation Area and to the south along South Sylamore and Roasting Ear Creeks. The Cason "button shale" is known from only one locality in Blanchard Spring Recreation Area. The westernmost region of outcrop is centered on the community of Gilbert on the Buffalo River in Searcy County. Outcrops of Silurian extend from about six miles east of Gilbert west to where the Searcy-Newton county line crosses the Buffalo River north of the community of Snowball. The best exposures of Brassfield are in this westernmost region, where it is confined to the vicinity of Gilbert. The St. Clair and Lafferty are more extensive, but the Cason "button shale", whose westernmost occurrence is at Blanchard Spring, is absent. All Cason Shale mapped in the Gilbert area is equivalent to the Ordovician age phosphatic beds of the Batesville district.

In recent years, different authors (Maher and Lantz,1952, 1953a, 1953b; Frezon and Glick, 1959; Glick and Frezon,1965; Wise and Caplan, 1967, 1979: Craig, 1969; Sgouras, 1979) have included information on these Silurian strata in reports that were either broad in scope and/or confined to the rocks of one of the three outcrop regions. The purpose of this report is to compare the Silurian stratigraphy between the three outcrop regions and to identify and provide lithic detail on reference sections that illustrate this stratigraphy (Fig. 2).

I gratefully acknowledge Norman F. Williams, Director of the Arkansas Geological Commission, for the logistical support and many courtesies extended to me and my students over the years. I am equally appreciative of the reception that our group has received from his geological staff, particularly Orville A. Wise, whose generous exchange of knowledge

FIGURE 1. Location of Silurian reference sections, Arkansas Ozarks



on Ozark geology, both in and out of the field, has added much to the conclusions presented here.

THE CASON PROBLEM

The Cason Shale was named by H. S. Williams (1894) for an exposure of red and green shale containing abundant algal "buttons" at the Cason Mine just north of Batesville (CM, Fig. 1). The "buttons" are disc-shaped objects an inch or so across. E. O. Ulrich (in Miser, 1922) identified them as algal in origin and referred them to *Girvenella richmondensis,* a guide to the Richmond Group of the Cincinnati region. Many of the "buttons" at the type are replaced by manganese oxide, which has all but obliterated their internal details. Some, however, have retained a relict cellular structure that suggests blue-green algae (Plate 1, Figs. A, B); they are almost certainly not *Girvenella.*

At its type, the Cason occurs between the Upper Ordovician Fernvale Limestone (Polk Bayou Limestone of Williams' nomenclature) and the Middle Silurian St. Clair Limestone. Based on meager fossil collections, Williams (1894) contended that the Cason contains the beginning of the overlying St. Clair macrofauna and thus is Niagaran in age. Conodonts from the "button shale" support this assignment (Craig, 1969). Ulrich, whose opinion has prevailed, assigned the unit to the Upper Ordovician, mainly based on the presence of the *"Girvenella* buttons". He did this in spite of the occurrence of similar "buttons", different only in that they are spherical rather than flattened, in the base of the overlying St. Clair.

Purdue (1907), in a study of the phosphate deposits along East and West Lafferty Creeks in the western part of the Batesville district, extended the name Cason to a sequence of phosphatic, conglomeratic sandstone and shale occurring above the Fernvale Limestone. Miser (1922), in his landmark study of the geology of the district, followed Purdue's lead and reported the Cason as "shale, sandstone, and phosphate rock". Although he reported "buttons" from many localities, he made no distinction between the "button" -bearing shale like that of the type Cason and the characteristically phosphatic beds included in the formation by Purdue.

During Miser's study, a development occurred that until recently obscured the correct interpretation of the physical relationships between, and the true stratigraphic position of, the different lithic types now known to be present in the Cason interval. Miser made a collection of macrofossils (manganese replaced) from a residual soil above the Cason Shale at the Montgomery Mine north of Batesville. Ulrich identified these fossils as characteristic of the Brassfield Limestone, with which he was familiar from outcrops in the Gilbert area to the west. Inasmuch as the Brassfield had not been found in outcrop in the rest of the Batesville district, it was Ulrich's opinion, conceded by Miser, that the limestone was once present in the region, but for the most part had been removed by pre-St. Clair erosion.

Miser's stratigraphic section thus included, in ascending order, the Fernvale Limestone, the Cason Shale, and the Brassfield and St. Clair Limestones. The Cason was considered a variable unit of Late Ordovician age occurring between the Fernvale below, and in the absence of the Brassfield, the St. Clair above. In fact, Miser assigned all rocks within this interval, including some oolitic material and a manganiferous limestone he observed in several places during a later study (Miser, 1941), to the Cason.

The major key to unravelling the complicated internal stratigraphy of the Cason has been an exposure at the Love Hollow Quarry located on Penters Bluff in the western part of

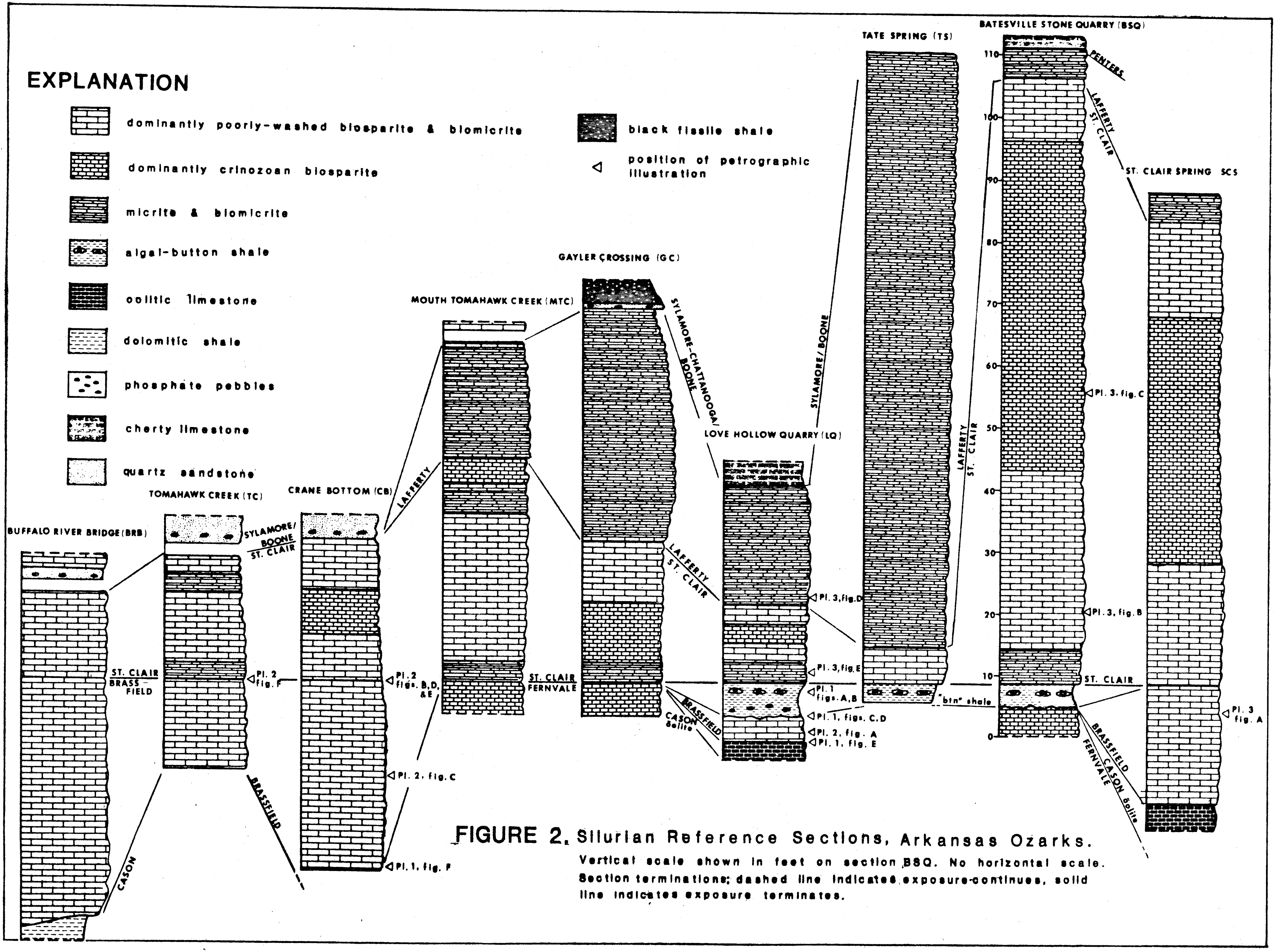


FIGURE 2. Silurian Reference Sections, Arkansas Ozarks.
Vertical scale shown in feet on section BSQ. No horizontal scale.
Section terminations; dashed line indicates exposure-continues, solid line indicates exposure terminates.

the Batesville district (section LQ). Here the Cason interval is bounded, as usual, by the Fernvale and St. Clair. During the 1960's, all the major Cason lithic types were exposed in the upper quarry face, allowing examination of their superposition and physical relationships (see Plate 5). The details of the succession of units have been presented elsewhere (Craig, 1969, 1975a) and need not be covered here. Suffice it to say that the exposure clearly showed the "button shale" to be the topmost unit in the Cason and that it rested unconformably on all underlying units. The base of the interval was the phosphatic sandstone and shale, resting unconformably on the Fernvale. A reddish crinozoan limestone, now unequivocably known (Amsden, 1968; Craig, 1969) to be the unit that produced the Brassfield fossils identified by Ulrich from the Montgomery Mine, was present between these two detrital rocks. Unfortunately, quarrying activity in recent years has removed the Brassfield (as well as an oolitic limestone present at the top of the phosphate beds), which was present apparently only as a localized erosional remnant (Plate 5, Fig. B).

This presents the unfortunate stratigraphic situation in which one formal lithostratigraphic unit (Brassfield) is incorporated entirely within another (Cason). The lower phosphatic unit, which contains some well-indurated sandstone layers, is the Cason lithic type most commonly encountered in outcrop. For this reason, the name Cason Shale is most commonly associated with these beds. This is unfortunate because the only unit of the Cason interval present at the Cason Mine is the "button shale". If the name Brassfield is to be retained in Arkansas stratigraphy, then the name Cason needs to be restricted to the "button shale" unit and the lower phosphatic beds and their associated oolitic limestone given a new name. Otherwise, use of the name Brassfield needs to be abandoned in Arkansas, and the entire heterogeneous post-Fernvale-pre- St. Clair interval referred to as the Cason Shale. The former alternative would serve the greatest utility because the rock types now known to occur between the Fernvale and St. Clair are easily distinguishable lithically and a division of them into separate lithostratigraphic units would better reflect what appears to be their rather complicated depositional-erosional history.

STRATIGRAPHY

BRASSFIELD LIMESTONE (LOWER SILURIAN)

The Brassfield Limestone was named by Foerste (1905) for exposures in Madison County, Kentucky. The Brassfield exhibits only broad lithic similarity to the Brassfield as it appears in outcrop in its type area and the surrounding Cincinnati region. The name was extended into northern Arkansas on paleontology by Ulrich (1911, p. 558), who applied it to exposures of coarse-grained limestone between the Cason Shale and St. Clair Limestone in the Gilbert area of outcrop.

The circumstances surrounding the inclusion of the Brassfield in the Batesville district have been given above. Miser (1922) reported that he had not seen the formation in the district, but it's likely that he had and did not recognize it. The best exposure (section SCS) of the Brassfield in the district is low on the hillside 200 yards north of St. Clair Spring. Here approximately 20 feet of the unit crops out between the St. Clair above and the oolitic limestone belonging to the lower Cason below. Miser (1922) seems to have mapped this exposure as Fernvale, and later Straczek and Kinney (1950) apparently mapped it as St. Clair. This mistaken identity is understandable because of the close overall similarity of the bioclastic limestones in the Ozark Paleozoic.

The Brassfield is also present along the western margin of the district as thin, discontinuous erosional remnants. Its occurrence at

Love Hollow Quarry (section LQ) has already been mentioned. The Brassfield is also present at the old phosphate workings on the west valley wall of Lafferty Creek just south of the junction of East and West Lafferty Creeks (PM, Fig. 1). It was recorded from this locality in a section measured by Purdue (1907) as 15 inches of manganiferous iron ore. In his later work in the district, Miser (1941) reported the presence in the Cason of a manganese carbonate up to 3 feet thick between Cushman and Penters Bluff. This limestone is no doubt the Brassfield occurring between the lower phosphatic beds and the "button shale". Miser did not recognize it, of course, because it was contained within rocks assigned to the Cason.

The best exposures of the Brassfield are in the Gilbert area, where it occurs discontinuously along the bluffs of the Buffalo River from the Highway 65 bridge east to the mouths of Tomahawk, Bear, and Brush Creeks, and along the lower stretches of these creeks (sections BRB, CB and TC).

The Brassfield is dominantly a bioclastic limestone of variable texture. Its chief constituent is fragmented crinozoans, but it also contains abundant debris of other common Paleozoic fossil groups. Its color varies from light gray to deep red, the latter occurring in irregular layers and patches that give the rock a mottled aspect in places. This color and its distribution is probably the single most valuable feature in distinguishing the Brassfield from other Ozark Paleozoic bioclastic limestones. Glauconite is a minor constituent, particularly in the lower layers of the formation.

The high percentage of coarse fossil fragments gives the Brassfield the appearance of being reasonably free from fine-grained constituents. This is misleading because the limestone contains an important component of lime mud, which in some zones is admixed with detrital mud that adds the deep red coloration to the rock. The lime mud has three types of distribution in the rock. It occurs as discontinuous stringers or layers up to several inches thick (Plate 1, Fig. F; Plate 2, Fig. C), as mud resting on the surface of shell fragments that floor interstices (Plate 2, Figs. A, E), and as irregular patches, no doubt the result of burrowing, that disrupt normal lamination (Plate 2, Fig. D).

This distribution of lime mud suggests that the Brassfield collected in a semi-protected area where wave energy was not consistently strong enough to remove all the mud that accumulated. In the more protected parts of the area, or during prolonged periods of reduced wave action, stringers and layers were left undisturbed. In more exposed places, or during periods of slightly higher energy, lime mud was washed away or allowed to accumulate only in protected areas provided by irregularities on, within, and between bioclastic grains. Life was abundant, and the infauna moved some of the mud from its original site of deposition and redistributed it into irregular patches and streaks that cut across the sedimentary layering.

This set of conditions developed the Brassfield's most dominant texture, which is a poorly washed and sorted mixed (skeletons from many different faunal groups) biosparite (packstone/grainstone), which along with biomicrite (wackestone/packstone) and mud-free, mixed biosparite (grainstone) comprise most of the Brassfield's textures.

There is a significant, but subordinate, amount of mud-free, coarse-grained crinozoan biosparite (grainstone) in the Brassfield (Plate 3, Fig. A). This rock is close to equigranular and is dominated by fragmented crinozoans. It is apparent that the sediment forming this rock did not result from a cleaning by higher energy of the muddier sediment that was deposited in the semi-protected area, or else it would have a higher content of other skeletal material. It would appear that it matured in wave-dominated shoal zones and was introduced into the semi-protected area of major Brassfield accumulation during periods of higher energy, possibly storms.

Although secondary precipitation on the surface of exposures gives the impression that the Brassfield is thick bedded, it is, on the contrary, a thin-bedded unit. Individual textural types are almost never greater than a foot thick. They rarely separate along bedding planes, probably because of the disruption of normal bedding by burrowing animals. No distinct vertical or lateral trends in lithology seem apparent from the limited exposures of the Brassfield available for study.

The Brassfield ranges in thickness from 0 to a maximum of 38 feet (section BRB; Plate 4, Fig. B). It overlies the Ordovician part of the Cason Shale at all places examined. Because the Cason does not crop out well, the base of the Brassfield is seldom exposed. At the time that the Brassfield was exposed in the face of Love Hollow Quarry (section LQ), it was in contact with the oolitic limestone associated with the phosphatic beds of the Cason, but the contact was stylolitic along the full extent of the exposure (Plate 1, Fig. E). It also overlies the oolitic limestone at section SCS, but the contact is not exposed.

The only locality providing information on the relationship of the Brassfield with the underlying Cason is section BRB. Here the Cason is exposed at the base of the bluff in a reentrant beneath the Brassfield (Plate 4, Fig. B). The contact is obviously uneven, with the Brassfield truncating beds in the Cason. It is probable that this relationship holds true for the base of the limestone throughout northern Arkansas.

The upper contact of the Brassfield with the "button shale", or where the shale is absent, the St. Clair Limestone, has been examined in several places. In fact, both of these contacts have been collected in a single hand specimen (welded contact), and examination in thin section shows micro-irregularity and truncation of Brassfield fossils by the base of the overlying unit (Plate 1, Figs. C, D; Plate 2, Figs. B, D, E). Interpretation of these features, along with the irregular distribution of the limestone, suggests that within its outcrop belt the Brassfield is everywhere unconformably overlain by younger units.

Selected hand specimens of the Brassfield can be indistinguishable from the Fernvale and St. Clair. It is generally necessary to examine a thickness of strata to confidently separate the three limestones in the field. The Fernvale is a clean biosparite throughout its entire extent. It never contains mud stringers and patches that are characteristic of the Brassfield. The St. Clair, on the other hand, is characterized by a lime mud matrix, and texturally it and the Brassfield are very similar. Overall, the St. Clair contains a greater percentage of lime mud and is coarser grained, but the overlap between the two limestones in these characters is significant. The St. Clair does not exhibit the deep-red mottling characteristic of some zones in the Brassfield. The former is rather consistently light-greenish gray in color. The Brassfield-St. Clair contact is sharp, commonly stylolitic, and can be subtle. However, it can be picked confidently in most localities by noting the abrupt change from the biosparite and poorly washed biosparite of the Brassfield to the characteristically micritic base of the St. Clair (Plate 2, Fig. F).

The Brassfield contains abundant white calcite vugs of secondary origin. These features also are present, although less commonly, in the St. Clair. At section TC, both units possess vugs lined with dogtooth spar and cored with a sand that carries a Sylamore conodont fauna (Plate 4, Fig. D). The vugs apparently represent a post-St. Clair - pre-Sylamore period of subaerial exposure and solution that affected both limestones. Devonian fills also occur along vertical solution channels cutting the entire thickness of the St. Clair at section BSQ.

"BUTTON SHALE" OF THE CASON (LOWER/MIDDLE SILURIAN)

As discussed above, only the upper part of the Cason, the so-called "button shale" (type Cason), is Silurian, and thus will be the only part of the unit considered here.

The Cason "button shale" does not crop out well and thus knowledge of it comes primarily from man-made exposures. It has not been identified in the Gilbert area, where all Cason belongs to the phosphatic beds of Ordovician age (Craig and Lemastus, 1979).

The "button shale" has been identified at one locality (BS, Fig. 1) in the Blanchard Spring region (Lemastus, 1979). This is in a ravine paralleling the old road leading up from the spring to Highway 14 (NE 1/4, NE 1/4, Sec. 8, T. 15 N., R. 11 W., Fifty Six quadrangle). The Cason occurs beneath an overhang formed by the basal St. Clair. It consists of about 2 feet of dolomitic, silty shale and shaley micrite that contains phosphate nodules and scattered shells in its upper foot. Although no "buttons" seem to be present, there is great similarity in other details between these rocks and the upper part of the "button shale" in the Batesville district.

The major occurrence of the "button shale" is in the Batesville district. The unit is present throughout the district, but appears to be more continuous in the western part from Cushman west to the bluffs of the White River. It would appear from present knowledge that its occurrence in the eastern part of the district is discontinuous. Thicknesses for the unit are not accurately known, but judging from published reports of geologists who examined the Cason interval at times when it was well-exposed in mine workings, the "button shale" probably does not exceed 20 feet.

The most notable exposure in the eastern part of the district is at the type section of the old Cason Mine, an open pit workings .5 miles north of Batesville (CM, Fig. 1). About 16 feet of the shale is exposed in the mine face and floor. The base of the shale is not exposed, but directly adjacent exposures at section BSQ to the west and the large Midwest Lime Company quarry (MLQ, Fig. 1) to the east show the Fernvale Limestone to be the underlying unit. The Midwest Lime quarry provides a remarkably instructive view of the relationship of the "button shale" to the underlying Fernvale and overlying St. Clair. In the western face of the quarry the "button shale" is in distinct erosional contact with the Fernvale, its material filling fractures and irregularities in the upper portion of the limestone (Plate 4, Fig. E). The shale, which is only a few feet thick here, grades upward into the base of the St. Clair by a decrease in detrital and increase in carbonate material. Abundant "buttons" (uncompressed) are present in the basal few feet of the St. Clair. North along the quarry face the "button shale" and overlying St. Clair climb onto an expanded Fernvale section, the St. Clair eventually onlapping the shale and resting directly on the Fernvale (Plate 5, Fig. C). About half way across the north face of the quarry, the St. Clair wedges out between the Fernvale and the Devonian Penters Chert. This relationship indicates that the "button shale" is a basal detrital phase of the St. Clair transgression. Considerable topography apparently existed, at least locally, on the Fernvale erosional surface, and deposition of detrital material, which was in short supply, was confined to the topographic lows.

The shale at the Cason mine is difficult to decipher petrographically because of the heavy manganese and iron mineralization to which it has been subjected. Its overall appearance is that of a silty to fine-sandy, calcareous, red to red and green mottled clay shale. Dolomite rhombs occur scattered throughout. In addition to algal "buttons", other fossils include fragmented ostracodes, crinozoans, trilobites, and calcareous and phosphatic brachiopods. Calcite cement increases upward toward the St. Clair, with the upper few feet calcareous enough to dissolve in acetic acid. Acid residues from this upper portion produce *Ammodiscus*- and *Psammoshaera*- type agglutinated foraminifers exactly like those in the base of the overlying St. Clair Limestone. Also present in these residues are conodonts characteristic of the basal St. Clair, which corroborates Williams' (1900) conclusion (from brachiopods) that the Cason at the mine contains the beginning of the St. Clair fauna.

Phosphatic pebbles occur in the shale, but not in the abundance that they appear in the lower phosphatic beds of the Cason. However, the presence of phosphate in the "button shale" provides a certain similarity between the upper and lower Cason interval that has no doubt helped obscure the distinction between the two units. Phosphate can be present in any one of several Ozark Lower Paleozoic units (See Horner and Craig, this volume), and thus is not a good criterion for distinction between the upper and lower Cason. The lower phosphatic beds of the Batesville district differ from the "button shale" in that they are distinctly interbedded sandstone and shale in which all fossil debris is phosphatic. They do not contain "buttons" or calcareous shells, nor are they characteristically calcareous or dolomitic.

The cleanest exposure of the "button shale" occurs at the Love Hollow Quarry (section LQ). Although thinner than at the Cason Mine, its boundaries are better exposed and thus easier to interpret. The unit, which averages about 3 feet thick in the upper quarry face, has a sharp, slightly scoured erosional contact with the rocks below (Plate 5, Figs. A, B). This contact is most distinct when the Brassfield, or the oolitic limestone associated with the phosphatic beds, is the underlying unit (Plate 1, Figs. C, D, E). The contact is more subtle when the phosphatic beds are subjacent to the shale, as has been the case for the past few years. The basal foot or so of the "button shale" is fine-to-medium-grained, conglomeratic, red quartz sandstone with prominent pebbles of phosphate and chert and scattered, fragmented crinozoan parts. This basal sandstone unit grades upward into a fine-sandy, pyritic, calcareous red to gray-green clay shale with scattered dolomite rhombs and abundant algal "buttons"; sparse, fragmented ostracodes, brachiopods, trilobites, and crinozoans occur along with abundant agglutinated foraminifers.

The upper portion of the shale at the Love Hollow Quarry is similar in all respects to the type Cason at Cason Mine. Through a gradual decrease in detrital constituents and an increase in carbonate, the shale grades upward into a greenish-gray, button-bearing ostracode biomicrite in the base of the St. Clair (Plate 5, Fig. A).

The lithic gradation and faunal similarity between the "button shale" and overlying St. Clair limestone leaves little doubt that the two are conformable. As mentioned earlier, this was the view held by Williams at the time he named the Cason. Because Miser worked the district at a time when manganese mining was active, he probably saw more exposures of the Cason than anyone before or since. It is, therefore, important to note that he (Miser, 1922) also recognized the lithic similarity between the "button shale" and the St. Clair. He states (p. 27) that the contact is not suggestive of an unconformity "because there is no abrupt change in the character of rocks." However, the evidence then available that the Brassfield occurred between the two units left him little choice but to separate them with an unconformity.

ST. CLAIR LIMESTONE (LOWER/MIDDLE SILURIAN)

The St. Clair Limestone was named by Penrose (1891) for an exposure at St. Clair Spring, 5 miles north of Batesville, Independence County. The exposure at the spring is no longer satisfactory, but an adequate substitute exists along the hillside about one-third of a mile north of the spring (section SCS).

Penrose's St. Clair included the Kimmswick and Fernvale Limestones, the Cason Shale, and the St. Clair and Lafferty Limestones. Subdivision by later geologists (Williams, 1894, 1900; Ulrich, 1911; Miser, 1922) of this all-encompassing, and in many respects somewhat similar stack of bioclastic limestone, has been covered elsewhere (Craig, 1975).

The St. Clair, which occurs in all three of the Ozark outcrop areas, is the most widespread and continuous of the Silurian units. The limestone is best developed in the Batesville district, where it ranges in thickness from 0 to 100 feet. It is thickest in the eastern part of the district, but occurs at only isolated localities, its distribution apparently controlled by a combination of topography on the underlying erosional surface (see discussion of the Cason Shale) and preservation in structural downwarps. The easternmost occurrence of the St. Clair is along the eastern flank of a small anticline breached by Pfeiffer Creek (section SCS). In the western part of the district, in the vicinity of Cushman west to the White River, the formation is more persistent, but averages only 15 feet in thickness.

As indicated earlier, the spectrum of St. Clair textures strongly parallels that of the Brassfield. Both are dominantly bioclastic limestones characterized by fragmented crinozoans and with abundant lime mud matrix. The modes of distribution of lime mud in the St. Clair are like those described for the Brassfield, although the patchy occurrences created by burrowing are less common in the St. Clair. Thus, the St. Clair does not have the mottled appearance characteristic of the Brassfield. Vertical textural trends, not identified in the Brassfield, are present in the St. Clair. These trends are significant in the interpretation of the depositional history of the formation, especially when considered in conjunction with the conformably overlying Lafferty Limestone.

In the Batesville district, the St. Clair is a pinkish-gray to light-gray, coarse-grained crinozoan biosparite. Some layers, especially those at the base and top of the formation, are poorly washed and contain gray-green micrite. In all sections where the base of the St. Clair is exposed, the lower two to five feet is a pyritic, gray-green or red ostracode biomicrite (wackestone) that contains angular to subrounded quartz sand, minor amounts of fragmented crinozoans, corals, trilobites, and brachiopods, and large, well-formed dolomite rhombs (Plate 3, Fig. E). This lithic type grades up into crinozoan biosparite that is poorly washed in its lower and upper portions. The poorly washed biosparite (wackestone/packstone) contains abundant whole shells of brachiopods, ostracodes, cephalopods, and well preserved pygidia and cephala of trilobites (Plate 3, Fig. B). The middle portion of the St. Clair is dominantly crinozoan biosparite (grainstone) containing fragmented skeletal material of brachiopods, trilobites, and ostracodes (Plate 3, Fig. C).

This superposition of lithic types is generally present throughout the area, but is best developed in the thicker eastern sections, especially section BSQ. At this locality, the lower 5 feet of the St. Clair is an ostracode biomicrite, which is followed by 30 feet of poorly washed crinozoan biosparite, 53 feet of clean crinozoan biosparite, and a capping 10 feet of poorly washed biosparite. The sequence of lithic types at section SCS is similar, except that the central 35 to 40 feet of coarse biosparite contains patches and stringers of lime mud in places and the basal ostracode biomicrite is discontinuous, apparently preserved only in irregularities in the top of the underlying Brassfield.

Section LQ is representative of the thinner western sections (Plate 5, Fig. D). As a rule, the middle biosparite is not well developed in the western part of the district. It is expressed as a clastic limestone that is distinctly coarser and contains a smaller percentage of lime mud than the rock above and below.

Algal "buttons" like those in the "button shale" of the Cason occur in the basal few feet of the St. Clair at most localities. The "buttons" in the limestone differ from those in the shale in that they are spherical rather than flattened. Acetic-acid residues of the lower ostracode biomicrite contain a high percentage of terrigenous clay and agglutinated foraminifers. Clay and agglutinated foraminifers are rare in the crinozoan biosparite of the St. Clair.

The St. Clair is of erratic occurrence in the Blanchard Spring area. The best exposure of the limestone is along the north bank of South Sylamore Creek just downstream from Gayler Crossing (section GC). The section there exhibits a lithic sequence similar to that of the Batesville district. At the base of the formation, which directly overlies the Fernvale, is a 1 foot bed of ostracode biomicrite. Above this is 10 feet of clean crinozoan biosparite followed by approximately 10 feet of biomicrite and poorly washed biosparite directly below the Lafferty Limestone.

Exposures of the St. Clair in the Gilbert area of outcrop are known mainly from the works of McKnight (1935), Maher and Lantz (1953a), and Glick and Frezon (1965). The best exposures of the unit are along the Buffalo River from the U. S. Highway 65 bridge (section BRB) east to the mouths of Tomahawk (section MTC), Bear, and Brush Creeks, and along the lower extent of the creeks (sections TC and CB). Thin, scattered exposures of the limestone have been mapped in the Snowball quadrangle to the west (Glick and Frezon, 1965).

The St. Clair in the Gilbert area is overlain by the Lafferty Limestone, or where pre-Boone erosion has removed the Lafferty, by the St. Joe Member of the Boone Formation or its basal sandstone (see Horner and Craig, this volume). The St. Clair ranges in thickness from 0 to a maximum of 36 feet at section MTC.

The St. Clair of the Gilbert area differs from that of sections to the east in that it contains a greater percentage of lime mud. Cleanly washed biosparite, although present in certain layers, is not common. The limestone consists mainly of gradations between biomicrite (wackestone/packstone) and very fossiliferous, poorly washed biosparite (packstone/grainstone). Poorly washed biosparite is not necessarily confined, as in eastern exposures, to the upper and lower portions of the unit, and layers of biomicrite can occur at any level. The basal ostracode biomicrite (wackestone) is replaced by a more abundantly fossiliferous and faunally varied biomicrite (wackestone/packstone), as illustrated on Plate 2, Figs. B, D-F.

LAFFERTY LIMESTONE (MIDDLE/UPPER SILURIAN)

The Lafferty Limestone was named by Miser (1922) for an exposure at Tom Tate Spring (section TS) on the west side of the valley of West Lafferty Creek, 1¼ miles north of Penters Bluff, Izard County. Miser gave the name Lafferty to 85 feet of slabby-bedded, earthy, gray-green to red micrite and sparsely fossiliferous biomicrite overlying the St. Clair. Individual beds in this exposure can change laterally from red to gray-green. The red color is probably from oxidation of iron associated with the disseminated clay. Miser recognized the Lafferty only in the immediate vicinity of its type locality.

Beneath the type Lafferty, the upper 12.5 feet of St. Clair Limestone, as understood by Miser, is a slightly dolomitic, pyritic, gray-green ostracode biomicrite (wackestone) with scattered pink crinozoan fragments and other calcareous skeletal material, subrounded quartz sand, clay, and agglutinated foraminifers. This rock type (Plate 3, Fig. D), which occurs above the coarse-grained crinozoan rock of the St. Clair at most localities, was included in the St. Clair by Miser. There is little to distinquish it from the 85 feet of dominantly red limestone that Miser called Lafferty. Miser's Lafferty differs from the micrite he included in the St. Clair only in that the Lafferty is red from the inclusion of detrital clay and contains less micrite and fewer fossils in its upper portion.

In 1950, Straczek and Kinney remapped geology of the Batesville district, and included in the Lafferty all of the gray-green micrite and ostracode biomicrite above the coarse-grained crinozoan limestone of the St. Clair. This appears to me to be a more workable division between the two units and I have followed their definition of the Lafferty.

The lithic description given above for the Lafferty fits the unit across northern Arkansas. Its contact with the St. Clair is placed above the last appearance of abundant coarse-grained bioclastic allochems. The change from St. Clair to Lafferty results from a decrease in the percentage of allochems and an increase in micrite and detrital grains. The basal beds of the Lafferty contain pinkish crinozoan skeletal material as laminae and scattered grains. This change was apparently rapid, because the contact is sharp in most places (Plate 5, Fig. D). There is a gradual increase in detrital mud and decrease in carbonate, including fossil debris, upward in the unit. This is best seen in thicker sections, such as TS and GC, where the basal ostracode biomicrite (wackestone) grades upward through a micrite (mudstone) into a rock that approaches a calcareous claystone. With the addition of clay, the unit becomes characteristically slabby bedded (Plate 4, Fig. C).

The Lafferty everywhere rests conformably on the St. Clair, as evidenced by the transitional passage between the two units, and is overlain unconformably by younger units. There is a remarkable similarity between the Lafferty and the micritic limestones occurring within the St. Clair, especially the basal ostracode biomicrite. The Lafferty would appear to represent a recurrence of the same environment that produced this fine-grained limestone.

At its type section (section TS), the Lafferty is 97.5 feet thick, which includes the 85 feet that Miser named Lafferty and the 12.5 feet of ostracode biomicrite he originally included in the St. Clair Limestone. The formation is considerably thinner in other parts of the Batesville district. Along the bluffs of the White River in the western part of the district the Lafferty averages 20 feet in thickness. In the eastern part of the district its average thickness is 5 feet. The unusual thickness of the limestone at its type is probably structurally controlled.

Geologists mapping in the Gilbert area (McKnight, 1935, Maher and Lantz, 1953a), although recognizing the distinction between the St. Clair and Lafferty lithologies, have dealt with the two units as a single, undifferentiated formation. The presence in the St. Clair of micrite beds lithically similar to the Lafferty, and the absence of the Lafferty over most of the area, have no doubt contributed much to this style of handling the two units. It is probable that the Lafferty lithology was once present above the St. Clair throughout the Gilbert area, just as it is in the outcrop areas to the east, but has been removed by pre-Boone erosion. Present evidence indicates that it appeared about 35 to 40 feet above the base of the St. Clair, as at section MTC. Maher and Lantz (1953a) report Lafferty lithology above the St. Clair from several places in the Gilbert area. From a section three-quarters of a mile below section BRB they record 30 feet of Lafferty-type micrite above 35 feet of St. Clair.

GEOLOGIC HISTORY

Present evidence suggests that the Brassfield Limestone is a transgressive unit over an erosional surface developed mainly on Ordovician carbonates. It may have had a discontinuous basal detrital phase, like that of the overlying St. Clair (Cason "button shale"), derived from the underlying Ordovician portion of the Cason shale. However, no such phase is present in the few sections available for study. In the Gilbert area, which is the major region of Brassfield preservation, the limestone rests with irregular contact on the Cason. The upper layers of the Cason in this area, however, are dolomitic mudstone that possesses few coarse detrital grains.

The Brassfield lithology records a subtidal complex of wave-exposed shoal and wave-protected quiet subenvironments, all generally shallow water and abounding in life.

Prior to the deposition of the overlying Cason "button shale" and St. Clair Limestone, the region underwent broad epeirogenic uplift that exposed the surface to subaerial erosion. In the Batesville district, where the "button shale" is the basal unit of the overlying transgression, the unconformity on top of the Brassfield is apparent. In the Gilbert area, lack of a basal detrital phase has created a subtle contact with St. Clair directly on Brassfield. Truncation of Brassfield fossils at "welded" unconformities seems to be an expression of a planar erosional surface developed in the absence of much terrestrial vegetation. This surface is similar in appearance to the scalloped/planar erosional surfaces of Read and Glover (1977). However, the Arkansas surface does not seem to fit well into the Read-Glover model of development on a prograded, early-cemented tidal flat or tidal rock platform. Nor is it particularly reminiscent of a subtidal penecontemporaneously cemented hardground. Such models, if applicable to northern Arkansas, would have the St. Clair follow the Brassfield with little erosional vacuity. Any model would have to explain the extreme variation in thickness (0 to 40 feet) of the Brassfield over short distances. Presumably, either of the above models could do that in combination with the St. Clair onlapping the Brassfield over topography developed on the underlying Ordovician, which would confine Brassfield accumulation to topographic lows and place the St. Clair directly on the Ordovician (e.g., sections MTC and GC) over the highs. This explanation certainly seems suitable for the variation in thickness of the Cason "button shale" and St. Clair Limestone in localized places in the eastern part of the Batesville district. No such topography seems apparent on the Ordovician of the Gilbert area, in which the Brassfield has a very irregular distribution. Present evidence best supports an episode of subaerial erosion between the Brassfield and St. Clair. The irregular distribution of the Brassfield probably represents preservation of the unit in structural lows and complete removal of it on the highs created by gentle warping during post Brassfield-pre - St. Clair epeirogenic movement.

The St. Clair Limestone, with the Cason "button shale" as a basal detrital phase, transgressed over an erosional surface developed on the Brassfield and older units. In the Batesville district, particularly the eastern part, some topography existed on the surface as evidenced in the previously discussed relationship exhibited at the Midwest Lime Company quarry. Topography does not seem to be characteristic of the surface over most of northern Arkansas. At the Midwest Lime

Company quarry it probably is related to differential relief associated with the Batesville Graben directly to the south (O. A. Wise, pers. comm.). Fractures filled with "button shale" and St. Clair in the Fernvale Limestone, and with Devonian in the St. Clair Limestone, indicate that the structure was active during the Early Paleozoic. The St. Clair and "button shale" appear to have collected lower on the topography, pinching out altogether on the high, thus explaining their rapid disappearance. Throughout the district, the detrital material comprising the "button shale" was limited in quantity, no doubt derived locally from the underlying phosphate beds of the lower Cason, and thus has a discontinuous distribution. In the Blanchard Spring and Gilbert areas, where the lower phosphate beds have a much smaller percentage of sand and gravel-size grains, no detrital phase collected at the base of the St. Clair.

The "button shale" is the basal unit of a transgressive-regressive cycle of sedimentation that includes the overlying St. Clair and Lafferty Limestones. The shale represents terrigenous sediment that accumulated in a near-shore environment containing spherical algal growths, agglutinated foraminifers, and ostracodes. Seaward, lime mud was collecting in a lagoonal environment that abounded with agglutinated foraminifers and ostracodes. This lagoonal environment produced the ostracode biomicrite present at the base of most St. Clair sections. This lagoonal environment produced the ostracode biomicrite present at the base of most St. Clair sections. The lagoonal environment was restricted from the open ocean by a crinozoan sand bar, now represented by the coarse-grained St. Clair. On the lagoon side of this bar, marine invertebrate life abounded on a substrate composed of intermixed lime mud and skeletal debris. Currents in this area were strong enough to oxygenate the water and remove some of the finer material, but not strong enough to shift the substrate. Thus, finer particles tended to collect beneath shells that were concave downward or in the protected hollow of shells that were concave upward. In this environment was deposited the fossiliferous, poorly washed crinozoan biosparite in the lower part of the St. Clair. Above this the St. Clair is more clastic and is composed almost entirely of crinozoan debris. This lithology is interpreted as representing the ocean facing, high-energy portion of the bar. The poorly washed upper layers of the St. Clair represent the recurrence during the regressive phase of the conditions on the lagoon side of the bar. The ostracode biomicrite of the Lafferty marks the regression of the lagoonal environment.

If a sedimentary environment similar to that represented by the "button" shale recurred during regression, its record has been removed by pre-Penters erosion.

The vertical succession of lithic types illustrating the depositional environments now represented by the "button shale", St. Clair, and Lafferty is best seen in the Batesville district. The thick build-up of coarse-grained St. Clair in the eastern part of the district indicates close proximity to the major build-up of crinozoan sand that formed the protective barrier. Possibly the topography in this region aided in the accumulation of the thickness observed by concentrating wave and current energies. The western part of the district was far-removed from the barrier as demonstrated by the thin interval of coarse-grained crinozoan rock followed by a relatively thick Lafferty section.

In the Gilbert area, most of the accumulation of the St. Clair seems to have taken place on the lagooon side of the barrier, with occasional incursions of the barrier and lagoonal facies represented by the layers of crinozoan biosparite and biomicrite, respectively, found interbedded with the main body of poorly washed biosparite. The main portion of the barrier probably crossed northern Arkansas a few miles to the south of Gilbert.

EXPLANATION OF PLATE I

Figure A. Photomicrograph of algal "button", Cason "button shale" (X7.1). Note micro "algal heads" that result from irregular growth of laminae toward exterior of "button". Section LQ.

Figure B. Negative print of thin section, Cason "button shale". Shows complete "button" in silty, dolomitic matrix (X 44). Section LQ.

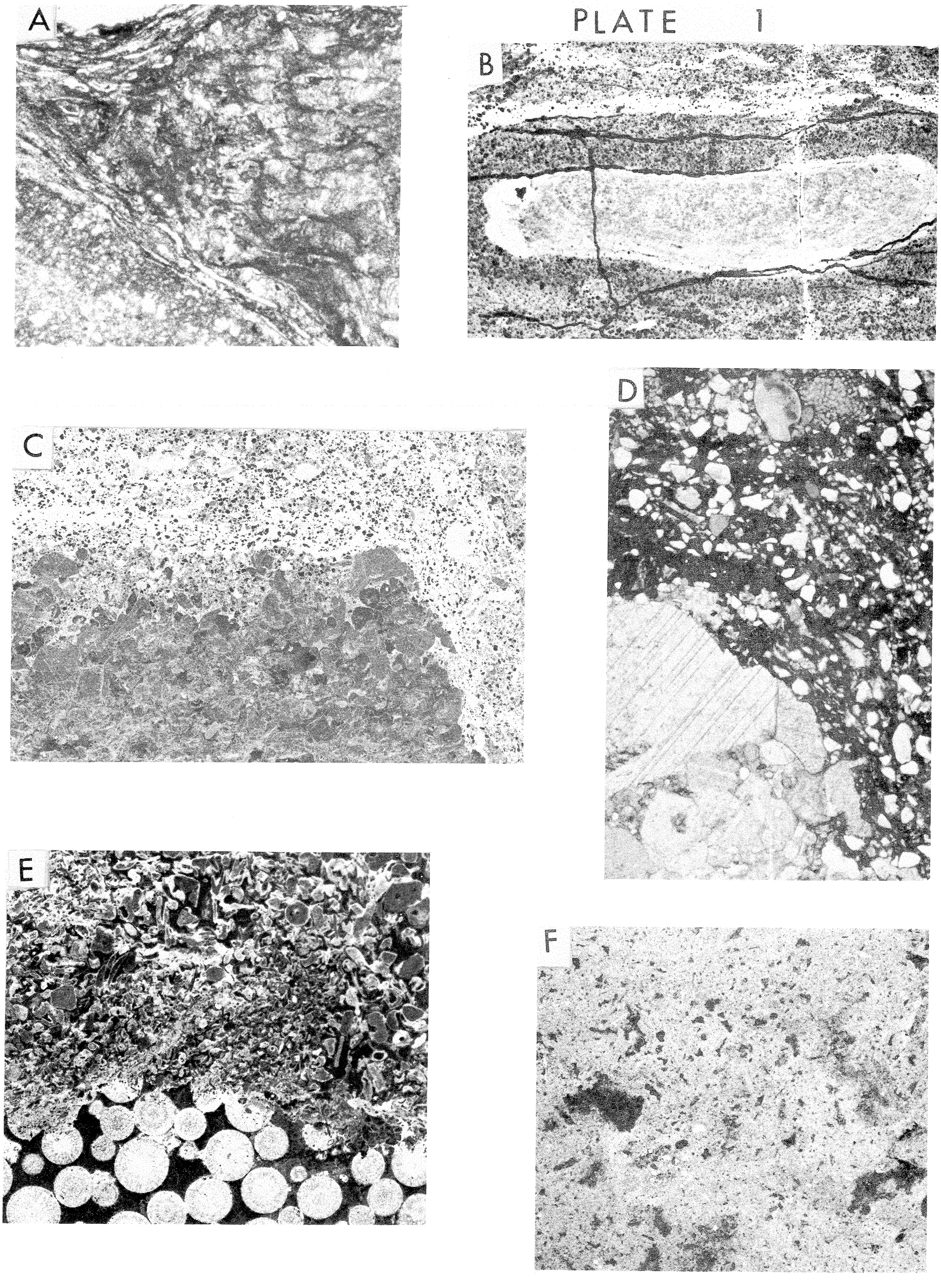
Figure C. Negative print of thin section, base of Cason "button shale" on Brassfield Limestone (X 7.1). Note irregularity of contact. Section LQ.

Figure D. Photomicrograph showing enlargement of contact zone illustrated in Figure C (X 30). Note truncation of crinozoan parts. Grains in "button shale" are quartz and phosphate in clay matrix. Section LQ.

Figure E. Negative print of thin section, Brassfield Limestone on oolitic limestone of lower Cason (X 7.1). Contact is stylolitic. Section LQ.

Figure F. Negative print of thin section, micritic Brassfield Limestone (X 7.1). Patches of spar (dark) represent burrows. Section CB.

PLATE 1



EXPLANATION OF PLATE II

Figure A. Negative print of thin section. Brassfield Limestone (X 10). Poorly washed mixed biosparite. Note "roofing effect" of long skeletal fragments. Skeletons include crinozoans, trilobites, brachiopods, bryozoans, and ostracodes. Areas of poor sorting are characteristic of Brassfield mixed biosparite and probably result from burrowing. Micrite (dense, white areas) is red in natural color. Section LQ.

Figure B. Photomicrograph of St. Clair Limestone on Brassfield (X 44). Note truncation of Brassfield fossils. Section CB.

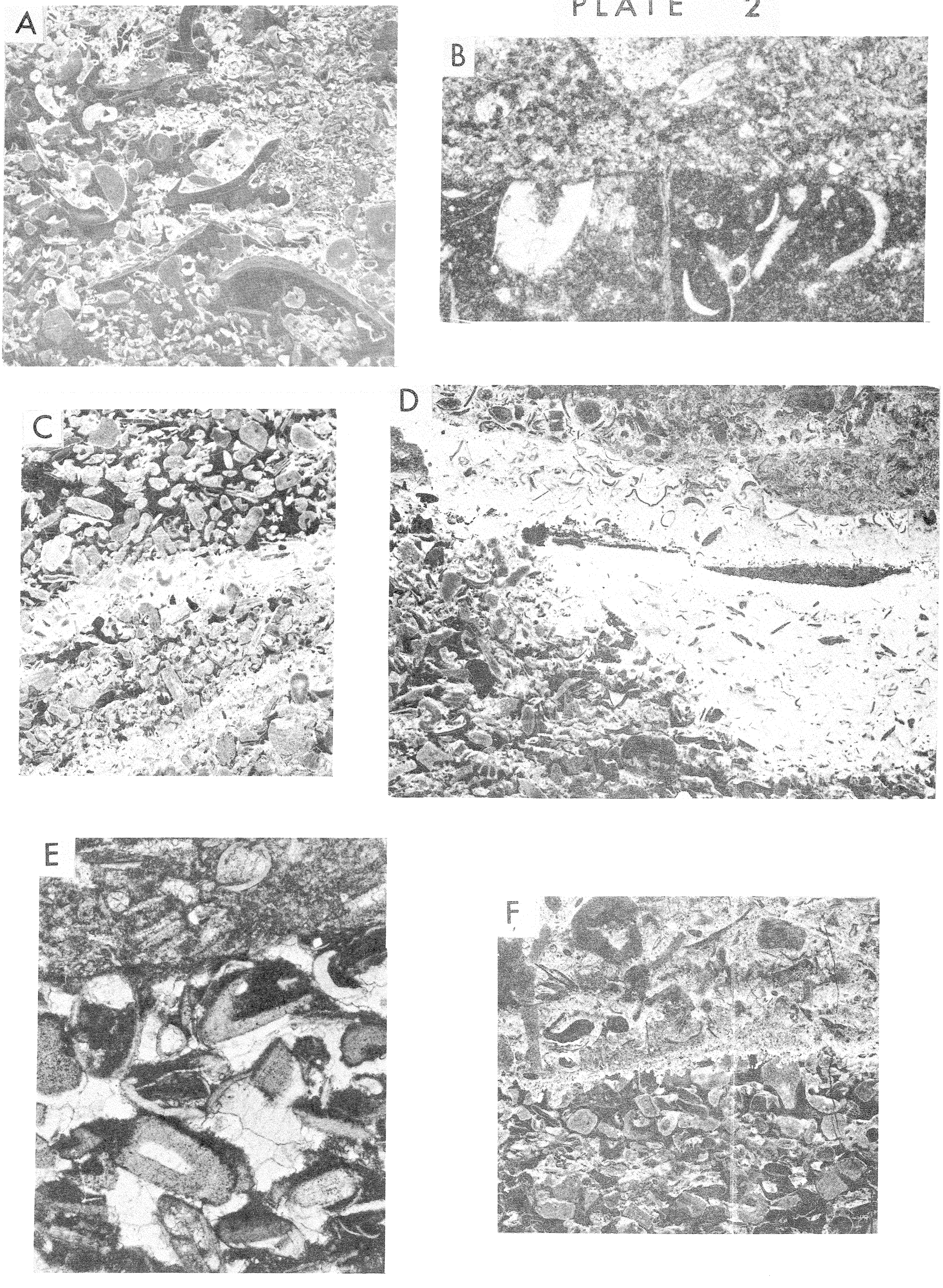
Figure C. Negative print of thin section, Brassfield Limestone (X 13.7). Poorly washed biosparite. Note micrite (light) stringers. Section CB.

Figure D. Negative print of thin section, St. Clair Limestone on Brassfield (X 7.1). Note truncation of micrite-filled burrow (large light area) at contact (scalloped). Micrite-filled burrow is dark red in natural color. Section CB.

Figure E. Photomicrograph of St. Clair Limestone on Brassfield (X 44). St. Clair shows characteristic biomicrite base; Brassfield is poorly washed biosparite. Note micrite (dark) flooring interstices in otherwise spar-cemented rock of Brassfield. Section CB.

Figure F. Negative print of thin section, biomicrite of St. Clair on biosparite of Brassfield (X 7.1). Contact is styolitic. Section TC.

PLATE 2



EXPLANATION OF PLATE III

Figure A. Crinoid biosparite of Brassfield Limestone (X 6). Note absence of skeletal debris of other fossil groups. Section SCS.

Figure B. Negative print of acetate peel, poorly washed biosparite of lower St. Clair Limestone (X 5.1). Unfragmented skeletons (note trilobites) are characteristic of this protected, back barrier zone. Note location of finer carbonate (mostly microspar) in protected areas. Section BSQ.

Figure C. Negative print of acetate peel, crinozoan biosparite of middle St. Clair (X 5.1). Note absence of micrite and dominance of crinozoans, both characteristic of St. Clair crinozoan barrier sand. Section BSQ.

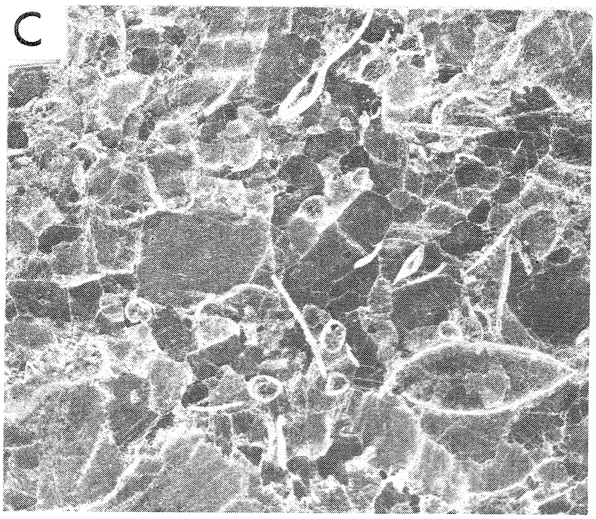
Figure D. Negative print of thin section, ostracode biomicrite of Lafferty (X 14.1). This lithic type is essentially the same as the basal ostracode biomicrite of the St. Clair (Figure E). Note crinozoans and other skeletal debris. Section LQ.

Figure E. Negative print of thin section, basal ostracode biomicrite of St. Clair (X 14.7). Rock carries high silt and clay content and is gradational downward with "button shale" of the Cason. Note dominance of ostracodes. Fine areas lacking fossils (lower right) probably result from burrowing. Section LQ.

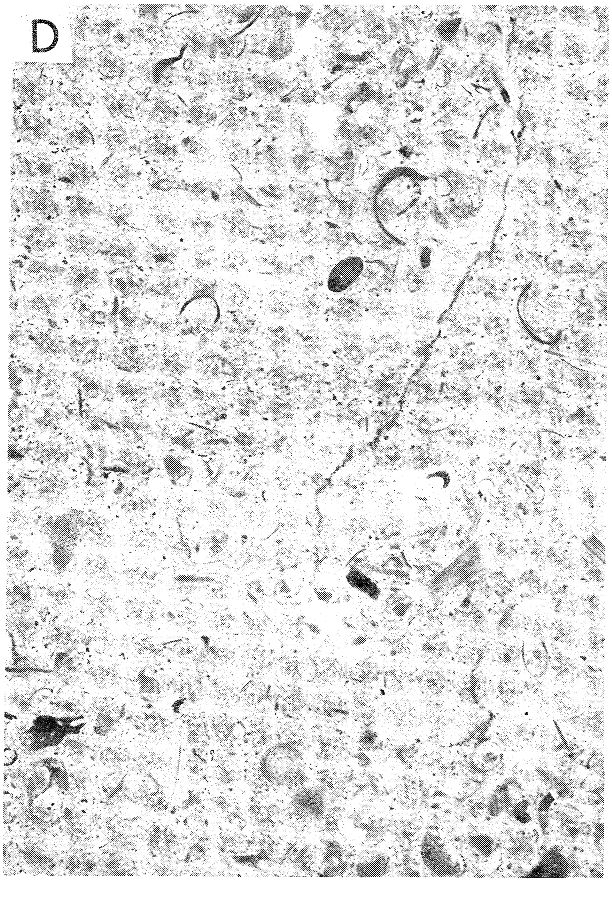
PLATE 3

A

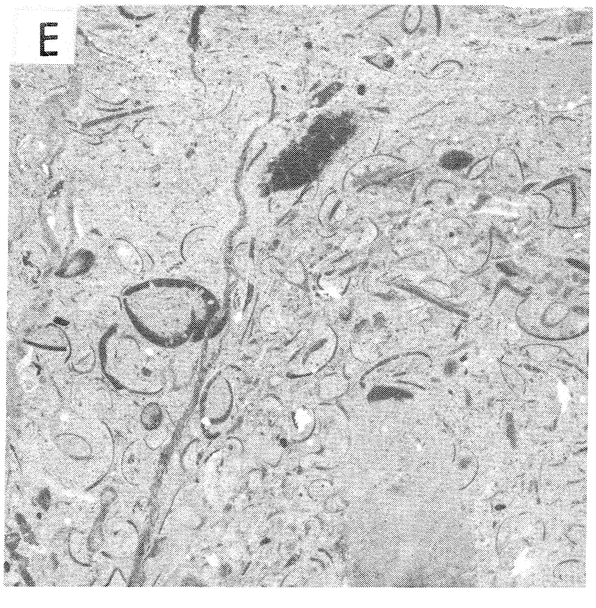
B



C



D

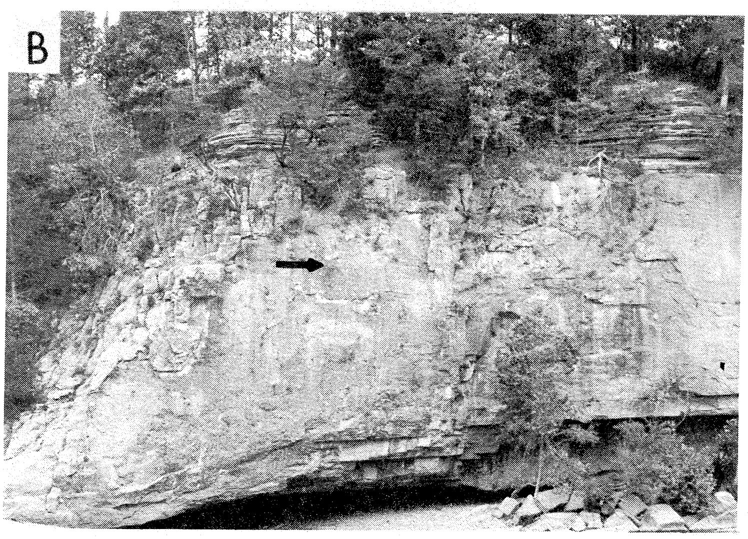


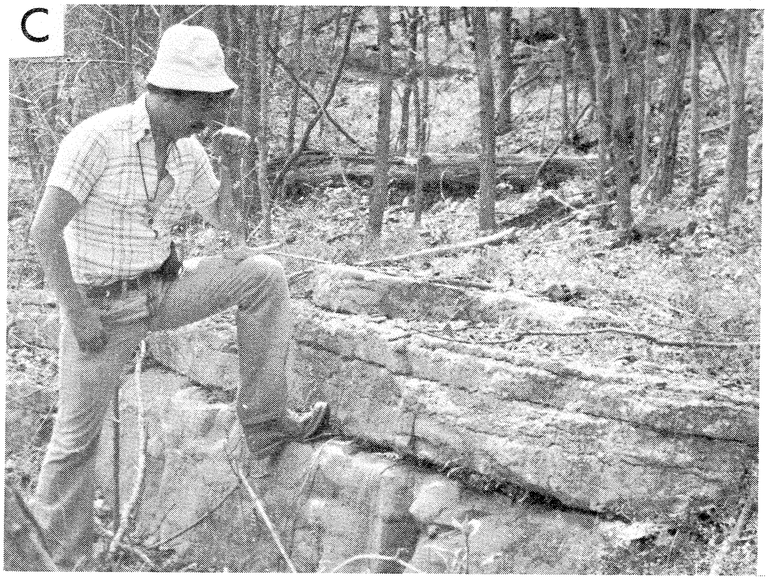
E

EXPLANATION OF PLATE IV

Figure A.	Reference section near Section TC (see text for location). Hammer rests on top of Brassfield. Geologist is standing at contact between St. Clair Limestone and sandy base (Sylamore Sandstone) of St. Joe Limestone.
Figure B.	Section BRB. Re-entrant at base of bluff (in Buffalo River) contains Ordovician Cason Shale. Contact between Brassfield and St. Clair Limestones visible as faint line a little over halfway up the bluff (arrow). Slabby beds on top are St. Joe Limestone with a thin sandy base (Sylamore Sandstone).
Figure C.	Thin-bedded micrite and biomicrite characteristic of Lafferty Limestone. Section GC.
Figure D.	Calcite spar-lined vugs in St. Clair Limestone at Section TC. Vug centers are filled with sand that carried a Sylamore (Late Devonian) conodont fauna.
Figure E.	"Button shale" filling in fractures in top of Fernvale Limestone at Midwest Lime Company Quarry (MLQ). Scale is O. A. Wise, Arkansas Geological Commission.

PLATE 4





EXPLANATION OF PLATE V

Figure A. Oolitic limestone of lower Cason Shale, Cason "button shale", and St. Clair Limestone at Section LQ. Hammer rests on light-colored oolitic limestone. To right of hammer, the overlying dark-colored "button shale" is channeled into the oolitic limestone. The lower arrows point to the base of the "button shale". Note the gradational change from "button shale" to basal fine-grained St. Clair. The contact between the two is located at about the level of the upper arrow.

Figure B. Phosphate beds of the lower Cason, Brassfield, and "button shale" at Section LQ. Hammer rests on thin layer of Brassfield. Layers below are phosphatic beds of lower Cason. Irregular contact of Brassfield and overlying "button shale" is marked by arrows.

Figure C. Fernvale Limestone, "button shale", and St. Clair Limestone at Midwest Lime Company quarry (MLQ). Dark band is "button shale". Note truncation of Fernvale by "button shale" and onlap of St. Clair onto topography on top of the Fernvale. To the north (right), the Cason is onlapped by the St. Clair, which then rests directly on the Fernvale.

Figure D. Phosphate beds of lower Cason, "button shale", St. Clair and Lafferty Limestones at Section LQ. Dark-colored rock at base of photograph is the phosphate beds and "button shale". Dark-colored arrow points to lower ostracode biomicrite at base of St. Clair. Above is coarse-grained St. Clair and Lafferty. White arrow points to sharp contact between St. Clair and Lafferty. Note difference in texture between the two formations.

PLATE 5

THE SECTIONS

ST. CLAIR SPRINGS (SCS)

location: Along the hillside about one-third of a mile northeast of St. Clair Spring in the SE¼ SW¼, Sec. 18, T. 14 N., R. 5 W. Sulphur Rock quadrangle, Independence County.

access: North on Highway 167, 2.3 miles north of Pfeiffer, then on foot 100 yards to east of highway across barnyard and pasture. Ask permission at residence just to west of highway.

exposure: Nearly complete exposure of St. Clair Limestone capped by 2 feet of Lafferty and underlain by 20 feet of Brassfield. Heavy moss growth but otherwise excellent exposure. Beds dip about 20° back into hillside (to east).

BATESVILLE STONE QUARRY (BSQ)

location: One-half mile (straight-line distance) north of Batesville in SW¼, SW¼, Sec. 34, T. 14 N., R. 6 W., Batesville quadrangle, Independence County.

access: North of Batesville on Highway 167 to sign marking entrance to Midwest Lime Company quarry on west side of highway. Ask permission at quarry office to cross over to the old cut stone quarry which is just to the west of the Midwest Lime quarry.

exposure: Best exposure of St. Clair available. Quarry is on three levels, lowest of which is in the Fernvale Limestone. This quarry, along with the operating Midwest Lime quarry and the intervening old Cason Mine, afford an excellent look at the "button shale", St. Clair, and Lafferty as developed in the eastern part of the Batesville district. The Penters Chert, Sylamore-Chattanooga, and Boone are also well exposed.

LOVE HOLLOW QUARRY (LQ)

location: East bluff (Penters Bluff) of White River in the SE¼, Sec. 4, T. 14 N., R. 8 W., Bethesda quadrangle, Izard County.

access: West out of Cushman on the Love Hollow Road. After crossing West Lafferty Creek, the road climbs Penters Bluff at a steep grade. Entrance to upper level of quarry is to the west (left) at top of hill. Road into quarry may be followed to quarry office to obtain permission. Exposure is on upper level of quarry near entrance.

exposure: Excellent exposure of "button shale" through Lafferty Limestone, showing well the gradational contacts of the units (Plate 5, Fig. D).

TATE SPRING (TS)

location: West side of valley of West Lafferty Creek about 1¼ miles north of Penters Bluff in SW¼, SW¼, Sec. 3, T. 14 N., R. 8 W, Mt. Pleasant quadrangle, Izard County.

access: Love Hollow Road just beyond the entrance to Love Hollow Quarry. Construction involved with an eastward shift of the Love Hollow Road because of encroachment of the quarry has passed directly through the type section. Section begins just to west (left) of road at the base of the steep hill descended just past quarry entrance.

exposure: Excellent exposure of upper part of "button-shale", thin St. Clair and thick Lafferty, which extends up the hillside into the woods. The upper clayey, red Lafferty is well displayed here.

TOMAHAWK CREEK (TC)

location: Two miles (straight-line distance) north of Gilbert where road from Gilbert to St. Joe crosses Tomahawk Creek in NE¼, SE¼, Sec. 19, T. 16 N., R. 16 W., Maumee quadrangle, Searcy County.

access: Gravel road north out of Gilbert, keeping left at forks, traveling toward St. Joe. Measured section was in small stream bed paralleling road (on east side) leading down to bridge from the south. Since this section was measured it has been covered by road widening. A suitable substitute, however, is present along the south bank of the Tomahawk just upstream from the bridge.

exposure: Substitute section is in a small draw that provides a complete exposure from the upper part of the Brassfield into the basal Boone (Plate 4, Fig. A). Beds dip about 6° to the north (down the draw).

CRANE BOTTOM (CB)

location. Located ½ mile southeast of Gilbert on the west bank of Bear Creek at Crane Bottom in the NE¼, SW¼, Sec. 32, T. 16 N., R. 16 W., Marshall quadrangle, Searcy County.

access: North out of Marshall on gravel road running on east side of Bear Creek to position opposite Crane Bottom, and from there on poor road (or on foot) into Bear Creek valley.

exposure: Good exposure of Brassfield and St. Clair. Very brushy and difficult access.

GAYLER CROSSING (GC)

location: On South Sylamore Creek 3.5 miles southeast of Fifty Six in the NE¼, NW¼, Sec. 21, T. 15 N., R. 22 W., Fifty Six quadrangle, Stone County.

access: On Arkansas 87 between Highway 14 and Mountain View passing through the community of Gayler. Located on north bank of South Sylamore Creek just downstream from low-water crossing.

exposure: Excellent exposure of St. Clair and Lafferty showing same superposition of lithic types as in the Batesville district. Upper clayey, thin-bedded Lafferty well exposed. Beds dip about 6° NW (upstream).

MOUTH OF TOMAHAWK CREEK (MTC)

location: 2.5 miles (straight-line distance) northeast of Gilbert ¼ of a mile upstream from the mouth of Tomahawk Creek in SE¼, NE¼, Sec. 21, T. 16 N., R. 16 W., Maumee quadrangle, Searcy County.

access: Along easternmost gravel road that crosses Tomahawk Creek traveling north out of Gilbert. From low-water crossing walk along south bank of Creek toward mouth.

exposure: Excellent exposure of thickest St. Clair-Lafferty in Gilbert area. Brushy during summer months.

BUFFALO RIVER BRIDGE (BRB)

location: Located 1 mile west of Gilbert on north bluff of Buffalo River just east of Highway 65 bridge in NE¼, SW¼, Sec. 36, T. 16 N., R. 17 W., Marshall quadrangle, Searcy County.

access: Access on dirt road from Highway 65 running east from north end of bridge. Walk or drive to top of bluff. Section is in bluff overlooking popular swimming hole.

exposure: Excellent exposure of complete Brassfield and St. Clair with contact approximately 3/4 of way up bluff (Plate 4, Fig. B). Steep, but climbable on foot with caution.

REFERENCES

Amsden, T. W., 1968, Articulate brachiopods of the St. Clair Limestone (Silurian), Arkansas, and the Clarita Formation (Silurian), Oklahoma: Paleontological Soc. Mem. 1, Jour. Paleontology, v. 42, no. 3, pt. 2, 120 p.

Craig, William W., 1969, Lithic and conodont succession of Silurian strata, Batesville district, Arkansas: Bulletin Geol. Society America, v. 80, p. 1621–1628.

————, 1975, History of investigations on the post-St. Peter Ordovician of northern Arkansas: the art of layer-cake geology: *in* Contributions to the Geology of the Arkansas Ozarks, Arkansas Geol. Commission, p. 1–17.

————, and Lemastus, S., 1979, Stratigraphy of the post-Fernvale (Ordovician) – pre-St. Clair (Silurian), northern Arkansas: Geol. Society of America, Abstracts with Program for 1979, v. 11, no. 2, p. 146.

Foerste, A. F., 1905, Silurian Clays: Kentucky Geol. Survey Bull. 6, p. 143–178.

Frezon, S. E., and Glick, E. E., 1959, Pre-Atoka rocks of northern Arkansas: U. S. Geol. Survey, Prof. Paper 314–H, p. 171–189.

Glick, E. E., and Frezon, S. E., 1965, Geologic map of the Snowball quadrangle, Newton and Searcy Counties, Arkansas: U. S. Geol. Survey, Geol. Quadrangle Map GQ 425.

Haley, Boyd R., 1976, Geologic Map of Arkansas: Arkansas Geological Commission, Little Rock.

Lemastus, Steven W., 1979, Stratigraphy of the Cason Shale (Ordovician-Silurian), northern Arkansas: unpublished Master's thesis, Univ. New Orleans, New Orleans, Louisiana, 94 p.

Maher, J. C., and Lantz, R. J., 1952, Correlation and Description of Lower Paleozoic at Gilbert, Carver, and Marshall, Arkansas: U. S. Geol. Survey, Circular 160, 21 p.

————, 1953a, Geology of the Gilbert area, Searcy County, Arkansas: U. S. Geol. Survey, Oil and Gas Investigations Map OM 132.

————, 1953b, Correlation of pre-Atoka rocks in the Arkansas Valley, Arkansas: U. S. Geol. Survey, Oil and Gas Investigations Chart OC 51.

McKnight, E. T., 1935, Zinc and lead deposits of northern Arkansas: U. S. Geol. Survey, Bull. 853, 311 p.

Miser, H. D., 1922, Deposits of manganese ore in the Batesville district, Arkansas: U. S. Geol. Survey, Bull. 734, 273 p.

————, 1941, Manganese carbonate in the Batesville district, Arkansas: U. S. Geol. Survey, Bull. 921–A, 97 p.

Penrose, R. A. F., Jr., 1891, Manganese – its uses, ores, deposits: Arkansas Geol. Survey, Ann. Rept. for 1890, v. 1, p. 99–137; p. 166–198; p. 586–595.

Purdue, A. H., 1907, Developed phosphate deposits of northern Arkansas: U. S. Geol. Survey, Bull. 315, p. 463–473.

Read, J. F., and Glover, G. A., Jr., 1977, Scalloped and Planar erosion surfaces Middle Ordovician limestones, Virginia: Analogues of Holocene exposed karst or tidal rock platforms: Jour. Sed. Petrology, v. 47, p. 956–972.

Sgouras, John D., 1979, Stratigraphy and conodont biostratigraphy of the Silurian St. Clair-Lafferty Limestones (undifferentiated), Searcy and Stone Counties, Arkansas: unpublished Master's thesis, Univ. New Orleans, New Orleans, Louisiana, 98 p.

Straczek, J. A., and Kinney, D. M., 1950, Geologic map of the central part of the Batesville manganese district, Independence and Izard Counties, Arkansas: U. S. Geol. Survey, Mineral Invest. Field Studies Map MF 1.

Ulrich, E. O., 1911, Revision of the Paleozoic Systems: Bulletin Geol. Society America, v. 22, p. 281–680.

Williams, H. S., 1894, On the age of the manganese beds of the Batesville region of Arkansas: American Jour. Sci., 3rd ser., v. 48, p. 325–331.

————, 1900, The Paleozoic faunas of northern Arkansas: Arkansas Geol. Survey, Ann. Rept. for 1892, v. 5, p. 268–362.

Wise, O. A., and Caplan, W. M., 1967, Silurian and Devonian rocks of northern Arkansas: *in* Symposium on Silurian-Devonian rocks of Oklahoma and Environs, Tulsa Geol. Society Digest, v. 35, p. 242–252.

————, 1979, Silurian and Devonian rocks of northern Arkansas: Arkansas Geol. Commission, Information Circ. 25, 14 p.

BASEMENT ROCKS IN NORTHERN ARKANSAS

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ABSTRACT

Northwest Arkansas is underlain by Precambrian rocks that are the continuation of the rhyolite-epizonal granite suite reported in adjacent states. East of a gravity and magnetic gradient, possibly marking the limit of the rhyolite-epizonal granite suite, one well penetrated a coarse, two feldspar, biotite granite cut by diabase in northcentral Arkansas. Granitic gneiss and metadiorite xenoliths brought to the surface in a carbonate-rich dike in Conway County are probably from deeper levels in the crust. A deep well in the Mississippi Embayment cut a thick section of unmetamorphosed red arkose beneath a grey arkose probably equivalent to the Lamotte Sandstone. The red arkoses overlie a regionally metamorphosed granitic gneiss. No ages have been determined from the basement rocks. The best age for the rhyolite-granite suite is 1300-1400 m. y., based on determinations on continuous rocks in adjacent states. The age of the rocks found in the Embayment is an interesting subject for speculation. The speculation here favors a Cambrian age for the red arkose and an age between 1500 and 1650 m. y. for the granitic gneiss.

INTRODUCTION

Samples of fifteen wells reaching basement in Arkansas have been described. In addition xenoliths of basement rock brought to the surface in an exotic carbonate-rich dike were examined from Conway County. Thirteen of the basement wells are in northwest Arkansas. One well is in northcentral Arkansas and the other is a deep test in far northeast Arkansas, the first known well to reach crystalline rocks in the Mississippi Embayment. No isotopic ages are available from these rocks but the age range for most of these rocks can be inferred with considerable confidence.

There has been a substantial amount of recent work published on the basement rocks of neighboring states. These studies in Oklahoma (Denison, 1981), Kansas (Bickford and others, 1981) and Missouri (Kisvarsanyi, 1979) have outlined the framework surrounding Arkansas.

The quality of samples varies greatly. Some wells have cuttings that do not exceed one mm in maximum dimension. Others have chips exceeding 5 mm. Fortunately the fine grain size of the igneous rocks allow definition of rock type and texture even in the finest chips. Only two of the wells were cored. A general description of the rocks examined is given in the following text. Descriptions of individual rock samples are found in the Appendix.

Northwest Arkansas

Denison (1981) grouped three basement rock units into the Northeast Oklahoma Province. This is a sequence of rhyolites, generally micrographic granites and microgranites believed to be very closely related in time and space. Examination of the samples from northwest Arkansas indicates that this area is underlain by the eastward extension of this province.

Four of the wells penetrated rhyolite or metarhyolite. Miser and Ross (1925) reported rhyolite from a well in Madison County for which samples are no longer available. In these samples phenocrysts of plagioclase and perthite are set in a finely crystalline groundmass of quartz-feldspar. Quartz as phenocrysts is present in some samples but always in lesser amounts. Iron oxides as microphenocrysts and disseminate hematite are always present. Biotite was seen in only one sample (Lg-1) but may well have been present in others before conversion to chlorite. Apatite and zircon most commonly in association with iron oxides are invariable primary minerals. Secondary minerals include clay-sericite-epidote feldspar alterations, calcite, chlorite, sphene-leucoxene and iron sulfides. The phenocrysts are set in a quartzfeldspar groundmass of differing texture. The unmetamorphosed rhyolites have a finely crystalline felted groundmass with deep reddish color. The metarhyolites have a finely granoblastic groundmass averaging 0.1 to 0.05 mm in the mosaic.

The distribution of rhyolite shown in Figure 1 is subject to substantial revision with additional well control. The long pretransgression erosional period and the relatively flat lying character of the volcanic rocks has surely led to a highly irregular pattern of preservation.

Nine wells penetrated granite and microgranite. The rocks are composed of perthite, quartz and plagioclase with lesser to trace amounts of iron oxides, biotite, hornblende, pyroxene, sphene, apatite and zircon. Secondary feldspar alterations, calcite, chlorite, sphene-leucoxene, epidote and actinolite are also found. The perthite is generally fresh, lightly dusted with hematite and vacuoles. Plagioclase is in lesser amounts and contains locally extensive clay-sericite as alterations.

The texture is extremely variable but tends to be delicately micrographic around feldspar phenocrysts. The microgranites have textures gradational with the rhyolites. The curious rod-like appearance of quartz in thin section (see Denison, 1981, p. 14 and p. 78) is characteristic of several of the finest microgranites.

The two core samples (Bn-3 and Ca-1) give the best definition of gross texture and bulk mineralogy. Both samples show rather large (3-7mm) phenocrysts of irregularly altered plagioclase with a thin optically continuous rim of perthite. The pattern of feldspar phenocrysts, accounting for between a fifth and a third of the rock, set in an irregularly micrographic matrix is probably representative of granites in northwest Arkansas.

Quartz accounts for about a fourth of the volume and is generally intergrown with perthite. Perthite and plagioclase are generally in subequal amount (an exception is Sr-1). The color index ranges between 5 and 10 but may be lower in some of the finer microgranites.

The rocks found in wells in northwest Arkansas are almost certainly the extension of granites and rhyolites found in contiguous states.

The age of these rocks is well known. They yield Rb/Sr ages of about 1290 ± 20 m. y. with an initial ratio of 0.7043 ± .0008 (Denison, 1981). The zircons from these rocks yield systematically older ages. Bickford and Lewis (1979) and Bickford and others (1981) obtained U/Pb concordia ages of from 1361 ± 6 to 1408 ± 21 m.y. on rock originally grouped into the Northeast Oklahoma Province. Thus, by comparison of zircon ages, the age of the Province is equivalent to the younger igneous event in the St. Francois Mountains (Bickford and Mose, 1975).

The granites and rhyolites found in Arkansas are the continuation of those in northeast Oklahoma but there are some small petrographic differences that should be noted. Most of the samples are considerably more

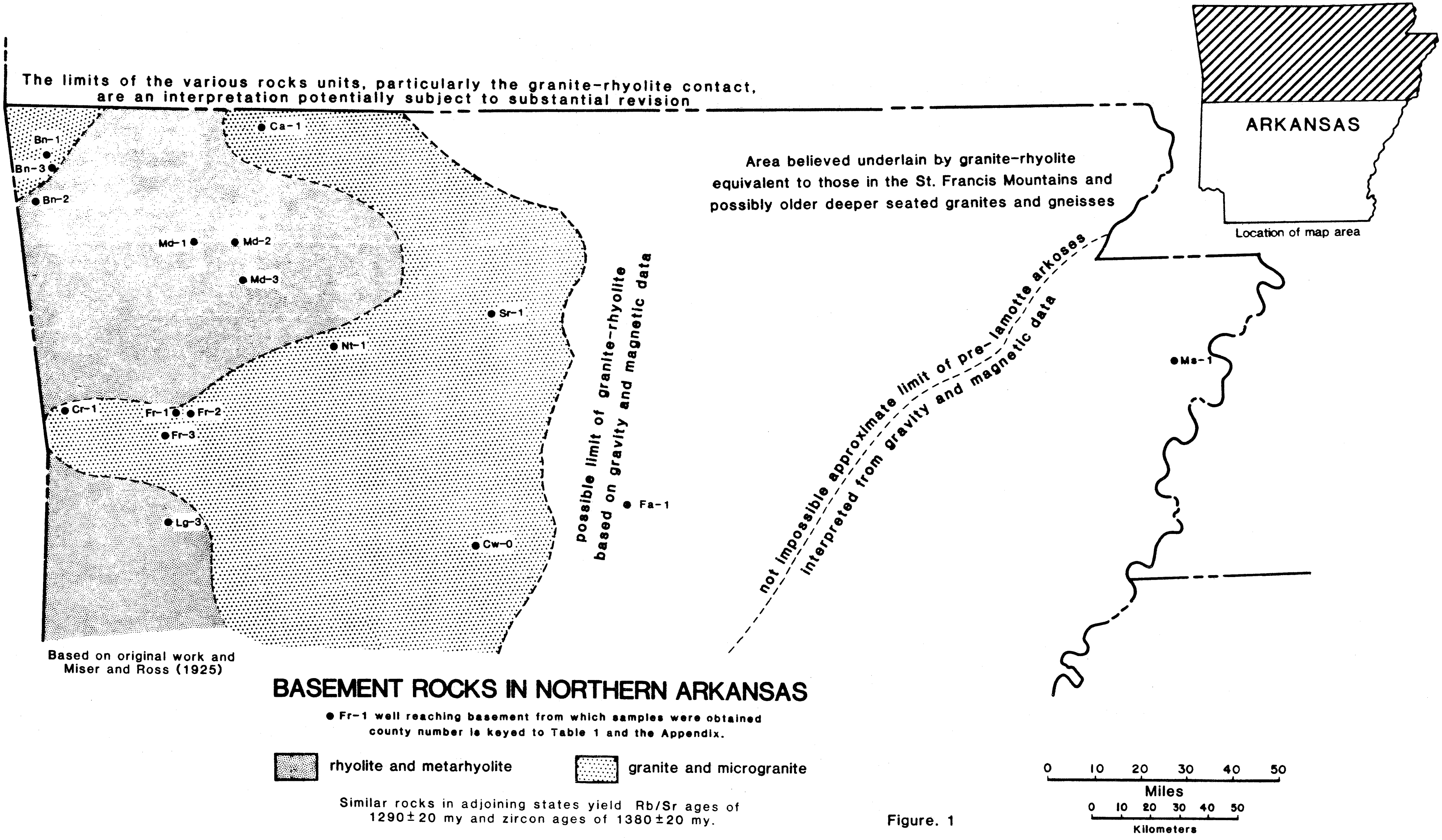
The limits of the various rocks units, particularly the granite-rhyolite contact, are an interpretation potentially subject to substantial revision
Bn-1
Bn-3
Bn-2
Ca-1
Md-1
Md-2
Md-3
Nt-1
Sr-1
Cr-1
Fr-1
Fr-2
Fr-3
Lg-3
Cw-0
Fa-1
Ms-1
possible limit of granite-rhyolite based on gravity and magnetic data
Area believed underlain by granite-rhyolite equivalent to those in the St. Francis Mountains and possibly older deeper seated granites and gneisses
not impossible approximate limit of pre-lamotte arkoses interpreted from gravity and magnetic data
ARKANSAS
Location of map area
Based on original work and Miser and Ross (1925)
BASEMENT ROCKS IN NORTHERN ARKANSAS
● Fr-1 well reaching basement from which samples were obtained county number is keyed to Table 1 and the Appendix.
rhyolite and metarhyolite
granite and microgranite
Similar rocks in adjoining states yield Rb/Sr ages of 1290±20 my and zircon ages of 1380±20 my.
0
10
20
30
40
50
Miles
0
10
20
30
40
50
Kilometers

Figure. 1

siliceous than typical samples from contiguous Oklahoma. Several of the granites and rhyolites carry primary biotite and/or hornblende; both are extremely rare to the west. These differences can be explained as expectable over a large area. Indeed, there is a systematic difference in the composition of granites and rhyolites in areas in northeast Oklahoma (Denison, 1981, p. 9-10). These probably represent tapping of different parts of an evolving magma suite over a broad area.

The metamorphism found in several samples is probably due to the proximity of a younger deeper seated plutonic granitic mass. Metamorphism in northeast Oklahoma is present only where the rhyolites and granites are in close association with slightly younger mesozonal granites (Denison, 1981, p. 11-12). It is a typical hornfelsic recrystallization with no accompanying shearing.

These granites and rhyolites are a relatively thin veneer and must be almost flat lying because of the large area over which they occur. The microgranite porphyrites are probably sills within or at the base of the volcanic pile. Perhaps the gneisses found as xenoliths in Conway County represent the plutonic basement on which the granite-rhyolite suite was extruded and intruded.

The area of drill control does not contain any significant regional gravity (Hendricks and others, 1981) or magnetic (Zietz, 1981) gradients. The gravity and magnetic data may indicate where the eastern edge of the basement province might lie (it is not present in northcentral Arkansas). The gravity and magnetic gradients are used to terminate the granite-rhyolite suite on the east (Fig. 1). To the south, the rapid deepening of the basement into the Arkoma Basin complicates gravity and the closest magnetic gradient is some sixty miles south of the last well control. The strong positive circular gravity and magnetic in far northwest Arkansas has a drill hole (Bn-2) very close to the crest of the anomaly. The micrographic microgranite porphry found in the well is clearly not the cause of the anomaly. The gravity and magnetic data suggest a circular high density and high susceptibility rock mass, probably a gabbroic intrusion that has not been exposed by erosion at the basement surface.

Northcentral Arkansas

One well (Fa-1) has been drilled to the east of the gravity and a magnetic gradient believed to mark the approximate limit of the rhyolite-epizonal granite association of northwest Arkansas. This well penetrated rock in sharp contrast to those to the west suggesting the gradient may well have some validity.

Beneath a slightly sandy dolomite (there is no basal sandstone) the well penetrated a snowy white, coarse, biotite-rich granite cut by a diabase interpreted as a dike or sill of Precambrian(?) age. The granite is a two feldspar type typical of mesozonal batholithic intrusions.

The granite is composed mostly of slightly perthitic microcline, quartz and irregularly altered oligoclase. Fresh biotite flakes are common in the cuttings. A dark strongly pleochroic common hornblende is found in lesser amounts. Minor to trace amounts of iron oxides, sphene, epidote, apatite, leucoxene and zircon are also present. The texture cannot be determined with any surely due to small cuttings and a coarse grain size but appears, on the basis of the available evidence, to be unaffected by a later metamorphism.

The diabase is typical of dike rocks. It is composed of fresh plagioclase laths with relict subophitic primary and secondary amphiboles, biotite and iron oxides as the major minerals. It does not show evidence of later metamorphism.

The granite is clearly unrelated to the rhyolite-epizonal granite suite found to the west. It represents a deeper erosional level in the basement. In Missouri (Kisvarsanyi, 1979; Bickford and others, 1981) the eastern limit of the rhyolite-epizone granite is marked by the same contrast. In Missouri these mesozonal rocks yield older ages, ca. 1450-1500 m.y. on zircons. The same may hold true here although there is no direct evidence to support either an older or younger age.

The diabase is unusual. No diabase intrusions have been documented from the rhyolite-epizone granite suite from Arkansas or contiguous states. They are reported, but are by no means common from wells in Missouri (Kisvarsanyi, 1975). There is no direct evidence for the age of this dike but it is probably Precambrian based on the rather altered character of the samples near the basement surface.

The gravity and magnetic gradient used to draw the eastern limit of the rhyolite-epizone granite is at least partly valid within the rather loose control limits available. The petrographic data would suggest that the rocks east of the gradient are a deeper erosional level and may well be older than those to the west.

Xenoliths

The xenoliths found in the exotic carbonate dike in Conway County contain no basement material of the granite-rhyolite suite. All the clasts examined are from the rocks formed at a deeper crustal level.

Igneous dikes, sills and breccias were described from this area by Croneis and Billings (1929), although not from the locality sampled here. They reported medium grained, semiporphyritic, "alkaline syenite and aegirite granite" in addition to more abundant fragments of shale, sandstone, and ouachitite in the Oppello breccia (N½, SE¼, sec. 2, T. 5 N., R.17 W.) From their description it is unclear if the aegirine granite is a fenitized basement xenolith or a variation of the characteristically undersaturated syenitic rocks of Cretaceous age.

The most conspicuous basement clasts are of a granitic gneiss. This rock is composed of plagioclase, quartz and microcline with biotite, hornblende and iron oxides as prominent minerals and zircon, apatite, and sphene in trace amounts. Secondary minerals include calcite, feldspar alterations and chlorite. The texture is completely reorganized into a clearly metamorphic fabric. There is a faint preferred orientation of platy minerals and the quartz-feldspar has a granular homeoblastic texture. The rocks appear to be granitic to granodioritic in composition and the product of regional metamorphism. The assemblage is not diagnostic but is probably in the intermediate almandine-amphibolite facies but the mineralogy does not preclude hornblende granulite facies.

Another type of fine grained xenolith may also be from the basement. This is a metadiorite composed almost entirely of biotite, plagioclase and lesser iron oxides. The texture is relict igneous and appears to have been recrystallized during a hornfelsic metamorphism.

The xenoliths are distinctly different from the igneous rock found at the basement surface to the north. They were formed at a much greater depth. There is a good possibility that the granite-rhyolites are at the basement surface here but the dike tapped xenoliths from a much greater depth. The closest known gneisses that resemble those found as xenoliths occur in the eastern Arbuckle Mountains of Oklahoma. Phases of the Blue River gneiss (Denison, 1973) are very similar to the gneiss xenoliths. The metadiorite xenoliths are not similar to any known surface or subsurface rocks.

Northeast Arkansas

One well examined reached crystalline rocks in the Mississippi Embayment. The crystalline gneisses are overlain by a thick sequence of red arkosic sandstones of uncertain age and these in turn by a gray arkosic sandstone that is probably equivalent in stratigraphic position to the Lamotte Sandstone of the St. Francois Mountains.

The Dow Chemical No. 1 Lee Wilson and Company, Mississippi County, penetrated about 537 feet (14,305′ to 14,842′ TD) of granitic gneisses with a curious mineralogy. The rock is interpreted as a light colored granitic gneiss with intervals of dioritic gneiss originally as dikes or sills in the granitic host. The granitic gneiss is composed mostly of plagioclase and quartz with lesser but substantial amounts of microcline and muscovite. Smaller to trace amounts of chlorite, iron oxides, anhydrite, sphene, carbonate, sillimanite, leucoxene, epidote, biotite, apatite and zircon also occur.

The proportions of minerals appear to vary considerably in cuttings (no cores were taken) suggesting the rock is rudely banded. Even in cuttings there is well defined preferred orientation of the micas. The texture of the quartz-feldspar is an equigranular mosaic averaging about 0.5 to 1.2 mm. The unusual and erratic mineralogic composition (discussed more fully in the appendix) suggests the granitic gneiss was derived from a sedimentary rock.

The dioritic gneiss is composed mostly of plagioclase and hornblende with lesser to trace amounts of biotite, epidote, sphene, iron oxides, quartz and apatite. The plagioclase appears to be in the andesine composition range and is extensively altered to sericite. The hornblende is a mildly pleochroic, pale green variety with a slightly bluish tint. The biotite is olive-green and partly altered to chlorite. Epidote is chiefly disseminated crystals but is also found as minor veinlets. The grain size averages 0.15 to 0.3 mm and was never observed to exceed 0.7 mm. The dioritic gneiss has a rather consistent bulk composition that suggests an igneous origin.

Overlying the gneiss, from about 12,620 to 14,305 feet is a red arkosic sandstone of uncertain age. The arkose is generally well sorted, medium to fine-grained with irregular rounding. Argillaceous material is common in some chips but the interval as a whole is well within the sand size range. Feldspars are common with microcline, plagioclase and mesoperthite clasts locally abundant. Lithic clasts are present in lesser amounts, mostly rhyolite and micrographic quartz-perthite fragments. Overgrowths are conspicuous around many quartz grains, outlined by a hematite rim. Other detrital minerals include muscovite, iron oxides, tourmaline, and zircon. In most intervals the sand is cemented with secondary silica and minor carbonate. No anhydrite was identified. Some cutting chips in the samples were porous. The grain size varies from about 1mm to 0.2 mm in cutting chips. There is no evidence of metamorphism.

The clasts in the arkose indicate a mixed source. The microcline and muscovite are indicative of the underlying gneiss. The rhyolite and micrographic quartz-perthite are from either the rocks of the Northeast Oklahoma Province or from rock equivalent to those found in the St. Francois Mountains or from both. Those in the St. Francois area are demonstratably closer.

Overlying the red arkose is a grey arkosic sandstone, probably equivalent to the Lamotte Sandstone of the St. Francois area. The sand is remarkably similar to the underlying red arkose except for a grey to off-white color, more abundant carbonate and a more quartzose composition. The sandstone is gradational at about 12,290 feet with an overlying dolomite. This is probably equivalent to the

Bonneterre Formation of the St. Francois area. This light colored, shallow water carbonate is overlain by a very thick sequence considered to be Late Cambrian and Lower Ordovician in age by Apache Corporation (successors to Dow Chemical Oil and Gas). Much of this extensive interval has a dark basinal look but received only a cursory inspection by me during sample examination. Lamprophyric dikes cut parts of the Lamotte, Bonneterre and basement sequences. These are almost certainly the equivalents to the Cretaceous dikes found in the Ouachita Mountains.

The Embayment well was drilled close to the center of the Mississippi Valley graben as outlined by the gravity and magnetic interpretation of Kane and others (1981). They estimated an average depth to magnetic basement of 5.0 km (16,400 feet). Ginsburg and others (1983) calculated a thickness of "Precambrian and early Paleozoic sediments" of 3.0 km (9840 feet) using refraction seismic data. These two papers provide an excellent summary of the previous work and ideas concerning the deep Embayment.

This is probably the most important well drilled in the Embayment. I believe the biggest surprise is that the well did not penetrate St. Francois type rocks (i. e. rhyolites and epizonal granites) that are known to occur to the north and eastward in Tennessee. The type of crystalline rocks found in the well demonstrate that if these rocks were ever present in the area they have been removed by erosion. No isotopic data is available on the age of crystalline rocks so one is free to speculate. The St. Francois rocks yield zircon ages of formation of about 1485 m.y. and a lesser grouping around 1385 m.y. (Bickford and Mose, 1975). These rocks show no evidence of regional metamorphism petrographically, although the Rb/Sr systems are disturbed. The environment of the granite-rhyolite suites is almost certainly anorogenic. Yet the regional metamorphism seen in the crystalline gneisses was almost certainly during an orogenic episode. The question is whether the metamorphism of the gneisses could have occurred at great depth without disturbing the nearby granite-rhyolite suite. The granite-rhyolite suite must have been close based on the abundance of clasts in the overlying arkose. It is my opinion, based on the pristine textures and the tectonically undisturbed character of the volcanic rocks, that these rocks are either younger than the metamorphism or that the metamorphism was restricted geographically. If the metamorphism is older than the St. Francois rocks the age is most probably in the 1500–1650 m.y. range (the oldest ages known from the Missouri-Kansas area, Bickford and others, 1981). If the metamorphism is younger than the granite-rhyolite suite the age is almost certainly older than about 1000 m. y., the youngest ages of regional significance in the southern continental interior. I tend to think of them as older than 1500 m.y. It will be no easy task to resolve. A mica age would probably yield a minimum for the time of metamorphism. The few zircons seen in the gneiss, the probable sedimentary origin, and the limited amount of sample will all but preclude a U/Pb age that can be compared with the St. Francois outcrop. The rock found in the well is unusual, in any case, because it represents a deep erosional cut into the crust in an area characterized by surface and near surface igneous rocks and a higher grade of metamorphism than is common in the southern interior.

The age of the overlying red arkose is an even knottier problem--not amenable to any known analytical solution. Clearly the arkose is younger than the gneiss and the granite-rhyolite suite and older than grey arkose, probably equivalent (in stratigraphic position if not strictly in time) to the Late Cambrian Lamotte Sandstone. Thus, the red arkose is older than about 510 m.y. and younger than about 1500 m.y. Arkoses of this type are known from other wells in the Embayment (Lidiak and others, 1982). To my knowledge, the 1685 feet penetrated in this well is the thickest and only complete section drilled to date. It probably does not represent normal marine sedimentation. Does it represent the sediments of the rift opening of the Reelfoot aulacogen? If it does and the Reelfoot was caused by the same forces as the Southern Oklahoma aulacogen then the most likely time of sedimentation is Cambrian. The red and grey arkoses are exceptionally similar in grain size, sorting, diagenetic effects and provenance. The small differences in the two could easily be accounted for by the transition from a non-marine to marine environment. For this reason, I find the young age of sedimentation most attractive.

The Cockrell No. 1 Carter (sec. 4, T. 4 N., R. 1 E.) was drilled in St. Francois County, some sixty five miles to the southwest of the Dow No. 1 Wilson. The well penetrated a reported 531 feet of Precambrian granite wash, topped at 14,350′ (−14,145′), according to records of Sun Exploration and Production Company (Charles Vertrees, Jr. personal communication). If correct, and the granite wash is the same unit as the red arkose found in the Wilson well, the potential area underlain by these rocks would be substantial.

Based on fragmentary data in the Embayment and a closely constrained understanding of southern Oklahoma, the Reelfoot and Southern Oklahoma aulacogens do not appear to share a particularly analogous geologic history during the Paleozoic.

Conclusions

The wells drilled in northwest Arkansas are a further documentation of the extensive granite-rhyolite complexes that typify the basement of the southern continental interior (Muehlberger and others, 1967).The occurrence of granitic gneisses as xenoliths and in a well in far northeast Arkansas suggest that the granite-rhyolite may be a relatively thin veneer on a more shield-like crust. The granitic gneiss found in the Mississippi Embayment well, in contrast to surface and near surface igneous rocks in the northwest, is estimated to have formed at depths of between 15 and 20 km. This indicates a much deeper level of erosion than is known either east or west of the Embayment. A thick red arkose overlies the crystalline gneisses and below probable Lamotte Sandstone equivalent. This unmetamorphosed sedimentary section, found in several Embayment wells, may have been deposited during the rift phase of the Reelfoot aulacogen.

Acknowledgements

The samples for most of the wells in northwest Arkansas were supplied by W. M. Caplan of the Arkansas Geological Commission. He also provided most of the data used in Table 1. Dr. R. C. Cohoon, Arkansas Tech University, collected the xenolith sample from Conway County. C. A. Baird, Apache Corporation, Houston, made the samples of the Dow No. 1 Wilson in Mississippi County available for examination and was very generous in discussing the results of the Embayment drilling. Dr. Eva B. Kisvarsanyi, Missouri Department of Natural Resources, tracked down the basement core samples described in the study of Kurtz and others (1975). Charles Stone provided invaluable assistance in all phases of the study. The Arkansas Geological Commission provided the drafting for the text figure.

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Table 1. — List of wells to basement from which samples have been examined.

County Number	Well Name	Location	Basement depth (ft) Subsea (ft)	Total depth (ft) Penetration (ft)	Rock type
Bn–1	Ozark No. 1 Curry	33–18N–33W	2205 –1015	2236 31	Rhyolite porphry
Bn–2	Layne Western No. 2 Decatur	11–19N–33W	2104(?) –882	2125 21	Microgranite porphry
Bn–3	St. Joe Minerals[1] No. BE–2	25–19N–33W	1952 –627	1974 22	Micrographic granite
Ca–1	St. Joe Minerals[1] No. AK–CA–1	30–21N–25W	2070 –630	2093 23	Micrographic granite
Cr–1	Gulf No. 1 Mowrey	14–10N–32W	5528 –4622	5701 173	Micrographic granite
Fa–1	ARCO No. 1 Edgmon	6– 7N–12W	12095	12171 76	Granite and diabase
Fr–1	Tenneco No. 1 Conaster	17–10N–27W	8234 –7764	8280 46	Microgranite porphry
Fr–2	SEECO No. 5 Lessley	21–10N–28W	7585 –6890	7746 161	Granite
Fr–3	Arkla No. 1 Ark. Valley Trust	9– 9N–28W	8430 –7656	8636 206	Micrographic granite
Lg–1	Exxon No. 1 Tanner	15– 6N–28W	12225 –11635e	12349 24	Metarhyolite
Md–1	Independent No. 1 Banks	6–16N–27W	2392 –847	2515 123	Rhyolite porphry
Md–2	Layne Western No. 1 Huntsville WW	3–16N–26W		3290	Metarhyolite
Md–3	War Eagle No. 1 Brener	13–15N–26W	2286 –880	2320 34	Rhyolite[2]
Ms–1	Dow Chemical No. 1 Wilson	14–12N– 9E	12620* –12368	14842 2222	Arkose and granitic gneiss
Nt–1	Pan American No. 1 USA	28–13N–22W	4946 –3250	4996 50	Microgranite porphry
Sr–1	Petroplex No. 1 Wheeler	18–14N–16W			Granite

Most data supplied by W. M. Caplan, Arkansas Geological Commission.

* base of Lamotte, granitic gneiss at 14305′ (–14053′)
e estimated

[1] Data from Kurtz and others (1975)

[2] Determination by Miser and Ross (1925), samples not examined

APPENDIX

Thin sections of each of the samples examined are described and arranged alphabetically within the county keyed to Table 1 and Figure 1. The mineral percentages are based on point counts where enough sample was available. In other descriptions the minerals are listed in estimated order of abundance.

Ozark No. 1 Curry, sec. 33, T. 18 N., R. 33 W., Benton County, Bn-1-1, 2210–22 feet

RHYOLITE PORPHYRY. Groundmass, plagioclase, perthite, iron oxides, chlorite, feldspar alterations, epidote, apatite, zircon.

The groundmass appears to be exceptionally quartz-rich intergrown with feldspar. Local coarsenings accompany quartz rich portions. Perthite and plagioclase phenocrysts are partly altered. A perthite rim surrounds some of the plagioclase which is not otherwise zoned. Hematite dust is more abundant in perthite. Chlorite is mostly disseminated in the groundmass as weakly birefringent masses and shreds. Grain size: phenocrysts, 2.0 mm; groundmass, 0.05 mm average. Texture: porphyritic-felted.

The well penetrated a rhyolite that must be close to the edge of the present limits of the volcanic field.

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Layne Western No. 2 Decatur WW, sec. 11, T. 19 N., R. 33 W., Benton County, Bn-2-1, 2120–25 feet

MICROGRAPHIC MICROGRANITE PORPHRY: Micrographic quartz-feldspar, perthite, quartz, iron oxides, feldspar alterations, chlorite, leucoxene, apatite, zircon.

All cutting chips are less than 1.0 mm in size. Phenocrysts of perthite and lesser quartz are set in a very delicately micrographic groundmass. Hematite dust and vacuoles are very common in the feldspars, obscuring the composition. Feldspar alterations are found as fine sericitic masses. Chlorite is the only accessory mineral present in significant amounts. Grain size: 0.7+ mm. Texture: porphyritic-micrographic.

The very fine grained quartz-perthite intergrowth is typical of chilled granites near the contact with the host rhyolite. The quartz phenocrysts, even in small amounts, are somewhat unusual. The well was drilled close to the crest of a very large gravity and magnetic anomaly. Clearly the rock found in the well is not the cause of the anomaly and a deeper source is the more likely explanation.

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St. Joe Minerals Hole No. BE–2, sec. 25, T. 19 N., R. 33 W., Benton County, Bn-3-1, 1967 feet

MICROGRAPHIC GRANITE PORPHYRY. Perthite, 30.8%; plagioclase, 24.9%; quartz, 22.5%, plagioclase alterations, 10.8%; amphibole, 5.3%; chlorite, 2.6%; epidote, 1.4%, iron oxides, 1.3%; sphene-leucoxene, 0.3%; apatite, 0.1%; zircon, tr.; carbonate, tr.

Phenocrysts of plagioclase rimmed by perthite are set in a generally micrographic quartz-feldspar matrix. The plagioclase contains nearly a third clay-sericite-epidote replacement. The perthite is rather uniformly clouded with vacuoles and hematite dust. The amphibole is mostly a pale brownish green common variety with lesser actinolite. Chlorite is a common associate with the amphibole. Epidote is found as well formed crystals in an intergranular position and smaller crystals in feldspar. Carbonate is found in a thin veinlet. Grain size: phenocrysts, 6 mm; groundmass 0.8–1.2 mm. Texture: porphyritic-micrographic.

The well was drilled close to the others that penetrated similar rocks. It is one of only two cores available from the basement of northwest Arkansas and is believed to be representative of the type of granites found in this area.

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St. Joe Minerals Hole No. Ak-CA-1, sec. 30, T. 21 N., R. 25 W., Carroll County. Ca-1-1, 2085 feet

MICROGRAPHIC GRANITE PORPHYRY. Perthite, 40.0%; plagioclase, 25.1%; quartz, 23.6%; chlorite, 4.2%; plagioclase alterations, 4.2%; carbonate, 1.6%; iron oxides, 1.0%; sphene-leucoxene, 0.3%; apatite, tr.; zircon, tr.

Phenocrysts of plagioclase rimmed by perthite are set in a micrographic quartz-feldspar matrix. The former femic minerals have all been replaced by chlorite. The chlorite is in clots with abundant sphene-leucoxene and iron oxides. Feldspar alterations are mostly at the expense of plagioclase. The perthite contains abundant evenly disseminated vacuoles and hematite dust. Carbonate is found as veinlets cutting the rock and as a replacement is some accessory mineral clots. Grain size: phenocrysts, 3.8 mm; groundmass 1.5 mm. Texture: porphritic-micrographic.

Ca-1-2, 2091 feet

MICROGRAPHIC GRANITE PORPHYRY. Perthite, 37.9%; plagioclase, 31.4% quartz, 22.1%; plagioclase alterations, 3.3%; chlorite, 2.0%; hornblende, 1.2%, iron oxides, 1.0%; sphene, 0.6%; epidote, 0.3%, carbonate; tr.; zircon, tr.; apatite, tr.

Phenocrysts of plagioclase, some quite large, are rimmed by perthite. These are set in a generally micrographic quartz-feldspar matrix. The perthite is dusted with extensive and uniform hematite and vacuoles. Plagioclase is calcic oligoclase and contains locally intense sericite-epidote-clay alterations. The amphibole is in clots with other accessories and both a pale common green variety and actinolite are present. Chlorite is associated with these clots. Sphene is a reddish variety in anhedral crystals. Both apatite and zircon are found as numerous small crystals. Grain size: phenocrysts, 7+ mm; groundmass 1.0–1.8 mm. Texture: porphyritic-micrographic.

Two thin sections were cut from each interval. Point counts showed substantial mineralogic abundance differences on the scale of a thin section. This is almost certainly due to the large plagioclase phenocrysts. These samples are the only ones to show changes in alteration with depth. The upper sample is more weathered. This is reflected in the conversion of sphene to leucoxene and the absence of hornblende. Carbonate as veins and a replacement of former femic mineral is conspicuous in the upper sample. The general texture and mineralogic abundances are probably fairly typical of all the epizonal granites in northwest Arkansas.

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Outcrop SE¼ SE¼ NE¼, sec. 34, T. 6 N., R. 17 W.
Conway County

Xenoliths brought to the surface by an exotic carbonate dike yield evidence of basement compositions in this area. Some of the xenoliths are granitic gneisses of almost certain Precambrian age. The depth from which these xenoliths were derived cannot be determined.

The carbonate contains smaller xenocrysts of a variety of minerals that indicate it is related to the Cretaceous lamprophyric suite. The dike sample, collected by R. R. Cohoon of Arkansas Tech University, contains xenoliths as large as 4 cm in length. The dike is probably one of those studied by Jackson and Steele (1976), Mitchell and Jackson (1979) and McCormick and Heathcote (1979).

Several of the xenoliths are granitic in composition and have a well defined metamorphic texture. These are almost certainly the product of regional metamorphism and not contact hornfels from the dike. This is based on a gneissic preferred orientation of certain minerals and a decided difference in mineralogy between the same mineral phases found in the host and xenolith. For example, both biotite and hornblende in the xenoliths have much different properties than the same minerals in the dike. There is, however, some fenitization near the margins of the xenoliths.

The gneisses are composed mostly of plagioclase, quartz and microcline. There are a variety of accessory minerals. Hornblende, biotite and iron oxides are most common with zircons, apatite and sphene in trace amounts. Plagioclase is most common occupying 40 - 50% of the volume. It is in the oligoclase composition range and is locally extensively altered but is generally quite fresh. Microcline accounts for a quarter to a third of the rock and is generally quite fresh. Quartz is unstrained and accounts for about a fifth of the volume. The percentage of dark minerals varies considerably in the xenoliths. The color index ranges between 5 and 15.

The texture is homeoblastic among the light constituent and faintly gneissic with a poorly defined preferred orientation in dark minerals. Biotite is found in several xenoliths as rather large reddish brown poikiloblastic crystals. Grain size is fairly even between 0.7 and 2.0 mm average and rarely exceeding 3 mm.

The xenoliths are far from pristine. Calcite has invaded around virtually every grain boundary along some cleavage. These films of carbonate are rarely wider than 0.03 mm, but are much wider and better developed near the contact with the host carbonate. Some changes in mineralogy are also seen with the crystalloblastic growth of fine grained minerals identical to those carried in the host. As pervasive as these changes are, the basic framework and original mineralogy remain clear.

Another darker xenolith type may also represent basement rock. This rock is composed almost entirely of plagioclase, biotite and iron oxides. The texture of the plagioclase is relict igneous but the the biotite is completely recrystallized. There is a suggestion of preferred orientation in the small biotites but it is poorly defined, at least in the plane of the thin section. Most plagioclase laths are about 0.3 to 0.7 mm in length, are poorly twinned and carry flakes of disseminated biotite. Aggregates of larger biotite are crowded between plagioclase laths. The biotite is a strongly pleochroic pale reddish brown to dark olive-green variety. The biotite averages 0.1 to 0.3 mm in length and accounts for almost exactly half of the rock with another 5% of iron oxides making the color index greater than 50. This type of xenolith is essentially a metadiorite and is more likely the product of hornfelsic metamorphism.

No xenoliths of the granite-rhyolite terrane found to the north were identified and they should be readily recognizable even in small size. No clasts were recognized that are similar to the boulders found in the Blakely Sandstone (Denison and others, 1977). These rocks could be present at the basement surface here, simply not picked up during the passing of the carbonate rich host.

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Gulf No. 1 Mowrey, sec. 14, T. 10 N., R. 32 W.
Crawford County, Cr-1-1 5680–90 feet

MICROGRAPHIC GRANITE: Perthite, quartz, plagioclase, iron oxides, feldspar alterations, chlorite, hornblende, sphene-leucoxene, pyroxene, apatite, zircon.

Most of the cutting chips are about 0.7 mm in length, none exceed 1.0 mm. The feldspars show considerable alteration to fine sericite. A number of chips show a delicate micrographic intergrowth around feldspar phenocrysts. Small amounts of relict pyroxene are present. The most common femic mineral is a fresh olive-green hornblende. Chlorite with attendant sphene has replaced portions of the femic minerals. Grain size: 1+ mm. Texture: micrographic-porphyritic.

The micrographic granite is typical of those associated with a rhyolitic volcanic field. The relict pyroxene is somewhat unusual but is reported in the Spavinaw Granite outcrop and also in scattered samples from the subsurface. The olive-green primary hornblende is very rare to the west in Oklahoma.

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Arco No. 1 Edgmon, sec. 6, T. 7 N., R. 12 W.
Faulkner County, Fa-1-1 12090–100 feet

DIABASE. Plagioclase, actinolite, biotite, iron oxides, hornblende, chlorite, feldspar alterations, apatite.

Plagioclase is in fresh, well twinned laths. There is a very even zoning from labradorite near cores to andesine or possibly oligoclase at margins. The original femic minerals included a common brownish green hornblende and a reddish brown biotite. Most have been replaced by a pale apple green actinolite and a brownish secondary biotite with minor chlorite. Apatite is found as abundant fine needles. Feldspar alterations are abundant in only a few cutting chips. Grain size: 0.8–1.2 mm. Texture: relict subophitic.

Fa-1-2 12160–70 feet

GRANITE: Microcline, plagioclase, quartz, biotite, hornblende, iron oxides, feldspar alterations, sphene epidote, apatite, leucoxene, zircon.

Microcline is partly perthitic, well twinned, very fresh and contains small poikilitic inclusions of quartz and plagioclase. Plagioclase is mottled with erratic alterations and is intermediate oligoclase in composition. Quartz is in single, essentially unstrained crystals. Biotite is abundant as large, fresh reddish brown books. Hornblende is a strongly pleochroic dark green variety. Sphene is in well formed but not euhedral crystals. Leucoxene is found around some iron oxides. A few chips appear mildly sheared. Grain size: 2+ mm. Texture: hypidiomorphic(?).

The diabase was identified in the 12090–100′ sample. The snowy white granite was first seen in the sample at 12130–40′. The granite is in marked contrast to those found to the west. This rock is typical of mesozonal batholithic mass rather than the fine-grained epizonal granite associated with the rhyolites. The grain size exceeds the cutting chip size (2 mm) but based on the size of the biotite flakes the rock is quite coarse. The texture is difficult to determine because of this but appears to be an igneous texture essentially unaffected by any metamorphism. The diabase clearly cuts the granite as a dike or sill and is the only rock of this composition found in northern Arkansas. It is most probably Precambrian in age but strict evidence for this is lacking. The overlying sedimentary rock is a slightly sandy dolomite showing no evidence of recrystallization.

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Tenneco No. 1 Conatser, sec. 17, T. 10 N., R. 27 W.
Franklin County
FR-1-1 8260–70 feet

MICROGRAPHIC MICROGRANITE PORPHRY: Micrographic quartz feldspar, 62.7%, plagioclase, 14.8%; perthite, 8.7%; chlorite, 5.2%; feldspar alterations, 5.2%; sphene-leucoxene, 1.1%; quartz, 1.1%; epidote, 0.3%; iron oxides, 0.3%; actinolite, 0.3%, calcite, 0.1%, zircon, tr.; sphene, tr.; apatite, tr.

Phenocrysts of perthite and plagioclase are set in a partially micrographic groundmass of quartz-feldspar. Feldspar alterations are extensive in the plagioclase phenocrysts. These are mostly sericite and epidote. Chlorite is probably after biotite and contains granular sphene-leucoxene. Sphene is also found in small well formed primary crystals. Actinolite is found in one chip associated with chlorite. Apatite is as rather numerous tiny crystals. Grain size: phenocrysts, 2+ mm; groundmass, 0.2 mm. Texture: porphyritic-micrographic.

The granite is unusually quartzose compared to those found to the west in Oklahoma. The occurrence of well formed sphene is not typical. The very fine chilled texture suggests the well was drilled close to the contact with the rhyolites.

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SEECO No. 5 Lessley, sec. 14, T. 10 N., R. 28 W.
Franklin County
FR-2-1 7730–40 feet

GRANITE: Perthite, quartz, plagioclase, chlorite, feldspar alterations, hornblende, iron oxides, sphene, biotite, apatite, flourite, zircon.

The cuttings are very fine (few exceed 1 mm) and almost all chips are single minerals. A few are clearly micrographic quartz-perthite but this is probably only a local development in an essentially hypidiomorphic texture. Hornblende is a common type. Sphene is well crystallized with a dark reddish color. Plagioclase is in the oligoclase composition range and is locally extensively altered. Grain size: 1+ mm. Texture: partly micrographic.

The granite is typical of those found associated with the rhyolites. The texture suggests it is more removed from contact with the rhyolite than FR-1-1.

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Arkla No. 1, Arkansas Valley Trust, sec. 9, T. 9 N., R. 28 W., Franklin County
FR-3-1 8620–30 feet

MICROGRAPHIC GRANITE: Perthite, quartz, plagioclase, iron oxides, feldspar alterations, chlorite, hornblende, biotite, sphene-leucoxene, zircon, apatite.

Cutting chips rarely exceed 1 mm. Most of the rock is composed of quartz, plagioclase and perthite. There is some well developed micrographic intergrowth but there is also some quartz as discrete crystals. The hornblende is an apple-green variety and is partly altered to chlorite. There is minor brownish green biotite. The feldspar contains hematite dust and vacuoles but is generally fresh. Grain size: 1+ mm. Texture: micrographic-porphyritic.

The granite is similar to that found in Fr-2-1. The hornblende and minor biotite are unusual in granites to the west.

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Exxon No. 1 Tanner, sec. 15, T. 6 N., R. 28 W.
Logan County
Lg-1-1 12340–50 feet

METARHYOLITE: Groundmass, 53.8%, microcline perthite, 30.0%; quartz, 13.1%; plagioclase, 1.6%; biotite, 0.8%; iron oxides, 0.3%; iron sulfides, 0.2% chlorite, tr.; apatite, tr.; feldspar alterations, tr.; sphene, tr.; flourite, tr.; carbonate, tr.

Very numerous phenocrysts of plagioclase quartz and microcline perthite are set in a hornfelsic groundmass of mostly quartz-feldspar. A few phenocrysts of biotite, now much altered to chlorite and attendant sphene, are also present. Biotite is also found as a crystaloblastic mineral in the groundmass. The groundmass is erratic in grain size and is generally fresh. Feldspar alterations of sericite are mostly restricted to the plagioclase phenocrysts. Some of the phenocrysts have a rim of seived groundmass crystals. Grain size: phenocrysts, to 5 mm; groundmass, 0.01–0.05 mm. Texture: relict porphyritic-granoblastic.

The metamorphism reflected by the groundmass recrystallization is typical of contact metamorphism. Several wells to the west in far east central Oklahoma are also metamorphosed but the recrystallization found here is more extensive. Biotite as relict phenocrysts or as a metamorphic mineral is not found in Oklahoma.

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Independent O & G No. 1 Banks, sec. 6, T. 16 N., R. 27 W., Madison County
Md-1-1 2500–04 feet

RHYOLITE PORPHRY: Groundmass, plagioclase, perthite, chlorite, iron oxides, feldspar alterations, sphene-leucoxene, epidote, apatite, zircon.

Phenocrysts of altered plagioclase and sparse perthite are set in a quartz-feldspar groundmass. The texture of the groundmass is not well defined – parts appear to have optically oriented rodlike quartz. Feldspars are altered and contain hematite dust. Plagioclase contains sericite flakes and epidote granules. Chlorite masses replace portions of some feldspar phenocrysts and is in the groundmass as probable pseudomorphs after a femic mineral and disseminated shreds. The iron oxide granules are partly surrounded by sphene-leucoxene. Grain size: phenocrysts, 2 mm; groundmass, 0.2 mm. Texture: porphyritic-felted.

The rhyolite is typical in mineralogy but somewhat coarser grained than average.

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Layne Western No. 1 Huntsville WW, sec. 3, T. 16 N., R. 26 W., Madison County
Md-2-1 3285–90 feet

METARHYOLITE PORPHRY: Groundmass, perthite, plagioclase, quartz, feldspar alterations, iron oxides, chlorite, zircon, apatite.

Phenocrysts of feldspar and lesser quartz are set in a finely granoblastic groundmass composed of the same minerals. The largest cutting chip is about 1.5 mm in length. Hematite dust and vacuoles cloud feldspars, obscuring the composition. Chlorite is found as the replacement of former femic minerals and as fine shreds in the groundmass. The groundmass is composed almost entirely of a granoblastic mosaic of quartz-feldspar. Grain size: phenocrysts, 1.3+ mm; groundmass, 0.03 mm. Texture: relict porphyritic-granoblastic.

The metamorphism of the rhyolite is unexpected but is very clear in thin section. It suggests a nearby plutonic mass that was not penetrated in any well, to be found only at depth.

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War Eagle No. 1 Brener, sec. 13, T. 15 N., R. 26 W. Madison County
Md-3-1 2286–2320 feet

Miser and Ross (1925) report "the rock is composed of phenocrysts of plagioclase and orthoclase in a very fine grained groundmass of feldspar and quartz. The rock would be described as a porphyritic rhyolite and the structure indicates that it may be a volcanic flow or an intrusive of limited size that crystallized at shallow or intermediate depths."

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Dow No. 1 Lee Wilson & Co., sec. 14, T. 12 N., R. 9 E. Mississippi County
Ms-1-1 12790–00 feet, Ms-1-2 13160–70 feet,
Ms-1-3 13460–70 feet, Ms-1-4 13820–30 feet,
Ms-1-5 14100–10 feet, Ms-1-6 14300–10 feet,
Ms-1-7 14470–80 feet, Ms-1-8 14590–00 feet,
Ms-1-9 14700–10 feet, Ms-1-10 14780–90 feet,
Ms-1-11 14840–50 feet.

About 2222′ of rock was penetrated beneath Lamotte Sandstone – a red arkose of uncertain age and a granitic gneiss of almost certain Precambrian age. The top of the arkose was picked on a color change from light grey and off white to a dark red. Six intervals were examined in thin section.

The arkose is red due to hematite found as detrital grains, disseminated dust in feldspars, a coating of quartz grains and in a locally argillaceous matrix. The sorting, grain size, and rounding vary considerably. The finer intervals tend to be poorly sorted and less rounded than the coarser sands. An argillaceous matrix is most common in those chips having fine sand or silt size clasts.

The sand size clasts are mostly quartz as single grains. The content ranges from about 50% in the finer sands to about 80% in the coarser, better sorted sands. Mesoperthite, microcline, and plagioclase are all common with perthite being most abundant. Lithic clasts are all of fine grained igneous rocks. Rhyolite and micrographic quartz-perthite clasts from a granite make up almost all the lithic debris. In the better sorted, originally porous sands, quartz overgrowths are well defined by a hematite film around the original grain.

Heavy minerals are most common in the finer sands. These include minor to moderate amounts of hematite, with lesser to trace amounts of muscovite, tourmaline and zircon. Secondary minerals include rather common quartz as overgrowths and as a cement and minor carbonate as veinlets and small blebs between sand grains and rare replacement of feldspar grains.

In the six intervals chosen for thin section there is no recognizable systematic change in grain size, sorting, rounding, or mineralogy. The section is surprisingly uniform. There is absolutely no evidence of metamorphism. The diagenetic changes seen in the arkose are those expected from simple burial. The occurrence of porous sandstone in the samples supports simple diagenesis.

The arkose is underlain by a granitic gneiss, topped at about 14305'. Five intervals were examined in thin section. The sequence varied considerably in bulk composition and the occurrence of accessory minerals. Both a light colored granitic gneiss and a darker, dioritic gneiss are found in the cuttings, suggesting a crude banding in the rock. Plagioclase and quartz make up most of the granitic gneiss with varying but substantial microcline and muscovite. Smaller to trace amounts of chlorite, feldspar alterations, anhydrite, iron oxides, sphene, carbonate, sillimanite, leucoxene, biotite, epidote, apatite, and zircon make up the remainder of the gneiss.

Plagioclase is in the oligoclase composition range, is generally fresh but locally mottled with extensive clay-sericite-epidote as an alteration. Microcline is well twinned, variable in amount and is generally very fresh. Quartz is mildly to moderately strained and abundant. The point counts of the cuttings, if truly representative of bulk composition, vary considerably. Plagioclase accounts for 30–40%, quartz 35–40% and microcline about 5 to 15%. Accessory minerals represent 10–25% of the volume.

Most conspicuous of the accessory minerals is muscovite, found both as well formed books and aggregates of small shreds. It varies in volume from about 5 to 20% and is inversely abundant to microcline. The most unusual mineral is anhydrite. It occurs as disseminated, well formed crystals showing good pseudocubic cleavage and in intimate association with muscovite. It accounts for about 1 to 5% of the volume where present. The occurrence is not systematic with depth – the topmost crystalline sample contains none. The distribution and textural position of anhydrite suggests it is not a secondary mineral.

Sillimanite was tentatively identified in at least one sample. It occurs as slender needlelike crystals about 0.006 mm wide and as long as 0.8 mm. The extremely small size precludes confident identification. Some of the finely divided muscovite may be after sillimanite. Zircons occur erratically; some are very small and euhedral and others are larger, more rounded crystals. Biotite is an erratically distributed olive-green variety. Chlorite replaces portions of the biotite but is also a primary mineral in some intervals.

Cutting chips of dioritic composition are found in almost every sample. This rock is composed mostly of plagioclase, hornblende and iron oxides with locally abundant biotite and epidote. Small to trace amounts of sphene, apatite, chlorite and plagioclase alterations are also present. Quartz is present in some chips in moderate to small amounts. The rock is quite fine grained, averaging 0.15 to 0.3 mm. Preferred orientation in many chips is well defined. The dark minerals account for a bit more than half the rock volume.

The erratic mineral abundances and curious mineralogy of the granitic gneiss are suggestive of a sedimentary origin for part of all the sequence. Anhydrite is most unusual as an unevenly distributed mineral. Normally anhydrite occurs in veinlets and other replacements in crystalline rock, where overlain by an evaporitic sequence. Veinlets of carbonate are present in the gneiss but these carry no anhydrite. There were no evaporite minerals identified in the overlying red arkose. The dioritic gneiss could have been derived from a common dioritic igneous rock. The metamorphic grade is most likely to be in the sillimanite almandine-muscovite subfacies of the almandine-amphibolite facies of regional metamorphism. The mineralogy would suggest depth of burial of not less than 15 km and more likely deeper during metamorphism with temperatures in the 600–700^{o} C range.

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Pan Am No. 1 USA, sec. 28, T. 13 N., R. 22 W.
Newton County
Nt-1-1 4996 cir

MICROGRANITE PORPHRY: Quartz-feldspar groundmass, 83.3%; plagioclase, 10.6%; perthite, 2.9%; feldspar alterations, 1.5%; sphene-leucoxene, 1.0%; chlorite, 0.6%; iron oxides, tr.; apatite, tr.; zircon, tr., calcite, tr.

Phenocrysts of perthite and plagioclase are set in a quartz-rich groundmass showing a curious acicular appearance in thin section. This is caused by rodlike quartz crystals that have optical continuity over areas as large as 1.0 mm. Chlorite is found as pseudomorphs after a femic mineral as well as fine shreds scattered in the groundmass. Accessory minerals are in clots, mostly centered around iron oxides and sphene-leucoxene. The plagioclase phenocrysts are rather altered and locally contain poorly crystalline epidote. The texture has an incipiently crystallized appearance. Grain size: phenocrysts, 3+ mm; groundmass, 0.5–0.8 mm. Texture: porphyritic-acicular.

The curious rodlike shapes of quartz in thin section are common in certain units in Oklahoma. The texture is transitional with that found in the rhyolites. The fine grain size and texture seen here suggest the well was drilled close to the contact with rhyolite.

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Petroplex No. 1 Wheeler, sec. 18, T. 14 N., R. 16 W.
Searcy County
Sr-1-1 4830–35 feet

GRANITE: Microcline perthite, 43.4%; quartz, 27.5%; plagioclase, 18.9%; feldspar alterations, 3.4%; biotite, 2.8%; sphene, 2.0%; iron oxides, 1.0% hornblende, 0.5%; chlorite, 0.3%; zircon, tr.; apatite, tr.; epidote, tr., calcite, tr.

Microcline perthite is fresh and partly intergrown with quartz in a rather rude micrographic intergrowth. Plagioclase is much less abundant, is mildly zoned, is in the oligoclase composition range, and carries locally extensive feldspar alterations of sericite and lesser epidote. Quartz is essentially unstrained. Biotite is a red-brown variety, is partly converted to chlorite and is associated with other accessory minerals. These accessory clots contain zircon, apatite, iron oxides, sphene and a green common hornblende. Grain size: 3–4 mm. Texture: hypidiomorphic, partly micrographic.

The samples recovered from the well drilled fartherest from the main group of granite-rhyolite basement wells are the most unusual of any of the samples seen in northwest Arkansas. The occurrence of primary biotite, sphene and hornblende are decidedly atypical of rocks from their province. Indeed it is placed together with wells to the west more as a matter of convenience than conviction. The gravity and magnetic data suggest it is probably in the same province but close to the edge, which may account for its unusual petrographic character.

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REFERENCES

Bickford, M. E., Harrower, K. L., Hoppe, W. J., Nelson, B. K., Nusbaum, R. L., and Thomas, J. J., 1981, Rb–Sr and U–Pb geochronology and distribution of rock types in the Precambrian basement of Missouri and Kansas: Geological Society of America Bulletin, v. 92, p. 323–341.

Bickford, M. E., and Lewis, R. D., 1979, U-Pb geochronology of exposed basement rocks in Oklahoma: Geological Society of America Bulletin, v. 90, p. 540–544.

Bickford, M. E., and Mose, D. G., 1975, Geochronology of Precambrian rocks in the St. Francois Mountains, southeastern Missouri: Geological Society of America Special Paper 165, 48 p.

Croneis, Carey, and Billings, Marland, 1929, New areas of alkaline igneous rocks in central Arkansas: Journal of Geology, v. 37, p. 542–561.

Denison, R. E., 1973, Basement rocks in the Arbuckle Mountains in Ham, W. E., and others, Regional geology of the Arbuckle Mountains, Oklahoma: Oklahoma Geological Survey, Guidebook for Field Trip No. 5, Geological Society of America 1973 Annual Meeting, Dallas, Texas, p. 43–49.

Denison, R. E., 1981, Basement rocks in northeast Oklahoma: Oklahoma Geological Survey Circular 84, 84 p.

Denison, R. E., Burke, W. H., Otto, J. B., and Hetherington, E. A., 1977, Age of igneous and metamorphic activity affecting the Ouachita foldbelt, in Stone, C. G., ed., Symposium on the geology of the Ouachita Mountains: Arkansas Geological Commission, v. 1, p. 26–40.

Ginzburg, A., Mooney, W. D., Walter, A. W., Lutter, W. J., and Healy, J. H., 1983, Deep structure of the northern Mississippi Embayment: American Association of Petroleum Geologists Bulletin, v. 67, p. 2047–2052.

Hendricks, J. D., Keller, G. R., and Hildenbrand, T. G., 1981, Bouger gravity map of Arkansas: U. S. Geological Survey Map GP–944.

Jackson, K. C., and Steele, K. F., 1976, New data on some Arkansas igneous rocks: Geological Society of America Abstracts with Programs, v. 6, No. 1, p. 25.

Kane, M. F., Hindenbrand, T. G., and Hendricks, J. D., 1981, A Model for the tectonic history of the Mississippi Embayment and contemporary seismicity: Geology, v. 9, p. 563–567.

Kisvarsanyi, E. B., 1975, Data on Precambrian in drill-holes of Missouri including rock type and surface configurations: Missouri Department of Natural resources, Geological Survey. Report of Investigations 56, 24 p.

Kisvarsanyi, E. B., 1979, Geologic map of the Precambrian of Missouri: Missouri Department of Natural Resources, Geological Survey, Contribution of Precambrian Geology, Map No. 7.

Kurtz, V. E., Thacker, J. L., Anderson, K. H., and Gerdemann, P. E., 1975, Traverse in the Late Cambrian strata from the St. Francois Mountains, Missouri to Delaware County, Oklahoma: Missouri Department of Natural Resources, Geological Survey, Report of Investigations No. 55, 118 p.

Lidiak, E. G., Kersting, J. J., and Hinze, W. J., 1982, Basal sandstones in the subsurface of the central midcontinent, United States: Geological Society of America Abstracts with Programs, v. 14, No. 7, p. 547.

McCormick, G. R., and Heathcote, R. C., 1979, Mineralogy of the Morrilton alvikite dike, Conway County, Arkansas: Geological Society of America Abstracts with Programs, v. 11, No. 2, p. 163..

Miser, H. D., and Ross, C. S., 1925, Pre-Cambrian rhyolite rhyolite discovered in well in northwestern Arkansas: American Association of Petroleum Geologists Bulletin, v. 9, p. 115.

Mitchell, J. R., and Jackson, K. C., 1979, Fenitized crustal crustal xenoliths from small carbonatite bodies, Morrilton-Perryville area, Arkansas: Geological Society of America Abstracts with Programs, v. 11, No. 2, p. 163.

Muehlberger, W. R., Denison, R. E., Lidiak, E. G., 1967, Basement rocks in the continental interior of the United States: American Association of Petroleum Geologists Bulletin, v. 51, p. 2351–2380.

Zietz, Isidore, (compiler), 1981, Composite magnetic anomaly map of the United States, Part A: conterminous United States: U. S. Geological Survey Map GP–954–A.

THE SYLAMORE SANDSTONE OF NORTH–CENTRAL ARKANSAS, WITH EMPHASIS ON THE ORIGIN OF ITS PHOSPHATE

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ABSTRACT

Phosphate-bearing sandstone conformably underlying the Mississippian Boone Formation and unconformably overlying strata ranging in age from medial Ordovician to early Devonian is herein assigned to the Sylamore Sandstone. In north-central Arkansas the Sylamore can occur interbedded with black, Chattanooga-type shale or positioned at the base of the St. Joe Limestone Member of the Boone. Phosphate occurs as pebbles and sand-size grains within a white to light-gray, siliceous or calcareous, dominantly medium-grained, well-rounded quartz sand matrix. Based on its three major grain types, quartz, phosphate, and chert, the Sylamore is a phosphatic quartz arenite, subphosphoarenite, cherty phosphoarenite, and phosphoarenite. Composition and textural trends of the Sylamore are consistent regionally regardless of its thickness or whether it is interbedded with Chattanooga-type shale. This lithic consistency, in combination with the conformability of these rocks with the overlying Boone, supports the conclusion of Swanson and Landis (1962) and Freeman and Schumacher (1969) that the Sylamore/Chattanooga strata represent a detrital phase that collected at the base of the Boone transgression.

Phosphate pebble concentrations are either in the base and/or the top, on bedding planes, or in a homogenous distribution throughout the sandstone. In general, these pebbles have a rounded to well-rounded form with a few gentle reentrants and a smooth, shiny surface. Sylamore phosphate pebbles possess textures that indicate they were derived from older phosphatized limestones. Phosphatized allochems, namely ooids, superficial ooids, and crinozoan skeletal fragments, in cryptocrystalline fluorapatite matrix, comprise the grains. Accessory grains are fragmented phosphatic brachiopods and quartz. Based on the abundance of these allochems, the pebble lithologies are: phosphomicrite, biophosphomicrite, oophosphomicrite, crinozoan biophosphomicrite, oophosphomicrite/crinozoan biophosphomicrite, and quartzose phosphomicrite.

The phosphate-bearing sand and shale of the basal Cason Shale (Ordovician) have been confused by many geologists with the Sylamore, particularly where these Cason strata are preserved as erosional remnants beneath the Sylamore. In comparison to the Sylamore, phosphatic sandstone of the Cason is some shade of green, dolomitic, fine-grained, and texturally immature. In addition to phosphatic pebbles, layers of phosphatized oolitic limestone can occur within the Cason. The Sylamore contains no layered phosphate. Cason phosphatic pebbles are less mature texturally and are of fewer lithologic types than the Sylamore pebbles.

Apparently two phosphogenic episodes occurred prior to the deposition of the Sylamore. The initial episode phosphatized portions of the crinozoan Ordovician Fernvale Limestone prior to the deposition of the Cason. Detrital phosphate eroded from an exposed Fernvale surface was incorporated into the sediment of the basal Cason beds. The second episode of phosphogenesis affected the Cason sometime following the deposition of its oolitic limestone. Portions of the oolite and probably some additional Fernvale were replaced by phosphate during this episode. Detrital phosphate eroded from these older phosphatized limestones, as well as second-cycle phosphate pebbles from the Cason, collected in the Sylamore as a lag concentrate during the initial part of the Boone transgression.

INTRODUCTION

While conducting field reconnaissance in northern Arkansas in the late 1890's, J. C. Branner, then state geologist of Arkansas, became intrigued with the thin and lithically variable strata occupying the interval between the Silurian and Mississippian rocks of the region. In the northwest part of the state he found the interval occupied by black and green shale, which he named the Eureka Shale (now called the Chattanooga). To the south and east, the shale was generally absent, and in its place at the base of the Mississippian Boone Formation was a sandstone containing abundant phosphate pebbles at many localities. In most places the sandstone was only a few feet thick, but along South Sylamore Creek he found 40 feet of it. He named it the Sylamore Sandstone for these exposures. Because of a delay in the publication of Branner's findings (Branner, 1897), the name was used in print first by Penrose in 1891.

After its naming, several geologists working in north central Arkansas (Penrose, 1891; Hopkins, 1893, Branner, 1897; Branner and Newsom, 1902; Adams and Ulrich, 1904) identified the Sylamore and discussed its stratigraphic relations. These discussions are misleading and difficult to follow because it is apparent that many of the exposures reported by these workers were actually of the Ordovician Cason Shale. Their confusion is understandable because it is not uncommon for Mississippian rocks to rest directly on the Cason throughout northern Arkansas. It was not until Purdue (1907) discovered phosphatic sandstone below Silurian rocks in western Independence County that geologists became aware of the existence of two phosphate-bearing units.

Branner was unable to determine the relationship between the sandstone and shale, which he found together only along South Sylamore Creek. In additon to Purdue's work, investigations by several geologists helped define the character of the interval occupied by the Chattanooga–Sylamore strata (Ulrich and Adams, 1905, Purdue and Miser, 1916; McKnight, 1935; Maher and Lantz, 1953; Frezon and Glick, 1959; Swanson and Landis, 1962; Freeman and Schumacher, 1969). In general, the rocks of the interval have three distinct stratigraphies. In Washington and Crawford Counties in northwest Arkansas, a thin, glauconitic, phosphatic sandstone occurs at the base of a relatively thick Chattanooga Shale. This sandstone has been referred to as the Sylamore Sandstone Member of the Chattanooga Shale. In Newton and Searcy Counties, the Chattanooga is absent, and the interval is occupied by a thin, phosphatic sandstone that lies directly below the St. Joe Limestone Member of the Boone Formation. Geologists working in this part of Arkansas have generally considered this sand to be a basal detrital phase of the St. Joe Limestone. In some isolated localities, most notably in the type area of the Sylamore along South Sylamore Creek, black Chattanooga-type shale is found between sandstone beds.

To add to the confusion rising out of the geographic variability of the stratigraphy of the interval, the base of the sandstone has produced diagnostic conodonts from different places that range in age from Late Devonian to Early Mississippian.

In an excellent summary of the lithostratigraphy and biostratigraphy of the interval, Swanson and Landis (1962) proposed a model that relates all of the interval's rocks to one genetic unit deposited during a single transgression of the sea (Fig. 1). In their model, the Chattanooga was deposited as a shelf mud and the Sylamore Sandstone represents a shallow-water, nearshore accumulation. Occurrences of the shale between sandstone beds (Swanson and Landis identified six such occurrences, not counting the type area) represent an interfingering of the two lithotopes. The different ages

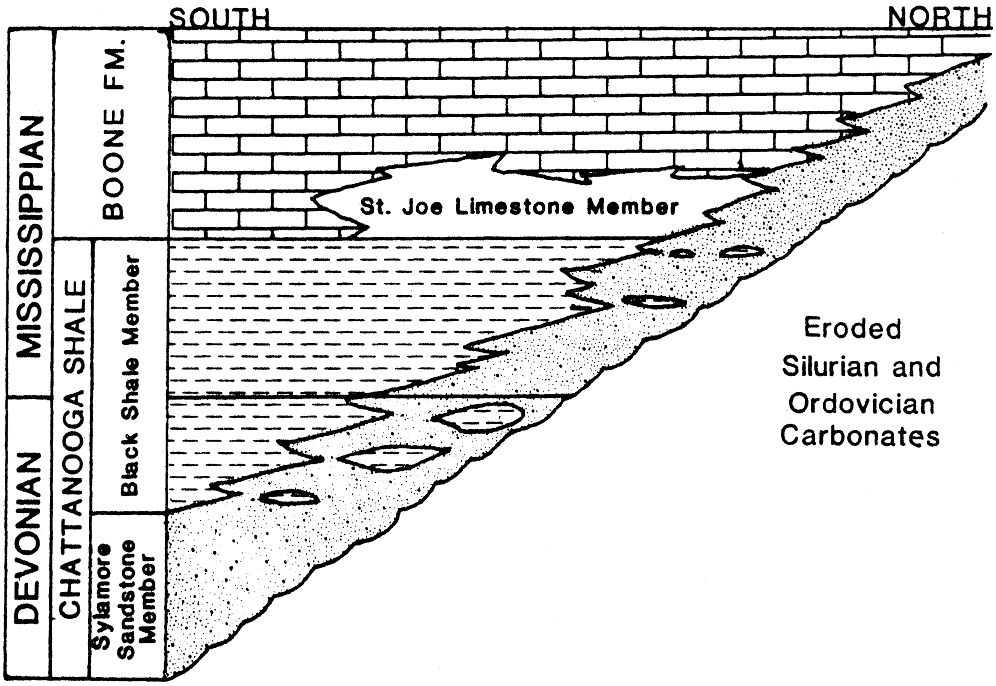


FIGURE 1

Diagrammatic sketch showing temporal relationships within the Devonian-Mississippian transition of Northern Arkansas.

(Freeman and Schumacher, 1969,Figure 1, as modified from Swanson and Landis, 1962, Figure 2).

determined for the base of the sandstone result from its position as a nearshore deposit in a transgressing sea. Freeman and Schumacher (1969) built on the model by constructing a qualitative paleophysiographic map based on the age of the base of the sandstone in different places.

Most geologists (including Branner) who have been involved with the stratigraphy of the interval have agreed that, wherever examined, its rocks lie unconformably on subjacent strata (ranging in age from medial Ordovician to medial Devonian) and grade conformably upward into the base of the Mississippian Boone Formation. This observation, coupled with the knowledge that black Chattanooga-type shale occurs as layers and lenses within phosphatic sandstone, strongly suggests that the Swanson and Landis model, as elaborated by Freeman and Schumacher, is correct. None of the observations that we have made during the course of our investigation of the interval over a broad geographic region suggests otherwise. Therefore, we accept the trangressive, single-sand lithosome model. We are aware that formal rock-stratigraphic units should not interfinger, and that if the single-sand model proves true, a revision of the interval nomenclature is in order. For the sake of simplicity of discussion, however, we refer to all sand in the interval as Sylamore and all black shale as Chattanooga.

The objectives of our study are threefold. One, to characterize the lithology of the sandstone over a larger area than has heretofore been attempted. Two, to establish petrographic and field criteria useful in distinguishing the Cason Shale from the Devonian-Mississippian sandstone of the interval. These two distinctly separate groups of rocks are similar in that they both contain phosphate and are the only detrital deposits in an otherwise carbonate section. This similarity, plus a complicated depositional-erosional history that can cause either one or both of them to occupy a thin stratigraphic interval directly beneath the Boone Formation, makes their separation a major problem in Ozark stratigraphy. Finally, we have addressed ourselves to the occurrence of the phosphate and the description of its petrography with the objective of elucidating its origin.

We gratefully acknowledge the logistical support of the Arkansas Geological Commission and aid that we have received from its geological staff.

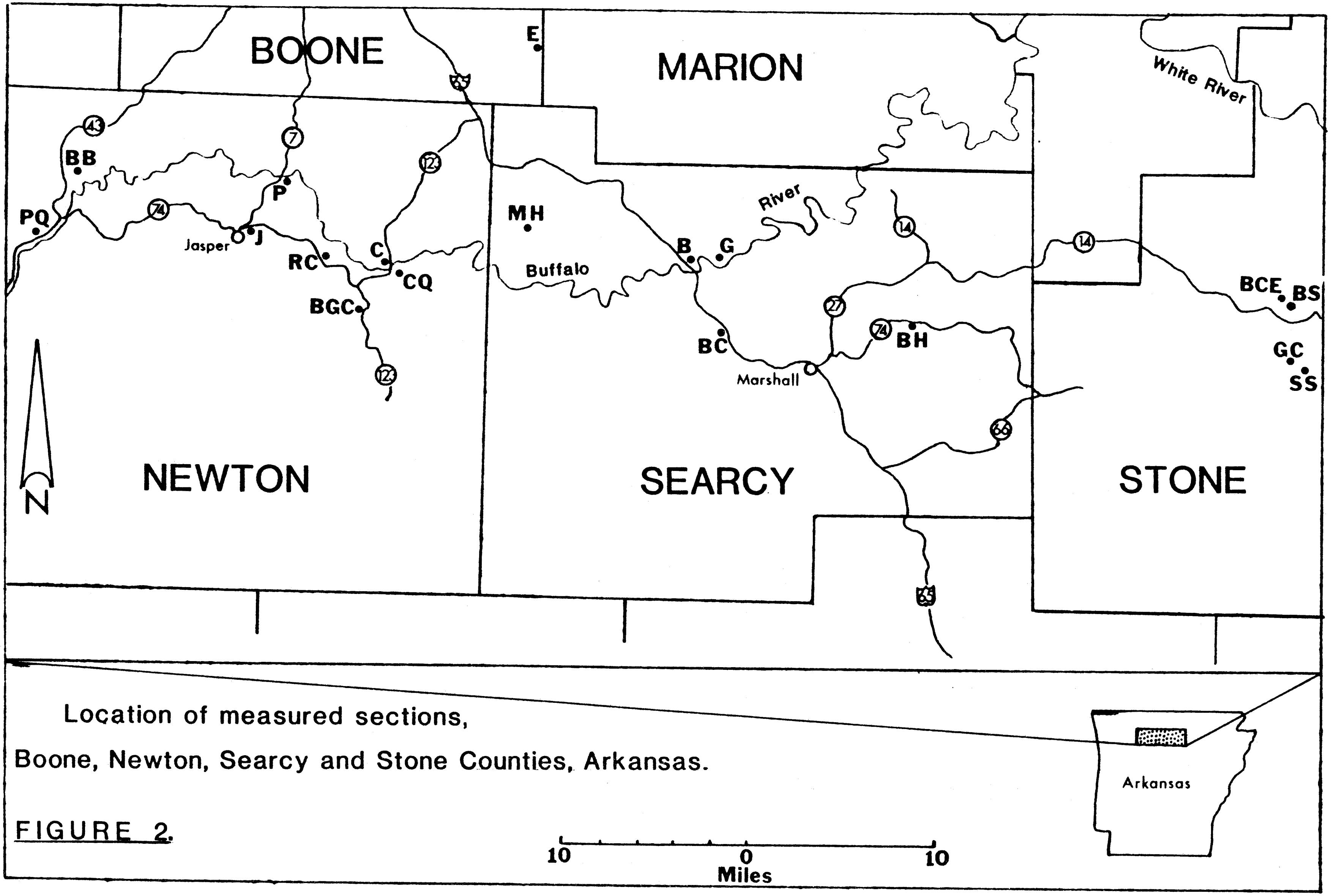
Study Area

Outcrops of the Sylamore were observed from East Lafferty Creek in eastern Arkansas to the extreme northwest corner of the state. Because of the abundance of phosphate pebbles in north-central Arkansas, this area was chosen for detailed stratigraphic and petrographic study. Chosen stratigraphic sections lie along a sinuous, generally east-west outcrop belt across northern Stone, Searcy, southern Boone, and northern Newton Counties from along South Sylamore Creek 4.1 miles southeast of Fifty-Six, Stone County, westward to 1.35 miles south-southwest of Ponca, Newton County. A total of eighteen Sylamore sections were chosen from within the study area Fig. 2).

Lithology of the Sylamore

Sandstone

The phosphatic quartz sandstone of the Sylamore exhibits color and textural variations from white to dark-gray and from fine-grained sandstone to sandy, pebble conglomerate. The color of the sandstone is dependent upon the content of sand-size phosphate and detrital clay, whereas the texture primarily is dependent upon the size and amount of pebble-size phosphate. The sandstone has a diagnostic salt and pepper appearance due to the dark-gray to



Location of measured sections,
Boone, Newton, Searcy and Stone Counties, Arkansas.

FIGURE 2.

black phosphate against a white to light-gray matrix of quartz sand and cement. Rounded to well-rounded, medium-grained monocrystalline quartz, phosphate, and chert, plus trace amounts of feldspar and glauconite, constitute the sandstone framework (Plate 1, Figs. A, B). This framework is cemented primarily by quartz overgrowths, poikilotopic calcite, some chert and rare, fine-grained dolomite. Where the Sylamore is thin it is cemented by poikilotopic and subequant sparry calcite, but where it is five feet or more thick it is cemented by quartz overgrowths and chert. The Sylamore grades into overlying St. Joe Limestone through the addition of cronozoan skeletal parts in its upper few inches.

Our sandstone classification of the Sylamore is modified after McBride (1963) and Folk (1974) and is based primarily upon the mineralogic percentages of the sand-size framework grains in the matrix sandstone only. Phosphate larger than sand-size was considered as a single grain in the mineralogic classification of the sandstone. Petrographic classification of the Sylamore yields phosphatic quartz arenite, subphosphoarenite, cherty phosphoarenite, and phosphoarenite according to relative abundances of the three basic framework grains of quartz, phosphate, and chert (Fig. 3). More than one sandstone type can fit the Sylamore at any one section, depending upon the distribution of phosphate within the outcrop.

Rounded quartz and phosphate sand packed in a tangential arrangement dominates the sandstone framework. The presence of phosphate pebbles can cause the matrix sandstone to have a looser packing in the vicinity of the pebbles because shale clasts incorporated into the sandstone with the pebbles have been compacted, filling pore space between framework grains and forcing them apart. In general, the Sylamore is texturally submature with less than 5% clay matrix and a moderately well-sorted arrangement with a mean sorting value of 0.65ϕ. The Sylamore can possess a variety of grain sizes, which elevates the mean sorting values (Plate 1, Fig. A). With a median grain size of 0.30 mm the Sylamore matrix sand is medium-grained, but it possesses a range in mean grain size from 0.20 mm to 0.42 mm.

Petrographic comparisons were made between two separate beds of sandstone that occur above and below black Chattanooga-type shale at Gayler Crossing (Fig. 4) on South Sylamore Creek. Frezon and Glick (1959, p. 178) considered these two sandstones to be lithologically different. Both are subphosphoarenites with rounded to well-rounded grains and relatively high chert contents. Both beds are comprised of supermature to submature sandstone with detrital clay contents of less than 5% and mean sorting values of 0.48ϕ (well-sorted) to 0.66ϕ (moderately well-sorted). Quartz overgrowths and chert cement the framework of the upper sandstone bed, but only small patches of quartz cement are visible in the lower bed due to extensive leaching. Grain size is the only significant difference between the two beds with the lower sandstone having a mean grain size of 0.42 mm and the upper sandstone having a mean grain size of 0.21 mm. However, fining upward is a ubiquitous characteristic of thicker Sylamore sections and is compatible with a transgressive strandline sand model.

Comparisons were made between these two layers and the sandstone in the South Sylamore section on the opposing side of the stream valley from Gayler Crossing. No intervening black shale occurs in this twenty-five foot thick Sylamore section, but the sandstone possesses the same general petrographic characteristics. The base of the sandstone is a conglomeratic, calcite-cemented, cherty phosphoarenite. Two feet above its base it becomes a quartz-cemented subphosphoarenite with a mean grain size of 0.43 mm and a sorting value of 0.63ϕ (moderately well). Furthermore, grain size decreases up-section. Thus, the sandstone of the South Sylamore

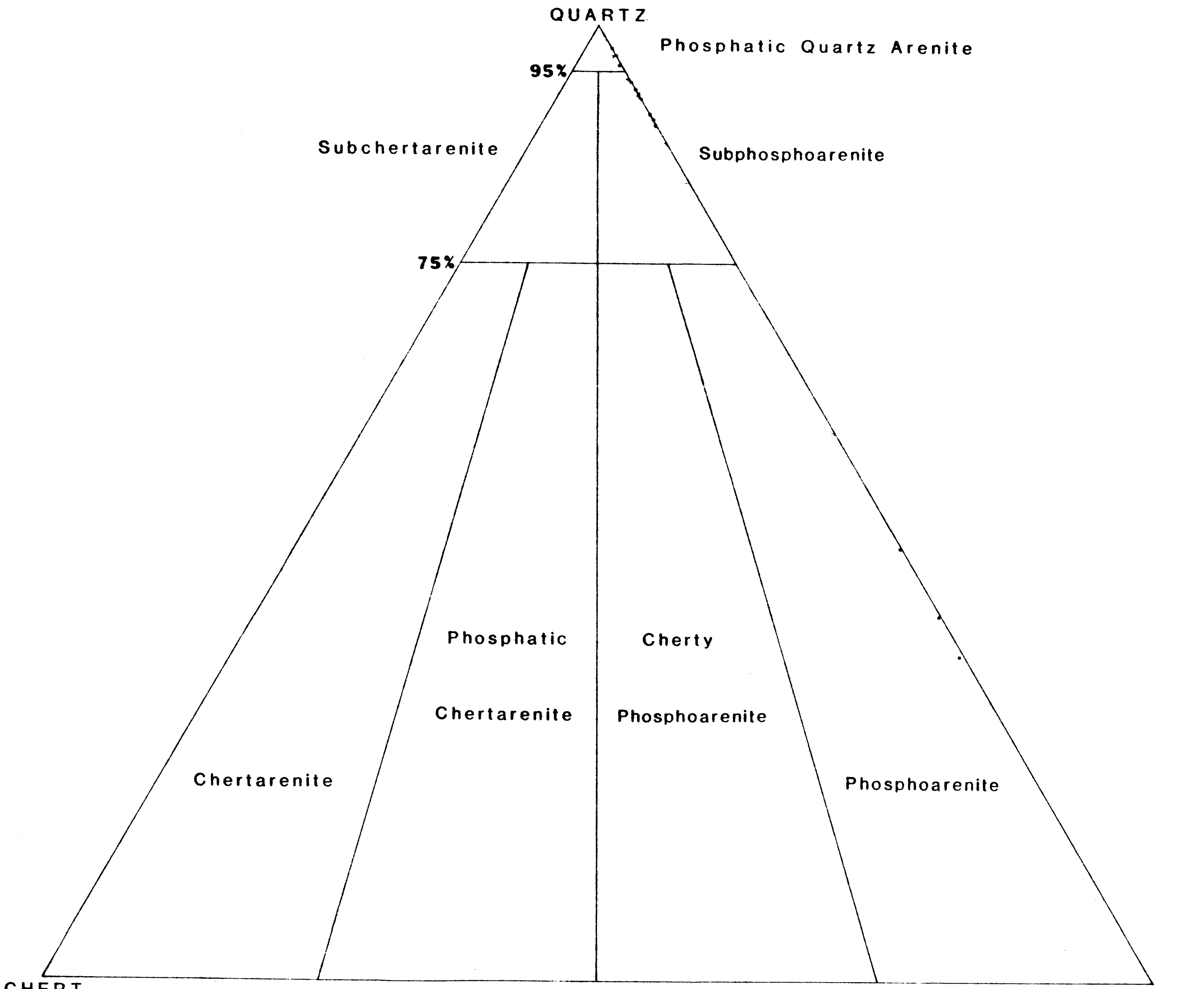

QUARTZ
Phosphatic Quartz Arenite
95%
Subchertarenite
Subphosphoarenite
75%
Phosphatic
Chertarenite
Cherty
Phosphoarenite
Chertarenite
Phosphoarenite
CHERT
PHOSPHATE

FIGURE 3

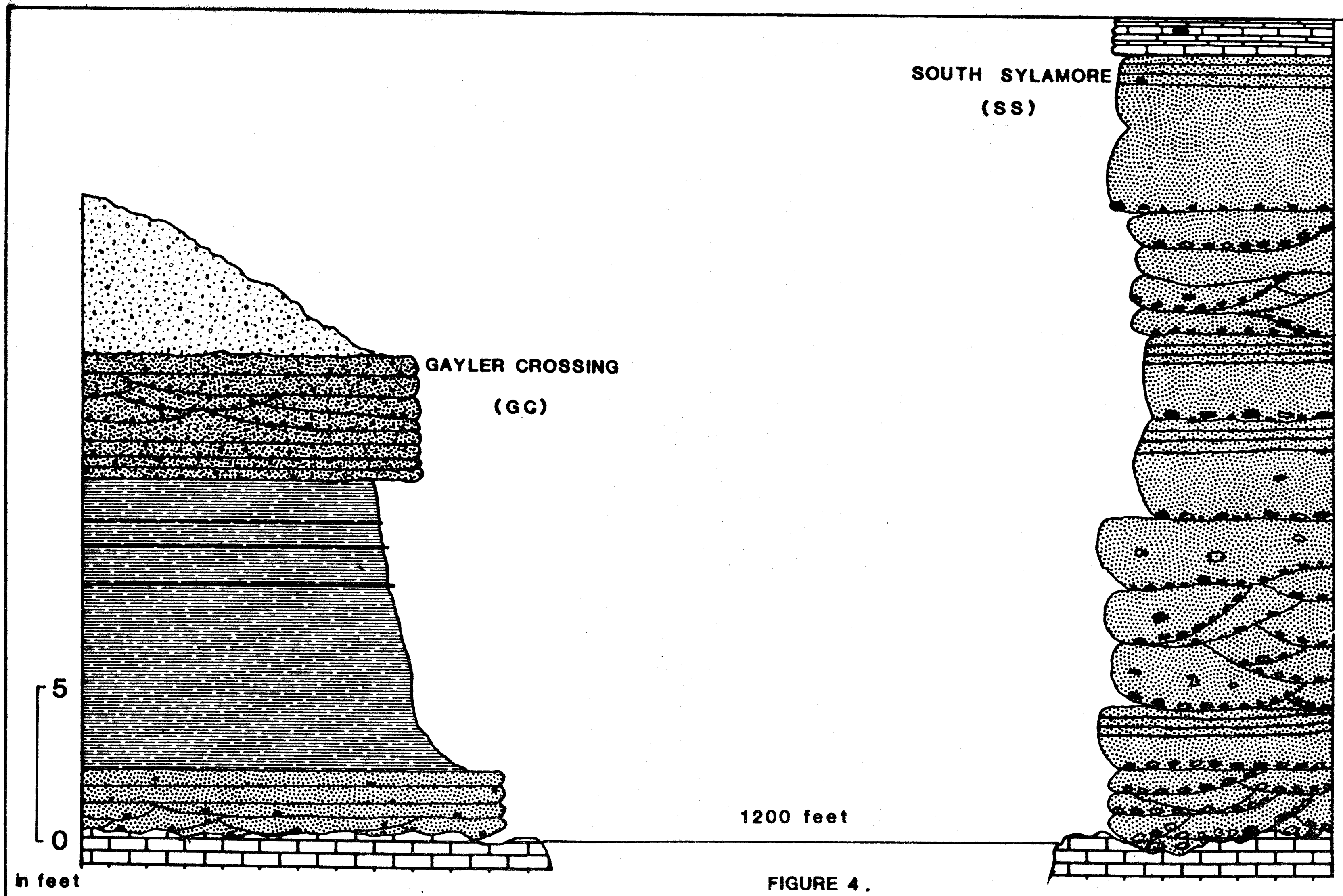


FIGURE 4.
SYLAMORE SECTIONS ON OPPOSING SIDES OF SOUTH SYLAMORE CREEK,AT GAYLER CROSSING.

section possesses a lithology that is similar to that of the sandstone beds above and below the black shale at Gayler Crossing.

Comparisons were made between the sandstone in the Sylamore type area with the basal Boone sandstone. Both thick (greater than one foot) and thin sections of basal Boone sandstone were examined. The thicker sections (Gilbert, Everton, Big Bluff) of this basal Boone Sylamore possess a mineralogy similar to that of the Sylamore in its type area. They also exhibit the same textural trends of upward decrease in mean grain size and sorting values. The sandstone in these thicker sections is characterized by mediun-grained (0.25 mm—0.36 mm), moderately well to moderately-sorted (0.60ϕ—0.94ϕ) subphosphoarenite in its lower part, grading upward into medium-to fine-grained (0.30 mm—0.22 mm), well-sorted to moderately well-sorted (0.47ϕ--0.52ϕ) phosphatic quartz arenite in its upper part. In thinner sections (Barren Hollow, Carver, Rock Creek) the sandstone compares well with the lower portion of the sandstone in thicker sections. These thinner sandstones are medium-grained (0.28 mm—0.34 mm), moderately well-sorted (0.52ϕ—0.64ϕ) calcite-cemented subphosphoarenite (Plate 1, Figs. A, B, E,). Variations in grain size and sorting that occur between sandstone at different localities apparently reflect the influence of local detrital sources.

Thus, regardless of whether the sandstone of the Sylamore is thick or thin, whether positioned at the base of the Boone or is interbedded with Chattanooga-type shale, it possesses a remarkable similarity in mineralogy and textural trends. These aspects, combined with the gradational transition of sand to carbonate of the overlying St. Joe Limestone, reinforces our belief in the Swanson and Landis single transgressive sand model.

Phosphate Pebbles

In order to determine the characteristics of phosphate pebbles found within the Sylamore, detailed examinations were conducted of pebble surface morphologies and petrographic textures. In addition, the gross mineralogy of the phosphate pebbles was obtained through x-ray diffraction, which identified the phosphate mineral as calcium fluoride phosphate or fluorapatite (Fig. 5), a cryptocrystalline variety of apatite known as collophane. When examined in thin section, these dark-gray to black phosphate pebbles are honey brown to dark brown under plain light and psuedo-isotropic under crossed nichols because of the cryptocrystalline nature of the apatite.

The pebbles generally possess a smooth, shiny surface with a few gentle reentrants. Sand-size, shallow circular pits and numerous deeper borings penetrate the pebble surface. The shallow pits are of unknown origin, but Balson (1980) has suggested that similar shallow pits in phosphate pebbles are distal parts of borings into a former pebble surface. We believe that these shallow pits could have been created during compaction of the Sylamore, as harder quartz (H=7) of the matrix sand was impressed into the softer phosphate (H=5) pebbles. The deeper borings are filled with well-rounded quartz sand and calcite cement or phosphatic silt. Crystals of pyrite occasionally are found within borings or on the pebble surface.

In general, the phosphate pebbles have forms that are somewhat flattened parallel to the longest dimension, which averages between 30 mm and 50 mm. The measurements of the three dimensions of each pebble were used to calculate shape, roundness, and sphericity. Plots of pebble shapes according to Zingg's method of pebble-shape classification (Pettijohn, 1975) show the most common

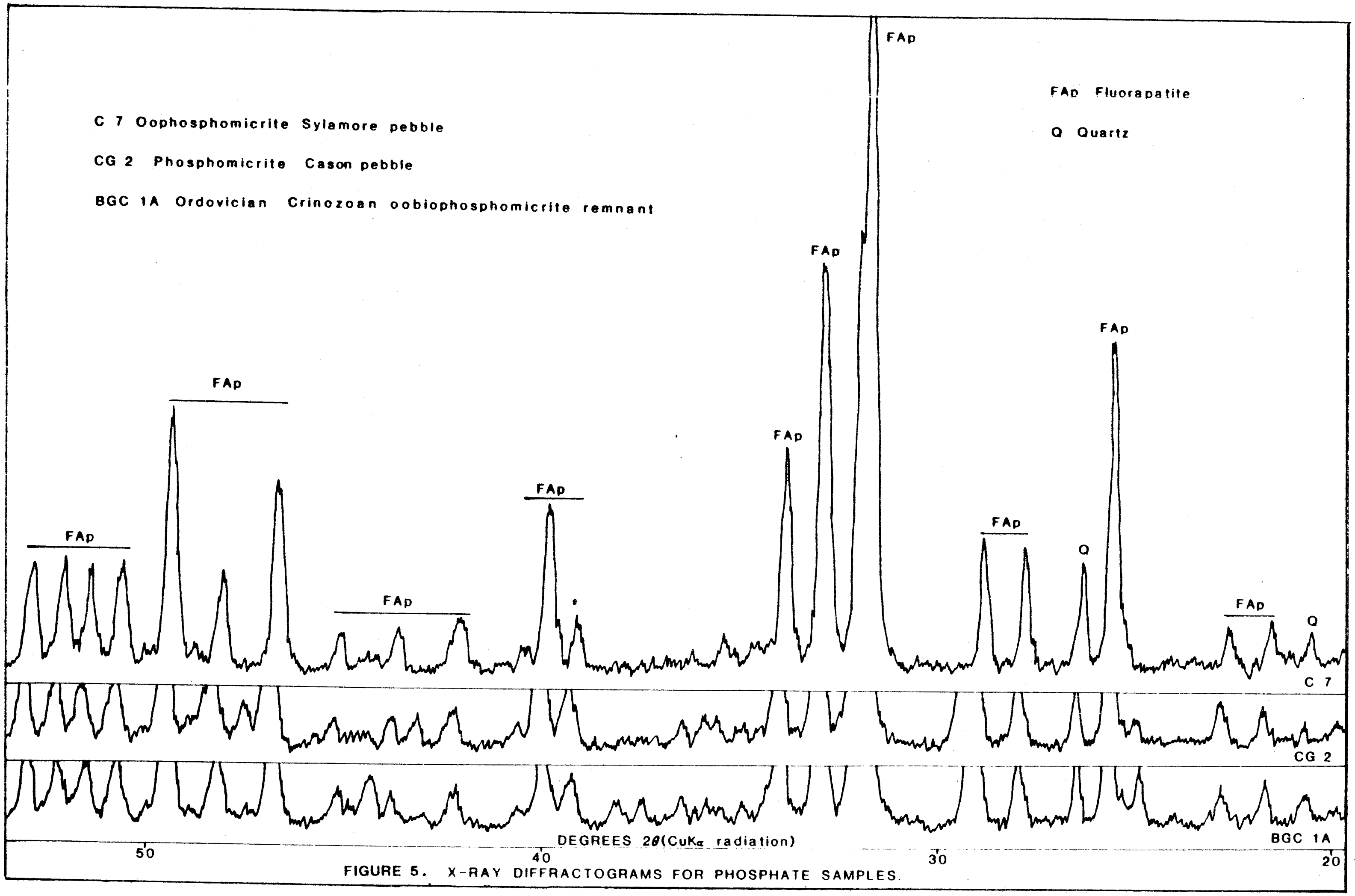


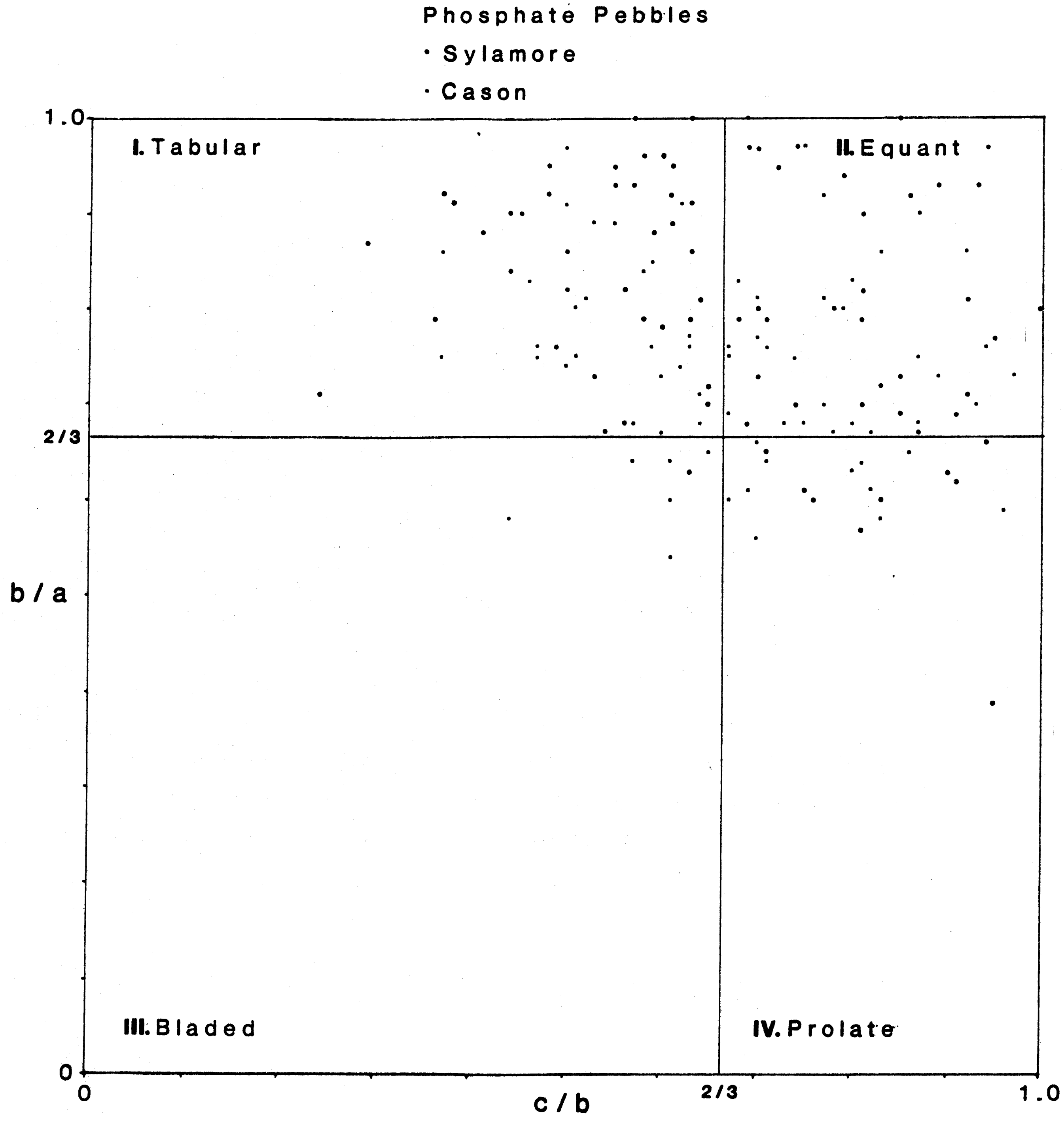
FIGURE 5. X-RAY DIFFRACTOGRAMS FOR PHOSPHATE SAMPLES.

forms to be tabular and equant (Fig. 6). Pebble roundness was quantified by coupling the quantitative roundness method of Krumbein (1940) with Pettijohn's (1975) roundness parameters. The average quantitative roundness for the chosen Sylamore pebbles is 0.61, which corresponds to a well-rounded form (Fig. 7). Pebble sphericities were calculated by the sphericity-form method of Folk (1974), which places most of the Sylamore pebbles within the compact-platy/compact-bladed/compact-elongate and compact sphericity classes (Fig. 8).

The pebbles are of lithologies distinctly different from the surrounding sandstone matrix. Thin-section examination shows them to be composed of grains and textures characteristic of limestone. These carbonate rock constituents, now completely phosphatized, are ooids, superficial ooids, crinozoan skeletal fragments, and rare skeletons of other invertebrate groups, all in a cryptocrystalline fluorapatite matrix, which is a fundamental component of all the pebbles. Grains occur either randomly dispersed through the matrix, in clusters probably localized by burrowing, or loosely-packed. Fragmented phosphatic brachiopods and quartz occur as accessory grains.

Because of the similarity of pebble lithologies to those of limestones, we have followed Blatt's (1982) suggestion and used modified carbonate terminology in their descriptions. Six lithologies based on abundance of allochems have been recognized. These are: a) phosphomicrite, b) oophosphomicrite, c) biophosphomicrite, d) crinozoan biophosphomicrite, e) oophosphomicrite/crinozoan biophosphomicrite, and f) quartzose phosphomicrite (Plate 2, Figs. A–F). Phosphomicrite pebbles, such as those found in the Sylamore at Jasper and in the Cason at the Gilbert city limits are composed solely of burrowed phosphomicrite matrix. Oophosphomicrite pebbles, found at Carver, Carver Quarry, Jasper, Pruitt, and Ponca Quarry, contain well-developed ooids, compound ooids, and superficial ooids (Plate 1, Figs. E, F, Plate 2, Figs. A, E). In addition, some silicified ooids are present in these pebbles. Ooids commonly reflect the shape of the nucleus. These nuclei are phosphatic crinozoan fragments, peloids, other ooids, fragments of phosphatic brachiopods, detrital monocrystalline quartz and chert. Biophosphomicrite pebbles, found in the Sylamore at Jasper and the Cason at the Gilbert city limits, contain a few phosphatic brachiopod fragments *(Lingula)* and some conodonts in phosphomicrite matrix (Plate 2, Fig. B). Crinozoan debris, with a characteristic porous, granular microtexture, plus snails and fragments of phosphatic brachiopods, comprise the ubiquitous crinozoan biophosphomicrite pebbles, (Plate 2, Fig. D). Oophosphomicrite/crinozoan biophosphomicrite pebbles, found at Carver and Big Creek, have grains of ooids and crinozoans on opposing sides of a sharp contact (Plate 2, Fig. E). The oolitic rock is younger since it obviously truncates the crinozoan biophosphomicrite. In phosphatized limestone directly beneath the Sylamore at the Big Creek section (Fig. 9) there exists a contact between phosphatic oolitic and crinozoan limestones identical to the contact in the oophosphomicrite/crinozoan biophosphomicrite pebbles (Plate 2, Fig. F). It is believed that the oophosphomicrite/crinozoan biophosphomicrite pebbles represent fragments of phosphate rock from the contact between these lithologies. Lastly, quartzose phosphomicrite pebbles contain clustered or homogeneously distributed sand-sized well-rounded quartz in Sylamore pebbles at Bear Creek, Jasper, and South Sylamore, or silt-sized angular quartz in phosphomicrite within Cason pebbles at Gilbert city limits (Plate 2, Fig. C).

The texture and form of the pebbles indicate that they were derived as rock fragments from older limestones that had been replaced by phosphate. Riggs (1979a) refers to these



a, length: b, breadth: c, thickness.

FIGURE 6

PLOT OF SYLAMORE AND CASON PEBBLE VALUES UPON ZINGG'S CLASSIFICATION OF PEBBLE SHAPES DIAGRAM.

(taken from Pettijohn,1975)

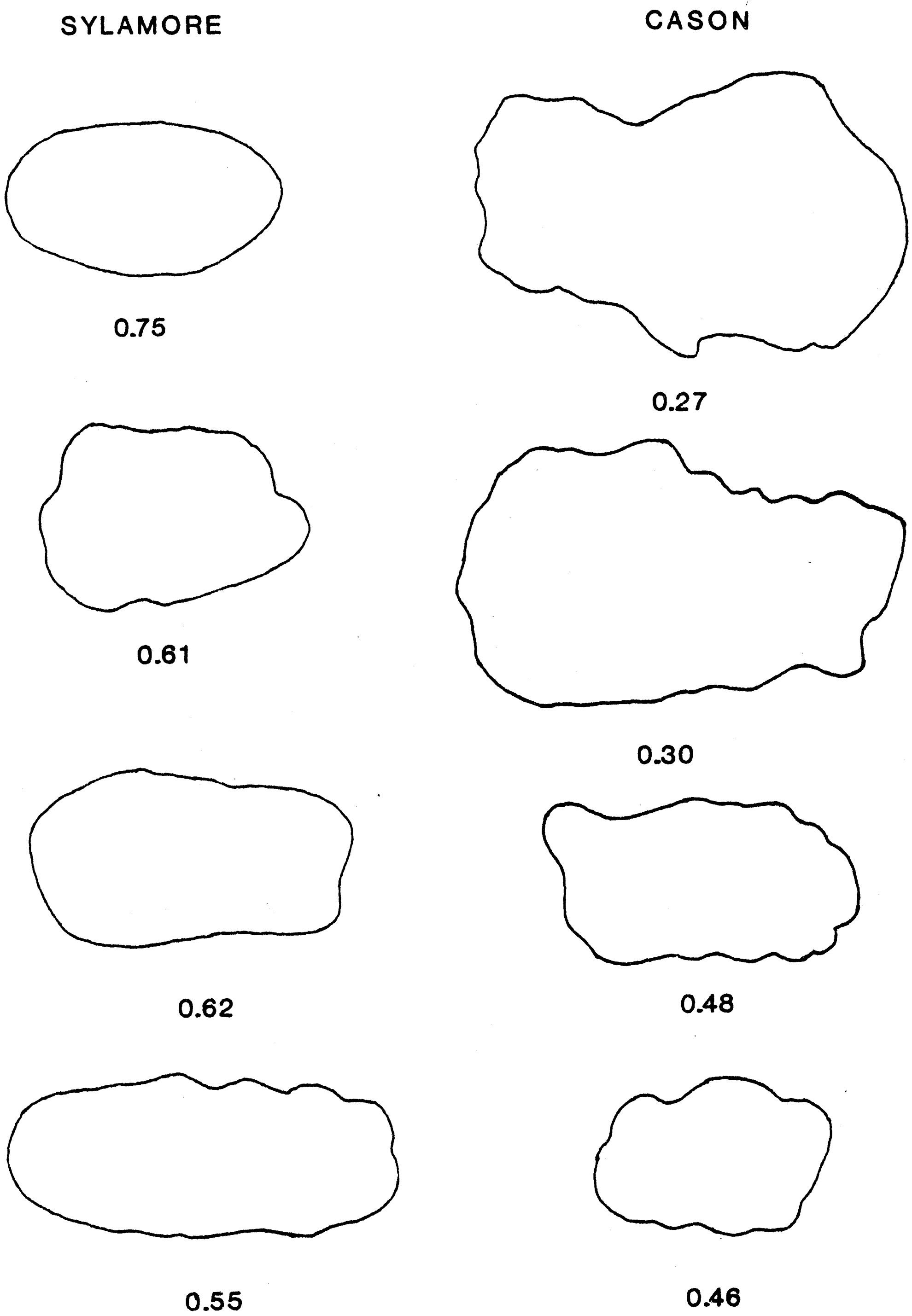


FIGURE 7

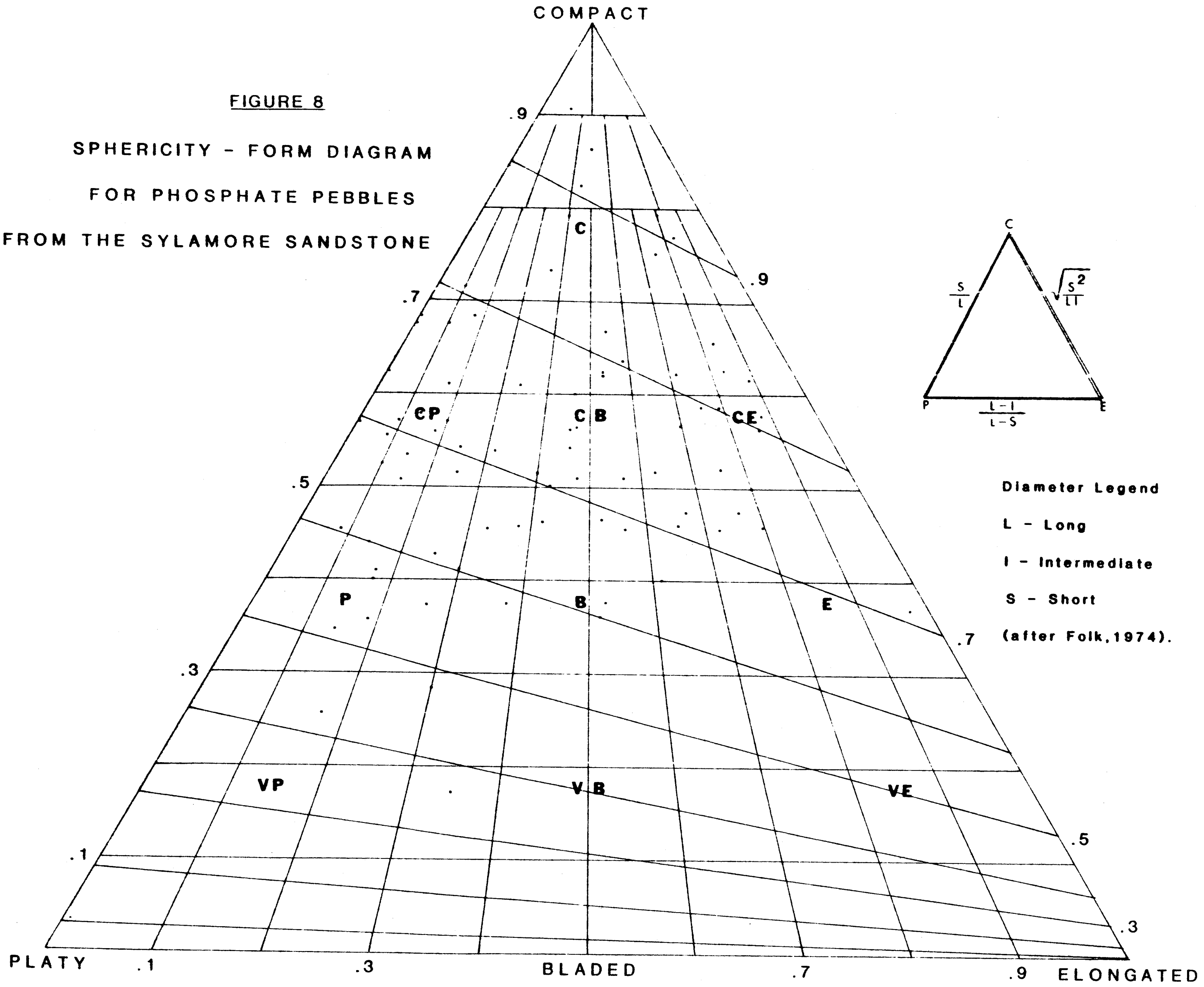
DEGREES OF ROUNDNESS AMONG PHOSPHATE PEBBLES

FROM THE SYLAMORE SANDSTONE AND THE CASON SHALE.

(Pettijohn, 1975)

FIGURE 8

SPHERICITY - FORM DIAGRAM
FOR PHOSPHATE PEBBLES
FROM THE SYLAMORE SANDSTONE



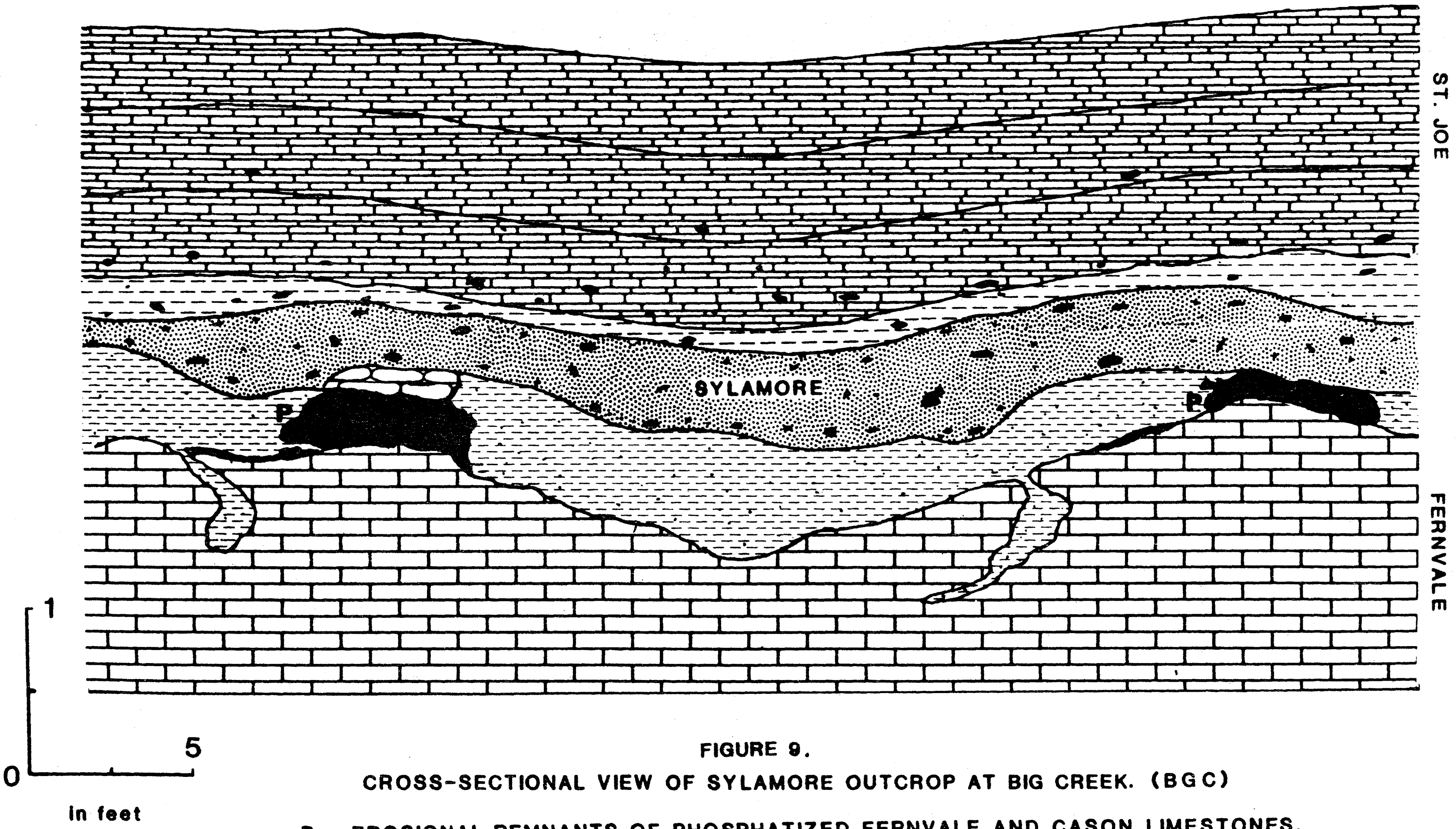


FIGURE 9.

CROSS-SECTIONAL VIEW OF SYLAMORE OUTCROP AT BIG CREEK. (BGC)

P - EROSIONAL REMNANTS OF PHOSPHATIZED FERNVALE AND CASON LIMESTONES.

reworked fragments of phosphate rock as lithochems. The surficial polish and smoothness they exhibit are indicative of reworking in the sedimentary regime (Pevear and Pilkey, 1966), as is the well-rounded shape of the average pebble. Furthermore, the truncation of ooid (Plate 1, Fig. E) and crinozoan (Plate 1, Fig. C) allochems along pebble margins leaves little doubt that the pebbles were originally a part of phosphatized older limestone strata.

Phosphate pebble distribution within the sandstone varies between concentrations in the base and/or the top, homogeneous distribution throughout the entire bed, or distribution upon bedding planes with a few pebbles widely scattered within the matrix sandstone. All of these distributions apparently are indicative of how and when the pebbles were incorporated into the sandstone matrix. The basal, superior, and homogeneous distributions all may be seen within the Sylamore at Jasper (Fig. 10; Plate 3, Figs. D, E). At Carver Quarry (Fig. 14), where the densest concentration of phosphate pebbles was seen, they are concentrated along the base and the top of the sandstone. The greatest concentrations of pebbles occur where the sandstone is thin. In thicker sections, such as South Sylamore (Fig. 13), Gilbert, Buffalo, Everton (Fig. 14), and Big Bluff (Fig. 15), pebbles are characteristically fewer and concentrated along bedding planes with a few scattered occurrences throughout the main body of sandstone.

Lithology of the Cason

The Cason Shale was named from an exposure just north of Batesville, Independence County. In the Batesville district, the Cason consists of two parts: a lower sandstone and shale (Ordovician) that contains abundant phosphate grains and an upper less phosphatic sandy shale (medial Silurian) characterized by abundant spherical (now flattened) algal growths (see Craig, this volume). In the Batesville district, the lower unit has above it, in an apparently conformable relationship, an oolitic limestone with associated intraclastic and dismicritic textures (Craig, 1969, 1975b). Over almost all of the district, this carbonate rock has been removed by post-lower Cason erosion.

To the west of the Batesville district, in Stone County and farther west, the Cason has a different development. The upper unit is known from only one locality in Stone County. In the major area of Sylamore outcrop only the lower unit (Ordovician) is present. The lower unit over this region contains a generally thin (few inches to a few feet) basal section of phosphatic, dolomitic, slightly conglomeratic sandstone and shale. In the thicker sections, these basal beds grade up into very dolomitic clayey micrite or calcareous shale that contains sparse grains of phosphatic sand and silt and in places scattered larger phosphate pebbles.

Comparison with the Sylamore

Because of its detrital nature and phosphate content, the Cason of this western region, particularly its basal beds, has often been confused with the Sylamore. This is especially true when the Cason is preserved as an erosional remnant between the Sylamore and the underlying Fernvale Limestone. There are, however, several characters that aid in separating the two phosphatic horizons.

Remnants of the detrital rock of the basal Cason, preserved in small topographic lows on the top of the Fernvale, occur beneath the Sylamore in two of our sections, Barren Hollow (Figure 11; Plate 1, Figs. C, D) and Ponca Quarry (Plate 4, Figs. A–D). In comparison to the sandstone of the Sylamore, the detrital matrix of the Cason is finer grained and texturally immature. The Cason color is some shade of light green, either grayish

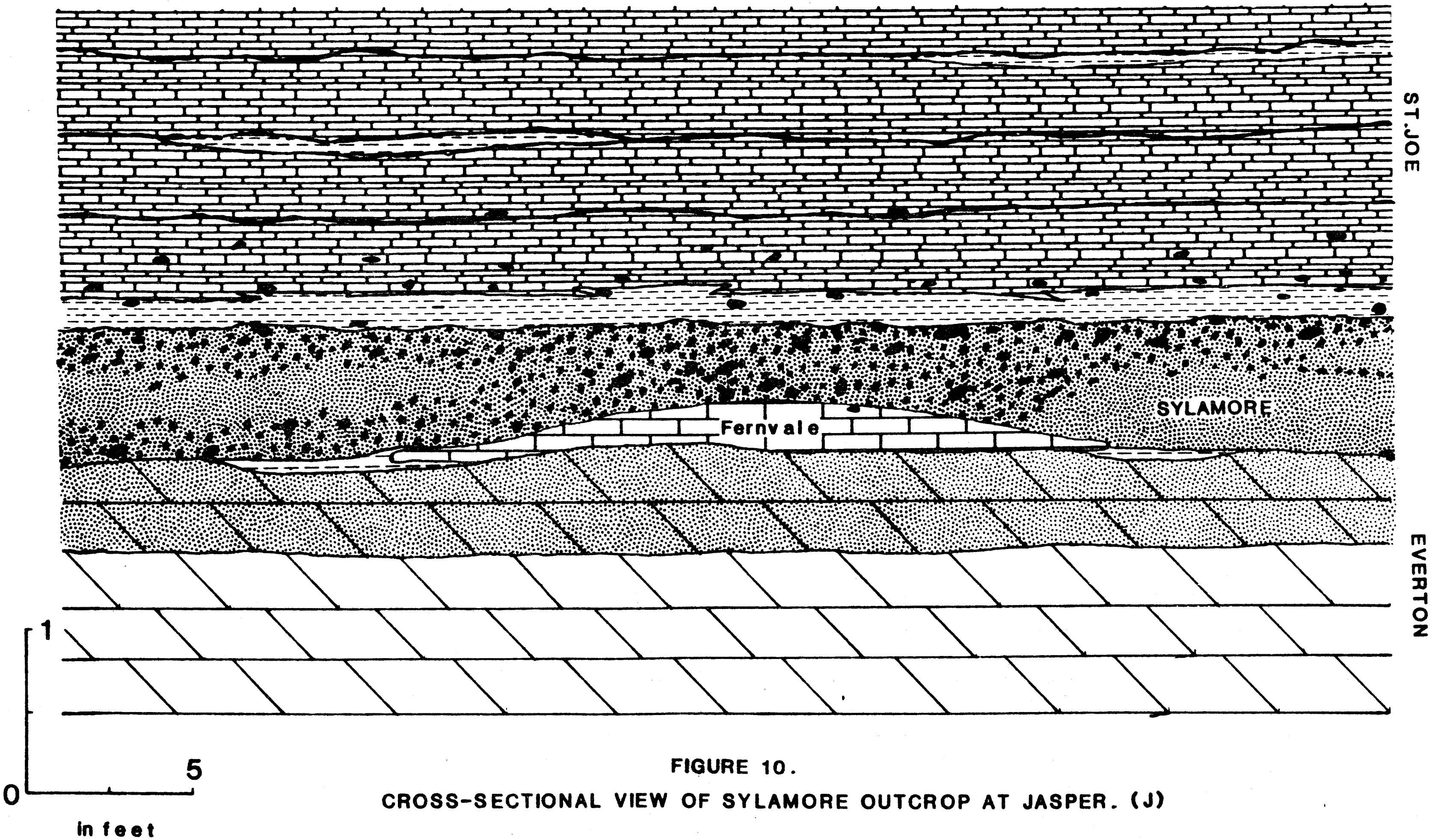


FIGURE 10.

CROSS-SECTIONAL VIEW OF SYLAMORE OUTCROP AT JASPER. (J)

NOTICE THE VARIABILITY OF PHOSPHATE PEBBLE CONCENTRATIONS WITHIN THE SANDSTONE.

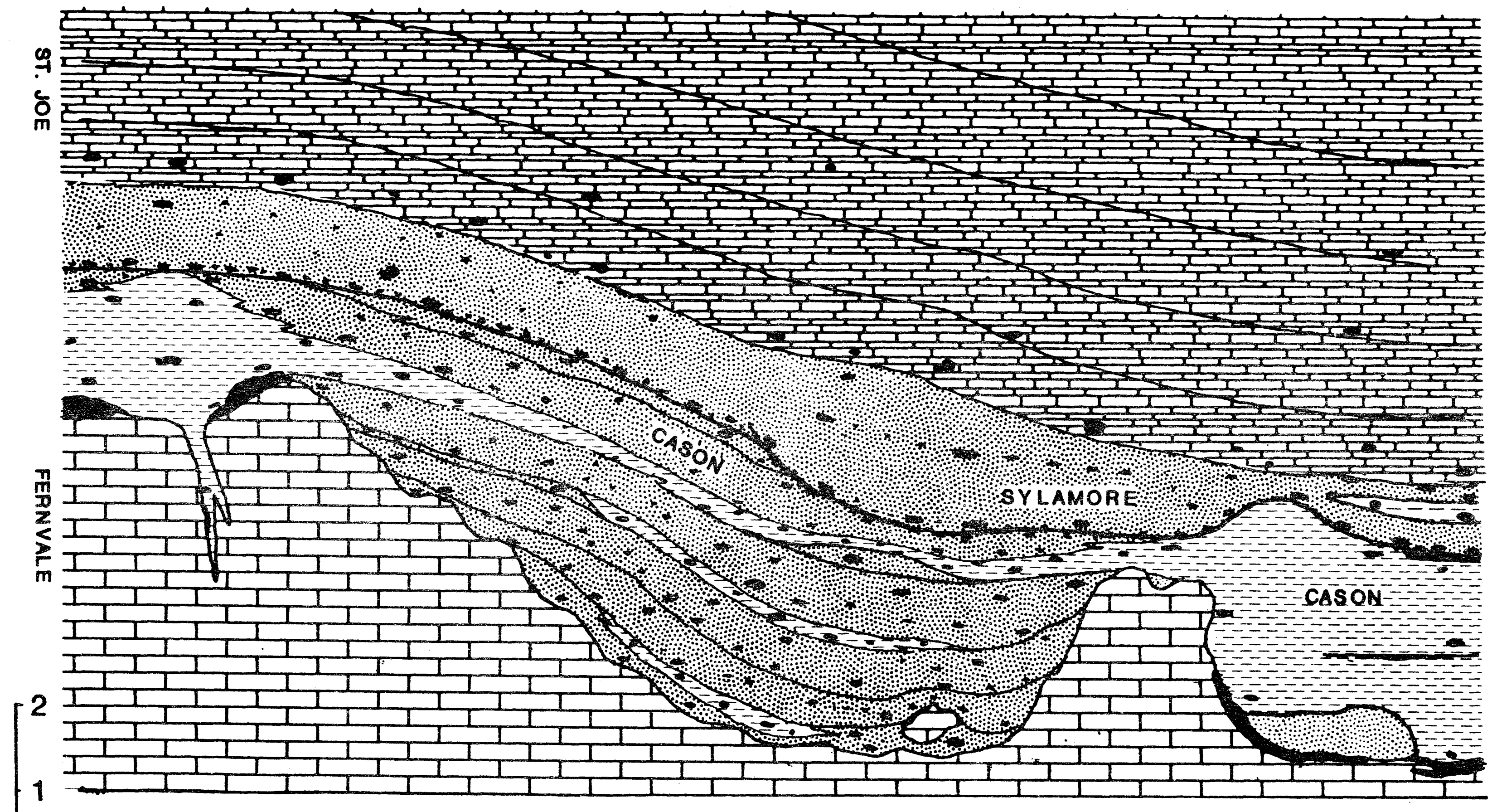


FIGURE 11.

CROSS-SECTIONAL VIEW OF SYLAMORE OUTCROP AT BARREN HOLLOW. (BH)

TROUGH-SHAPED DEPRESSION IN FERNVALE PRESERVES DETRITAL CASON STRATA.

green or yellowish green, as opposed to the light to dark gray color of the Sylamore. The Cason is usually thinner bedded and much more argillaceous, with numerous discontinuous interbeds of green shale. Dolomitic, patchy-calcareous, and fine-siliceous cements bind the rock along with the abundant silt and clay content. Framework grains are moderately well-sorted (0.60ϕ) in tangential contact or else "free floating" in a wispy dolomitic silt and detrital clay matrix. The sandstones have a median grain size of 0.17 mm composed of subrounded to angular grains of quartz, detrital phosphate, feldspar, fragments of phosphatic brachiopods and trace amounts of chert (Plate 1, Figs. C, D). Glauconite also is abundant locally within the Cason sandstones (Lemastus, 1979). Subrounded phosphate pebbles are randomly scattered within the sandstone and often are concentrated along bedding planes as seen in the exposure at Barren Hollow (Figure 11). In general, the sandstone of the Cason is a dolomitic subphosphoarenite with small percentages of microcline and microperthitic feldspar. Thus, whereas the Sylamore is a light-gray, relatively clean, submature, medium-grained phosphatic sandstone with well-rounded quartz, the phosphatic sandstones of the Cason are green, clayey or immature, and fine-grained with subangular to angular quartz.

In addition to color and textural differences of its detrital fraction, the Cason of these western sections differs from the Sylamore by possessing oolitic material at different stratigraphic levels. The oolitic material is in layers and is not to be confused with ooid-bearing pebbles present in the Sylamore. We have observed oolite layers in the Cason at two of our sections, Big Creek and Ponca Quarry. At both localities it is entirely replaced by phosphate. At Big Creek (Fig. 9) it appears to be preserved as discontinuous erosional remnants resting directly on the Fernvale, whose upper few inches have also been replaced by phosphate (Plate 3, Fig. C). The contact of the crinozoan biophosphomicrite of the Fernvale and the oophosphomicrite layers of the Cason within the hard phosphatic layer is sharp, with crinozoan skeletal material truncated beneath it (Plate 2, Fig. F). As mentioned previously, it is probably rocks from stratigraphic situations like this that were reworked to produce the oophosphomicrite/crinozoan biophosphomicrite pebbles found in the Sylamore. At Ponca Quarry, the oophosphomicrite layers seem to form a hard crust that has served to protect the underneath layers of detrital Cason in topographic lows on top of the Fernvale. The contact of the oophosphomicrite with the overlying Sylamore is sharp and irregular (Plate 4, Figs. B–D). We believe that the production of this oolitic material represents the same conditions that produced the oolitic limestone above the phosphatic beds in the Batesville district, and that the two deposits are essentially coeval. For reasons unknown to us, the Batesville oolitic limestone escaped later phosphate replacement.

An additional field characteristic that could help to distinguish the Cason sandstones from those of the Sylamore is that the Cason only overlies the pinkish crinozoan limestone of the Ordovician Fernvale, whereas the Sylamore may overlie numerous different lithologies, including the Fernvale.

Phosphate Pebbles

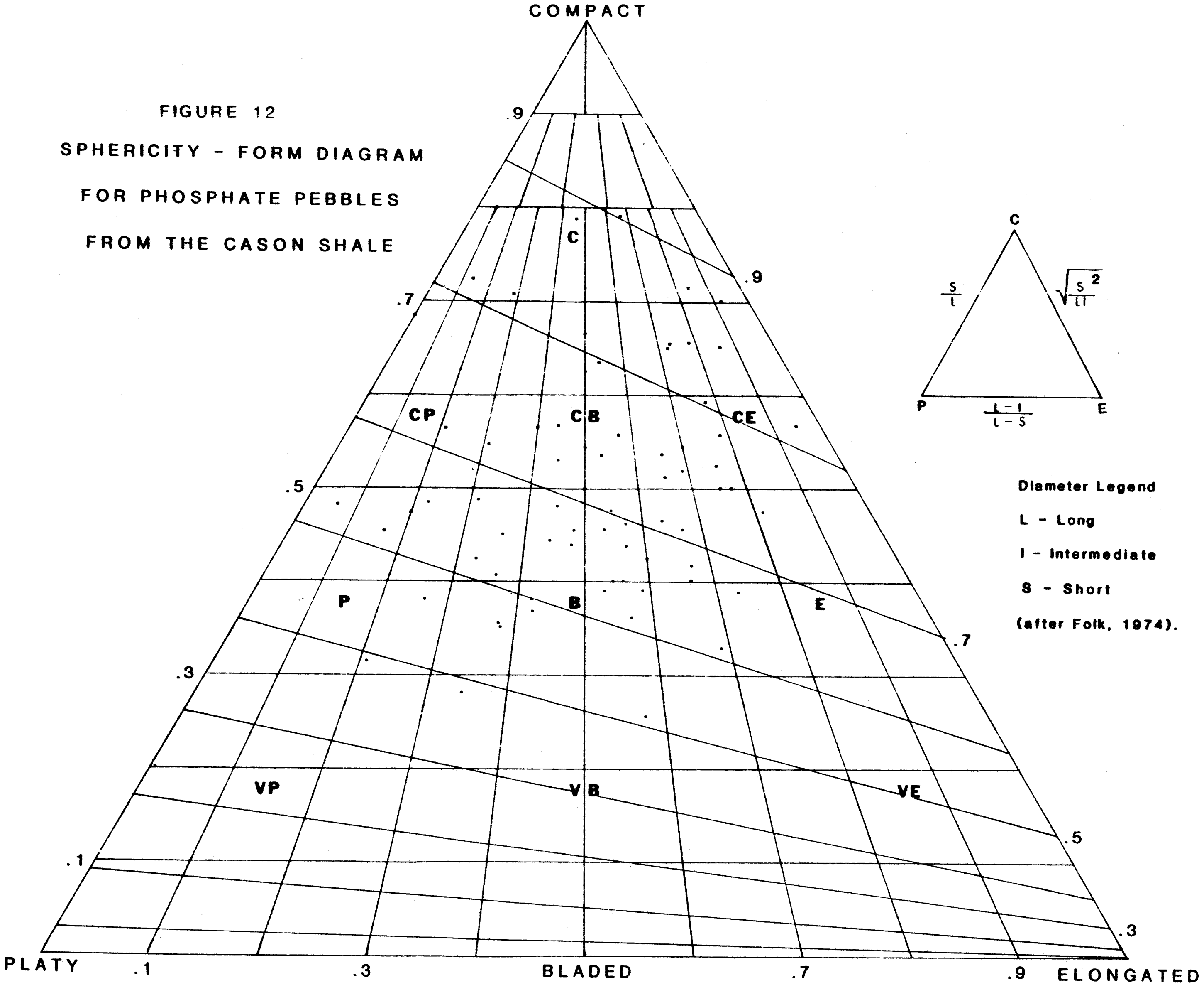
Our examination of Cason phosphate pebbles has not been extensive. Our observations of them have been confined mainly to their occurrences in the Cason in our measured sections (i.e., Ponca Quarry and Barren Hollow). These occurrences are in the basal portion of the unit. For comparison, we also examined from two localities the larger phosphate pebbles that occur in the stratigraphically higher dolomitic micrite/shale. One of these localities is an exposure of the shale in

FIGURE 12

SPHERICITY - FORM DIAGRAM

FOR PHOSPHATE PEBBLES

FROM THE CASON SHALE



Diameter Legend

L - Long

I - Intermediate

S - Short

(after Folk, 1974).

the road ditch of Arkansas Highway 333 at the city limits of Gilbert, Searcy County (NE¼, Sec. 31, T. 16 N., R. 16 W.). The other is a similar occurrence in Stone County along Arkansas Highway 87 leading from Gayler Crossing to Mountain View as it climbs out of the valley of South Sylamore Creek (NE¼, Sec. 21, T. 15 N., R. 11 W.).

Phosphate pebbles that occur within the sandstone and dolomitic shale of the Cason at the Barren Hollow section (Fig. 11) have uneven, dull to slightly polished, bored surfaces upon flattened forms. The sand-size phosphate grains in the sandstone and shale are subspherical and higly polished. The lithology of these pebbles and smaller grains is limited to phosphomicrite and crinozoan biophosphomicrite (Plate 1, Fig. C.) Pebble and sand-sized phosphate of similar texture are present within the dolomitic siltstone and sandstone of the lower portion of the Cason at the Ponca Quarry section. Large phosphomicrite and biophosphomicrite pebbles (Plate 2, Fig. B) with few widely scattered ooids characterize the Cason pebble populations in the stratigraphically higher dolomitic shale exposed at the Gilbert city limits and along Arkansas Highway 87.

Cason phosphate pebbles have generally the same sizes, tabular and equant shapes, and abundance of borings as pebbles from the Sylamore. In contrast to those of the Sylamore, the Cason pebbles are characterized by numerous salients and reentrants about hackly surfaces and dull to slightly polished lusters. The sand-size pits are absent from Cason pebbles and borings are filled solely by equant sparry calcite cement or finely crystalline dolomite and phosphatic silt with some glauconite. Quantitative roundnesses of Cason pebbles are lower than those of the Sylamore; they average 0.39, which corresponds to a subrounded form (Fig. 7). In addition, Cason pebbles are somewhat less spherical with sphericity values concentrated in the platy/bladed/elongate to compact-platy/compact-bladed/compact-elongate sphericity classes (Fig. 12).

Origin of Sylamore Phosphate

It is apparent that the phosphatic grains in the Sylamore have been reworked from older phosphatized carbonates. There appears to have been at least two periods of phosphate replacement. The first affected the Fernvale sometime before the deposition of the Cason. Irregular patches of phosphate in the top of the Fernvale at Barren Hollow might record this earlier period of *in situ* replacement of the limestone (Plate 3, Figs. A, B). The chief evidence, however, for a pre-Cason period of phosphate replacement of the Fernvale is the occurrence of detrital phosphate in the Cason. The ubiquitous crinozoan biophosphomicrite, the most common lithology among phosphate pebbles in the basal Cason, is texturally similar to the areas of phosphatized Fernvale (Plate 3, Fig. B).

The phosphomicrite and biophosphomicrite pebbles in the Cason are from an unknown source. No suitable parent lithology is known from the Fernvale. Their origin is probably one or a combination of the following. 1) crinozoan biophosphomicrite pebbles whose texture has been completely obliterated by phosphate replacement; 2) pebbles derived from rocks deposited after those now represented by the uppermost Fernvale in the region, but whose layers were removed by pre-Cason erosion; 3) (especially for phosphomicrite with sparse ooids) pebbles that were originally intraclasts derived from depositional sites of ooid-bearing lime mud like those known to have been associated with the deposition of the Cason phosphatic beds in the Batesville district (Craig, 1975b). Replacement of such intraclasts could have been accomplished during the second period of phosphogenesis.

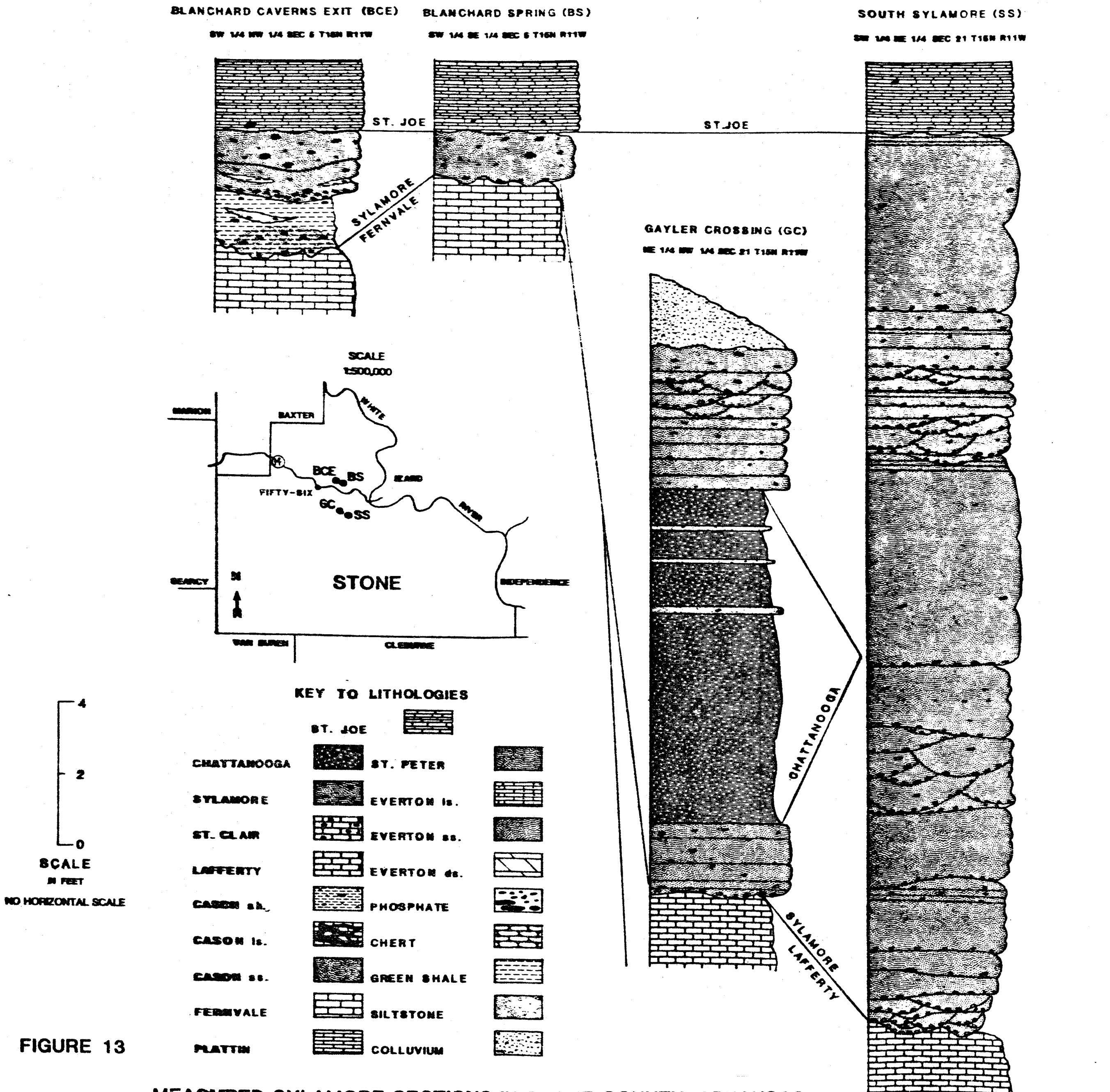


FIGURE 13

MEASURED SYLAMORE SECTIONS IN STONE COUNTY, ARKANSAS

The second period of phosphate replacement affected the Cason oolitic limestone as evidenced by the phosphatized layers of this lithology in the Cason at Big Creek and Ponca Quarry. Additional phosphate replacement of the Fernvale could have taken place also during this time, producing phosphate layers containing oophosphomicrite above crinozoan biophosphomicrite like that found at the Big Creek section.

The Sylamore phosphate pebbles show a greater variety of types than those of the Cason because the second period of phosphate replacement, which affected the Cason, provided additional lithologies for reworking (oophosphomicrite and oophosphomicrite/crinozoan biophosphomicrite). Thus, sources for the Sylamore phosphate include both Fernvale and Cason bedded phosphate, which would provide first-cycle grains, and Cason phosphatic pebbles, which would provide second-cycle grains. The slightly greater sphericity and polish of the Sylamore pebbles in comparison with the Cason pebbles might reflect the second-cycle source, but it is more likely that the Sylamore environment of deposition simply provided a setting more conducive to textural maturation.

The phosphate pebbles in the Sylamore represent a lag concentrate that collected at the base of a transgressive unit (Mississippian Boone Formation) passing over a surface of weathered Ordovician limestone, some of which previously had been replaced by phosphate. As a result, the lithology of the pebbles reflect the phosphate lithology in the underlying unit. For example, at Barren Hollow both the basal Cason and Fernvale are present beneath the Sylamore, and the Sylamore pebbles are composed mainly of crinozoan biophosphomicrite. At Ponca Quarry, where the Sylamore is underlain by beds of phosphatic oolite in the Cason, Sylamore pebbles are dominantly oophosphomicrite.

The degree to which phosphate pebbles were concentrated in the Sylamore sand depended on the availability of phosphate and sand at different points along the transgressing strandline. Abundant phosphate supply (i. e., shoreline exposure of phosphatized Fernvale and/or Cason) and low sand supply (i. e., position relatively distant from stream mouths) resulted in a thin detrital section with abundant pebbles. Thus we find that in general, phosphate pebbles are more concentrated in thinner sections [e. g., Carver, Jasper (Plate 3, Figs. D, E), Ponca Quarry (Plate 4, Figs. A–C)] than in thicker ones (e. g., South Sylamore, Gilbert, Buffalo, Everton, and Big Bluff). Of course, low sand supply and paucity of weathered phosphate rock could lead to a thin sandstone section with few or no phosphate pebbles, as is the case at the base of the Chattanooga in northwest Arkansas. Middle Ordovician supermature quartz arenite of the St. Peter and Everton seems to have provided the majority of quartz grains for the Sylamore matrix sand.

Concentration of pebbles above the base of the sandstone, such as at Jasper and Carver Quarry, resulted from transportation of strandline concentrations to off shore positions during periods of increased wave activity, probably storms. These concentrations generally have irregular bases that could be interpreted as scour fills.

Summary

Although the texture of the Sylamore may range between fine-grained sandstone and conglomerate depending upon the presence of phosphate pebbles, the mineralogy of the sandstone is remarkably consistent. The sand-size fraction is dominantly well-rounded quartz and detrital phosphate grains that are fragments of phosphatized Ordovician limestones. The

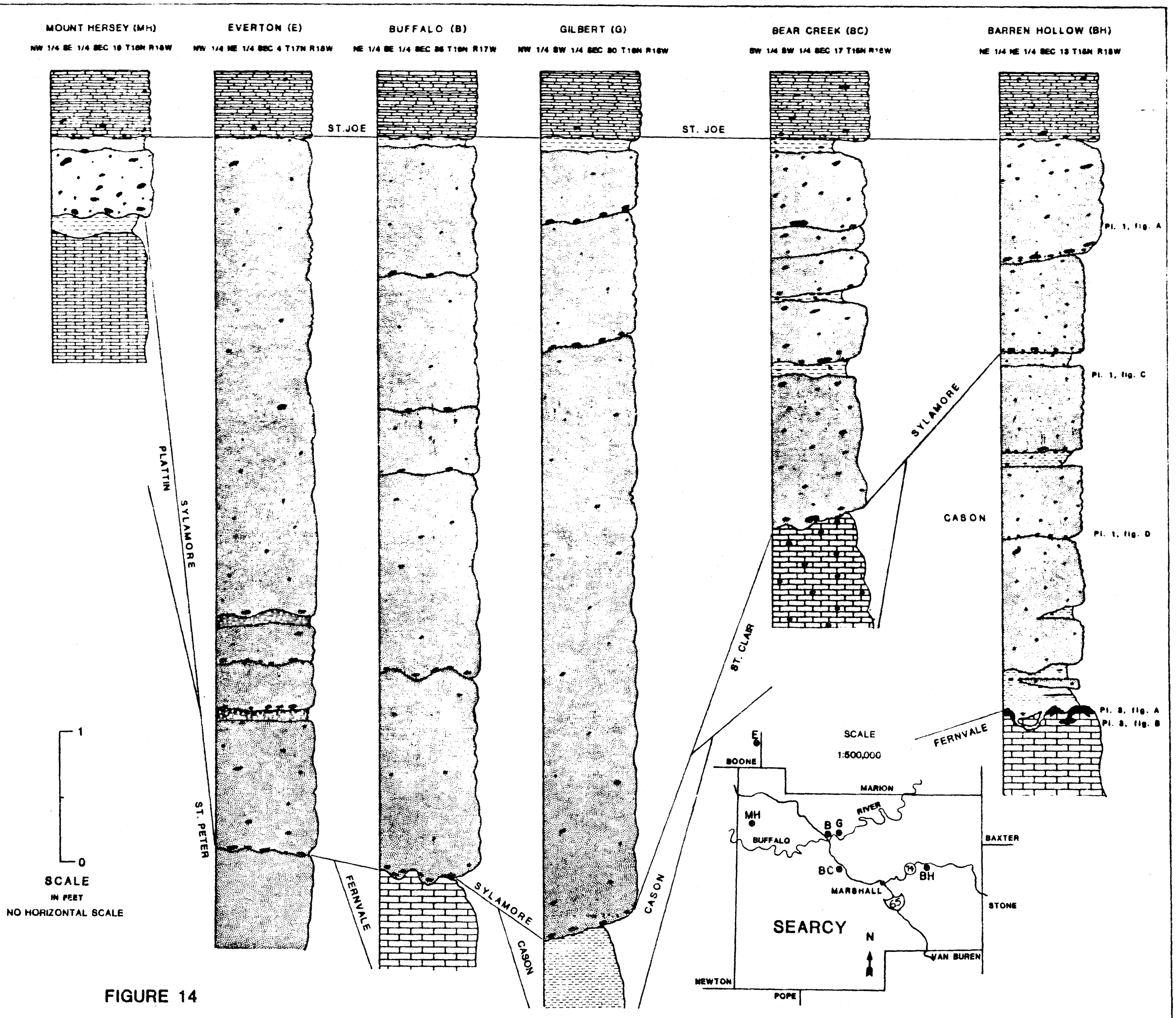


FIGURE 14

MEASURED SYLAMORE SECTIONS IN SEARCY AND BOONE COUNTIES, ARKANSAS mineralogy remains consistent even where the sandstone interfingers with black Chattanooga-type shale. The detrital grains that comprise the framework of the Sylamore were derived from a source that seems to have been mineralogically consistent throughout northern Arkansas, namely the supermature quartz arenites of the St. Peter Sandstone and Everton Formation and phosphatized carbonates of the Fernvale Limestone and Cason Shale. The Sylamore-Chattanooga strata grade conformably into the overlying Boone Formation, and the sandstone of the interval fines upward consistently throughout the region. It is our conclusion that the regional petrography of the Sylamore is supportive of the single transgressive sandstone model of Swanson and Landis (1962).

The overall consistency of the petrography of the Sylamore enables it to be separated from the sandstone of the Cason. Whereas the Sylamore is a white to light-gray, siliceous or calcareous, moderately well-sorted, submature, medium-grained subphosphoarenite with well-rounded quartz, the sandstone of the Cason is grayish green to yellowish green clayey, dolomitic, moderately well-sorted, immature, fine-grained subphosphoarenite with subangular to angular quartz grains. It possesses only minor amounts of well-rounded, medium-grained quartz. The Cason is noticeably more feldspathic and glauconitic, and can contain layers of oolitic limestone. The Cason contains a lesser variety of phosphate than does the Sylamore. Its pebbles are dominantly of crinozoan biophosphomicrite and phosphomicrite. It does not contain oophosphomicrite pebbles as does the Sylamore. In addition, the Cason is always superjacent to the Fernvale, whereas the Sylamore may be superjacent to seven different Ordovician and Silurian formations. Distinguishing between the pebbly green shale of the Cason and green shale that in places occurs in the Sylamore is difficult in the absence of sandstone. In such examples biostratigraphic support with conodonts may be necessary.

Petrography of the Sylamore phosphate pebbles clearly indicates that they are reworked fragments of phosphatized carbonate rock from the Fernvale Limestone and Cason Shale. Identical x-ray diffractograms from Fernvale, Cason, and Sylamore phosphate supports this conclusion (Fig. 5). At lease two separate episodes of *in-situ* phosphatization occurred: and earlier episode that phosphatized portions of the Fernvale and later episode that phosphatized the oolitic limestone of the Cason and possibly again affected the Fernvale. The global tectonic orientation of the present site of northern Arkansas during the Late Ordovician apparently was conducive to phosphatization (Cook and McElhinny, 1979; Sheldon, 1980). During each phosphogenic episode this site was between approximately 20° and 32° south of the paleoequator (Ross, 1976) and shallow epicontinental seas were adjacent to predominately carbonate terrains that provided only slight terrigenous material. A warm equatorial current around this mostly southern hemisphere landmass could have forced deeper, cooler, nutrient-rich, therefore phosphate-laden water, to upwell onto shelves subject to westerly winds (Ross, 1976) through a latitudinally directed seaway (Cook and McElhinny, 1979). These factors in confluence would facilitate phosphate replacement of originally calcium carbonate sediment (Manheim, 1975; Sheldon, 1980).

After each episode of phosphatization, subaerial exposure provided phosphatic debris that collected as lag deposits at the base of the overlying transgression. The lithologic variety of phosphatic pebbles in the Sylamore is greater than that in the Cason because the Sylamore had as its source phosphate-replaced carbonates that formed during both periods of phosphogenesis.

We conclude that the source rock for the Cason was for the most part the Fernvale Limestone and exposed Precambrian rock of the St. Francois Mountains, the latter of which

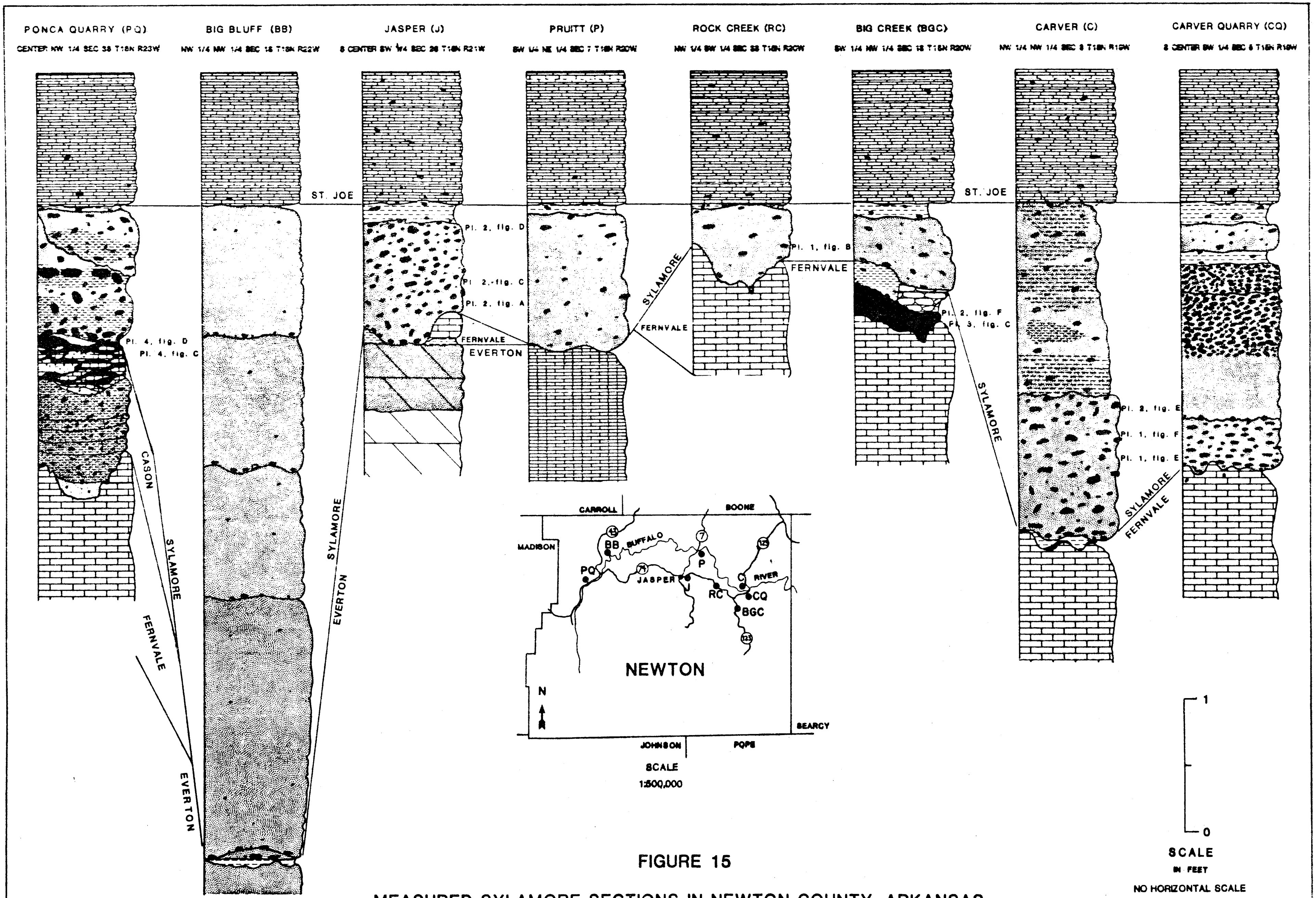


FIGURE 15

MEASURED SYLAMORE SECTIONS IN NEWTON COUNTY, ARKANSAS contributed first-cycle detrital grains. Middle Ordovician supermature quartz arenites provided only a minor source, apparently because they were covered by younger Ordovician strata. Deeper erosion prior to the Sylamore transgression removed these younger strata from broad areas of northern Arkansas, exposing the Middle Ordovician quartz arenites, which, along with remnants of the phosphatized Fernvale and Cason to the south, served as source material for the Sylamore Sandstone.

REFERENCES

Adams, George I., and Ulrich, E. O., 1904, Zinc and Lead Deposits of Northern Arkansas: U. S. Geological Survey, Professional Paper No. 24, 121 p.

Balson, P. S., 1980, The Origin and Evolution of Tertiary Phosphorites from Eastern England: Jour. of Geological Society of London, V. 137, p. 723–729.

Blatt, Harvey, 1982, Sedimentary Petrology, W. H. Freeman and Company, New York, 564 p.

Branner, John C., 1897, The Phosphate-Deposits of Arkansas: Trans. American Inst. of Mining Engineers, V. 26, p. 580–598.

Branner, J. C. and Newsom, J. F., 1902, The Phosphate Rocks of Arkansas: Arkansas Agriculture Experiment Station, Bulletin 74, p. 61-123.

Cook, Peter J. and McElhinny, M. W., 1979, A reevaluation of the Spatial and Temporal Distribution of Sedimentary Phosphate Deposits in the Light of Plate Tectonics: Economic Geology, V. 74, p. 315–330.

Craig, William W., 1969, Lithic and Conodont Succession of Silurian Strata, Batesville District, Arkansas: Bulletin Geol. Society of America, V. 80, p. 1621–1628.

———; 1975b, Stratigraphy and Conodont Faunas of the Cason Shale and the Kimmswick and Fernvale Limestones of Northern Arkansas; Headrick, K. N. and Wise, O. A. ed., Contributions to the Geology of the Arkansas Ozarks, Arkansas Geological Commission, p. 61–94.

Folk, Robert L., 1974, Petrology of Sedimentary Rocks: Univ. of Texas, Austin; Hemphill Publishing Company, 182 p.

Freeman, Tom and Schumacher, D., 1969, Qualitative Pre-Sylamore (Devonian-Mississippian) Physiography Delineated by Onlapping Conodont Zones, Northern Arkansas: Bulletin Geol. Society of America, V. 80, p. 2327–2334.

Frezon, Sherwood E. and Glick, E. E., 1959, Pre-Atoka Rocks of Northern Arkansas: U. S. Geological Survey, Professional Paper No. 314-H, p. 178–179. and Plate 24.

Hopkins, T. C., 1893, Marbles and Other Limestones: Arkansas Geological Survey, Annual Report for 1890, V. 4, p. 112–114.

Krumbein, W. C., 1940, Flood Gravel of San Gabriel Canyon, California: Bulletin Geol. Society of America, V. 51, p. 636–676.

Lemastus, Steven W., 1979, Stratigraphy of the Cason Shale (Ordovician-Silurian), Northern Arkansas: unpublished Masters thesis, University of New Orleans, 94 p.

Maher, J. C. and Lantz, R. J., 1953, Geology of the Gilbert Area, Searcy County, Arkansas: U. S. Geological Survey, Oil and Gas Investigation Map OM 132.

Manheim, F., Rowe, G. T., and Jipa, D., 1975, Marine Phosphorite Formation Off Peru: Jour. of Sedimentary Petrology, V. 45, No. 1, p. 243–251.

McBride, Earl F., 1963, A Classification of Common Sandstones: Jour. of Sedimentary Petrology, V. 33, P. 664–669.

McKnight, Edwin T., 1935, Zinc and Lead Deposits of Northern Arkansas: Bulletin U. S. Geological Survey, No. 853, p. 66–73.

Penrose, R. A. F., Jr., 1891, Manganese – Its Uses, Ores, and Deposits: Arkansas Geological Survey, Annual Report of 1890, V. 1, p. 112–115.

Pettijohn, Francis J., 1975, Sedimentary Rocks; Third Edition: Harper and Row Publishers, New York, p. 52–61, 628 p.

Pevear, David R. and Pilkey, O. H., 1966, Phosphorite in Georgia Continental Shelf Sediments: Bulletin Geol. Society of America, V. 77, p. 849–858.

Purdue, A. H., 1907, Developed Phosphate Deposits of Northern Arkansas: Bulletin U. S. Geological Survey, No. 315, p. 463–473.

Purdue, A. H. and Miser, H. D., 1916, Description of Eureka Springs and Harrison Quadrangles (Arkansas-Missouri): U. S. Geological Survey, Geol. Atlas, Folio 202, 21 p.

Riggs, Stanley R., 1979a, Petrology of the Tertiary Phosphorite System of Florida: Economic Geology, V. 74, p. 195–220.

Ross, Rueben J., Jr., 1976, Ordovician Sedimentation in the western United States; *in* Bassett, M. G., ed., The Ordovician System: Proceedings of a Paleontological Assoc. Symposium, Birmingham, September, 1974, p. 73–105; Univ. of Wales Press and National Museum of Wales, 696 p.

Sheldon, Richard P., 1980, Episodicity of Phosphate Deposition and Deep Ocean Circulation — a Hypothesis: Bentor, Y. K. ed., Marine Phosphorites, Society of Economic Paleontologists and Mineralogists Special Publication No. 29, p. 239—247.

Swanson, Vernon E. and Landis, E. R., 1962, Geology of a Uranium-Bearing Black Shale of Late Devonian Age in North-Central Arkanas: Arkansas Geol. and Conservation Commission, Information Circular No. 22, 16 p.

Ulrich, E. O. and Adams, G. I., 1905, Description of the Fayetteville Quadrangle (Arkansas-Missouri): U. S. Geological Survey, Geol. Atlas, Folio 119, 20 p.

EXPLANATION OF PLATE 1

(All figures are photomicrographs of thin sections.)

Figure A. Section BH, x34. Subphosphoarenite from a thin Sylamore section, with bimodal distribution of well-rounded quartz, poikilotopic calcite and small, subordinate patches of fine-grained dolomite. Arrows point to crinozoan biophosphomicrite fragments.

Figure B. Section RC, x34. Subphosphoarenite from a thin Sylamore section with well-rounded quartz and well-rounded to angular phosphomicrite grains in anhedral sparry to poikilotopic calcite cement.

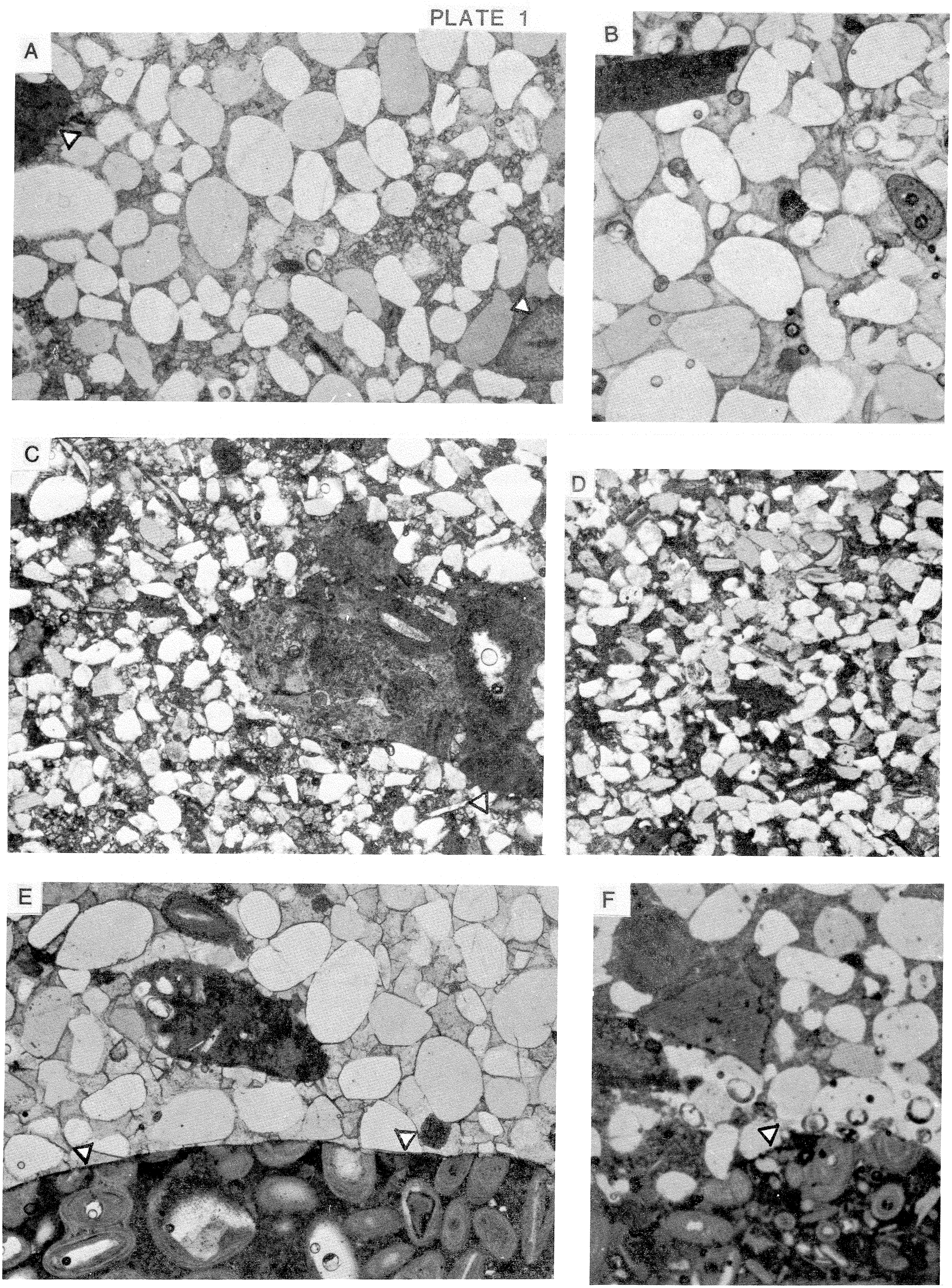
Figure C. Section BH, x34. Cason subphosphoarenite with angular and some rounded quartz in dolomitic, clayey matrix and cement. Contains angular crinozoan biophosphomicrite granule; note arrow points out truncated crinozoan skeletal fragment along margin of clast.

Figure D. Section BH, x34. Cason subphosphoarenite with angular to subrounded quartz in dolomitic, clayey matrix and cement. Contains numerous fine sand-size phosphomicrite or crinozoan biophosphomicrite fragments and fragmented phosphatic brachiopods.

Figure E. Section C, x34. Pebbly subphosphoaranite of thin Sylamore section with well-rounded quartz, oophosphomicrite pebble, solitary phosphate ooid, and crinozoan biophosphomicrite granule in poikilotopic calcite cement. Notice arrows point out pebble-margin truncation of ooids in oophosphomicrite pebble.

Figure F. Section C, x34. Pebbly subphosphoarenite of Sylamore with well-rounded quartz, oophosphomicrite pebble, and granule-size phosphomicrite and crinozoan biophosphomicrite fragments within detrital clay matrix and fine sparry calcite cement. Notice detrital clay in lower left adjacent to oophosphomicrite pebble with truncated allochems. Arrow indicates margin of pebble.

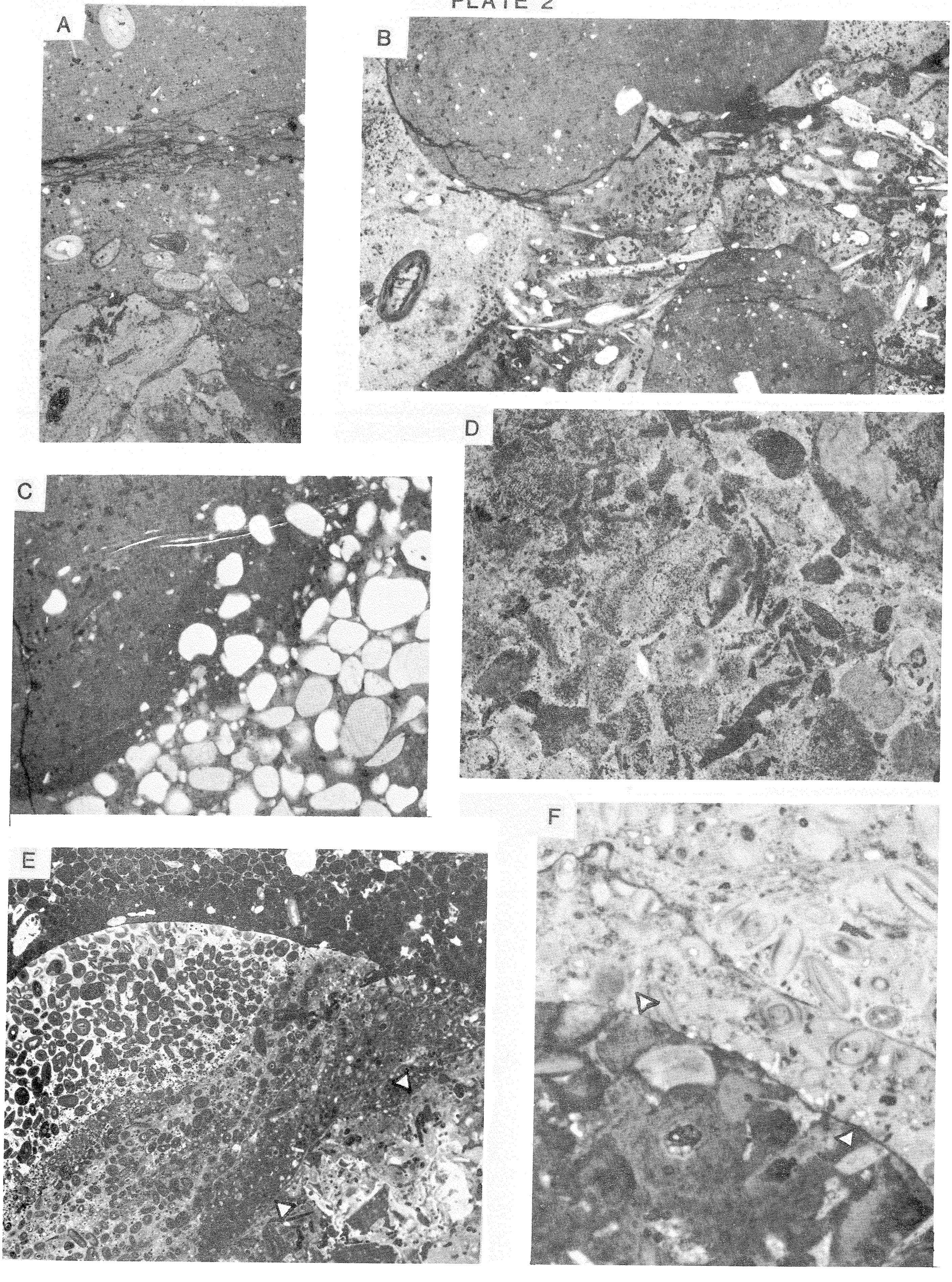
PLATE 1



EXPLANATION OF PLATE 2

Figure A.	Photomicrograph of Sylamore phosphate pebble. Section J, x28.8. Lithology of oophosphomicrite pebble that contains a subrounded crinozoan biophosphomicrite granule.
Figure B.	Photomicrograph of Cason phosphate pebble. x34. Lithology of biophosphomicrite pebble from Cason exposure at Gilbert city limits. Note burrow-mottled texture of phosphomicrite matrix and fragments of phosphatic brachiopods, silt-size quartz, and solitary ooid.
Figure C.	Photomicrograph of Sylamore phosphate pebble. Section J, x34. Lithology of quartzose phosphomicrite pebble. Note burrow-mottled texture with clustered well-rounded quartz.
Figure D.	Photomicrograph of Sylamore phosphate pebble. Section J, x34. Lithology of crinozoan biophosphomicrite pebble. Crinozoan framework recognized by the porous, granular microtexture.
Figure E.	Negative print of Sylamore pebble thin section. Section C, x8. Pebbly subphosphoarenite of Sylamore with oophosphomicrite/crinozoan biophosphomicrite pebble. Crinozoan biophosphomicrite lithology and oophosphomicrite lithology in sharp contact in lower right. Note three distinct ooid layers. Arrows indicate oophosphomicrite/crinozoan biophosphomicrite contact.
Figure F.	Photomicrograph of oophosphomicrite/crinozoan biophosphomicrite, in an erosional remnant of phosphatized limestone on top of the Fernvale. Section BGC, x34. Sharp contact between phosphatized oolitic limestone of the Cason and the phosphatized crinozoan limestone of the Fernvale. Notice truncation of crinozoan bioclasts along contact. Arrows indicate contact.

PLATE 2



EXPLANATION OF PLATE 3

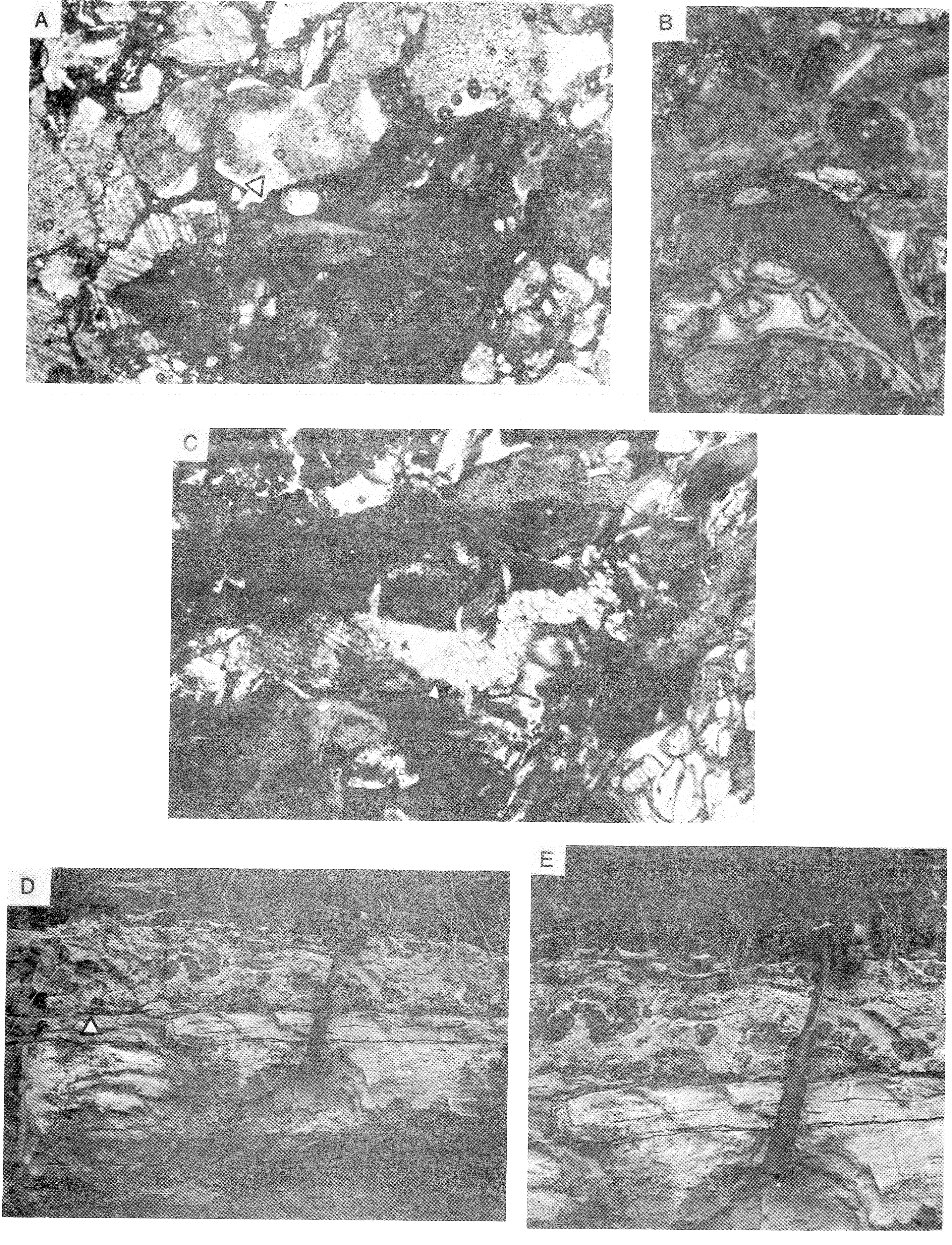
Figure A. Photomicrograph of Fernvale, Section BH, x34. Cemented rubble zone in the upper portion of the Fernvale caused by karstification of the limestone. Note crinozoan biophosphomicrite fragment of *in situ* phosphatized Fernvale and the interstitial phosphatic silt mixed with coarsely crystalline crinozoan fragments of unphosphatized Fernvale. Arrow indicates crinozoan biophosphomicrite fragment.

Figure B. Photomicrograph of phosphatized Fernvale, Section BH, x34. Crinozoan biophosphomicrite at the top of eroded Fernvale surface directly beneath the Cason at Barren Hollow. Coarsely crystalline calcite fills vugs surrounded by cryptocrystalline isopachous phosphate cement.

Figure C. Photomicrograph of phosphatized Fernvale, Section BGC, x34. Arrow indicates coarsely crystalline calcite. Crinozoan biophosphomicrite of phosphatized Fernvale with coarsely crystalline replacement calcite filling vugs in phosphatized limestone. From phosphatized limestone of the Fernvale directly beneath truncation by phosphatized oolitic limestone of the Cason (Plate 2, Fig. F).

Figure D&E. Roadcut exposure of Sylamore Sandstone at Jasper. Sylamore in disconformable contact with dolomitic quartz arenite/sandy dolostone of the Everton. Note various distributions of phosphate pebbles within Sylamore. Arrow in Figure D points to Sylamore/Everton contact.

PLATE 3



EXPLANATION OF PLATE 4

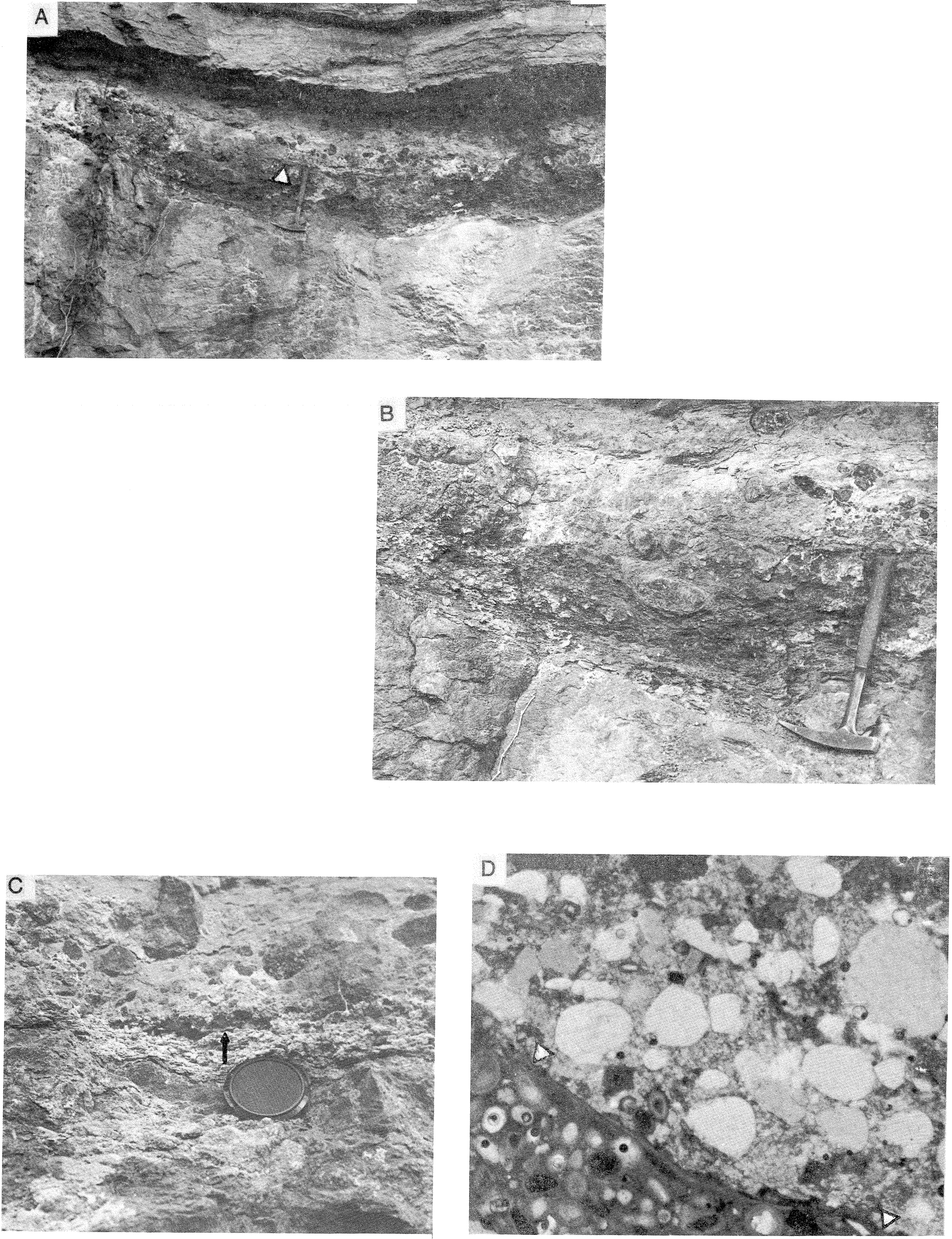
Figure A. Section PQ. Quarry face exposure at Ponca of detrital interval which includes strata of both the Sylamore and Cason. Karstified, undulose surface of the Fernvale is visible directly beneath the detrital interval. Cason strata is preserved in lows upon this surface. Conformable contact with overlying St. Joe. Arrow indicates Sylamore/Cason contact.

Figure B. Section PQ. Close-up of the Cason-Sylamore interval. Sylamore contact directly above the end of the hammer handle. Cason is composed of conglomeratic, dolomitic siltstone, sandstone, and shale overlain by phosphatized oolitic limestone. Hard crust of oophosphomicrite protects the underlying detrital strata of the Cason. Superjacent Sylamore, above the hammer handle, is an argillaceous, conglomeratic subphosphoarenite.

Figure C. Section PQ. Close-up of Sylamore-Cason contact. Conglomeratic Sylamore contains numerous oophosphomicrite pebbles in addition to other pebble textures. Top of Cason oophosphomicrite layers approximately one inch above lens cap.

Figure D. Photomicrograph of Sylamore-Cason oophosphomicrite contact. Section PQ, x34. Subphosphoarenite of Sylamore with well-rounded quartz, few solitary phosphate ooids in calcite cement in contact with Cason oophosphomicrite layers in lower left hand corner. Arrows point out contact.

PLATE 4



THE GEOLOGIC SIGNIFICANCE OF TEXTURES OF PALEOZOIC CHERT AND NOVACULITE IN THE OUACHITA MOUNTAINS OF ARKANSAS AND OKLAHOMA

By

W. D. Keller[1], Charles G. Stone[2], and Alice L. Hoersch[3]

Textures of chert and novaculite observable in scanning electron micrographs, (SEMs) are useful as a practical geologic thermometer for estimating the maximum temperature to which those silica rocks have been subjected to during and since deposition. Such information may then be applied during exploration for hydrocarbons, i. e., estimating temperatures of maturation or degradation, and in search for metallic and non-metallic minerals whose deposition or concentration may be temperature related.

The most extensively studied large geologic area, known to us, that illustrates the occurrence and application of chert-novaculite textures in relation to their thermal history is in the Ouachita Mountain fold belt in Arkansas and Oklahoma (Keller, Viele, and Johnson, 1977). Within a combined regional and contact metamorphic setting these Paleozoic cherts and novaculites show a sequential range in textures (in the same formation) from cryptocrystalline, anhedral quartz in the nonmetamorphosed chert and novaculite (Fig. 1) to coarse euhedral, polygonal, triple-point quartz crystals, as in Arkansas honestone (Fig. 2) and even coarser texture in xenoliths where the crystal diameters may exceed 100 μm.

These textures, moreover, can be morphologically correlated with the textures and crystal sizes developed in chert recrystallized at a contact metamorphic aureole on the Isle of Skye, Scotland, where classic metamorphic mineral suites from talc, through tremolite, diopside, and forsterite grades are represented (Hoersch, 1981; Keller, Stone and Hoersch, manuscript in preparation). Based on textural morphology and crystal sizes, representative SEMs of textures in rocks from Arkansas that are textural counterparts to those from Skye, are illustratively shown in Figures 3 to 7. These range in mean apparent diameters progressively from approximately 0.9 μm in talc grade, to 1.5 μm in tremolite grade, 5 μm in diopside grade, 7.3 μm in forsterite grade, all approximate values, to 100 μm or more in periclase grades.

To further show the areal distribution of the crystal sizes in the chert-novaculite Ouachita fold belt, a map was prepared using numerical data obtained from SEMs of many specimens of Paleozoic chert-novaculite collected from outcrops in the Ouachita fold belt. From the SEMs, the mean apparent diameters of the triple-point quartz crystals were measured along transects across the SEMs, and were plotted by isopleths, thus showing the textural sizes in the fold belt. The map, Figure 8, defines a linear 25–65 km (15–40 miles) wide belt that extends from Little Rock, Arkansas, for about 250 km (155 miles) west to Broken Bow, Oklahoma. Within this belt, the crystal sizes increase from the margins to the core. Two anomalies of coarse crystallinity are also present, one near Little Rock, where the crystals are about 35 μm in diameter, and

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another near Broken Bow, where the crystals are 15 μm in diameter. This textural belt encompasses the most intense predominantly Late Paleozoic structural deformation, and Mesozoic vulcanism, in the Ouachita Mountains. Previous studies have interpreted the rocks in the core of the regional fold belt to have attained a maximum metamorphic grade in the zeolite to lower green schist facies.

Cherts and novaculites adjacent to Magnet Cove, site of a Cretaceous age pluton in the eastern Ouachita Mountains of Arkansas, illustrate an overprinting of polygonal triple-point texture. Crystal sizes range from a background value of about 5 μm (talc grade) in chert about 305 m (1000 feet) from the pluton, to over 100 μm (periclase grade) from near the contact. Private drilling operations indicate that the pluton contact dips about 45^{o} beneath much of the sedimentary rock that locally exhibits anomalous crystallinity. Homogenization temperatures of vein quartz determined by Jackson and Nichols (1973, personal communication) show a temperature gradient along this profile of slightly above 200^{o} C in Quartz 4500 m (14,500 feet) from the pluton to about 440^{o} C near the contact. Novaculite xenoliths in the adjoining Potash Sulphur Springs intrusive are also very coarse in texture, 100 μm or larger, as found in periclase metamorphic grade (temperature approximately 760^{o}C) at Crestmore, California (Carpenter, 1967).

The triple-point texture and coarseness of chert and novaculite are obviously related to the degree of thermal metamorphism brought about by various heating events (Spry, 1969). They were developed in both regional and contact metamorphism, but the changes due to individual agents, namely temperature, physical deformation, time, depth of burial, and mineralizers have not yet been separately resolved.

In addition to the use of triple-point texture in chert-novaculite for estimating rock temperatures, some of the coarsely recrystallized rocks themselves have special economic value. The world-famous Arkansas honestone owes its remarkable blade-sharpening properties to the natural cutting edges of its microcrystals of quartz as in Figure 2, when the stone is processed to a plane working surface. Moreover, the edges are continually renewed with undiminished efficiency as the working surface of the stone is worn down during use which thereby exposes new edges. Alternatively, where weathering has partially disintegrated the novaculite it becomes a potential deposit of "soft-lump" ore (Figs. 9 and 10) from which industrial tripoli powder can be produced by further disaggregating either with, or without, crushing the individual grains, and separating them into size fractions (Keller, 1978).

In summary, by using SEM techniques, very small quantities of chert-novaculite from the Ouachita Mountains can serve to identify areas that have undergone elevated rock temperatures resulting from deep burial, mechanical stresses, intrusions, hydrothermal exhalations, and other thermal events, and can be used as a guide in exploration for metallic and non-metallic mineral deposits in those areas.

ACKNOWLEDGEMENT

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REFERENCES

Carpenter, A. B., 1967, Mineralogy and petrology of the system $CaO-MgO-CO_2-H_2O$ at Crestmore, California: Amer. Mineral., v. 52, p. 1341–1362.

Hoersch, A. L., 1981, Progressive metamorphism of the chert-bearing Durness limestone in the Beinn au Dubhaich aureole, Isle of Skye, Scotland: a re-examination: Amer. Mineral., v. 66, p. 442–506.

Keller, W. D., 1978, Textures of tripoli illustrated by scanning electron micrographs: Econ. Geol., v. 73, p. 442–446.

Keller, W. D., Viele, G. W., and Johnson, C. H., 1977, Textures of Arkansas novaculite indicate thermally induced metamorphism: Jour. Sed. Petrol. v. 47, p. 834–843.

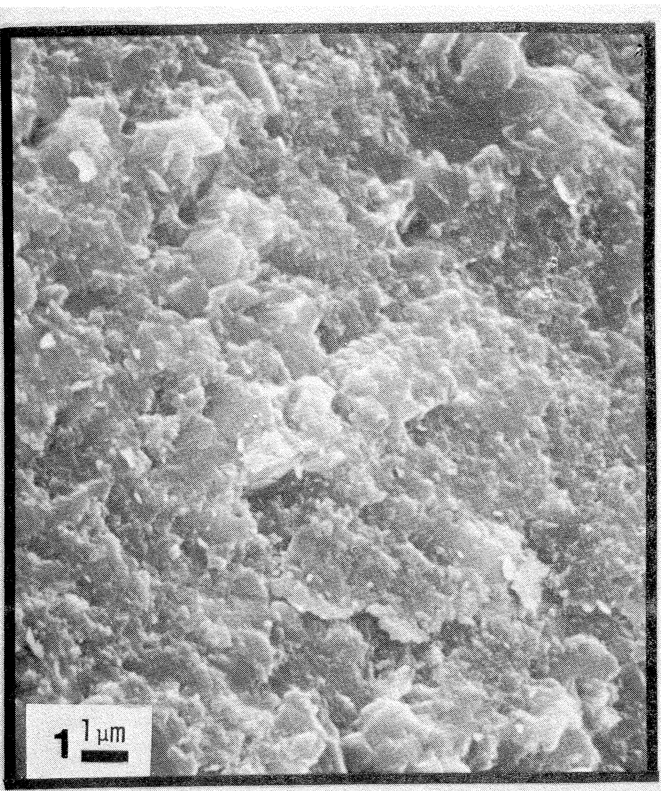
Spry, Alan, 1969, Metamorphic Textures: Oxford Press, 350 p.

Figure 1. Scanning electron micrograph (SEM) of Arkansas Novaculite at Atoka, Oklahoma, that shows fine-grained, anhedral mosaic quartz typical of sedimentary chert. The scale bar represents 1 μm, magnification 6000X.

Figure 2. SEM of honestone-quality Arkansas Novaculite showing polygonal, triple-point texture. Collected from a quarry near (east) Hot Springs, Arkansas, 2000X.

Figure 3. SEM of talc-grade texture (Arkansas counterpart to Skye), Arkansas Novaculite from Highway 375, SE NE , sec. 11, T. 3 S., R. 30 W., Polk County, Arkansas, 5000X.

Figure 4. SEM of tremolite-grade texture, Arkansas Novaculite, south of Shady Lake, in the NW SE , sec. 31, T. 4 S., R. 28 W., Polk County, Arkansas, 4800X.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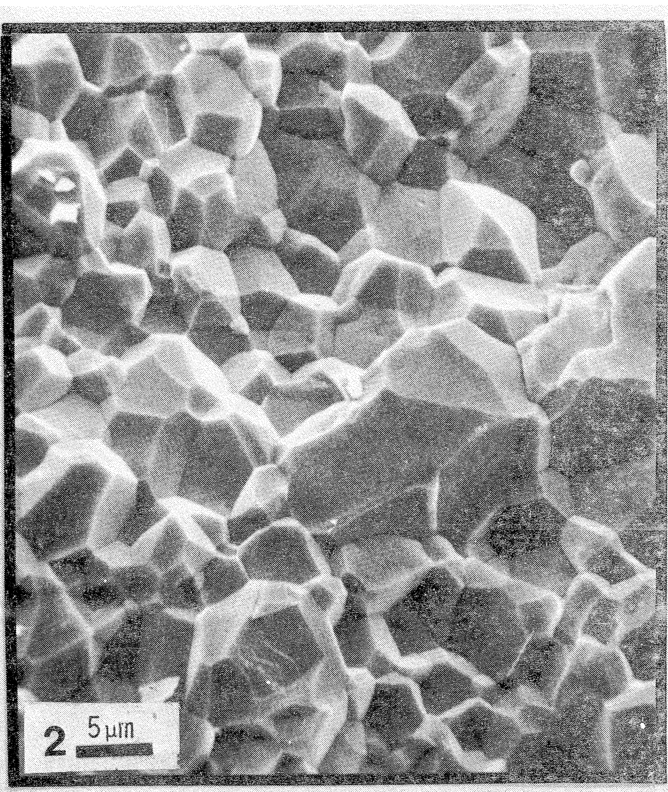
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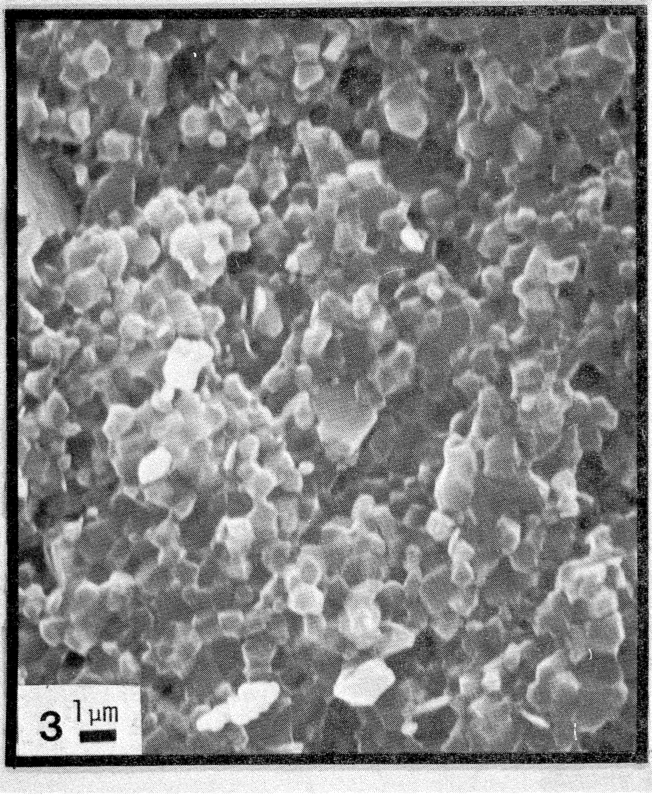
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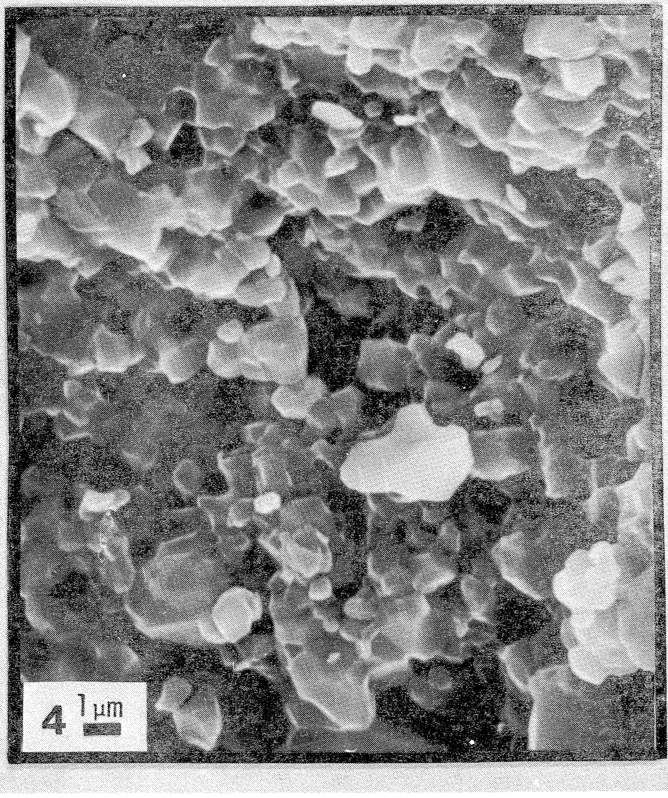
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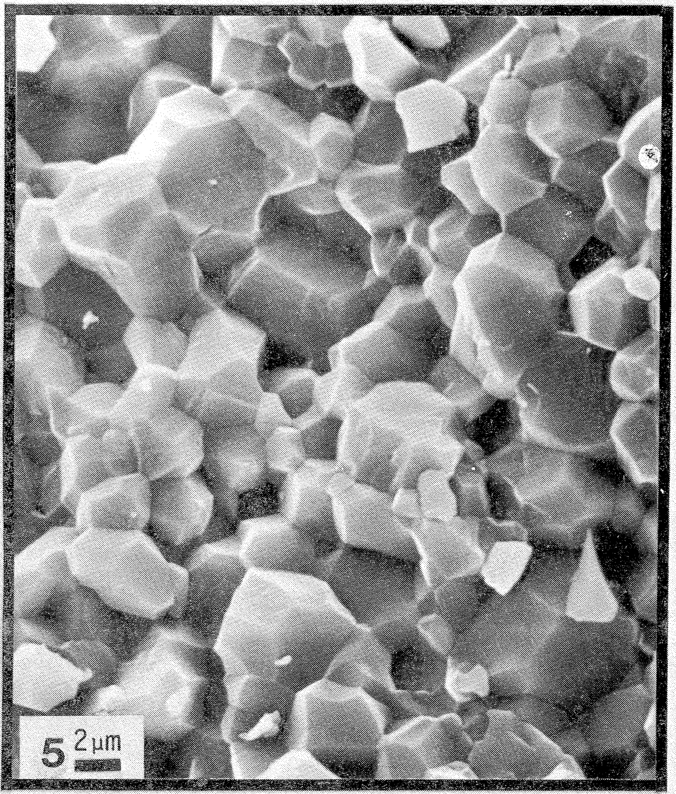
Figure 5. SEM of diopside-grade texture, Arkansas Novaculite, from a quarry in the NW SW sec. 9, T. 3 S., R. 16 W., Saline County, Arkansas, 3000X.

Figure 6. SEM of forsterite-grade texture, Arkansas Novaculite, Reyburn Creek gap in the center of SE¼, sec. 13, T. 3 S., R. 17 W., Hot Spring County, Arkansas, 2000X.

Figure 7. SEM of periclase-grade texture, xenoliths in the intrusion at the Union Carbide Vanadium mine, Potash Sulphur Springs, Arkansas, 200X.

Figure 10. SEM of partially weathered and disintegrated, "soft-lump" ore from the tripoli deposit of the Malvern Minerals Company, 3000X.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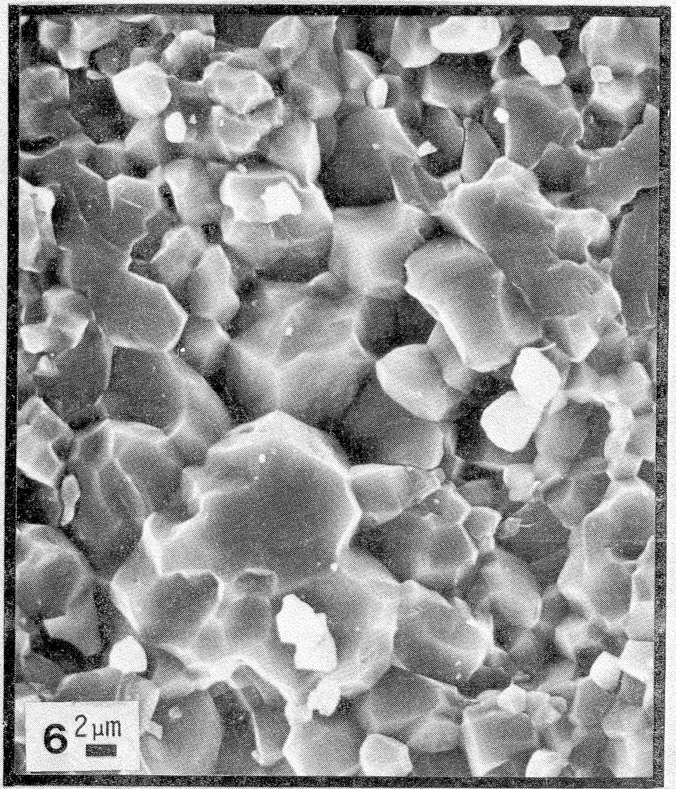
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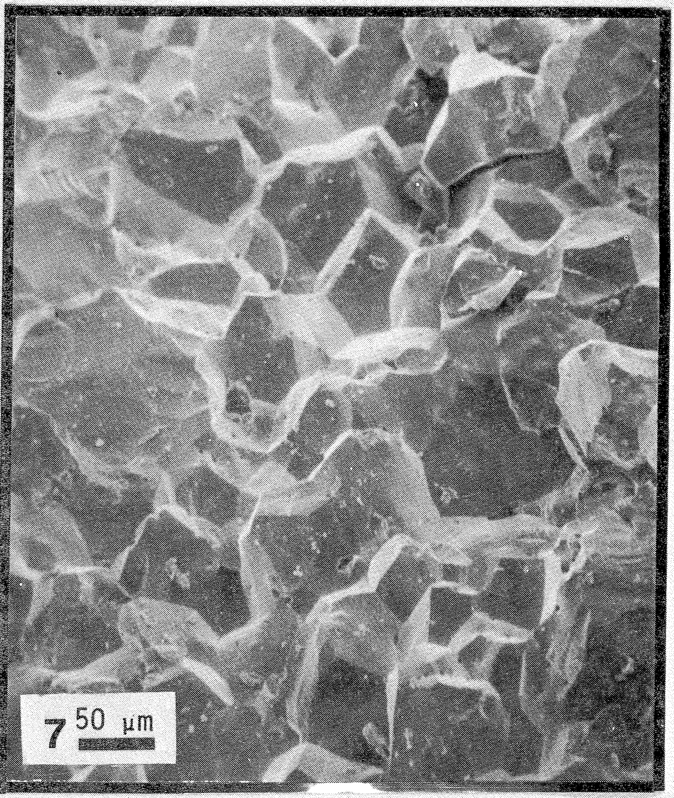
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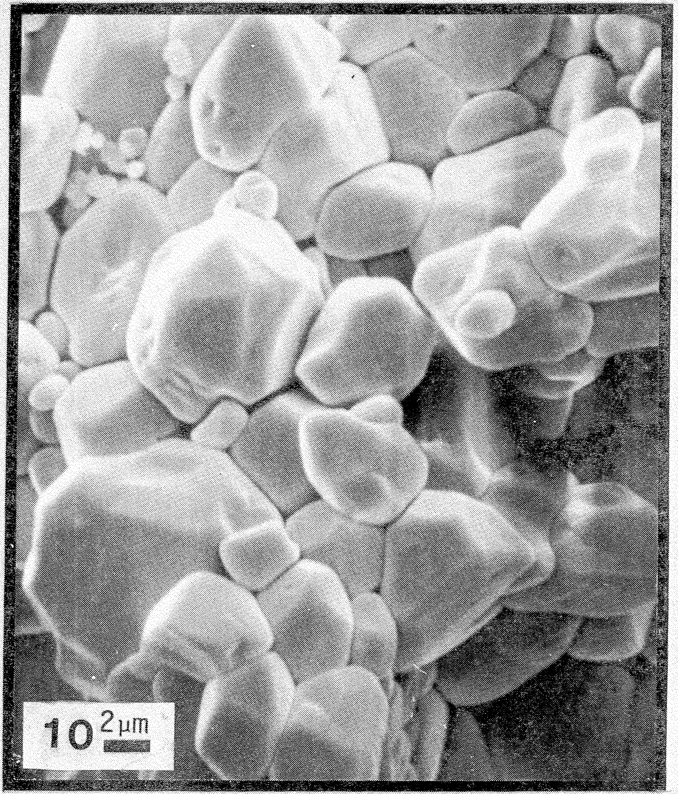
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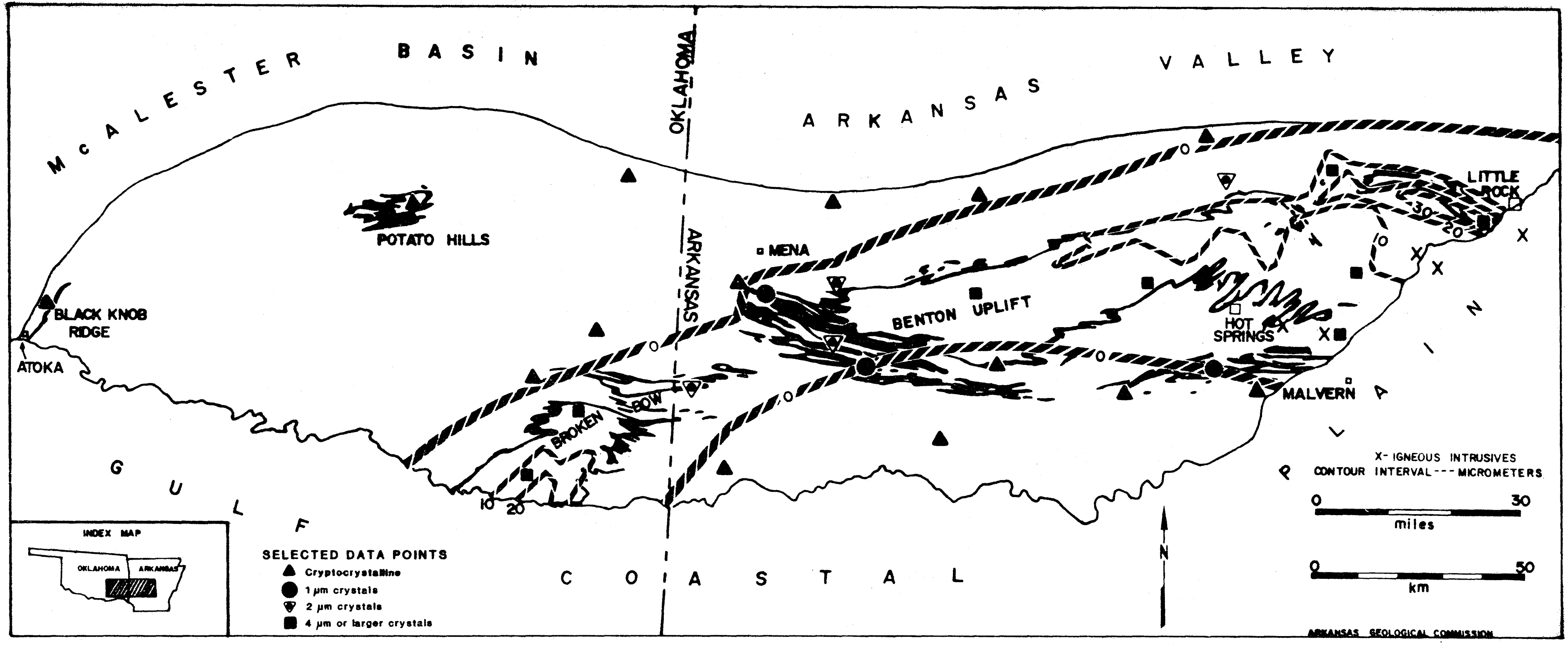

Figure 8. Generalized map showing distribution, by isopleths, of the mean apparent diameters of crystals in cherts and novaculites of the Ouachita Mountains, Arkansas and Oklahoma. The "0 μm boundary" is only transitional, and includes some triple-point crystals ⟨ 1 μm in apparent diameter.

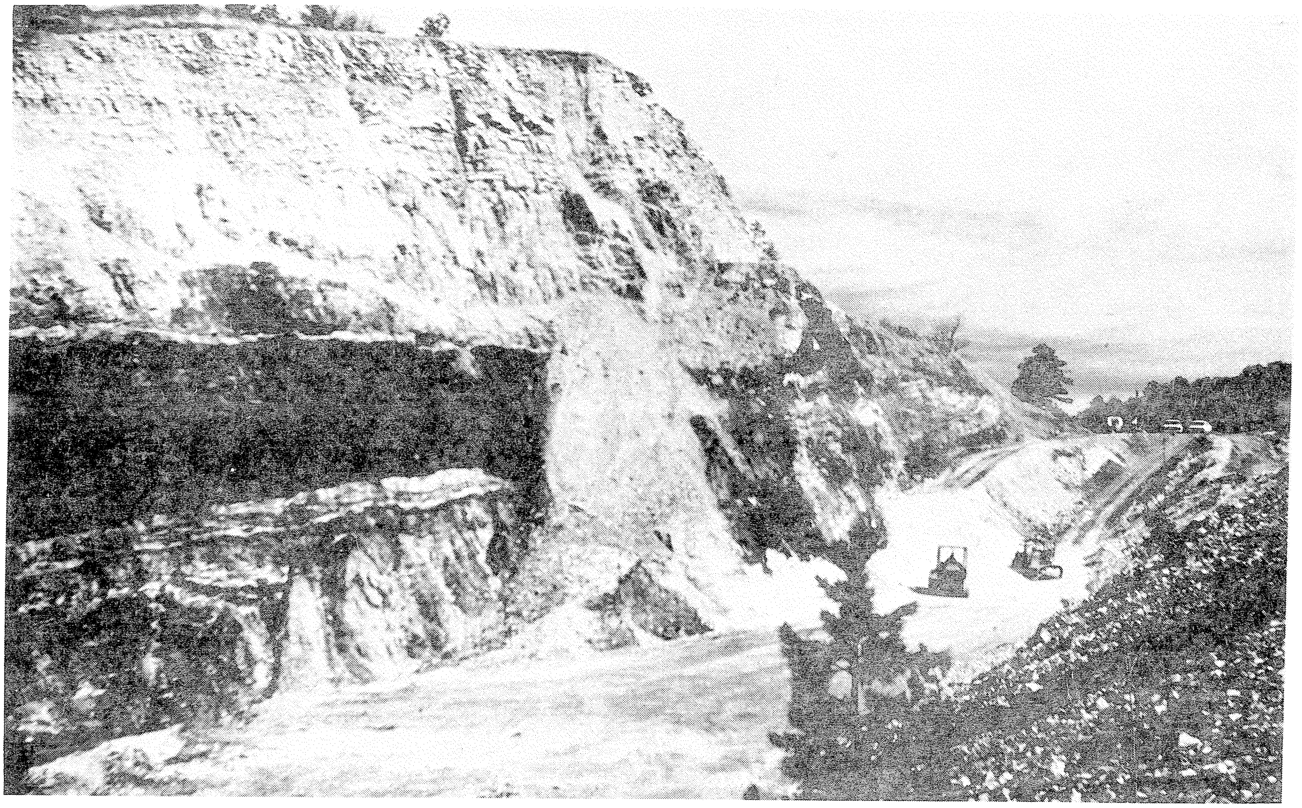

Figure 9. Open-pit, tripoli ("Novacite") mine of the Malvern Minerals Company, Garland County, Arkansas. The stratigraphic section is inverted here; the white tripolitic, upper member of the Arkansas Novaculite is at the base of the deposit, the dense novaculite of the lower member is at the top, and the dark shale of the middle member is between them.

MISERITE, A REVIEW OF WORLD OCCURRENCES WITH A NOTE ON INTERGROWN WOLLASTONITE

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ABSTRACT

Miserite with variable composition but essentially $K(Ca, Ce)_4 Si_5O_{13}(OH)_3$ was first found in Arkansas and has since been reported from Quebec, Wisconsin, and six localities in the (Asiatic) USSR. It is generally associated with calcareous sediments metamorphosed by alkalic intrusives; and in Arkansas and the USSR is intergrown with wollastonite. New analyses of the Arkansas miserite and wollastonite are given, with X-ray powder patterns. Optical, chemical, and X-ray data from the literature of other occurrences show the similarities and differences of miserite from diverse localities. The recent and definitive crystal structure determination of miserite from Quebec, by J. Douglas Scott has defined the chemical and physical parameters of miserite.

INTRODUCTION

Miserite, essentially a hydrous fluoro-silicate of calcium and potassium with minor and varying rare-earths, niobium and other cations, has been a problematic mineral from its first discovery in Arkansas by J. F. Williams (1891) almost a century ago. The first major revision of its composition (and change of name) was made by W. T. Schaller sixty years later (1950) and only recently has the crystal structure of miserite been finally determined by Scott (1976). For many years it was known only from the one Arkansas locality; but in the last quarter century it has been found in Quebec and Wisconsin and in six localities in the USSR. This paper presents a summary of what is now known about miserite, with new analyses and X-ray powder data for the Arkansas miserite and its associated wollastonite; and of the two newly discovered occurrences in North America. Because of the relative inaccessibility of much of the literature relating to the Russian occurrences, extensive reference to the latter is included in this report.

This report, it is hoped, will serve two purposes: first, as a compendium of the literature about miserite, which may be of use to mineralogists concerned with this mineral, and second, to make known to mineral collectors, especially those in Arkansas, the history and properties, in some respects unique, of one of the more noteworthy of Arkansas minerals.

HISTORICAL REVIEW

Miserite was first observed by J. Francis Williams (1891) at Potash Sulfur Springs, Arkansas. Williams reported it as a "pink variety of wollastonite." Based on erroneous analytical data (Table 6) he proposed a formula $5(Ca,Na_2)SiO_3$ + aq. with $Ca:Na_2 = 9$, and the name natroxonotlite.

The name miserite was given by W. T. Schaller (1950) in honor of Hugh D. Miser.

Schaller re-analyzed the mineral (Table 6) and derived a formula $KCa_4Si_5O_{13}(OH)_3$, therefore discrediting the name natroxonotlite. He also presented the first X-ray powder diffraction pattern (Table 3).

Prior to 1955, I. V. Kupriyanova discovered the first occurrence of miserite in the USSR in the Aldan Massiv, east of Lake Baikal (Fig. 1, Locality 2). Kupriyanova and Vasil'eva (1961) described it as being rich in rare earths (7.52 wt. %) and termed it therefore "rare-earth miserite," none having been then reported in the Arkansas miserite. It is found in metasomatic rocks composed of microcline, albite, and aegirine augite, in sheaf-like or parallel aggregates, also as tabular crystals; and alters to an aggregate of calcite, quartz, and allanite. X-ray powder data and chemical analysis were given. Later Rudnitskaya (1970) described miserite from Aldan Mesozoic intrusions, giving a new analysis. It is not quite clear whether this occurrence is the same as Kupriyanova's or not. However, Demchenko (1973) did restudy Kupriyanova's miserite, giving a new analysis, but no further X-ray pattern.

The second discovery of miserite in the USSR was by Ruzhov and Moleva (1960) in 1955, in the Khodzha Ahchan alkalic massiv, Alai Range, Kirghiz SSR (Central Asia) (Fig. 1, Locality 3). They gave chemical and X-ray data. The occurrence was re-studied by Ulyanova and Ilinskii (1964).

The next discovery was in 1958 by P. S. Kozlova (1962) in the Talaask Range, east of the Aral Sea in western Siberia (Fig. 1, Locality 4). She gave chemical and X-ray data.

Kravchenko and Bykova (1967) reported the discovery in 1959 of miserite in Southern Yakutia, eastern Siberia (Fig. 1, Locality 5). Besides chemical and X-ray data, they give an infrared absorption spectrum.

Arkhangel'skaya (1968) reported miserite from an albite-rich quartz microcline granite southwest of Lake Nichatka in the Olekma-Vitim Highlands, in eastern Siberia (Fig. 1, Locality 6). X-ray powder data (Table 5) are given, but no analysis. It forms rosettes and sheaves of radial spindle-shaped grains, and less commonly tabular crystals; sp. gr. 2.85, H $\sim$ 5. It shows replacement by an unknown white powdery substance.

About 1970 L. G. Berry and several of his students in Queens University, Ontario, identified miserite from Kipawa, Quebec (Fig. 1, Locality 7).

In 1972 Berry, Lin, and Davis published an abstract giving preliminary data on the Kipawa miserite. Slightly revised, this data appeared on JCPDS (X-ray powder diffraction file) Card No. 22–806. All this was made obsolete by the new data of Scott (1976).

Semenov, Duzmatov and Khomyakov (1973) described miserite from skarns of the Darei Piez alkalic massiv in Turkestan (Fig. 1, Locality 8), giving an analysis (Table 8), and physical properties but no X-ray data.

Scott gives crystallographic data for the Quebec, Arkansas and three USSR miserites, with slight variation in cell constants and density, and rare earth content varying from 0.0 (Ruzhov and Moleva, 1960) to 7.47 percent (Kipawa, Quebec). As determined by Scott, the symmetry of miserite is triclinic, space group $P\bar{1}$. To to solve the nearly century-old enigma of the miserite structure, Scott devoted "four years of painful trials" before success was achieved.

D. E. Appleman (1978, personal communication) re-refined the X-ray powder data of Scott on the Quebec miserite, and that of Milton (this paper) on the Arkansas and Wisconsin miserites (Table 3), obtaining closely agreeing unit cells for all three (Table

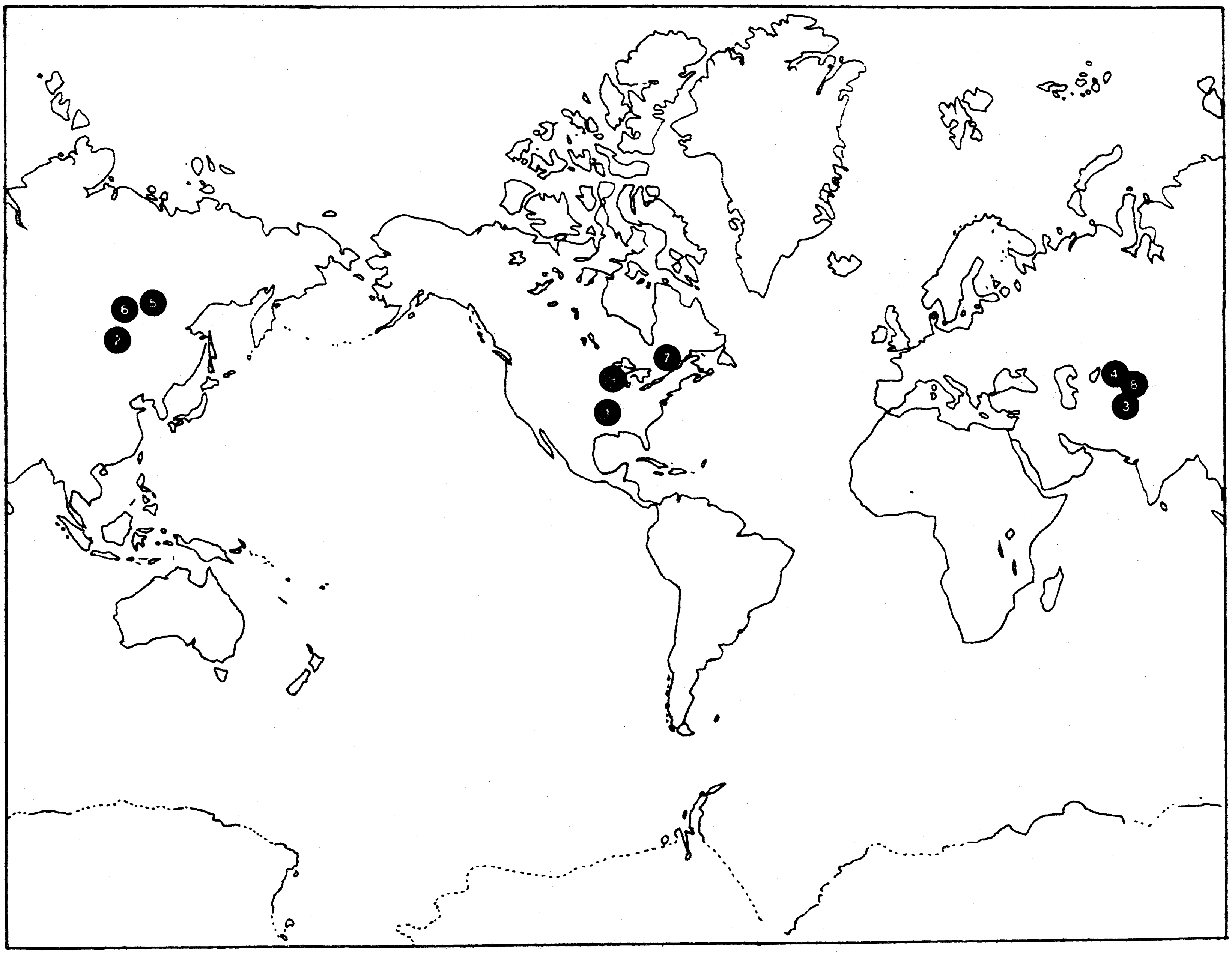

Figure 1 -- Map showing 9 occurrences of miserite.

4); and similar close agreement is shown by the three USSR miserites cited by Scott.

In the last thirty years then miserite has been recognized in six widespread USSR localities, and most recently in two more North American localities, in Quebec and Wisconsin. The unusually large crystals from Quebec afforded Scott cleavage fragments which made possible his structural study which has finally yielded precise knowledge of the unit cell. Until its complex structure was finally unraveled, each new discovery of miserite led to generally futile attempts to establish the correct chemical formula for the mineral. But now that the structure is understood, with each possible substituting atom assigned its proper place, it may be assumed that, although further discoveries of miserite will undoubtedly occur, the fairly copious controversies in the literature will have been resolved.

PHYSICAL PROPERTIES – MORPHOLOGY

Because of the fibrous structure of the mineral and its intimate intergrowth with associated minerals, wollastonite especially, it was for many years difficult, if not impossible, to obtain a pure sample for analysis or definitive crystallographic study. However, with great pains, suitable material for chemical and X-ray powder pattern was obtained from the Arkansas rock; and in recent years, massive miserite has been found in Quebec, Wisconsin, and possibly at least one USSR locality, permitting single-crystal X-ray study, if not a representation (crystal drawing) of the external morphology.

Optical data are given in Table 1; and fluorescence and phosphorescence, in Table 2.

Professor L. G. Berry (written communication, Feb. 10, 1978) notes that in the Kipawa, Quebec miserite no terminated crystals were found; there are cleavage pieces up to nearly two inches (5 centimeters) in maximum dimension, but (the rock) is nowhere vuggy, and the miserite is in tight contact with amphibole, eudialyte, and other minerals. Dr. J. Douglas Scott (written communication, Feb. 17, 1978) also notes that he knows of no euhedral free crystals of Kipawa miserite. Crystal outlines can be seen in rock slabs, but only cleavage fragments can be removed. Nor have any of the Russian authors given any information as to external morphology differing from that of the North American localities (fibrous or massive), or given any picture of single crystals.

Some difficulty was found in precisely locating the six USSR miserite localities from their published descriptions. Figure 1 (Map), locates three in south central Siberia, and three in eastern Siberia.

Fluorescence and phosphorescence of Miserite and Wollastonite

The author of the brochure listed in the References as Anonymous (1969) refers to the fluorescence of the Arkansas miserite as follows (p. 18):

> "This mineral is probably the rarest authenticated fluorescent mineral discovery of recent years . . . Some specimens have the typical green response of the rare earths with unfiltered ultraviolet."

However, somewhat different results were obtained by Mr. Pete Dunn, mineralogist of the U. S. National Museum, who found that (personal comm.) the miserite fluoresces feebly or doubtfully in violet pink both in long and short wave, and wollastonite pink cream in short wave, strong creamy yellow in long wave. Neither mineral shows phosphorescence, short or long wave.

TABLE 1. —— OPTICAL DATA

Indices of Refraction					
α	β	γ	2V	Locality	Source
1.58	——	1.59	60°	Arkansas	Schaller, 1950
1.587	1.589	1.5940	65°	Quebec	Berry, Lin, and Davis, 1972
				Wisconsin	Stephen S. Guggenhein (personal communication, 1976)
1.579	——	1.588	78°	Alai Range, Kirghiz SSR	Ruzhov and Moleva, 1960
1.588	1.593	1.600	68° (average)	Eastern Siberia	Kupriyanova and Vasileva, 1961
			65°–69° (range)	Aldan Massiv	
1.583	——	1.593	78°–79°	Talaask Mt. Range, USSR	Kozlova, 1962
1.583	——	1.589	60° ± 1°	Southern Yakutia, USSR	Kravchenko and Bukova, 1967
1.589	——	1.602	62°–65°	Olekma–Vitim Highlands USSR	Arkhangel'skaya, 1968
1.587	——	1.595	——	Aldan Intrusive, USSR	Rudnitskaya, 1970, 1972
1.577	——	1.588	80°	Khodzha Achkan, Alai Range, Kirghiz, USSR	Ulyanova and Illinskii, 1964
1.586	1.593	1.599	52°–78°		Demchenko, 1972, 1974
1.58		1.59			Semenov, Duzmatov, and Khomyakov, 1973

TABLE 2. (from Anonymous, 1969)

Fluorescence and phosphorescence of miserite and wollastonite

		Miserite	Wollastonite
Short wave			
	Fluorescence	cream to cream yellow	"shades of orange" (p. 31)
		with a slight greenish cast (p. 18)	"shades of yellow" (p. 32)
		also shades of red (p. 31)	
	Phosphorescence	----	dull orange (p. 18)
Long wave			
	non fluorescent (p. 18)		lavender pink to dull red
	phosphorescence	----	dull orange (p. 18)

Schaller (1950) briefly describing the Arkansas miserite, did not refer to fluorescence. However, his note was given some years later to D. L. Hughes, Union Carbide geologist, who gave it to the editor of Rocks and Minerals (Anon., 1972) who noted a "bright yellow" fluorescence. Possibly both he and others may have observed the yellow fluorescence of intergrown wollastonite.

CRYSTALLOGRAPHY

The unit cell constants of the Kipawa Lake Quebec miserite have been determined from Debye-Scherrer X-ray study (Berry, Lin, and Davis, 1972). Scott (1976) found a 3 mm. long cleavage fragment from a crushed specimen of Kipawa miserite; and succeeded in using this for a crystal structure determination. Its chemical composition is given in Table 6. Its structure is described by Scott (1976) as follows: Its structure is "characterized by the presence of three-membered sub-chains of silica tetrahedra, similar to those of wollastonite, but having all tetrahedra sharing three corners. Four such sub-chains link to produce a closed quadruple $Si_{12}O_{30}$ chain composite, infinite in extent parallel to *c* and centered on each corner in [001] projection. These pipes display almost perfect *mmm* symmetry and are crossed-linked parallel to *a* by K (potassium) atoms, adjacent pipes being displaced vertically by exactly c/2. Edge sharing columns of Ca polyhedra bonded to each other and to independent Si_2O_7 groups form a slab infinite in extent in *a* and *c*, which falls between the pipes and connects them in the *b* direction. . . . The miserite structure is a pyroxenoid derivative, the structural framework of which is described by the ideal formula $KCa_5(Si_2O_7)(Si_6O_{15})(OH,F)$." Two such units, (the unit cell) repeated in three dimensions, form the miserite crystal structure.

TABLE 3. — INDEXED X-RAY POWDER PATTERNS, THREE NORTH AMERICAN LOCALITIES

Arkansas						Wisconsin				Quebec					
Schaller (1950)		Milton* (1978)		Appleman* (1978)		Milton* (1978)		Appleman* (1978)		Scott (1976)				Appleman*(1978)	
I	d	I	d_{obs}	d_{calc}	hkl	I	d_{obs}	d_{calc}	hkl	I	d_{obs}	d_{calc}	hkl	d_{calc}	hkl
s	16.0	w	17.3	15.549	010	vs	15.531	15.593	010	100	15.42	15.57	010	15.559	010
—	—	vw	8.735	8.780	110	w	8.802	8.793	110	6	—	—	110	—	—
—	—	vvw	7.932	7.775	020	w	7.803	7.964	020	5	7.775	7.784	020	7.780	020
—	—	—	—	—	—	—	—	—	—	5	7.266	7.295	$\bar{1}10$	7.291	$\bar{1}10$
—	—	—	—	—	—	vw	6.852	6.875	001	—	—	—	—	—	—
—	—	w	5.172	5.183	030	mw	5.195	5.198	030	10	5.193	5.189	030	5.186	030
—	—	—	—	—	—	w	5.051	5.073	021	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	10	5.018	4.985	130	4.983	130
—	—	s	4.688	4.694	210	—	—	—	—	45	4.681	4.691	210	4.689	210
—	—	—	—	—	—	—	—	—	—	5	4.594	4.608	200	4.606	200
—	—	—	—	—	—	ms	4.496	4.480	$\bar{2}\bar{2}1$	—	—	—	—	—	—
—	—	—	—	—	—	mw	4.433	4.463	$1\bar{1}1$	—	—	—	—	—	—
—	—	vvw	4.408	4.390	220	—	—	—	—	8	4.374	4.383	220	4.382	220
—	—	—	—	—	—	ms	4.211	4.214	$0\bar{3}1$	—	—	—	—	—	—
—	—	vw	4.150	4.156	$\bar{1}30$	—	—	—	—	15	4.175	4.168	$\bar{1}30$	4.165	$\bar{1}30$
—	—	w	3.893	3.887	040	ms	3.908	3.898	040	35	3.887	3.892	040	3.890	040
—	—	vs	3.445	3.453	$\bar{1}\bar{2}2$	w	3.484	3.478	$\bar{1}\bar{2}2$	50	3.469	3.471	$\bar{1}\bar{2}2$	3.473	$\bar{1}\bar{2}2$
—	—	vs	3.339	3.343	211	ms	3.361	3.356	211	45	3.357	3.352	211	3.350	211
—	—	—	—	—	—	—	—	—	—	80	3.179	3.177	$\bar{1}22$	3.175	$\bar{1}22$
—	—	—	—	—	—	—	—	—	—	80	3.166	3.169	$\bar{1}\bar{3}2$	3.170	$\bar{1}\bar{3}2$
vs	3.15	vvs	3.144	3.140	310	vvs	3.149	3.147	310	90	3.140	3.141	310	3.139	310
—	—	—	—	—	—	—	—	—	—	64	3.106	3.113	050	3.112	050
—	—	vs	3.090	3.089	022	—	—	—	—	60	3.100	3.109	022	3.108	022
m	3.07	—	—	—	—	s	3.033	3.076	300	55	3.072	3.073	300	3.071	300
s	2.94	—	—	—	—	vs	2.942	2.931	330	57	2.922	2.922	330	2.921	330
—	—	vs	2.870	2.875	250	—	—	—	—	25	2.873	2.870	250	2.870	250
mw	2.82	vs	2.800	2.810	032	s	2.804	2.824	032	55	2.825	2.825	032	2.825	032
mw	2.78	—	—	—	—	—	—	—	—	40	2.776	2.780	$\bar{1}50$	2.778	$\bar{1}50$
m	2.68	—	2.663	2.669	$\bar{3}20$	s	2.682	2.677	$\bar{3}20$	45	2.677	2.675	$\bar{3}20$	2.673	$\bar{3}20$
—	—	—	—	—	—	—	—	—	—	5	2.641	2.644	160	2.643	160
—	—	w	2.481	2.504	$1\bar{5}1$	vw_b	2.501	2.515	$1\bar{5}1$	40	2.509	2.511	$1\bar{5}1$	2.510	$1\bar{5}1$
—	—	—	—	—	—	—	—	—	—	5	2.454	2.456	350	2.455	350
—	—	s	2.373	2.373	$\bar{4}\bar{1}2$	vw	2.367	2.382	$\bar{4}\bar{1}2$	65	2.377	2.377	$\bar{4}\bar{1}2$	2.377	$\bar{4}\bar{1}2$
—	—	—	—	—	—	—	—	—	—	5	2.362	2.363	$\bar{2}50$	2.361	$\bar{2}50$
—	—	—	—	—	—	—	—	—	—	8	2.346	2.351	410	2.350	410
—	—	vvw	2.301	2.302	400	ms_{vb}	2.308	2.308	400	10	2.304	2.304	400	2.303	400
—	—	—	—	—	—	—	—	—	—	8	2.289	2.289	430	2.288	430
—	—	vvw	2.242	2.237	$\bar{1}52$	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	5	2.227	2.224	070	2.223	070
—	—	—	—	—	—	—	—	—	—	2	2.220	2.222	360	2.221	360
—	—	—	—	—	—	—	—	—	—	4	2.195	2.200	$\bar{3}40$	2.198	$\bar{3}40$
—	—	vw	2.159	2.160	$1\bar{5}2$	—	—	—	—	—	—	—	—	—	—
mw	2.10	s	2.097	2.096	$0\bar{6}2$	ms	2.104	2.107	$0\bar{6}2$	43	2.103	2.102	$0\bar{6}2$	2.102	$0\bar{6}2$
—	—	s	1.964	1.964	322	ms	1.974	1.972	322	40	1.968	1.970	322	1.969	322
—	—	vw	1.910	1.917	332	—	—	—	—	35	1.922	1.922	332	1.921	332
—	—	vs	1.832	1.831	$\bar{2}\bar{1}4$	ms	1.848	1.844	$\bar{2}\bar{1}4$	60	1.845	1.844	$\bar{2}\bar{1}4$	1.844	$\bar{2}\bar{1}4$
—	—	vw	1.761	1.759	352	—	—	—	—	—	—	—	—	—	—
—	—	vvw	1.715	1.713	$\bar{5}20$	—	—	—	—	—	—	—	—	—	—
m	1.672	s	1.667	1.667	$\bar{2}\bar{9}2$	s	1.675	1.673	$\bar{2}\bar{9}2$	66	1.667	1.667	$\bar{2}\bar{9}2$	1.667	$\bar{2}\bar{9}2$
—	—	—	—	—	—	w	1.642	1.642	$\bar{1}82$	20	1.641	1.642	$\bar{1}82$	1.641	182
—	—	vw	1.620	1.621	$\bar{5}41$	—	—	—	—	—	—	—	—	—	—
—	—	s	1.579	1.580	542	w	1.589	1.585	$\bar{5}42$	50	1.589	1.585	$\bar{5}42$	1.584	$\bar{5}42$
—	—	vw	1.555	1.565	630	w	1.566	1.568	630	10	1.564	1.563	630	1.563	630
—	—	—	—	—	—	w	1.544	1.544	$\bar{3}91$	—	—	—	—	—	—
—	—	—	—	—	—	w	1.530	1.530	$\bar{2}73$	—	—	—	—	—	—
—	—	vvw	1.522	1.522	432	—	—	—	—	—	—	—	—	—	—
—	—	vvw	1.517	1.517	1101	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	vw	1.444	—	—	—	—	—	—	—	—
—	—	vw	1.417	—	—	vw	1.423	—	—	—	—	—	—	—	—
—	—	vw	1.322	—	—	vvw_b	1.310	—	—	—	—	—	—	—	—
—	—	vw	1.289	—	—	vw	1.295	—	—	—	—	—	—	—	—
—	—	vvw	1.261	—	—	w_{vb}	1.265	—	—	—	—	—	—	—	—
—	—	vvw	1.215	—	—	—	—	—	—	—	—	—	—	—	—
—	—	vvw	1.201	—	—	—	—	—	—	—	—	—	—	—	—
—	—	vvw	1.171	—	—	—	—	—	—	—	—	—	—	—	—

* -- Unpublished X-ray data

TABLE 4. — Unit cell data (D. E. Appleman,[1] 1978

	Quebec		Arkansas	Wisconsin
	(Can. Min. 1976) Scott (unpublished)*	(This study) Scott (refined)	(This study)	(This study)
a	10.100 (5) A	10.098 (5) A	10.104 (9) A	10.13 (2) A
b	16.014 (7)	16.007 (6)	16.015 (13)	16.05 (3)
c	7.377 (5)	7.376 (4)	7.323 (6)	7.38 (1)
	96°25′ (3)	96°27′ (3)	96°35′ (4)	96°38′ (10)
	111°09′ (3)	111°10′ (3)	111°10′ (5)	111°11′ (11)
	76°34′ (3)	76°32′ (3)	76°16′ (5)	76°22′ (12)
	1081.88	1080.7 (7) A^3	1073. (1)	1086. (2)

[1] D. E. Appleman has kindly computed unit cell data for the three Northern American miserites; his data agree closely with those of Scott for the Quebec miserite, and all three give indeed very similar unit cells.

* Limits in parentheses and cell volume from L. G. Berry (Personal comm. June 23, 1982).

These major elements form a complete, self-consistent framework of slabs and pipes . . . the miserite structure, in effect, functions as a kind of atomic sieve, accepting and, by the different natures of the assorted sites, potentially sorting the minor elements which percolate through it."

Further Scott (1976) notes that the miserite structure is the first known representative of a Zoltai Type 5 structure. Zoltai (1960) proposed a classification of minerals with five types of tetrahedral structures, basically built up from a unit consisting of four oxygen ions tetrahedrally enclosing a central cation, Si, Al, B, etc. Such structures in nature were known for only four recognized types.

(1) Isolated groups of tetrahedra such as olivine, garnet, sphene, apatite, gypsum, epidote

(2) One-dimensionally non-terminated (chains) such as pyroxenes, amphiboles, wollastonite, tourmaline

(3) Two-dimensionally non-terminated (sheets) such as apophyllite, micas, chlorite, datolite

(4) Three-dimensionally non-terminated (frameworks) such as scapolite, sodalite, analcite, feldspars, leucite, beryl, natrolite, quartz, nepheline, sphalerite, fluorite

and

TABLE 5. — X–RAY POWDER PATTERNS, 5 MISERITE LOCALITIES, USSR (SIBERIA)

Aldan Massiv (Intrusive) east of Lake Baikal				Alay Range Kirgiz SSR				Talassk Range		Olekma-Vitim			
										Southern Yakutia		Highlands	
(Kupriyanova and Vasil'eva 1961)		(Rudnitskaya, 1970)		(Ruzhova and Moleva, 1960)		(Ulyanova and Ilinskii, 1964)		(Kozlova, 1962)		(Kravchenko and Bukova, 1967)		(Arkhangels'kaya, 1968)	
I	d	I	d	I	d	I	d	I	d	I	d	I	d
—	—	9	15.4	—	—			—	—	—	—	—	—
—	—	—	—	—	—			—	—	—	—	—	—
—	—	—	—	—	—			—	—	—	—	—	—
—	—	—	—	—	—			—	—	—	—	—	—
5	5.068	3	5.07	1	5.09			—	—	—	—	—	—
5	4.651	3	4.65	1	4.57			1	4.59	3	4.65	—	—
—	—	—	—	—	—			—	—	—	—	—	—
—	—	—	—	1	4.01			1	4.19	—	—	—	—
3	3.811	4	3.84	4	3.81			2	3.87	3	3.90	—	—
—	—	—	—	—	—			2	3.66			—	—
2	3.618	—	—	—	—	—	—	—	—	—	—	—	—
4	3.446	5	3.44	—	—	—	—	—	—	5	3.50	4	(3.47)
8	3.307	—	—	5	3.31	—	—	10	3.33	6	3.34	10 w	3.13
10	3.112	10	3.13	10	3.10	10	3.13	10	3.14	10	3.13		
4	3.032	6	3.06	—	—			5	3.01	—	—		
7	2.911	8	2.94	—	—	5	2.93	8	2.91	9	2.92	5	2.93
—	—	—	—	8	2.87	—	—	—	—	—	—	—	—
3	2.814	—	—	—	—	5	2.788	—	—	—	—	6	2.80
3	2.761	8	2.77	9	2.73	—	—	5	2.77	6	2.78	—	—
5	2.662	8	2.65	8	2.63	7	2.656	5	2.66	9	2.65	6	2.68
2	2.485	—	—	1	2.45			3	2.44	2	2.47	1	2.49
4	2.377	—	—	—	—			—	—	6	2.36	3	2.38
—	—	6	2.34	4	2.33			3	2.34	—	—	—	—
2	2.280	4	2.29	4	2.27			4	2.28	5	2.28	1	(2.28)
—	—	—	—	1	2.15			—	—	—	—	—	—
5	2.087	2	2.08	5	2.08	6	2.097	4	2.09	6	2.09	5	2.09
5	2.964	—	—	2	1.967	—	—	3	1.973	6	1.95	4	1.955
4	1.922	—	—	2	1.916	—	—	4	1.919	5	1.91	—	—
—	—	—	—	—	—	—	—	1	1.872	—	—	—	—
6	1.835	—	—	5	1.827	—	—	4	1.815	6	1.82	5	1.843
—	—	3	1.752	1	1.754	—	—	—	—	3	1.75	—	—
—	—	—	—	1	1.699	—	—	—	—	—	—	—	—
9	1.661	2	1.673	8	1.657	9	1.680	10	1.667	9	1.66	6	1.660
—	—	—	—	1	1.635	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—
5	1.578	7	1.588	1	1.574			3	1.581	6	1.57	3 w	1.580
4	1.557	—	—	1	1.562			—	—	—	—	—	—
3	1.537	—	—	—	—			3	1.542	—	—	—	—
—	—	6	1.520	—	—			1	1.522	—	—	2	1.526
—	—	—	—	1	1.514			—	—	2	1.514	—	—
—	—	—	—	—	—			1	1.456	—	—	—	—
—	—	8	1.440	1	1.440			1	1.442	3	1.443	—	—
5	1.415	2	1.413	4	1.413			5	1.415	3	1.408	2	1.412
5	1.367	—	—	1	1.382			7	1.371	3	1.374	—	—
—	—	—	—	1	1.351			—	—	—	—	2	1.347
5	1.320	—	—	2	1.318			3	1.321	3	1.317		
4	1.281	3	1.293	2	1.287			4	1.289	3	1.289		
—	—	—	—	1	1.262			—	—	—	—		
3	1.250	4	1.253	3	1.250			7	1.256	3	1.253		
—	—	—	—	—	—			1	1.230	—	—		
2	1.211	6	1.212	1	1.213			1	1.216	—	—		
6	1.197	—	—	3	1.200			7	1.199	3	1.199		
1	1.180	—	—	—	—			6	1.182	—	—		
1	1.166	—	—	2	1.164			—	—	2	1.166		
1	1.153	1	1.152	—	—			7	1.152	—	—		
—	—			1	1.123			—	—	—	—		
—	—			1	1.114			7	1.114	—	—		
3	1.103			7	1.098			3	1.102	1	1.102		
—	—			4	1.090			3	1.091	—	—		
—	—			1	1.083			7	1.082	—	—		
—	—			1	1.053			—	—	—	—		
2	1.036			6	1.043			7	1.044	1	1.043		
—	—			—	—			6	1.034	—	—		
—	—			5	1.010			4	1.011	—	—		
2	1.006				1.005			4	1.005	2	1.004		
										2	0.991		
										2	0.980		
										2	0.916		
										1	0.877		
										2	0.862		

Table 6.—Analyses of miserites from Arkansas (4), Wisconsin (1) and Quebec (1).

	Arkansas					Wisconsin	Quebec
	A	B	C		D	E	F
	Williams	Schaller	Milton		Medaris, et al.*	Medaris, et al. **	Scott ***
	(1891)	(1950)	(1974)		(1978)	(1978)	(1976)
SiO_2	50.96	49.25	49.2^{2}		50.3	48.4	50.18
Nb_2O_5	—	—	3.3^{3}		—	—	—
Al_2O_3	—	0.81	—		—	—	0.60
Fe_2O_3	—	0.42	0.25	(total Fe)^{3,4}	—	—	—
FeO	1.69	0.06	—		(total Fe) 0.29	0.31	0.14
TiO_2	—	0.43	$0.55^{3,5}$	Be 5 ppm	—	0.17	—
RE_2O_3	—	—	2.3	Bi 150 "	n.d	n.d	7.47
MgO	0.57	—	—	Cr 30 "	—	—	0.37
MnO	1.40	0.28	$0.54^{3,6}$	Cu 200 "	0.44	1.68	0.45
CaO	36.72	36.53	$33.8^{3,6}$	Pb 50 "	33.5	30.6	31.00
SrO	—	—	$0.68^{3,6}$	Sc 50 "	—	0.40	—
BaO	—	—	—	Sn 200 "	—	—	—
Na_2O	4.41^{1}	0.96	0.11	V 300 "	0.02	0.25	0.81
K_2O	0.90^{1}	6.09	$6.1^{3,6}$	Zr 70 "	7.03	6.55	5.58
H_2O	2.74 (ign.)	5.17	1.5^{7}	not detected	—	—	0.55
CO_2	—	—	1.0^{7}	Ag Au Cd Co Ge Hf	—	—	—
F	—	—	2.1^{8}	In Li Ni Pd Pt Re	—	3.77	2.04
Cl	—	—	—	Sb Ta Te Th U W	—	0.05	—
B_2O_3	—	—	0.02^{9}	Zn	—	—	—
Li_2O	—	0.02	—		—	—	—
P_2O_5	—	—	—		0.31	0.20	—
	99.39	100.02	100.42		91.9	92.2	99.19
		$-F_2=0$	0.9				$-F=0$ .87
			99.52				98.32
		Y_2O_3	0.94				3.34
		La_2O_3	0.43				0.59
		Ce_2O_3	0.43				(CeO_2) 1.00
		Pr_2O_3	—				(Pr_6O_{11}) 0.21
		Nd_2O_3	0.20				0.86
		Eu_2O_3	0.02				(EuO) 0.04
		Dy_2O_3	—				0.59
		Er_2O_3	—				0.31
		Tm_2O_3	0.02				0.03
		Yb_2O_3	0.08				0.47
		Lu_2O_3	—				0.04
		Ho_2O_3	0.03				—
		Sm_2O_3	0.14				—
		Gd_2O_3	—				—
		Tb_2O_3	—				—
			2.29				7.47
Density	2.84	2.89^{10}			—	—	2.926

* By electron probe (average of three analyses).

** By electron probe (average of four analyses).

*** This analysis supersedes earlier analyses by Berry, Lin, and Davis (1971). F=0 correction supplied by Charles Milton.

TABLE 6 NOTES

A. "Pink wollastonite," "natroxonotlite" (Williams, 1891, p. 457. Potash Sulfur Springs, Arkansas, Analyst, R. N. Brackett.

[1] As published; presumably a typographic error. As noted by Schaller, 1950, p. 914, the values for Na_20 and K_20 should be interchanged. The formula assigned was $5(\frac{9}{10}Ca + \frac{1}{10}Na_2)\ Si0_3 + aq$.

B. Miserite (Schaller, 1950, p. 916). Analyst, W. T. Schaller. Potash Sulfur Springs, Arkansas. The formula assigned was $KCa_4Si_50_{13}(0H)_3$. This analysis omitted fluorine and carbon dioxide, also rare earths (probably weighed as Ca0) and niobium oxide.

The analysis B (Schaller, 1950) is the "average of two closely agreeing recalculated analyses of miserite," one "by fusion with Na_2C0_3," the other "by acid decomposition." The former is the analysis cited by Semenov et al., 1973.

C. Miserite Potash Sulfur Springs, Arkansas (this report). B. Ingram and J. L. Harris, U. S. Geological Survey, Reston, Va., analysts.

[2] Determined spectrophotometrically as molybdenum blue after decomposition of 5-mg samples by fusion with Na0H.

[3] Approximately 100 mg of sample was decomposed with HF - H_2S0_4 and the solution was diluted to a definite volume. The Nb in the miserite sample was hydrolyzed and the solution filtered prior to dilution. Aliquots of these solutions were used to determine Ti, Fe, total R_20_3, Mn, Ca, Sr, Na, and K in each sample where applicable. The hydrolyzed Nb from the miserite sample was ignited and weighed as Nb_20_5.

[4] Determined spectrophotometrically, Fe with o-phenanthroline and Ti with Tiron.

[5] Total R_20_3 was determined gravimetrically after making a double precipitation with NH_40H. Total RE_20_3 was calculated by difference from total R_20_3 by subtracting $Ti0_2$ and Fe_20_3.

[6] Atomic absorption methods were used for these determinations.

[7] A 100-mg sample was mixed with V_20_5 and ignited in a tube furnace at 900°C in an atmosphere of 0_2. The H_20 and $C0_2$ released were absorbed and weighed.

[8] A 50-mg sample was fused with Na_2C0_3–Zn0 and the fusion mixture leached with H_20. F in the leachate was determined without separation using a fluoride specific ion electrode.

[9] This and the rare earth trace element determinations were spectroscopic.

[10] Determined by John Marinenko (U. S. Geological Survey) using a pycnometer and tetrachlorethylene as the displacement liquid.

The Arkansas miserite analysis computed to a formula $(K,Na)_{.778}\ (Ca, Sr, Mn, Fe^{+2}, R. E.)_{3.701}\ (Si,Nb,Ti)_{5.001}\ 0_{13,000}(0H,0,C0_2,F)_{2.963}$ which compares with Schaller's analysis, namely $(K,Na)_{.855}(Ca, Mn, Fe^{+2})_{3.531}(Si,Al,Fe^{+3},Ti)_{4.549}0_{13.000}(0H)_{3.094}$.

D. Miserite, Potash Sulfur Springs, Arkansas, Gordon Medaris, analyst.
Microprobe, average of 3 analyses.

E. Miserite, Wausau, Wisconsin, Gordon Medaris, analyst. Microprobe, average of 4 analyses.

F. Miserite, Temiscamingue County, Quebec

Two analyses, respectively "wet chemical" and "microprobe", are given in Berry, L. G., Lin, Hsi-Che, and Davis, G. C., 1972. Card 22–808 XRPD, gives "average of two analyses (in atoms per cell)." The formula is given as $(Ca,K,Na,Al,Y\ etc.)Si0_3$. These are superseded by a third analysis (Scott, 1976) cited in Table 6; a combination of "microprobe and wet analyses" gave an average composition of $Y_{0.56}\ R.E._{0.45}\ Na_{0.50}\ Mg_{0.18}Mn0_{0.12}\ Al_{0.22}\ K_{2.26}\ Ca_{10.54}\ Si_{15.92}\ 0_{44}\ (0H)_2\ F_2 \cdot 0.5\ H_20$ simplified (Scott, personal communication, 1976) to the "ideal formula" KCa $(Si_20_7)(Si_60_{15})(0H)F$ where denotes an "empty site" that may be partially filled by yttrium and other minor constituents.

(5) mixed (or which miserite is the first and only known representative).

Essentially, the Type 5 Zoltai structure shown in miserite exhibits both Type 2 (chains) and Type 1 (isolated groups) linked together to form the composite (mixed) Type 5 miserite structure.

In general, this description with but little modification is applicable to the Arkansas and other miserites, allowing for very minor differences in cell size and ionic substitutions.

Ghose and Wan (1979) described agrellite $Na(Ca,RE)_2Si_4O_{10}F$ as a "layer structure with silicate tubes," first recognized in 1971 in the mineral fenaksite $FeNaKSi_4O_{10}$. Ghose and Wan find such a structure in narsarsukite $Na_2(Ti,Fe^{+3})Si_4(O,F)_{11}$ and three other rare silicate minerals, besides in miserite. However, L. G. Berry (personal communication, June 23, 1982) observes that "it seems certain that agrellite does not represent a Zoltai type 5 structure, so I believe that Scott's statement is still true . . . ."

Agrellite, it may be noted, occurs with miserite in Quebec and Wisconsin, but has not been observed in Arkansas.

CHEMISTRY

Altogether, there are four analyses of the Arkansas miserite, one of the Wisconsin miserite, thee of the Quebec miserite and six from Siberia, USSR. These analyses are shown in Tables 6 and 7 which are followed by Notes on the analyses.

The Arkansas miserite is relatively high in Nb: 3.3 percent Nb_2O_5; others less than 1 percent. The rare earth content is relatively low, 2.3 percent, one other as low as 0.48 percent, others up to a reported 9.09 percent. Where rare earths are high (Talas Mt., USSR and Quebec) yttrian earths exceed the lanthanides. Otherwise, miserite from its various localities shows but minor variation in composition from locality to locality, by ionic substitutions; and accordingly for each locality, the unit cell dimensions and corresponding powder pattern, and physical properties, may vary slightly.

USSR occurrences of miserite

As shown on Figure 1, there are five reported localities for miserite in the Soviet Union, all in Siberia. Mrs. Janina Block (personal comm.) has kindly read the original accounts, and the map locations are hers.

Scott (1976, p. 524–5) cautions that the Russian analyses for OH and F are probably unreliable, because the conventional thermogravimetric methods of analysis employed would not distinguish between these expelled anions. For the Quebec miserite,an "improved" method was applied; and for Arkansas, the analysis of Ingram and Harris is probably as close to the actual truth as current analytical skill can produce.

WOLLASTONITE

The white prismatic wollastonite intergrown with miserite at Potash Sulfur Springs, Arkansas, was first analyzed by Brackett (Williams, 1891); and again recently by Ingram and Harris. Both analyses are shown in Table 9. An X-ray powder pattern is given in Table 10 with the pattern of pure $CaSiO_3$ (synthetic) for comparison; also, the pattern of a USSR wollastonite associated with miserite (Rudnitskaya, 1970). She does not give a complete chemical analysis of this wollastonite, but spectrographic determinations show Mn, Na o.n percent, Mg, Al, Cu, Sr, Ba, La, Ce o.on percent, Ti, Zr, Y Yb o.oon percent, and Ag o.ooon percent. Presumably the minor differences in the respective X-ray powder patterns reflect these differences

TABLE 7. –– ODDO-HARKINS RULE – RARE EARTH ABUNDANCE

It may be noted that in the Arkansas and Quebec miserites the rare earth abundances conform to the Oddo-Harkins Rule––that even atomic numbered lanthanides generally occur more abundantly with their odd numbered immediate neighbors. Thus in the Arkansas and Quebec miserites, we have:

Oxide	Element At No.	Wt. percent			
		Arkansas	Quebec		
La_2O_3	57	0.94	3.34		
		0.43	0.59		
Ce_2O_3	58	0.43	0.95	Ce_2O_3	$(1.00\ CeO_2)$
Pr_2O_3	59	––	0.20	Pr_2O_3	$(0.21\ Pr_6O_{11})$
Nd_2O_3	60	0.20	0.86		
Pm_2O_3	61	––	––		
Sm_2O_3	62	0.14	––		
Eu_2O_3	63	0.02	0.04	Eu_2O_3	(0.04 EuO)
Dy_2O_3	66	––	0.59		
Ho_2O_3	67	0.03*	––		
Er_2O_3	68	––	0.31		
Tm_2O_3	69	0.02	0.03		
Yb_2O_3	70	0.08	0.47		
Lu_2O_3	71	––	0.04		

*anomalous – error?

in minor element composition. The Arkansas wollastonite is $Ca_{.946}Mn_{.016}Fe_{.004}Sr_{.004}Na_{.021}K_{.003}Si_{1.004}O_{3.000}$.

Both in the Kipawa, Quebec, carbonatite (Scott, 1976) and at Wausau, Wisconsin (Stephen Guggenheim, personal communication, 1976) the massive miserite occurs with agrellite, a complex silicate ideally $NaCa_2Si_4O_{10}F$ (Gittins, Brown, and Sturman, 1977), whereas in the Arkansas, and probably all the reported Siberian localities, it is fibrous and associated with wollastonite, $CaSiO_3$.

Ulyanova and Iliinskii (1964) point out the chemical relationship of wollastonite, pectolite, and miserite thus

wollastonite	Ca_3	$[Si_3O_9]$
pectolite	$NaCa_2$	$[Si_3O_8OH]$
miserite	KCa_2	$[Si_3O_8OH]$

Acknowledgments

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TABLE 8. —Chemical analyses, five Siberian localities.

	Aldan Massiv east of Lake Baikal				Alay Range, Kirgiz USSR			Talassk Range			Yakutia	Turkestan	
	A		B	C	D			E			F	G	
	Kupriyanova and Vasileva (1961)	Kupriyanova (1961)*	Rudnitskaya (1970)	Demchenko (1973)	Ruzhov and Moleva (1960) Ulyanova and Illinskii (1964)			Kozlova (1962)			Kravchenko and Bykova (1967)	Semenov, Duzmatov and Khomyakov (1973)	
SiO_2	48.39	49.26	48.52	49.30	49.20			47.82			51.00	49.05	
Nb_2O_5	—		—	—	(o.n)			(o.n)			—	0.69	
Al_2O_3	0.82	0.82	0.57	1.02	0.20			1.32	(includes $Nb_2O_5 + ZrO_2$)		2.40	0.66	
Fe_2O_3	0.48	0.48	—	0.40	1.00^2			—			0.88	0.40	
FeO	—	—	—	—	—			—			—	—	
TiO_2	—	—	0.78	—	—			—			0.82	—	
RE_2O_3	7.52	7.52	1.15	6.48	—	Nb La Zr Y Mg	o.n	4.00	Nb La Zr	o.n	0.48	5.42	
MgO	0.58	0.58	—	1.60	(o.n)	Sr Pb	o.on	0.35	Sr Ti Sn Be Pb	o.on	0.35	2.09	
MnO	0.16	0.16	tr	1.72	1.33	Sn	o.oon	1.38	Ga Mo Bi	o.oon	0.57	0.20	
CaO	31.10	31.10	34.19	30.24	36.22	Be Cu Tl	o.ooon	33.50			31.14	33.30	
SrO	0.04	0.04	—	—	(o.on)			(o.on)			1.46	0.10	
BaO	—	—	—	—	0.20			—			0.32	—	
Na_2O	0.08	0.08	0.94	0.14	0.92			1.10			0.62	1.18	
K_2O	5.49	5.49	6.57	5.76	6.47			5.42			6.09	4.25	
H_2O	1.96	1.96	4.46	1.32	4.20			2.40			1.96	1.30	
CO_2	1.70	0.50	—	0.85	—			—			—	—	
F	2.50	2.50	4.07	2.40	0.90			4.32			3.00	2.05	
B_2O_3	0.10	0.10	—	—	—			—			—	—	
Li_2O	—	—	—	—	—			—			—	0.08	
	100.92	100.59	101.31	101.23	100.64			101.61			101.09	100.77	
$-F_2=O$	1.05	1.05	1.71	1.01	0.37			1.81			1.28	0.86	
	99.87	99.54	99.60	100.22	100.27			99.80			99.81	99.91	
Y_2O_3	2.99		0.02	1.57							0.01	1.44	Y_2O_3
La_2O_3	0.77		0.27	0.55							0.17	0.33	La_2O_3
Ce_2O_2	1.54		0.72	1.50							0.24	1.39	Ce_2O_3
Pr_2O_3	0.19		—	0.23							0.01	0.22	Pr_2O_3
Nd_2O_3	0.77		0.13	0.83							0.04	0.77	Nd_2O_3
Sm_2O_3	0.20		—	0.27							0.00	0.26	Sm_2O_3
Eu_2O_3	—		—	0.06							—	0.04	Eu_2O_3
Gd_2O_3	0.27		—	0.20							0.00	0.29	Gd_2O_3
Tb_2O_3	0.04		—	0.06							—	0.04	Tb_2O_3
Dy_2O_3	0.35		—	0.41							—	0.29	Dy_2O_3
Ho_2O_3	0.09		—	0.12							—	0.05	Ho_2O_3
Er_2O_3	0.27		0.00	0.33							—	0.15	Er_2O_3
Tm_2O_3	—		—	0.06							—	0.03	Tm_2O_3
Yb_2O_3	0.02		—	0.25							—	0.08	Yb_2O_3
Lu_2O_3	—		—	0.02							—	0.03	Lu_2O_3
	7.50		1.14	6.46							0.47	5.41	
D	2.899		2.82	2.914	2.85			2.84			2.783	2.9	

* As cited in Semenov, et al., 1973

TABLE 8 NOTES

A. Kupriyanova and Vasileva, (1961, p. 144); Aldan Massiv. Micheev (1967) is cited for the formula $HKCa_2Si_30_9$; they propose $(K_{0.86}Na_{0.13})(Ca_{3.68}Sr_{0.09}Mn_{0.05}Mg_{0.05}\,R.E._{0.04}\,Fe_{0.08})\,Si_{5.62}\,Al_{0.31}\,Ti_{0.07}\,(0_{15.00}\,F_{1.05})\cdot 0.72\,H_20$ essentially $KCa_4Si_60_{15}F\cdot nH_20$.

B. Rudnitskaya, L. S., (1970, p. 182); Aldan Massiv.
[1] The summation 101.31 appears to be incorrect; it should be 101.25, and likewise 99.60 (after oxygen correction for fluorine) should be 99.54. The formula proposed is $(K_{1.04}Na_{0.20})(Ca_{4.53}\;R.E._{0.04})(Si_{6.00}Ti_{0.07}Al_{0.07})0_{16}(0H_{1.40}F_{1.60})1.15\,H_20$.

C. Demchenko, V. S. (1973); Aldan Massiv.

D. Ruzhov and Moleva, (1960, p. 396).
Khodghaakchan Massiv, Alay Range, Kirgiz SSR. (39° 40'N 71°00'E)

[2] "due to mechanical admixture of iron oxide." The values in parentheses are spectrographically determined. The formula is given as: $K_{0.84}Na_{0.18}Ca_{4.07}Si_50_{13}(OH)_{2.85}F_{0.29}$ or $(K,Na)Ca_4Si_50_{13}(OH,F)_3$ corresponding to that proposed by Schaller, 1950.

E. Kozlova, (1962, p. 202); Talassk Mt. Range (42° 30' N, 70° 30' – 74° 00' E). The formula assigned is: $(K_{0.69}\,Na_{0.22})\,(Ca_{3.64}\,R.E._{0.34}Mn_{0.12}Mg_{0.05})(Si,Al)_5\,0_{13}\,(0H_{1.62}\;F_{1.38})$ or $(K, Na)\,(Ca, R.E., Mn, Mg)_4\,(Si,Al)_50_{13}(0H,F)_3$.

F. Kravchenko and Bykova, (1967, p. 165), Southern Yakutskia, Siberia (Yakokut 58°32'N 125°44'E). The proposed formula is $(K_{0.86}Na_{0.13})(Ca_{3.88}Sr_{0.09}\,Mn_{0.05}\,Mg_{0.05}\,R.E._{0.04}\,Fe_{0.08})\,(Si_{5.82}\,Al_{0.31}\,Ti_{0.07})\,(0_{15}F_{1.05})\cdot 0.72H_20$, or $KCa_4Si_60_{15}F\cdot nH_20$.

G. Semenov, Duzmatov, and Khomyakov, 1973. They suggest the formula $KCa_5\,Si_60_{17}\;F$ (Z = 3) or $KCa_4Si_60_{16}F$.

TABLE 9. —— 2 analyses of wollastonite from Arkansas

	(K) (Williams, 1891) R. N. Brackett, analyst		(this report) Blanche Ingram, analyst, U.S. Geol. Survey	Joseph L. Harris, U. S. G. S. (spectrographic) analyst		
SiO_2	51.93		51.4[2]	Si	10	percent
FeO	2.03		0.28 (total Fe)[3,4]	Fe	0.2	"
MgO	0.44		——	Mg	0.03	"
MnO	2.08		0.96[3,5]	Mn	20,000	ppm (2.0%)
CaO	42.55		45.2[3,5]	Ca	10	percent
ign	1.23	SrO	0.30[3,5]	Sr	5,000	ppm
		Na_2O	1.1[3,5]	Na	1	percent
	100.26	K_2O	0.24[3,5]	K	N.D.	
			99.5	Al	0.0015	"
		Sp.gr.	2.91[10]	Ti	0.015	"
				As[11]	N.D.	
				B	50	ppm
				Ba	100	"
				Be	3	"
				Pb	10	"
				Nb	700	"
				V	7	"
				Y	150	"
				Ce	100	"
				La	70	"
				Yb	15	"
				Cu	18	"

2, 3, 4, 5, 10 See corresponding notes

Analysis C (miserite, Arkansas)

Arsenic not detected. The 0.1% as reported previously in the Arkansas wollastonite (Milton, 1973) is an error.

TABLE 10. —— X–ray powder data, wollastonite from Arkansas, synthetic, and Siberia

Arkansas (Milton, this report)		Synthetic JCPDS 29–372			Siberia (Rudnitskaya, 1970)	
I	d	I	d	hk	I	d
s	7.724	50	7.68	100		
vvw	5.473	40	5.47	101		
vvw	4.966	——	——	——		
vvw	4.693	——	——	——		
——	——	20	4.48	111	——	——
vw	4.058	20	4.06	111	——	——
vs	3.850	60	3.838	200	8	3.76
vs	3.602	70	3.510	201	6	3.49
——		20	3.410	211	——	——
vvs	3.323	80	3.312	102	10	3.27
——	——	40	3.228	201	——	——
vvw	3.181	40	3.195	211	——	——
——	——	40	3.167	021	——	——
vvs	3.098	70	3.080	210	——	——
——	——	20	3.002	120	4	3.03
vs	2.989	100	2.976	220	10	2.94
w	2.919	20	2.893	112	——	——
vvw	2.806	20	2.804	121	——	——
——	——	40	2.790	221	——	——
s	2.730	20	2.750	211		
——	——	40	2.712	121,112+		
——	——	20	2.588	310		
		20	2.572	300		
s	2.562	50	2.553	122		
——	——	40	2.522	022	——	——
——	——	20	2.510	311	——	——
s	2.483	60	2.470	202	——	——
——	——	40	2.445	130	2	2.45
s	2.349	50	2.346	222	——	——
——	——	40	2.338	003	——	——
——	——	50	2.330	301	——	——
vs	2.304	60	2.295	103	3	2.31
——	——	20	2.274		5	2.28
——	——	40	2.209		——	——
s	2.183	70	2.179		——	——
vvw	2.093	40	2.161		4	2.14
vw	2.028	50	2.017		——	——
——	——	20	1.993		——	——
w	1.984	40	1.978		——	——
		——	——			
——	——	40	1.969			
——	——	20	1.933			
——	——	——	——			
w	1.922	40	1.914			
vw	1.886					
vw	1.860					
vw	1.834					
w	1.801					
s	1.760					
s	1.724				8	1.703
vvw	1.663				——	——
s	1.609				6	1.592
w	1.537				——	——
w	1.482				3	1.474
w	1.461				7	1.445
vw	1.392				——	——
s	1.364				7	1.351
vw	1.344				——	——
vvw	1.301				——	——
vw	1.282					
vvw	1.270					
vw	1.243					
w	1.216					
vvw	1.196					
w	1.176				6	1.168
vvw	1.151				——	——
vvw	1.143				——	——

114.6 mm camera, Cu/K , Ni filter

REFERENCES

(Anonymous) 1969, A Partial List of Minerals that Fluoresce Mineralight Lamps: Ultra-Violet Products, Inc., San Gabriel, California. Page 18.

(Anonymous) 1972, Miserite from Arkansas: a renaming of natroxonotlite: Rocks and Minerals 47, 14–15.

Arkangelskaya, V. V., 1968, Albitized rare metal granite found for the first time in the Olekma Vitim Highalnds:Doklady Akad. Nauk SSSR 183, p. 1166–1169. English translation – Am. Geol. Inst., 1969, p. 162–164.

Berry, L. G., Lin Hsi-Che, and Davis, G. C., 1972, A New Occurrence of Miserite from the Kipawa Lake Area, Temiscamingue County, Quebec: Can. Miner. V. 2, pt. 2, p. 569.

Cohee, George V., 1962, Hugh Dinsmore Miser (1884–1969), Bulletin Amer. Ass. Petroleum Geologist 53, No. 12, p. 2560–2562.

Demchenko, V. S., 1973, Rare earth miserite from metasomites: Vopr. Magmat. Metamorf. Orudeneniya Dal–Nego Vostoka Petrog. Reg. Soveshch., 2nd pub. 1972, p. 214–218, Chem. Abs. V. 81, No. 22 138536h (1974).

Ghose, Subrata and Che'ng, Wan, 1979, Agrellite $Na(CaRE)_2 Si_4 0_{10} F$: A layer structure with silicate tubes: American Mineralogist 64, p. 563–572.

Kemp, J. F., 1891, Memorial of John Francis Williams: Bull. Geol. Soc. Am. 3, p. 455–458.

Kemp, J. F., 1892, John Francis Williams: American Geologist IX, No. 3, p. 149–153.

Kozlova, P. S., 1962, Miserite from the Talassk Range: Trud. Mineral. Musei, Fersman Akad. Nauk, SSSR 13, p. 198–204.

Kravchenko, S. M., and Bukova, A. V., 1967, Southern Yakutia Miserite: Mineral. Pegmatitov Gideoterm. Shchelochnykh Massivov. Redk. Elem. Akad. Nauk, SSSR Inst. Min. Geokhim–Kristallokhim., p. 160–167.

————————, and Vlasova, E. V., 1962, Alkalic rocks of Central Aldan: Tv. IMGRE 196, No. (?) 14.

Kupriyanova, I. I., and Vasil'eva, V., 1961, On Rare-earths (in) Miserite: Geol. Mestorozhd. Redikh. Elementov (Geology of deposits of rare elements) 9, p. 139–147. English trans. (1963).

"Medaris et al., 1978" personal communication.

Micheev, V. I., 1957, Roentgenometric Mineral Data: State Geological Report (cited in Kupriyanova and Vasil'eva, 1961, p. 141).

Milton, Charles, 1973, New data on miserite (abstract) Program, South Central Section, Geol. Soc. Amer., Little Rock, Ark., p. 273.

Rudnitskaya, L. S., 1970, Miserite, pectolite and other minerals from the metasomatized exocontact zone of the Mesozoic alkalic Aldan intrusives: Minerl. Syre No. 22, p. 177–186; Chem. Abs. (1972) 76 (20) 115906.

Ruzhov, B. I. and Moleva, V. A., 1960, A discovery of Miserite in the U.S.S.R.: Dokl. Akad. Nauk, SSSR 131, p. 1420–1422. (Same locality as Ulyanova and Ilinskii, 1964).

Schaller, W. T., 1950, Miserite from Arkansas: A renaming of Natroxonotlite: Am. Min. 35, p. 911–921.

Scott, J. Douglas, 1976, Crystal Structure of Miserite, a Zoltai Type 5 Structure: Canadian Mineralogist, 14, p. 515–528.

Semenov, E. I., Duzmatov, V. D. and Khomyskov, A. P., 1973: Rare earth miserite: Mineral. Issled. 3, 42–45 (Chem. Abstr. 1350 in Geol. Chem. 83).

Ulyanova and Ilinskii, 1964, A New Discovery of Miserite from Khodzha Achkan, Alai Range, Kirgiz SSSR (Central Asia): Mineral Geokhim. Leningrad Gos. Univ. Sb. Statei 1, p. 40–46.

Williams, J. F., 1891, The igneous rocks of Arkansas: Ark. Geol. Sur. Annual Report 1890, 2: p. 1–391, 429–457. Maps, Little Rock, 1891.

————————, 1891 (with Kemp, J. F.), Tabulation of the dikes of igneous rocks of Arkansas: Ark. Geol. Sur. Annual Report 1890, 2: p. 407–427, Little Rock, 1891.

————————, 1890, Manganpectolith, ein neues Pectolith ahnlichen mineral Magnet Cove, Arkansas: Zs.f.Kr. 18, 386.

————————, 1890, Eudialite and Eucolite from Magnet Cove, Arkansas: Am. J. Sc. 40, 457.

Zoltai, Tibor, 1960, Classification of silicates and other minerals with tetrahedral structures: Amer. Mineral. 45, p. 960–973.

NORTH MOUNTAIN MINE – GOLD?

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ABSTRACT

The North Mountain Mine in Montgomery County, Arkansas has been mined intermittently for manganese since 1886. Recent interest in this property is for gold and silver. Assays done for a Forest Service, U. S. Department of Agriculture, validity examination of the property display gold and silver values characteristic of the background, relatively low values in the Ouachita Mountains.

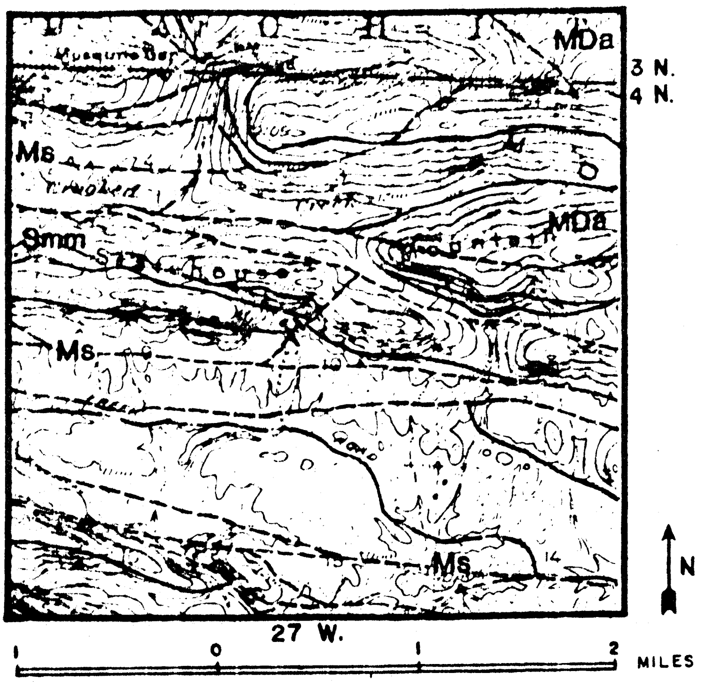
INTRODUCTION

The North Mountain Mine is in the Ouachita National Forest, Section 10, T. 4 S., R. 27 W., Montgomery County, Arkansas. Also known as the Stenger Mine, its mineral potential was vigorously promoted by Mrs. Mabel C. Stenger from about 1944 through the early 1960's. She and her husband, as well as some of their employees, lived in small residences just below the present adit. Remains of some of these residences are still in evidence.

Literature suggests that the earliest production from the mine was about four tons of manganese in 1886, with subsequent manganese production during World War I and for a brief period in 1958–1959 (Stroud, et al., 1969). In spite of Mrs. Stenger's promotion, any mineral production from the property other than manganese is undocumented. In recent years, the mine has become somewhat of a curiosity, a favorite for collectors and students. Additionally, groups and individuals have filed numerous, often conflicting mining claims on and around the property for copper and manganese, (which are known to occur), and gold or silver (whose occurrence is questionable). "Arkansas Geological Survey, County Mineral Report 3" lists the results of assays performed by W. F. Manglesdorf on samples from the North Mountain Mine. No silver values were detected; gold values ranged from none to 0.4 oz./ton, and copper values ranged from 0.02% to 9.5%. More recent assays on samples from this site by the Arkansas Geological Commission show no extraordinary gold or silver values, and none as high as the 0.4 oz./ton noted in the Manglesdorf assays.

VALIDITY EXAMINATION

Recent prospecting by Callahan, Exxon Minerals, Texasgulf-Western, and other companies in the Garland, Montgomery, Polk County area has stimulated a renewed public interest in the possibility of gold and other precious elements in the Ouachita Mountains. The presence of mining companies and activities of various promoters, coupled with gold market values often in excess of $400 per troy ounce, have piqued public awareness, even to the extent that an article on gold in the Ouachitas was published in the June 12, 1983 edition of the "Arkansas Magazine."



EXPLANATION

Ms	STANLEY SHALE	———	CONTACT LINE
MDa	NOVACULITE	— — —	FAULT LINE
Smm	MISSOURI MOUNTAIN SHALE	X	NORTH MOUNTAIN MINE

GEOLOGIC MAP OF NORTH MOUNTAIN MINE AREA

The North Mountain mine is one property purported by local miners to contain "high" gold and silver values. Since this property is on National Forest lands, it is subject to examination and evaluation for the purpose of mining claim validity determination. Forest Service geologists conducted a validity examination on May 24–26, 1983.

Sampling methods consisted of taking bulk samples from vertical trenches cut at three locations in the spoil pile at the mouth of the adit, and two bulk samples on a 15 foot square grid. About one-half pound of sample per sample point of the grid was taken across the surface of the spoil pile. American Interplex Corp., Little Rock, Arkansas, using standard sample preparation, atomic absorption, and induction furnace techniques assayed the two grid samples and two of the trench samples. Table 1 presents the results.

Table 1 – American Interplex Results
(ppm = mg./kg.)

	Grid 1	Grid 2	Trench 1	Trench 2
Gold, ppm	0.05	0.05	0.05	0.05
Silver, ppm	0.73	2.0	1.2	0.52
Lead, ppm	68	32	32	34
Zinc, ppm	120	73	100	78
Manganese,%	0.54	0.07	0.33	0.43
Total Sulfur,%	0.18	0.42	0.37	0.11

(Samples contain only a trace of sulfur in sulfide form.)

Lindroos Laboratories,Tombstone, Arizona assayed the third trench sample. Fifty pounds of sample were barrel-tested for cyanide amenability with controlled pH and controlled cyanide for 72 hours. Head and tail assays were done by both chemical and fire methods (Table 2).

Table 2 - Lindroos Results

	Head	Tail
Silver, oz./T.	.04	.02
Gold, oz./T.	.022	.018

The barrel-test indicates that approximately 50.0% of the silver and 18.2% of the gold are recoverable by cyanide leaching, or a poor amenability. Lindroos attributes this, in part, to the carbonaceous nature of the sample. The assays in Table 1 show gold values less than .0015 oz./T. and silver values ranging between .015 oz./T. and .058 oz./T. There is a favorable comparison between silver results from two different firms, using different analytical methods. Gold values differ by about one order of magnitude, not a significant departure considering methods and the extremely low gold content.

CONCLUSIONS

Assay results from five bulk samples of the spoil pile at the North Mountain Mine do not show any anomalous values for gold or silver. Additionally, amenability of the ore is so low that even a richer ore would have limited profit potential due to low recoverability. These results fall within the range of gold and silver values reported by Comstock (1888) and others at various locations throughout the Ouachita Mountains. The results are unremarkable.

REFERENCES

Arkansas Geological Commission, Bell, Carney and Cross, 1974, Record of Mineral Analysis (unpublished), 3 p.

Arkansas Geological Survey, 1942, County Mineral Report 3, Mineral Resources of Montgomery, Garland, Saline, and Pulaski Counties, p. 28–30.

Comstock, T. B., 1888, Report upon the Geology of Western Central Arkansas with special reference to gold and silver: Ark. Geol. Survey, V. 1, pt. 2, 320 p.

Fellone, Frank, 1983, Arkansas – Gold Rush of the '80's?, in Arkansas Magazine, June 12, 1983, 1 p.

Stroud, Raymond B. et al., 1969, U. S. Bureau of Mines Bulletin 645, Mineral Resources and Industries of Arkansas, p. 297–301.

U.S.D.A. Forest Service, 1983, Report on Validity Examination – Copper Lume Properties, 9 p.

ARKANSAS NOVACULITE: INDIANS, WHETSTONES, PLASTICS AND BEYOND

By

Charles T. Steuart[1], Drew F. Holbrook[2], and Charles G. Stone[3]

ABSTRACT

The Arkansas Novaculite of Devonian and Mississippian age is the most distinctive formation in the central Ouachita Mountains of Arkansas from the standpoint of both topography and lithology. The Formation typically forms precipitous, narrow ridges (hogbacks) which reflect the steepness of dip and the resistance and thickness of the rock units. The Arkansas Novaculite varies from less than 200 feet thick in the north to locally over 800 feet thick in the south and crops out over an east-west distance of 360 km (200 miles) in the Ouachita Mountains of Arkansas and Oklahoma. Most of the rocks of the Formation are highly siliceous with novaculite predominating, and lesser quantities of chert, siliceous shale, conglomerate and rarely sandstone. Novaculite is commonly defined as a homogeneous mostly white rock, translucent on thin edges with a dull to waxy luster and composed almost entirely of microcrystalline quartz. The word novaculite comes from the Latin word novacula, meaning razor hone. This term was first used by Richard Kirwan in his mineralogy text of 1784.

The origin of novaculite remains controversial but we believe it was a primary silica derived mostly from the alteration of submarine volcanic-rich materials, with numerous fine biogenic components and some minor clastics (derived mostly from foreland facies via turbidity currents) and deposited as an amorphous siliceous ooze in the deep Ouachita trough. It was subsequently converted to microcrystalline quartz during diagenesis by the compaction of the overlying rocks and the intense Late Paleozoic Ouachita Mountain tectonism.

In the central and southern portions of the region three divisions of the Formation have been defined: A Lower Division predominantly of light-colored novaculite; a Middle Division of interbedded dark chert and shale; and an Upper Division of massive often calcareous novaculite. In the northern exposures the Formation is thinner and contains more black chert and shale, chert-shale conglomerate, sandstone and less novaculite. The massive novaculite of the Upper and Lower Divisions is a source of high purity silica (99+%).

Indians were the first to use novaculite. They made their various tools and hunting implements from the weathered broken rock until they learned the quarrying techniques. The first white men in the area recognized the use of novaculite for sharpening their tools and knives; previously whetstones were obtained in small quantities from Europe. To this day whetstone is quarried mostly in areas which the Indians and possibly the Spanish previously worked. Some of the current mining practices have changed very little from those of the early quarryman.

In areas where the novaculite of the Upper Division is poorly cemented, this friable tripolitic material has found applications in such uses as an abrasive, as a filler or extender in paints, plastics and other purposes. Some of the dense novaculite is suitable as a filler in fire bricks and also in plastics. Because some of it has the tendency to decrepitate when heated not all novaculite can be used for these applications.

Several quarry operators are crushing novaculite for road, concrete, and other aggregate. This use was formerly rather limited due to the abrasiveness of the novaculite on crushing and screening equipment. Further potential uses of novaculite are speculative but include application of high purity silica in the production of silicon metal for solar cells, seed or "growing", and fusing quartz for electronics.

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INTRODUCTION

For many years siliceous rock in the vicinity of Hot Springs, Arkansas has been used as whetstones. In his book on the geology and physical features in Missouri, Arkansas and surrounding areas, Schoolcraft (1819, p. 183) named the rock novaculite, a term that had been in use for sometime to describe a fine quality of whetstone, In 1892 Griswold published a comprehensive report on the very fine-grained siliceous rocks of the Ouachita Mountain region in Arkansas entitled *Whetstones and the Novaculites of Arkansas.* He mapped these rocks and called them the Arkansas Novaculite, but he included what is now known as the Bigfork Chert in the unit. It was Purdue in 1909 that named the group of siliceous rocks lying between the Stanley Shale and the Missouri Mountain Shale as the Arkansas Novaculite. Other workers identified the Arkansas Novaculite in McCurtain County, Black Knob Ridge, and Potato Hills in southeast Oklahoma. The Arkansas Novaculite is comprised largely of novaculite, lesser amounts of shale and conglomerate, and a few thin sandstone beds. This report is concerned particularly with those deposits of novaculite that are the source of current tripoli, whetstone and rock aggregate production and that are a potential source of these and other industrial mineral products.

DEFINITION OF NOVACULITE

When the term novaculite was first applied to the rock found in the vicinity of Hot Springs, Arkansas, it referred to a white siliceous rock whose textural characteristics and hardness made it particularly suitable for whetstones. Griswold (1892, p. 89) in fact described the fine-grained whetstone known as Arkansas Stone as a "true novaculite satisfying all the necessary conditions regarding homogeneity, grittiness, finely granular structure and siliceous composition, it is translucent on the edges and has a marked conchoidal fracture." As mapping of these remarkable rocks was extended throughout the Ouachita Mountains, the term novaculite was retained to describe them despite the fact that they showed some variation in both physical and chemical characteristics from place to place. Keller et al. (1977, p. 843) proposed that the use of the word novaculite be restricted to a petrologic name for a thermally metamorphosed chert which shows polygonal triple-point texture.

To accommodate these variations, novaculite is here defined as a homogeneous, mostly white or light-colored rock, translucent on thin edges, with a waxy to dull luster, and almost entirely comprised of microcrystalline quartz.

ORIGIN OF NOVACULITE

The subject of the origin of the novaculite in the Arkansas Novaculite has intrigued geologists ever since the rock was first identified; not only because of its unusual texture, but also because it is essentially pure silica that occurs in massive layers over a very broad area. At the time that Park and Croneis (1969) published their version of the origin, thirteen alternatives had already been proposed. They grouped those theories into three categories; direct precipitation of silica by chemical and/or organic means; replacement of non-siliceous deposits; and recrystallization and silification of volcanic-rich material. As a result of their studies of both the Arkansas Novaculite and the Caballos Novaculite of Texas, they proposed that novaculite was formed from organically precipitated silica particles in the 1 to 10 micrometer range, and that the ultimate texture was the result of diagenetic processes. Lowe (1975) has agreed that siliceous organisms were the local source of the silica in the novaculite, but proposed further that volcanism away from the immediate sites of accumulation may have provided much of the primary silica. Most investigators, including McBride and Sholes (1975), have considered the Arkansas Novaculite as a deep-water marine deposit, whereas Folk (1973) and a few others give evidence that some, if not all of the Formation, represents a shallow marine (mostly peritidal) deposit. We are strongly in favor of a deep-water marine origin for the respective Arkansas Novaculite facies that crop out in Arkansas and Oklahoma.

DESCRIPTION OF ARKANSAS NOVACULITE

The Arkansas Novaculite crops out along the borders of the Benton-Broken Bow Uplift in Arkansas and southeastern Oklahoma and at Potato Hills and Black Knob Ridge in southeastern Oklahoma, a distance of some 320 km (200 miles) (Fig. 1). The Formation varies in thickness throughout its outcrop area. It attains its maximum thickness, 950 feet, near West Hanna Mountain in Polk County, Arkansas. It thins rather gradually both eastward and westward from this point, but it thins very rapidly northward. Miser (1917) subdivided the Formation into three lithologic divisions based on the exposures at Caddo Gap, Arkansas; a Lower Division of almost entirely massive white novaculite, a Middle Division of interbedded dark chert and shale, and an Upper Division of massive mostly calcareous novaculite (Fig. 2).

Very few fossils have been found in the Arkansas Novaculite. Findings thus far have been limited mostly to conodonts, sponge spicules, radiolarians, and spores. Hass (1951), on the basis of conodont studies at Caddo Gap, placed the Lower Division of the Formation, and all except the upper 28 feet of the Middle Division in the Early or Middle Devonian. The remainder of the Middle Division and all of the Upper Division were defined as Early Mississippian in age. The contact between the Arkansas Novaculite and both the underlying Missouri Mountain Shale and the overlying Stanley Shale are conformable.

The Ouachita Mountains are a product of extensive folding and thrust faulting followed by erosion. This intense deformation resulted in the folding and fracturing of the novaculite beds as well as the variation in the attitude of these beds at different places. Along the southern belt of the exposures of the Formation, particularly, the massive novaculite of the Lower Division forms the steep-sided ridges of the region. The Middle Division cherts and shales underlie very narrow adjacent valleys, the thinner bedded less resistant novaculites of the Upper Division are frequently expressed as a series of low knobs paralleling the main ridges.

The relative proportions of each constituent (novaculite, chert, and shale) of the Arkansas Novaculite vary throughout the outcrop area. The thicker shale-free sections of novaculite, which are the particular concern of this report, are restricted to the Upper and Lower Divisions along the southern and central outcrop belt, specifically that segment of the outcrop belt within the area outlined by the row of dots on Figure 1.

Lower Division

The Lower Division of the Formation contains the thicker sections of novaculite that form the most prominent ridges and outcrops. This division in the southern outcrop belt is comprised of massive novaculite in beds 4 to 30 feet thick with the thicker beds being near the top. The novaculite is white, gray, or light brown, with black and reddish-brown beds in a few places. Near the base of the division the massive novaculite is often slightly calcareous, and it is the leaching of the calcite rhombohedra that developed the Washita variety of whetstone texture found in some of the novaculite near Hot Springs. Laminations are present throughout the Lower Division novaculite but they are particularly prominent near the base. Jointing is common, the most pre-

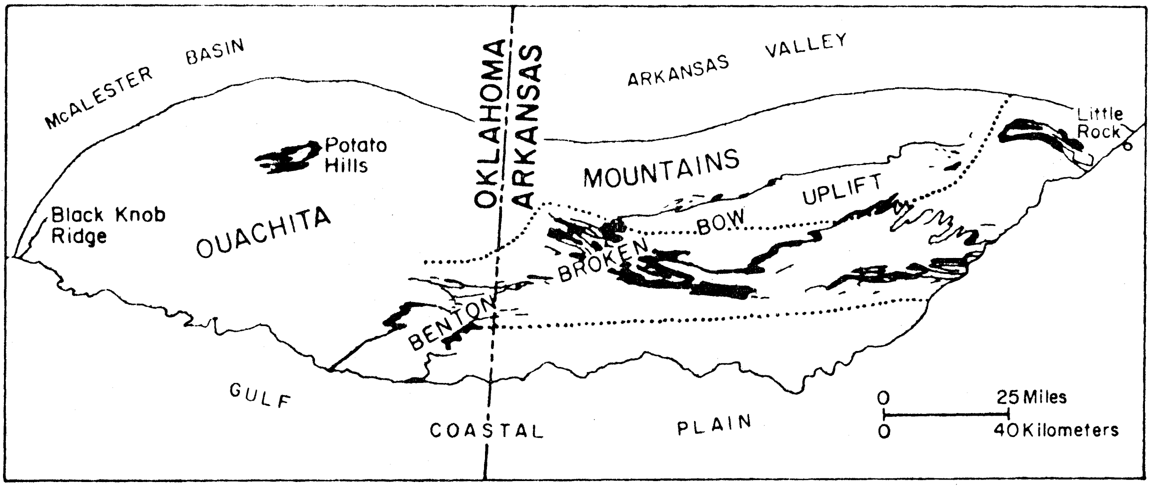

Figure 1. Map of Ouachita Mountains (stippled area) region showing outcrop of the Arkansas Novaculite (black lines) in Arkansas and Oklahoma. Thick, shale-free portions of the novaculite unit are within area outlined by dots.

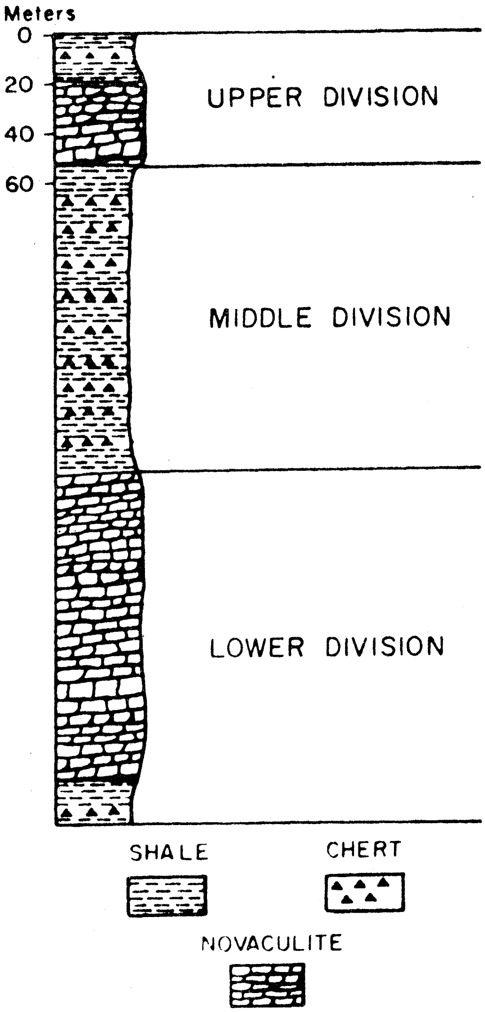

Figure 2. Diagrammatic columnar section of the Arkansas Novaculite (after Sholes and McBride, 1975).

valent set being normal to the bedding and joints are frequently filled with quartz. A few very widely spaced thin shale beds are present, and in places thin quartzitic sandstone beds occur near the base of the Lower Division. A submarine slurried conglomeratic interval is often present at the top of the Lower Division.

Upper Division

The Upper Division of the Formation is generally thinner than the Lower Division, attaining a maximum thickness of 120 feet at Hot Springs and thinning abruptly to the north and gradually to the south and west. This division is either absent or more likely represented by a more shaly and conglomerate facies in the northern outcrop belt except near Little Rock. Beds are even to irregular and up to four feet thick. It consists chiefly of massive novaculite that when fresh is light gray to bluish black and generally resembles the novaculite of the Lower Division. In much of the area, however, the Upper Division novaculite is calcareous, and it has weathered to a light-brown or buff-colored punky rock with a gritty texture giving it the appearance of a porous siltstone. Weathered novaculite varies from a firm, slightly porous rock to tripoli depending on the percentage of carbonate present in the original rock. The carbonate in the fresh novaculite, which can constitute up to 30 percent of the rock, may be rhodochrosite, calcite, manganiferous calcite, or ankerite. In fact, it is the manganese-bearing carbonates that are a prominent source of the manganese ores and associated minerals found in both the Upper and Lower Divisions of the Arkansas Novaculite in the southern outcrop belt.

NOVACULITE QUARRYING BY THE INDIANS

The first use of novaculite was by the various Indian tribes that quarried it in large quantities for making tools and weapons. The many thousands of tons of novaculite that were mined by the Indians came mostly from the southern Ouachita Mountains, but notably near Hot Springs, Arkansas, an area considered a neutral zone by the tribes because of its renowned healing waters. Much of the denser and harder novaculite was ideal for making tools because it typically broke into very sharp angular flakes. Mining techniques used by the Indians included: (1) the building of a fire adjacent to a suitable ledge or crevice and the ensuing heating of the rock would create flaking when doused with cold water, and (2) the hammering, prying and breaking of the rock with more resilient stones and other materials. This rough rock was then "blanked" at the site and then removed to their various encampments for further processing. Novaculite was a valuable trade item for the Indians who bartered it throughout the Mississippi Valley region.

TRIPOLI

The term tripoli is used in this report in the geologic sense rather than as a trade term, and it refers to a microcrystalline, finely particulate, more or less friable form of silica. In the Ouachita Mountain region of Arkansas and Oklahoma, the tripoli deposits have been formed by the leaching of calcium carbonate from the Upper Division of the Arkansas Novaculite. Tripoli deposits and prospects occur in four general areas of novaculite outcrop (Fig. 3). East and northeast of Hot Springs, tripoli float and outcrops occur on the flanks of the novaculite ridges in T. 2 S., R. 18–19 W., Garland County, Arkansas and the only currently active tripoli mines (Malvern Minerals Company) are located in this area. Another Malvern Minerals Company prospect is located to the west in Sec. 24, T. 3 S., R. 23 W., near Sunshine in southeastern Montgomery County. Further west at Blocker Creek are two quarries recently operated by Robert B. McElwaine for buffing compounds in Sec. 1, T. 5 S., R. 27 W., Pike County, Arkansas that mark the eastern end of

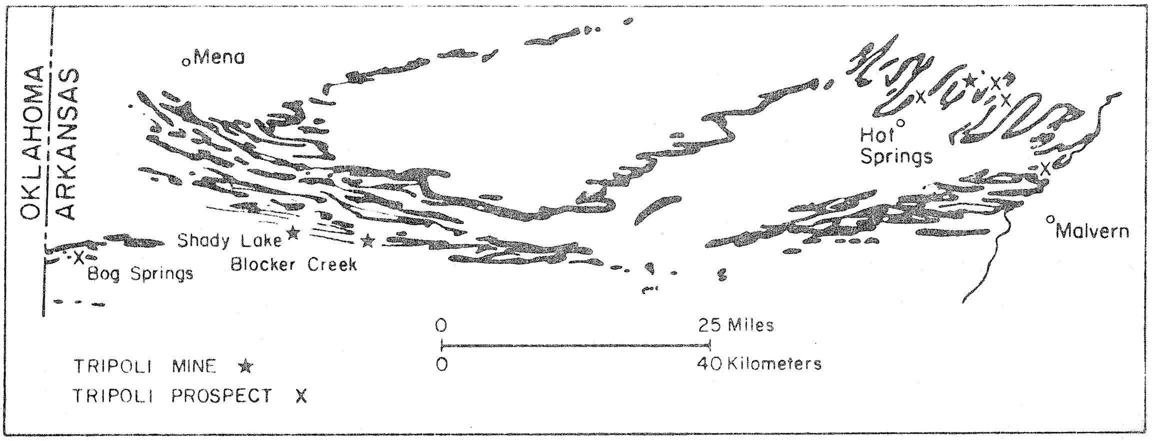

Figure 3. Outcrop map of novaculite (black lines) showing tripoli mines and prospects.

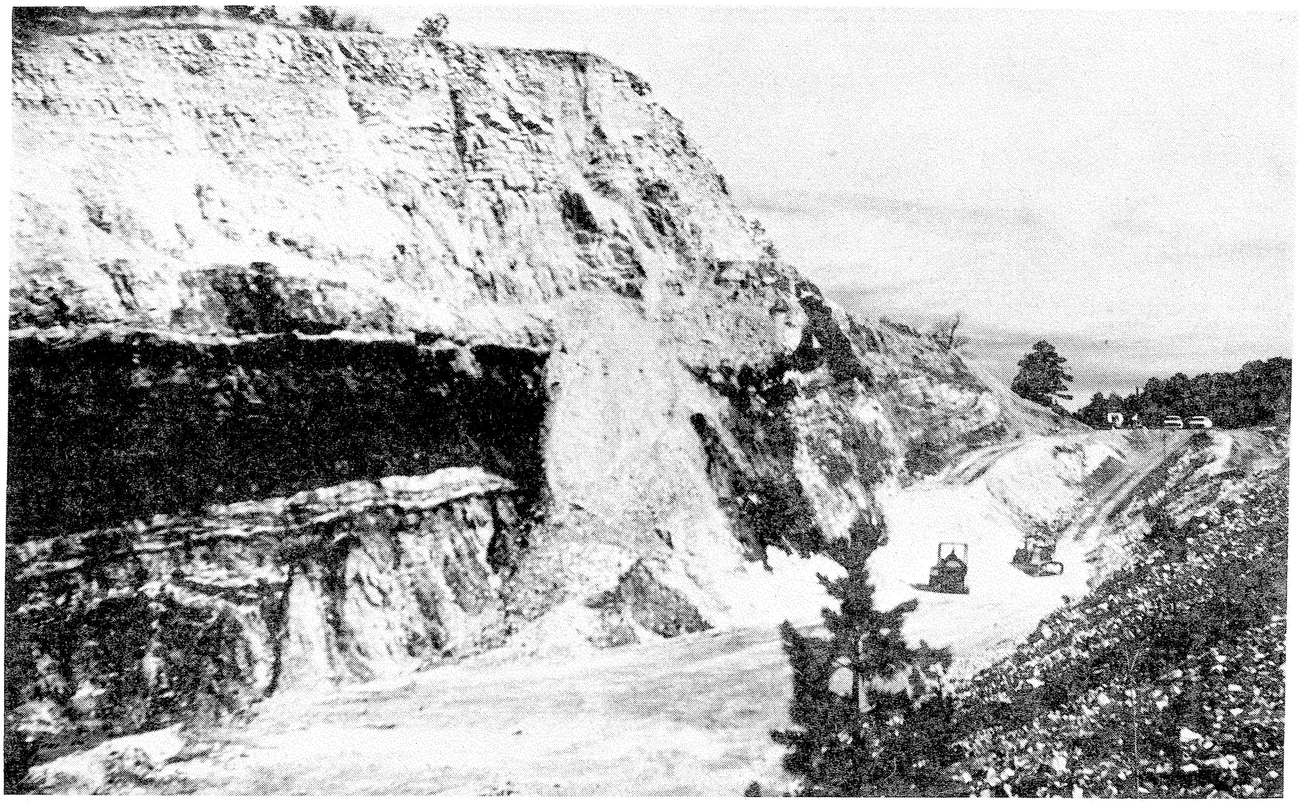

Figure 4. Malvern Minerals Company novacite (tripoli) mine in Garland County, Arkansas. The Arkansas Novaculite is inverted with the white tripolitic novaculite of the Upper Division at the base, the dense novaculite of the Lower Division at the top, and the black carbonaceous shale of the Middle Division occurring between them. Photograph by John David McFarland, III, Arkansas Geological Commission.

a tripoli trend that continues practically uninterrupted for 7½ miles to the west. In this same general area just west of the Blocker Creek deposits is a mile long trend from which a small tonnage of tripoli was produced at a quarry by Mr. McElwaine near Shady Lake in the NE¼ SW¼ Sec. 26, T. 4 S., R. 29 W., Polk County, Arkansas.

The fourth area is near Bog Springs, Arkansas in the NW¼ NW¼ Sec. 16, T. 5 S., R. 32 W., Polk County, near the Arkansas-Oklahoma border. This locality is the beginning of a tripoli occurrence that extends westward a distance of 5 miles into Oklahoma. Also in Oklahoma, tripoli float has been found along the outcrop of the Upper Division of the Arkansas Novaculite in Secs. 19, 20, and 21, T. 2 S., R. 25 E., just north of the Broken Bow Reservoir and east of the reservoir in Secs. 4, 5, 6, and 7, T. 4 S., R. 27 E.

Exploration and Mining

The tripoli beds are confined to a specific stratigraphic horizon, the Upper Division of the Arkansas Novaculite, that frequently has a characteristic topographic expression, a series of low knobs or ridges that parallel the higher main novaculite ridges. Exposures in stream beds and punky, vuggy, usually tan to red in color float blocks are relatively common, and all these factors combine to simplify exploration. The procedure generally used is to bulldoze off the shallow soil and vegetation cover to determine the width and quality of the material, and, if justified, drill to ascertain the depth of the deposit. Open-pit mining methods are used. Both overburden and tripoli must be drilled and blasted, and scrapers, front-end loaders (track hoes) are used to remove the overburden and mine the tripoli.

Character of the Tripoli

The tripoli deposits range from inches to over 100 feet in true thickness. There is a wide range of hardness in the individual beds that constitute the tripolitic zone reflecting the variation in carbonate content of the unweathered novaculite. The beds vary from material that can be crumbled by hand to hard unweathered non-calcareous novaculite layers. The tripoli consists of cryptocrystalline quartz grains or aggregates of grains more or less firmly bonded together.

The range of particle size of individual quartz grains is from 1 to 10 micrometers, and the quartz particles are essentially equidimensional. The color of the tripoli varies considerably even within the same deposit and includes white, cream, tan and brown, with white being the least prevalent and most desired. Discolorations in the tripoli mostly represent minor quantities of iron and manganese oxides formed during surficial weathering.

Malvern Minerals Company Tripoli Deposits

At this time Malvern Minerals Company is the only active producer of tripoli in Arkansas. They have one active mine and two active prospects. The active mine, located in the NE¼ NW¼ Sec. 21, T. 2 S., R. 18 W., Garland County, has been in operation since 1947. The deposit is situated on the flanks of an overturned fold so that the tripoli (Upper Division) is overlain by shale of the Middle Division and massive novaculite of the Lower Division of the Arkansas Novaculite (Fig. 4). The tripoli bed is 40 - 50 feet thick, dipping from 32 to 45 degrees to the northwest. The hard novaculite overburden of the Lower Division is drilled, shot and loaded by track hoes onto dump trucks. The shaly Middle Division member is stockpiled for future use. In most cases the tripoli can be dug with track hoes and does not need to be drilled or shot. All of the mining and trucking to the plant is done by contract. At the plant beneficiation consists of drying, grinding, and classification, both by air and screen.

The latest and most active prospect, referred to as "South Mine," is located in the SE¼ NE¼ Sec. 34, T. 2 S., R. 18 W., Garland County. The second prospect is referred to as "Sunshine Mine" and is located in the NW¼ SE¼ Sec. 24, T. 3 S., R. 23 W., Montgomery County. Interestingly the formations at the operating mine and the two prospects are overturned to the south.

The particle sizes range from about 1 to 7 micrometers according to Keith and Tuttle (1952). Seventy-five percent by volume of the tripoli produced by Malvern Minerals Company is used as a filler or extender. The remaining twenty-five percent is used in specialized abrasive applications. Bush and Stroud (1979) indicate there were about 15,500 short tons of tripoli produced that had an estimated value of over $600,000. Production in 1982 was approximately 14,000 short tons and no estimated value is available.

WHETSTONES

History

The word novaculite comes from the Latin word novacula, meaning razor hone. Richard Kirwan in his mineralogy text of 1784 is attributed with the anglicizing of this word (Griswold, 1892, p.27).

The first authentic mention of whetstones in America concerns those of Arkansas. Discussion is made in a letter written by Mr. Bringier of Louisiana in 1818 (Griswold, 1892, p. 19) about a quarry of honestone which had been worked for several years, a few miles about the "Cove of Wachitta" which is now Magnet Cove.

Another early account is given by Schoolcraft (1819) in which he stated "a quarry of this mineral (novaculite), three miles above the Hot Springs of Washita, has often been noticed by travelers for its extent and excellency of its quality. . . . It appears to me, from external character, to contain less alumine and more silex than the common novaculite, and, hence, perhaps, its superiority."

From 1891 - 1918 when Norton-Pike Company bought several hundred acres of land in the Ouachita Mountains near Hot Springs, there have been numerous quarries operated for several grades of whetstones. Most of the processing of the novaculite quarried to whetstone was done elsewhere in various out-of-state plants, mainly in Worcester, Massachusetts and Littleton, New Hampshire.

In 1962 Hiram A. Smith constructed a whetstone processing plant near Hot Springs, and at least one facility has been located in the area since then. The Smith's Whetstone Company is currently the largest producer of novaculite whetstones in the United States. There are some eight additional firms that produce and process whetstones in the Hot Springs area; these include: Pioneer, Arkansas, Wallis, Frontier, Halls, Dan's, Poor Boy, and Rigid. The Norton Company presently transports much of their raw stone to a New Hampshire plant for processing, but they also sell uncut whetstone to local processers. There are significant quantities of whetstone quarried in the area by individuals who process and sell some, but sell most of their product to local processing firms.

Location of Deposits

The main area of active whetstone mining is located just northeast of Hot Springs (Fig. 5). Most of the mining and exploration has been confined to two areas: (1) T. 2 S., R. 18 W., and (2) T. 2 S., R. 19 W., Garland County. Recently a small tonnage has been produced from deposits in Pike and Montgomery Counties. Nearly all of the commercial whetstone rock deposits are confined to massive novaculite in the lower part of the Lower Division of the Arkansas Novaculite.

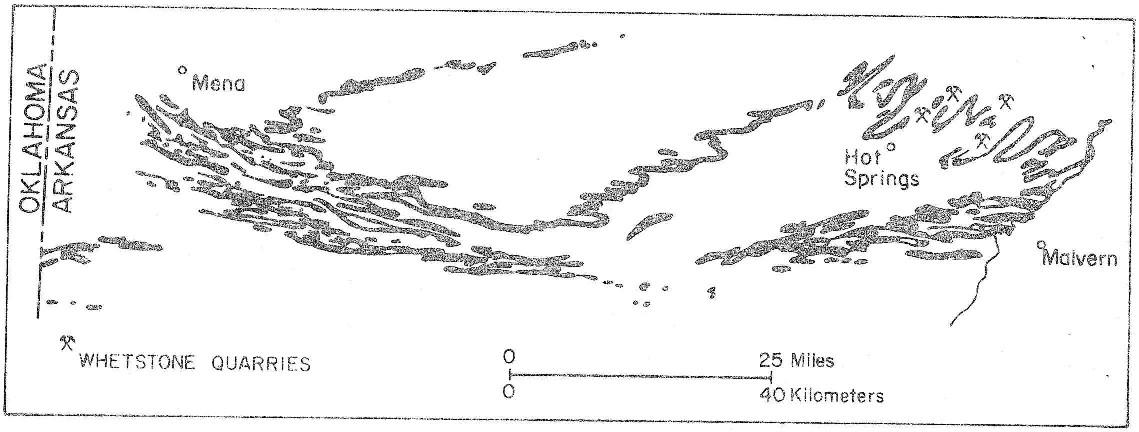

Figure 5. Outcrop map of novaculite (black lines) showing whetstone quarries.

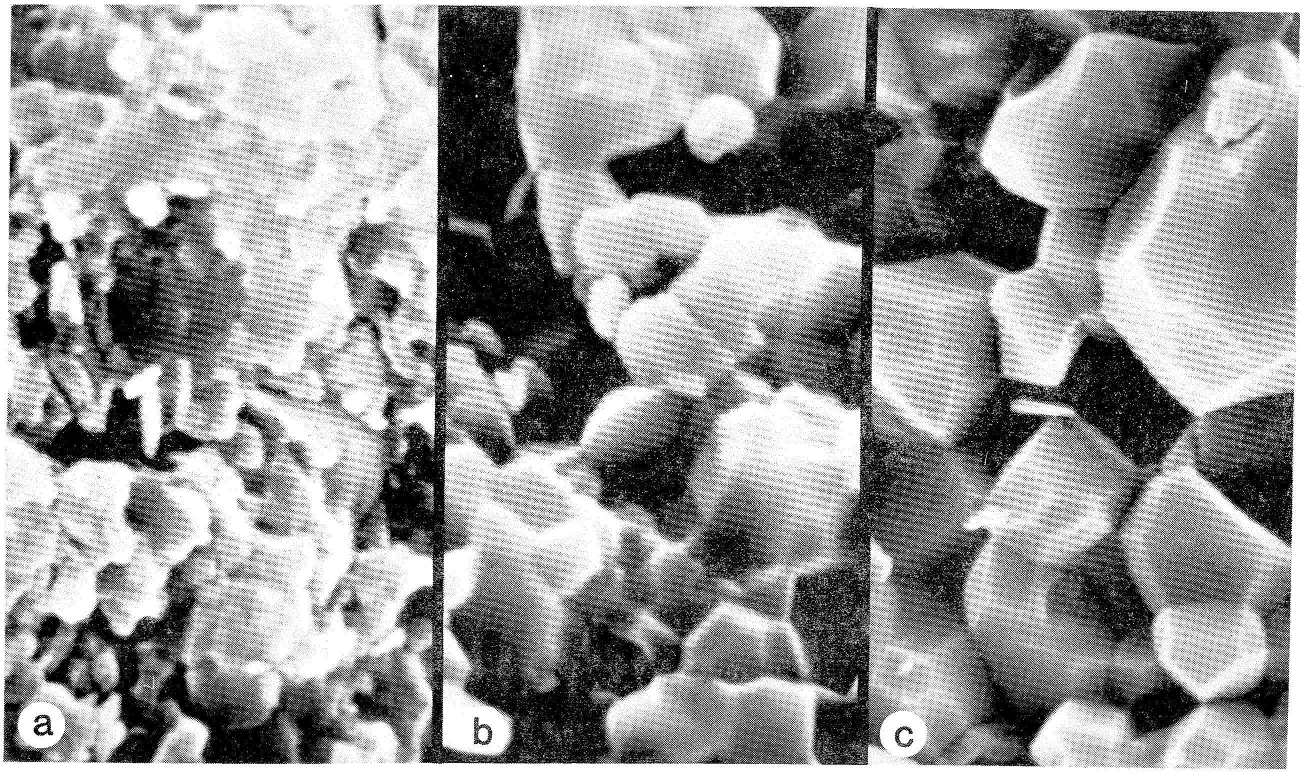

Figure 6. Variations in the textural types formed by progressive regional metamorphism of the novaculite in the Ouachita Mountains--(a) cryptocrystalline, (b) fine polygonal grains, and (c) medium polygonal grains. The sharp edges of the quartz grains of (b) and (c) afford many of the excellent qualities ascribed to the various Arkansas whetstones. Photomicrograph adapted by John David McFarland, III, Arkansas Geological Commission for Walter D. Keller, University of Missouri-Columbia.

Exploration and Mining

The area where whetstone rock is mined is one of steep, densely-wooded hills. Good outcrops are scarce, because the hillslopes where the whetstone rock layers occur are usually covered with novaculite debris. Furthermore, these whetstone layers are somewhat less resistant to erosion than the adjacent novaculite. From experience the miners know the approximate elevation on a particular hillside where the whetstone rock layers are likely to occur. They first clear the selected site with a bulldozer, and, if the rock looks promising, the exposure is drilled and shot. Quarrying follows if the quality of the rock persists. Quarrying consists simply of hand drilling and blasting using black powder, because of its low velocity. A follow-the-ore concept is used in quarrying the whetstone rock; following along the strike and dip of the bed until the quality of the rock deteriorates, or quarrying becomes too expensive.

The quality of a particular whetstone rock layer (lead) may vary both along the strike and the dip of that layer. Individual leads are narrow varying from 6 to 35 feet, but they may persist up to 900 feet along strike. After the rock is partially broken it is then carefully quarried usually by hand tools but track hoes and other equipment may be employed. Blocks are then trimmed to a roughly equidimensional shape weighing 10 to 20 pounds in most varieties. They are then hand sorted for further quality evaluation before shipment to the plant for processing. Quartz veins, cavities, cracks or fractures, laminations, styolites, and lack of uniform texture are all a basis for rejection.

Characteristics of the Whetstone Rock

Whetstone-grade novaculite or whetstone rock, as it is called here, is found in the lower part of the Lower Division of the Arkansas Novaculite. These particular layers of novaculite are especially suited for whetstones because of their porosity, uniformity of texture, and the sharp edges of the minute quartz grains making up the rock. The porosity is due to rhombic cavities from which carbonate has been leached during weathering. Uniformity of texture implies not only uniformity in the size of the individual grains but more especially a uniform distribution of the cavities throughout the rock. Individual quartz grains present sharp edges because the grains are not rounded but rather are closely packed in a mosaic texture, the individual grains exhibiting polygonal shapes in cross section (Fig. 6).

In the sharpening process small micro-quartz particles bounding the pore spaces break off preventing the pores from being filled with steel particles which would cause the stone to glaze over. Several varieties of whetstones are manufactured under different trade names, and the essential difference between them is a difference in porosity. Thus, the hard Arkansas Stone has a porosity of 0.07 percent, while the more rapid-sharpening Washita Stone has a porosity of 16 percent. The increased porosity in the Washita Stone is due not only to a greater frequency of rhombic cavities but also to an increase in the size of the individual cavities. This variation in porosity is reflected in a difference in luster in the two stones. Thus, the denser hard Arkansas Stone is characterized by a waxy luster while the more porous Washita Stone has the dead appearance of unglazed procelain.

Whetstone Processing

Quarry blocks are sawed into desired shapes with a diamond saw using cutting oil or solvent as a lubricant. These shapes are then finished on a lap with carborundum powder, given a final visual quality-control check, then packaged and shipped. Only about 5 percent of the blocks quarried end up as finished whetstones.

The five types of stones produced for market are listed in order of decreasing porosity:

Trade Names	Use
Washita Stone	for rapid sharpening
Soft Arkansas	for general all around sharpening
Hard Arkansas	for polishing a blade to a very fine edge
Translucent	for polishing to the finest possible edge
Black Hard	ditto, and miscellaneous

ROCK AGGREGATE

Stone for rock aggregate is widespread in the Ouachita Mountains. Novaculite has generally been bypassed for aggregate uses in past years mostly because of its abrasiveness in various crushing processes. Recently it has been quarried in significant quantities in Arkansas as a result of an increased demand for rock aggregate and the development of more effective crushing techniques (Fig. 7).

Novaculite has been mined and crushed for concrete aggregate, road construction, railroad ballast, high-silica refractories and other uses. Production of novaculite used for construction and other miscellaneous purposes has not been segregated from Arkansas's total output of stone except for the period 1941-1949 when 760,000 short tons valued at about $420,000 were produced. In recent years, as much as one million short tons has been quarried annually at a value from about $3 to over $5 per ton.

Five quarries are currently operated for commercial rock aggregate mostly in the massive, dense novaculite of the Lower Division of the Arkansas Novaculite. These quarry operations include: The Walker Stone Company in Sec. 6, T. 5 S., R. 31 W., at Hatton in Polk County; the Mid-State Construction and Materials Company in Sec. 21, T. 4 S., R. 20 W., north of Bismarck in Hot Spring County; in Sec. 9, T. 4 S., R. 17 W., north of Malvern in Hot Spring County; and in Sec. 14, T. 3 S., R. 18 W., near Lake Catherine in Garland County; and the West Construction Company in Sec. 9, T. 3 S., R. 16 W., near Glen Rose in southwestern Saline County. In recent years novaculite has also been mined for rock aggregate in Sec. 34, T. 3 S., R. 17 W., at Butterfield; in Sec. 8, T. 3 S., R. 16 W., in Hot Spring County; in Sec. 36, T. 2 S., R. 17 W., in Saline County; in Sec. 26, T. 2 S., R. 18 W., in Garland County; and in Sec. 19, T. 1 N., R 13 W., near Little Rock in Pulaski County.

Exploration and Mining

Most of the novaculite rock aggregates occur in the massive, dense, cryptocrystalline, highly fractured and often steeply tilted beds of the Lower Division of the Arkansas Novaculite which characteristically forms the tall jagged peaks or hogbacks in the southern and central portions of the region. Massive novaculites are initially examined and tested at the surface in areas adjacent to transportation and/or industrial facilities. Upon the establishment of a quarry site there is close attention directed initially towards drilling and blasting procedures to insure the desired breakage of the rock. Typically the rock has a primary crush to about a 6-inch size and a secondary crush to various smaller screened sizes that include some by-products. It is then stockpiled in storage bins and transferred to loading facilities. Typically conveyors connect the crushing and screening systems. It is reported that most crushed novaculite meets the various specifications set for highway aggregates-including-asphalt, concrete and road subbase. The high anti-skid characteristics of some novaculite adds considerable potential for partial application to other highway aggregates. Additional uses of novaculite are as railroad

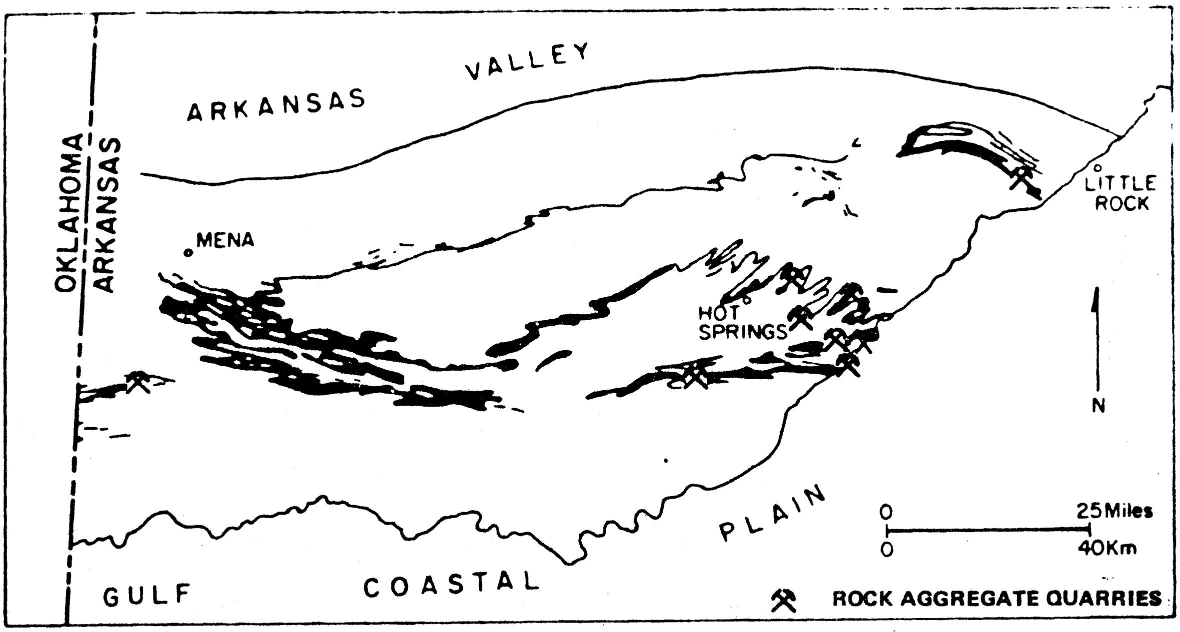

Figure 7. Outcrop map of novacullite (black lines) showing rock aggregate quarries.

Table 1. —Typical Chemical Composition of Ebony Novacite^R

		550 MESH	20 MICRON	10 MICRON
Silica	SiO_2	63.15%	58.00%	60.40%
Carbon	C	3.24%	3.09%	3.37%
Sulphur	S	0.13%	0.08%	0.07%
Aluminum Oxide	Al_2O_3	19.16%	21.06%	22.40%
Ferric Oxide	Fe_2O_3	2.29%	2.29%	2.15%
Titanium Oxide	TiO_2	1.80%	1.40%	1.70%
Calcium Oxide	CaO	0.25%	6.88%	2.00%
Magnesium Oxide	MgO	0.15%	0.45%	0.38%
Total Loss on Ignition		10.25%	8.75%	9.75%

ballast, rock aggregate fill in dams, dikes and many other constructional purposes.

Reserves and Outlook

Reserves of massive novaculite represent many billions of tons; however, novaculite located near railroads and other transportation facilities is less abundant. It is likely that exploitation of novaculite for rock aggregates will continue to expand in or near the areas presently established and that except for local needs the cost of transportation will be prohibitive for many other locations.

OTHER USES

The Ouachita Sand and Gravel Company obtains a dense translucent novaculite from their quarry in Sec. 6, T. 3 S., R. 16 W., north of Magnet Cove in Hot Spring County for a number of purposes. First a highly selected rock is used for whetstones, next large "rough" blocks are shipped to various dealers for unknown purposes, the remaining material is then crushed to 3/4" – 2 1/4" and sold mostly for use as a filler in fire brick. This rock is further crushed and screened by the brick manufacturers for application in various products. Not all novaculite in the region is ideal for fire brick uses since some of it decrepitates at high temperatures. The Ouachita Sand and Gravel Company also sells some crushed novaculite to the Buffalo Stone Corporation in Hot Springs who further process it in sizes from 1/16" – 2" for use as a polishing and grinding media.

The Malvern Minerals Company is currently mining, processing, and marketing quantities of a variety of black siliceous shale as Ebony Novacite^{R} from the Middle Division of the Arkansas Novaculite from their tripoli mine in the NE¼ NW¼ Sec. 21, T. 2 S., R. 18 W., in Garland County. This material is used in industry as a black silica pigment in various resinous mediums. Studies are being further conducted on its potential use as a black pigment or an extender since it offers superior rheological features as compared to many other black pigments that often have higher surface areas and sorption properties. The typical chemical composition of Ebony Novacite^{R} shown in Table 1 is published by permission of the Malvern Minerals Company.

An increased demand for novaculite in refractory and in polishing and grinding uses is likely. Since some of this rock is rather unique in occurrence, more selective mining and processing is required. An expanded use of Ebony Novacite^{R} as a pigment and filler in various products is anticipated by the Company.

FUTURE CONSIDERATIONS

Crushed dense novaculite has numerous other potential applications; namely, for any purpose requiring an essentially pure silica rock with but minor undesirable chemical impurities; and high physical properties.

In recent years some investigations have taken place concerning the application of crushed novaculite for interlayered sequences with plastics. It reportedly would provide a binding and add other physical qualities to the material upon application with rather intense heat. From our brief tests it has been determined that decrepitation takes place in some of the coarsely crushed novaculite but rarely in the finer mesh sizes. Studies have also revealed that some of the novaculite, especially recrystallized types, does not appreciably decrepitate or spall during the application of heat for processing. In an open report in August, 1980, Robert B. McElwaine states that testing done on the Blocker Creek tripoli indicates that it meets the qualifications for uses as semi-lightweight aggregate in concrete blocks, as a non-polishing skid-resistant aggregate in asphalt highways, and others.

Chemical analyses of quarry and outcrop samples (Fig. 8) reveal that novaculite has a vast potential as a high-silica rock resource (Table 2). The electronics applications are developing and the scope is tremendous. Surface modification of high-purity silica is now a reality and there are many potential uses. High-purity silica has application for making silicon metal and possibly in the growing of quartz crystals from seed. Billions of tons of dense novaculite have yet to be tapped by the rock aggregate industry. An expanded demand for microcrystalline quartz is anticipated in the next decade and the Arkansas Novaculite of Arkansas and Oklahoma should make important contributions to these expected needs.

ACKNOWLEDGEMENTS

We would like to express our deepest gratitude for the kind assistance provided by the personnel of the various companies and other individuals involved in the mining and processing of Araknsas Novaculite. A special thanks is extended to Hewitt Harlow and Jim Moreland of Malvern Minerals Company; Robert B. McElwaine, a consulting geologist; Bryan Willis of Smith's Whetstone Company; Marc Dilatush of Arkansas Whetstone Company; Charles Pate of Pioneer Whetstone Company; Owen Spickard of Norton Company; Clovis Wallis of Wallis Whetstone Company and Ouachita Sand and Gravel Company; Marvin Woodson of Ben M. Hogan Company; Charles Gibbs of Mid-State Construction and Materials Company; David S. Walker of Walker Stone Company; Dick Hedrick of the Hedrick Corporation; Buddy Blood of Arkansas Explosives Company; Ira West of West Construction Company; and Oliver Duty of Little Rock, Arkansas.

APPENDIX

COMMERCIAL USES OF BLACK HARD ARKANSAS STONE

By

Marc H. Dilatush
Arkansas Whetstone Company, Inc.

As with other types of Arkansas Whetstones, the primary uses of Black Hard Arkansas are for sharpening knives and honing metal surfaces. However, Black Hard Arkansas is unique because it has a very uniform density and honing quality. The specific gravity of 2.66 makes it the most dense of Arkansas Whetstones and the extremely fine recrystallized structure allows it to hone to an extremely smooth finish on metals. Because of the qualities it is suited for some applications better than any other material.

One of the interesting uses found in recent years is for polishing electronic microcircuit dice pedestals. The microcircuits are manufactured by duplicating a particular circuit thousands of times in one large sheet. The individual circuits are then cut from the large sheet and are called "dice". Although the method varies somewhat, the dice must be connected to other circuits or pin connections before encapsulation. Some methods require that the dice be placed on a small pedestal for manual connection under a microscope with ultrasonic welding devices. In order to minimize damage to the dice, the pedestal must be extremely smooth with no burrs. A small Black Hard Arkansas stone is excellent for removing burrs and polishing the pedestal without leaving grit or mineral deposits.

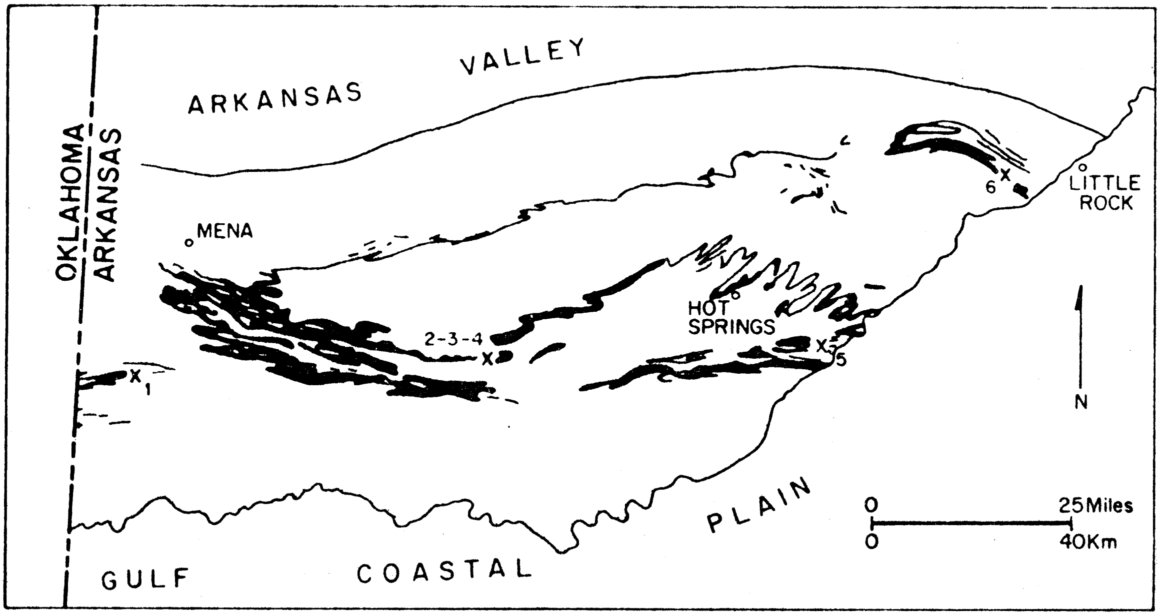

Figure 8. Map of eastern part of Ouachita Mountains (stippled area) showing outcrop of Arkansas Novaculite (black lines) and location of novaculite samples listed in Table 2.

Table 2. —Chemical Analysis (in Percent) of Arkansas Novaculite Samples and Additional Data On Samples (See Figure 8 for Location Map)

Sample Number	SiO_2	Al_2O_3	MnO	Fe_2O_3	TiO_2	CaO	MgO	Na_2O	K_2O	LOI
1	99.03	0.42	0.01	0.07	0.02	0.03	0.03	0.004	0.10	0.29
2	99.18	0.26	0.07	0.09	0.02	0.03	0.02	0.003	0.03	0.30
3	99.07	0.25	0.01	0.29	0.02	0.03	0.02	0.003	0.05	0.26
4	99.19	0.19	0.03	0.19	0.01	0.03	0.02	0.003	0.05	0.29
5	99.19	0.25	0.01	0.14	0.02	0.02	0.02	0.003	0.05	0.30
6	99.16	0.27	0.04	0.16	0.02	0.03	0.03	0.002	0.01	0.28

Sample Number	True Thickness (Feet)	Division of Arkansas Novaculite	Location
1	130	Lower	Abandoned quarry, NW¼NW¼ Sec. 6, T. 5 S., R. 31 W., Polk Co.
2	100	Bottom of Lower	Caddo Gap outcrop, SW¼SE¼ Sec. 18, T. 4 S., R. 24 W., Montgomery Co.
3	249	Remainder of Lower	Caddo Gap outcrop, SW¼SE¼ Sec. 18, T. 4 S., R. 24 W., Montgomery Co.
4	105	Upper	Caddo Gap outcrop, NW¼NE¼ Sec. 19, T. 4 S., R. 24 W., Montgomery Co.
5	210	Lower	Butterfield quarry, SW¼NE¼ Sec. 34, T. 3 S., R. 17 W., Hot Spring Co.
6	60	Lower	Lawson Rd. outcrop, NE¼SE¼ Sec. 19, T. 1 N., R. 13 W., Pulaski Co.

Another interesting application which has been used for many years is for testing the purity of gold. A particular variety of Hard Black Arkansas which is solid black with no discolorations or other non-uniformities is used to remove a very small amount of gold for testing. The gold is scratched on the stone and then acids are applied to the scratch and the reaction is observed to determine the purity. The stone is sold in a kit, with the proper acids and instructions.

Many stones are cut in special shapes for wood carving tools, gunsmiths, watch and clock repair and manufacture, as well as reciprocating and jet engine repair, and tool and die work. Any application which requires a highly polished lightly honed metal surface can use a Hard Black Arkansas stone.

REFERENCES

Bush, William V., and Stroud, Raymond B, 1980, Directory of Arkansas Mineral Producers and Production—1979. Arkansas Geological Commission, Miscellaneous Publication, 33 p.

Folk, Robert L., 1973, Evidence for Peritidal deposition of Devonian Caballos Novaculite, Marathon Basin, Texas: American Association Petroleum Geologists Bulletin, Vol. 57, p. 702–725.

Goldstein, August, Jr., and Hendricks, Thomas A., 1953, Siliceous sediments of Ouachita facies in Oklahoma: Geological Society of America Bulletin, Vol. 64, p. 421–442.

Griswold, Leon S., 1892, Whetstones and the Novaculites of Arkansas: Arkansas Geological Survey, Annual Report for 1890, Vol. 3, 443 p.

Hass, W. H., 1951, Age of Arkansas Novaculite: American Association Petroleum Geologists Bulletin, Vol. 35, p. 2526–2541.

Holbrook, Drew F., and Stone, Charles G., 1978, Arkansas Novaculite – A Silica Resource: 13th Annual Forum on the Geology of Industrial Minerals, Oklahoma Geological Survey Circular 79, p. 51–58.

Keith, M. L., and Tuttle, O. F., 1952, Significance of variation in high-low inversion of Quartz: from Bowen edition, American Journal Science, p. 203–280.

Keller, Walter D., Viele, George W., and Johnson, Clayton H., 1977, Texture of Arkansas Novaculite indicates thermally induced metamorphism: Journal Sedimentary Petrology, Vol. 47, p. 834–843.

Lowe, Donald R., 1975, Regional controls on silica sedimentation in the Ouachita System: Geological Society of America Bulletin, Vol. 86, p. 1123–1127.

Miser, Hugh D., 1917, Manganese deposits of the Caddo Gap and DeQueen quadrangles, Arkansas: United States Geological Survey Bulletin 660-C, p. 59–122.

Miser, Hugh D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles: United States Geological Survey Bulletin 808, 195 p.

Park, David E., and Croneis, Carey, 1969, Origin of Caballos and Arkansas Novaculite Formations: American Association Petroleum Geologists Bulletin, Vol. 53, p. 94–111.

Purdue, A. H., 1909, The Slates of Arkansas: Arkansas Geological Survey, 170 p.

Schoolcraft, Henry R., 1819, A view of the lead mines of Missouri, including some observations on the mineralogy, geology, geography, antiquities, soil, climate, population and production of Missouri, Arkansas and other sections of the western country: C. Wiley and Company, New York, 299 p.

Sholes, Mark A., and McBride, Earle F., 1975, Arkansas Novaculite *in* Briggs, Garrett et al., editors, Sedimentology of Paleozoic flysch and associated deposits. Ouachita Mountains—Arkoma Basin Oklahoma: Dallas Geological Society, Guidebook, April, 1975, p. 69–87.

Stroud, Raymond B., Arndt, Robert H., Fulkerson, Frank B., and Diamond, W. G., 1969, Mineral resources and industries of Arkansas. United States Bureau of Mines Bulletin 645, 418 p.

Stone, Charles G., and McFarland, John D., III, with the cooperation of Haley, Boyd R., 1981, Field Guide to the Paleozoic rocks of the Ouachita Mountain and Arkansas Valley Provinces, Arkansas: Arkansas Geological Commission Guidebook 81–1, 140 p.

GEOLOGY OF THE BLAKELY SANDSTONE IN EASTERN MONTGOMERY AND WESTERN GARLAND COUNTIES, ARKANSAS

By

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ABSTRACT

The Blakely Sandstone south of Lake Ouachita in eastern Montgomery and western Garland Counties comprises about 720 feet (220 m) of shale, sandstone and siltstone in that order of abundance. The formation can be divided into a lower (basal) sandstone division with maximum thickness of 150 feet (46 m), a middle shale division, and an upper sandstone division with maximum thickenss of about 200 feet (60 m). Thin, interbedded siltstones are typical of all three subdivisions. Sandstones, which comprise 20% to 25% of the formation are typically quartz-cemented or calcite-cemented quartz arenites, sublitharenites, or litharenites (sedarenites). Well rounded quartz grains form the dominant matrix constituent of most sednaenites; lithic clasts include fragments of chert, siltstone, sandstone, and micritic and pelletoidal limestone. Ages common to seven graptolite species from the middle shale division suggest a lower Middle Ordovician age for that part of the formation. Paleocurrent determinations, lateral facies changes, trace fossil evidence for a bathyal to abyssal ichnofacies, and the probable presence of partial Bouma sequences, grain flow and debris flow deposits offer strong collective evidence that Blakely sediments were derived from a northern source and were deposited in deep water, possibly as a submarine fan.

INTRODUCTION

The Blakely Sandstone (Middle Ordovician) crops out over much of the Benton Uplift as low, narrow, complexly folded ridges (Fig. 1). The focus of this paper is an area of outcrop south of Lake Ouachita in eastern Montgomery and western Garland Counties. Here, the formation is well-exposed, relatively continuous, and its structure comparatively simple, consisting of two sets of easterly plunging, macroscopic folds separated by linear, northeast and east-trending ridges. The area of investigation is about 14.5 miles (23.2 km) long and is, in part, bounded on the north by a sequence of structurally complex Lower Ordovician rocks mapped by Soustek (1979). The western edge of the study area coincides with the boundary of the Bonnerdale 7.5 minute quadrangle west of Little Mazarn Mountain in Sec. 18, T. 3 S., R. 23 W. The eastern limit is the eastern edge of Sec. 20,

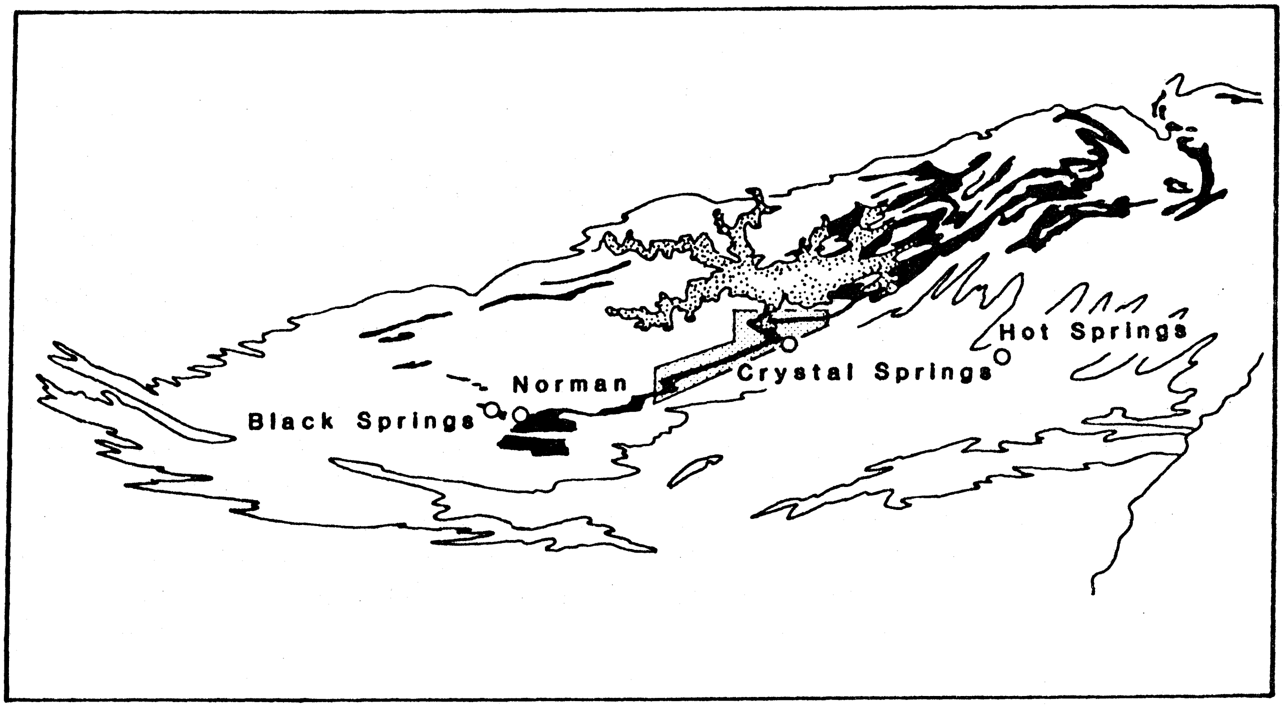

Figure 1. Generalized geologic map of the Benton Uplift and adjacent parts of Arkansas, showing the location of the area covered in this paper. The solid line represents the approximate outcrop limits of the Arkansas Novaculite (Devonian-Mississippian) and defines the boundaries of the Benton Uplift. Major outcrops of the Blakely Sandstone are shown in black. After Haley and others, 1976.

T. 2 S., R. 21 W. in the Crystal Springs 7.5 minute quadrangle. Within this area (Fig. 2) we have mapped the Blakely Sandstone and adjacent parts of the conformably underlying Mazarn Shale (Lower Ordovician) and the conformably overlying Womble Shale (Middle Ordovician).

Miser and Purdue (1929, p. 25, 26, 29) have outlined the early work that resulted in recognition of the Blakely Sandstone as a distinct unit and led to a major revision of the stratigraphy of Lower Paleozoic rocks in the Arkansas Ouachitas. More recently, extensive mapping of the Benton Uplift and adjacent areas has been conducted by personnel of the Arkansas Geological Commission and the United States Geological Survey (Stone and others, 1973; Haley and others, 1976; Arkansas Geological Commission, 1977). Detailed work on the Blakely Sandstone and adjacent units south of Lake Ouachita is limited to studies by Brieva (1963) and Buthman (1982a, 1982b, 1982c), although Pitt and others (1961) included these formations in their broader-scale paper on the rocks of the Ouachita core. Buthman's (1982a) investigation included the area between Black Springs (Fig. 2) and the western boundary of Little Mazarn Mountain. It should be considered a comparison study to Stolarz (1982), upon which this paper is based.

DISCUSSION OF MAJOR STRATIGRAPHIC UNITS

All rocks exposed in the study area are Lower to Middle Ordovician sedimentary rocks (Fig. 3) with the exception of minor occurrences of vein quartz of probable Pennsylvanian age (Miser, 1959) and a thin lampro-

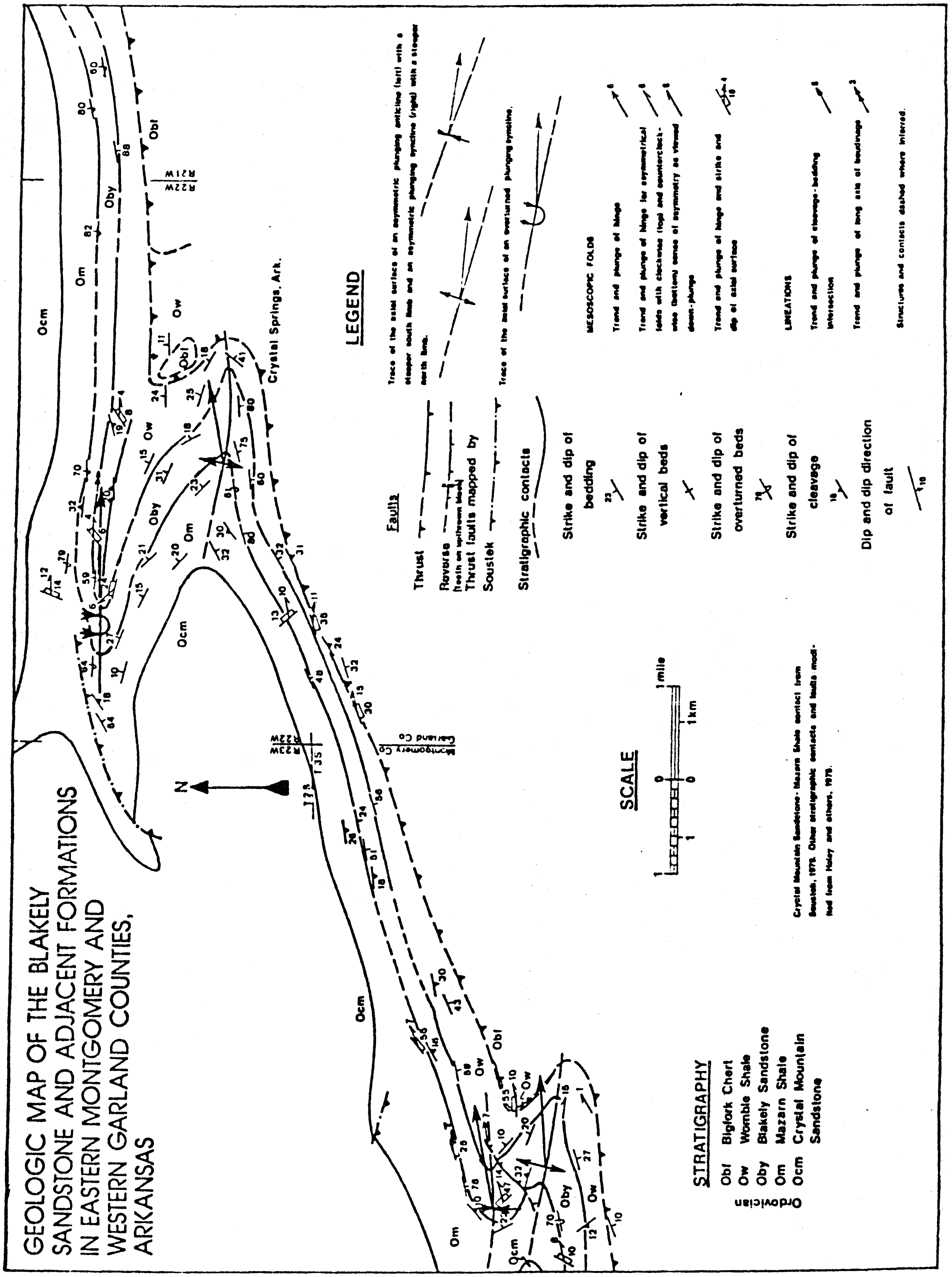

Figure 2. Geologic map of the Blakely Sandstone and adjacent formations in eastern Montgomery and western Garland Counties. Partly adapted from Soustek (1979) and Haley and others (1976).

AGE	ROCK UNIT and map symbol		THICKNESS (feet)
Pennsylvanian	Atoka Formation	Pa	27,000
	Johns Valley Shale	Pjv	1500
	Jackfork Sandstone	Pj	6000
Mississippian	Stanley Shale	Ms	8500
Devonian	Arkansas Novaculite	MDa	950
Silurian	Missouri Mountain Shale	Smm	250
	Blaylock Sandstone	Sb	1500
Ordovician	Polk Creek Shale	Opc	175
	Bigfork Chert	Obf	800
	Womble Shale	Ow	3500
	Blakely Sandstone	Oby	700
	Mazarn Shale	Om	1100
	Crystal Mountain Sandstone	Ocm	850
	Collier Shale	Oc	1000+

Figure 3. Paleozoic stratigraphic column for the Ouachita Mountains. Modified from Stone, Haley, and Viele (1973).

Figure 4. Small-scale crossbeds, resembling climbing ripples, in a thin siltstone layer, upper Mazarn Shale. Pencil for scale.

phyre sill probably intruded during early Late Cretaceous time (Stone and others, 1981).

Mazarn Shale

The Mazarn Shale underlies the Blakely Sandstone and overlies the Crystal Mountain Sandstone along conformable, gradational contacts. It comprises about 1,100 feet (335 m) of dark, locally banded shale with minor amounts of interbedded sandstone, siltstone and limestone. Where present, the banding is caused by alternating layers of black and gray-green shale, typically less than 1 inch (2.5 cm) to 3 inches (7.6 cm) thick. Thin section analysis of the layering indicates that the black layers contain more silt-size detrital quartz and more siltstone laminations than the green. Although common in the Mazarn, banded shales also occur in the Blakely and Womble Shale (Miser and Purdue, 1929, Plate 6A) which limits their use as a stratigraphic indicator.

Zones up to several feet thick containing thin (1 - 4 inches), cross-laminated, gray siltstones interlayered with shales are common in the upper parts of the Mazarn Shale (Fig. 4).

Mazarn limestones are micritic and partially recrystallized, black when fresh, and weather to dark blue or steel-gray. They are lensoidal to evenly bedded and range in thickness from less than 1 inch (2.5 cm) to 6 inches (15 cm). The limestone beds are interlayered with black calcareous shales up to several inches thick and with yellow-brown calcareous siltstones that average less than 1 inch (2.5 cm) thick. The limestones typically lack sedimentary structures but occasionally contain small-scale trough crossbeds in erosion resistant zones near the middle and upper parts of the formation.

Sandstones can be found throughout the Mazarn Shale. They are very fine to medium grained, gray to tan, and locally weather to a reddish-brown. Most beds are less than 6 inches (15 cm) thick, but occasionally reach 10 feet (3 m) in thickness. Typical contacts with the surrounding shales and siltstones are sharp. Thicker sandstone beds are massive and not laterally extensive. The thinner beds locally show parallel laminations, small-scale crossbeds, and sole marks such as flute and groove casts.

Blakely Sandstone

The principal reference section of the Blakely Sandstone is located at Blakely Mountain Dam 12 miles northeast of Hot Springs, Arkansas. The formation overlies the Mazarn

Shale and is overlain by the Womble Shale along a gradational contact.

According to various workers (Miser and Purdue, 1929; Pitt and others, 1961, Sommer, 1971; Stone, Haley, and Viele, 1973) the thickness of the Blakely Sandstone is 450 to 700 feet (137 to 214 m). The senior author measured one complete stratigraphic section through the formation along Walnut Creek in the SW¼, Sec. 32, T. 2 S., R. 22 W. and one partial section in the center of the NW¼, SW¼, Sec. 20, T. 2 S., R. 21 W., finding thicknesses of approximately 720 feet (220 m) and 510 feet (156 m), respectively (Fig. 5, sections A and C). A second partial stratigraphic section 67 feet (20 m) thick from the upper Blakely Sandstone has been measured in the E½, NE¼, SE¼, Sec. 17, T. 3 S., R. 23 W. (Fig. 5, section B).

Figures 5 and 6 indicate that sandstones are most abundant near the base and top of the formation, a fact noted by Miser and Purdue (1929, p. 29 and 30) and implicit in Buthman's (1982b) Figure 10. The presence of these sandstone-rich zones separated by a thick section of shale produces the "double hogback" ridge noticeable in several parts of the map area. For purposes of discussion we will refer to these zones informally as the lower sandstone, middle shale, and upper sandstone divisions.

The lower sandstone division is composed of thick beds of massive sandstone interbedded with shale and siltstone (Fig. 5). The division can reach thicknesses of about 150 feet (46 m) but is typically much thinner. Throughout the map area, the lower sandstone division is found to pinch out laterally and reappear, possibly due to depositional processes. When present at a low angle of dip, the lower division forms one of the double hogbacks of the Blakely ridge.

The middle shale division is composed predominantly of shale and siltstone with several sandstone units of variable thickness (up to 6 feet or 2 m) scattered throughout (Fig. 5). This division varies in thickness from 250 to 450 feet (76 to 137 m) and is found over the entire map area (Fig. 6).

The upper sandstone division is about 200 feet (60 m) thick and is the principal ridge former. This division consists of thick, massive sandstone with some interbedded shale, siltstone, and thin sandstone. West of the map area the upper sandstone division is interpreted to thin and eventually pinch out (Fig. 6).

The shales of the Blakely Sandstone comprise 75 to 80 percent of the formation. The shale occurs in alternating green and black layers, typically thicker than those of the Mazarn Shale. Green bands in the Blakely shales range up to 6 inches (15 cm) thick and black bands reach thicknesses of 1 to 3 inches (7.6 cm).

Petrographic study of the Blakely shales indicates that the black layers contain more silt-sized detrital quartz than the green layers. Some black layers also show thinly laminated layers of silt-sized quartz. Units of black shale up to tens of feet thick that contain thinly laminated interbeds of pyritic siltstone can be found in the middle shale division of the formation.

Siltstones are commonly found throughout all three divisions of the Blakely Sandstone (Fig. 5). The siltstone layers, similar to those in the Mazarn Shale (Fig. 4), are gray, quartz rich, up to 3 inches (7.6 cm) thick, and interbedded with shale. Numerous beds display flaser bedding typically defined by black laminations within crossbeds. The flaser beds, when well developed, serve as geopetal indicators.

The sandstones of the Blakely Sandstone comprise 20 to 25% of the entire formation. They are most abundant in the upper sand-

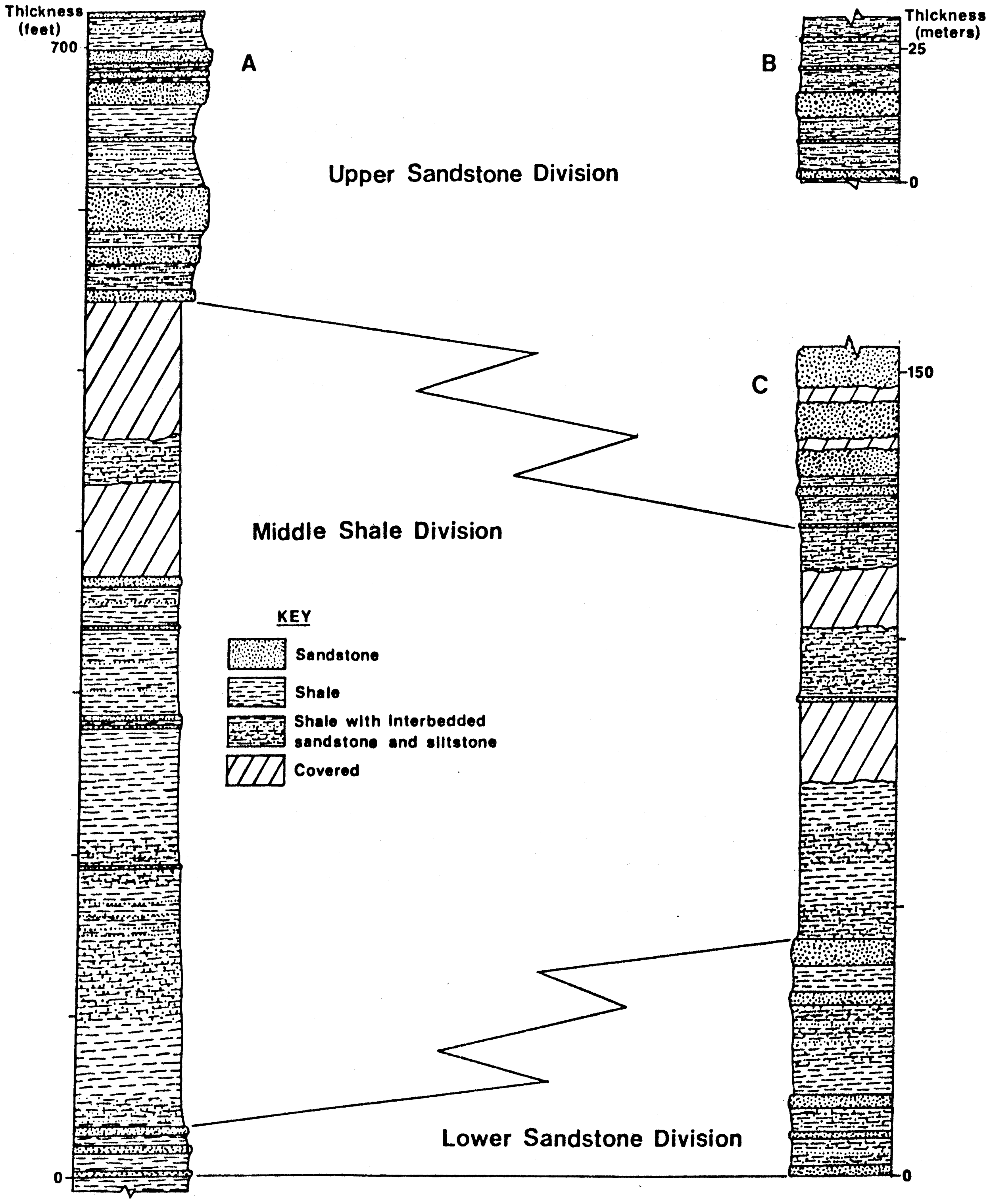

Figure 5. Measured sections in the Blakely Sandstone. A) Section along Walnut Creek in the center of the SW¼, Sec. 32, T. 2 S., R. 22 W. B) Partial section from the upper sandstone division, E½, NE¼, SE¼, Sec. 17, T. 3 S., R. 23 W. C) Partial section measured in the center of the NW¼, SW¼, Sec. 20, T. 2 S., R. 21 W.

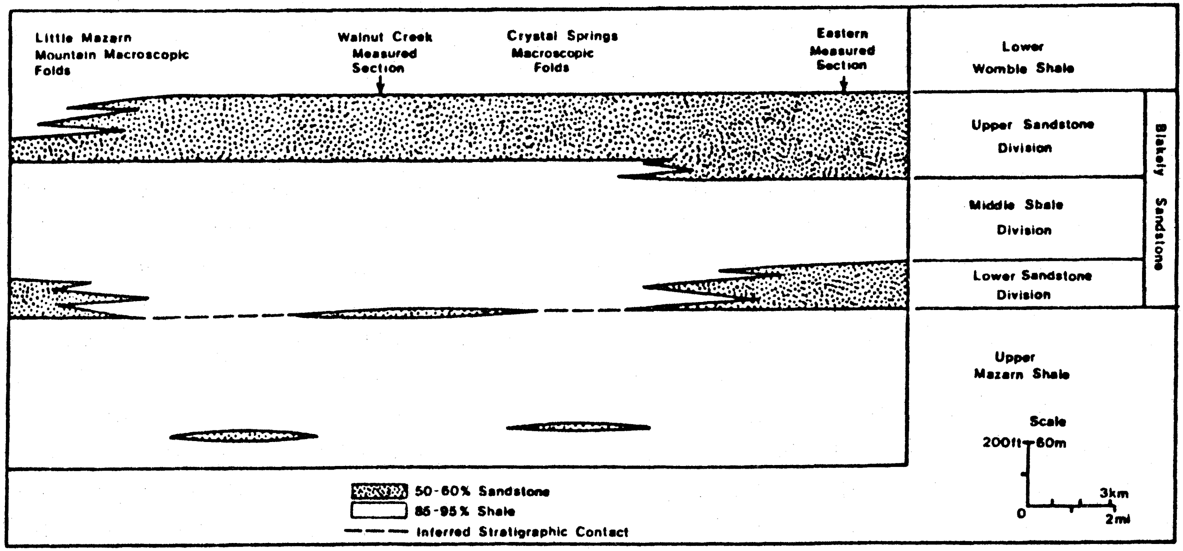

Figure 6. Generalized facies diagram showing inferred lateral variations of sandstone units in the upper Mazarn Shale and the three informal divisions of the Blakely Sandstone. Datum is the base of the Blakely Sandstone.

stone division and least abundant in the middle shale division. Sandstone beds range in thickness from several inches up to 8.0 feet (2.5 m), but occur in sets up to 27.0 feet (8.2 m) thick in the upper sandstone division (Fig. 5A). The majority of the sandstones are well indurated, silica cemented, gray or black quartzarenites that locally weather brown and look very much like metaquartzites. Other sandstones are highly weathered, brown, friable and were interpreted to have been originally cemented with calcite by Miser and Purdue (1929). These two varieties of sandstone, the silica- and carbonate-cemented arenites, are very fine to coarse grained, well sorted, and contain minor amounts of chert, secondary limonite, pyrite, and quartz vein fillings. These sandstones are generally massive, but sedimentary structures such as parallel laminations and trough and planar cross bedding are occasionally present.

The Blakely Sandstone also contains granule to cobbly sedarenites (Folk, 1974). Granule to pebbly sedarenites can be found throughout the formation, but cobbly sedarenites are restricted to the upper sandstone division (Table 1). Clasts of chert, siltstone, sandstone, and micritic and pelletoidal limestone comprise the coarse and some of the fine grain size fraction of the sedarenites (Figs. 7c and 7d). The larger clasts are angular, very elongate, and range up to several inches in length. Well rounded quartz is the dominant constituent of the matrix. These rocks are typically gray or black and weather to brown or bluish-black. The pebbly sedarentites can display both normal and inverse graded bedding and parallel alignment of clasts.

Petrography of Selected Sandstones

Nine samples (five from the lower division and four from the upper division) were collected from Blakely sandstone beds for detailed petrographic study. In each sample 100 to 200 grains were identified and measured for size, roundness and sphericity. The texture, cement type and quartz: feldspar: lithic fragment ratio of each rock was also determined and the results tabulated in Table 1. According to Folk's (1974) classification (Fig. 8) the samples include five quartz arenites, two sublithare-

Sample	Formation	Bed Thickness	Grain Size		Q:F:L of Framework Grains (%)	Cement	Texture		Rock Name (Folk Classification)
			Range	Size with most grains			Roundness	Sphericity	
190	Lower LSD	7.0'	Very fine to coarse	Medium	92:0:8	Quartz	Rounded to angular	Subequant to elongate	Sublitharenite
1853	Lower LSD	0.5' to 1.25'	Silt to medium	Fine	96.5:0:3.5	Quartz	Well rounded to subangular	Very equant to Very elongate	Quartz arenite
1853B	Lower LSD	5.0"	Silt to medium	Fine	97:1:2	Quartz	Rounded to subangular	Very equant to Very elongate	Quartz arenite
18531	Upper LSD	8.0'	Very fine to coarse	Medium	95.5:0.5:4	Quartz	Rounded to subangular	Very equant to Very elongate	Quartz arenite
185332	Upper LSD	3.5'	Very fine to coarse	Medium	97.5:0:2.5	Quartz	Well rounded to angular	Very equant to elongate	Quartz arenite
124A	Lower USD	3.3'	Silt to very coarse	Medium	49:0:51	Calcite	Well rounded to very angular	Very equant to Very elongate	Sedarenite
221	Middle USD	1.7'	Very fine to pebbles	Medium	35:0:65	Quartz and calcite	Well rounded to angular	Very equant to Very elongate	Sedarenite
154	Upper USD	3.0"	Silt to fine	Silt	99.5:0.5:0	Quartz	Rounded to angular	Equant to Very elongate	Quartz arenite
72	Upper USD	2.0"	Silt to medium	Very fine to fine	94.5:3:2.5	Quartz	Rounded to angular	Very equant to elongate	Quartz arenite

Table 1. Summary of petrographic analyses of 9 sandstones from the Blakely Sandstone. LSD = lower sandstone division; USD = upper sandstone division. Sample localities listed in Stolarz (1981, Appendix 1).

nites and two litharenites (listed as sedarenites in the table). The photomicrographs in Fig. 7 illustrate the typical textures of two quartz arenites (Figs. 7a and b) and one sedarenite (Fig. 7), a litharenite containing sedimentary rock fragments. Figure 7d shows a large clast of pelletoidal limestone in a calcareous sandstone matrix.

Graptolites and Trace Fossils

Black graptolite-bearing shale collected from the upper part of the middle shale division in the center of the NW¼, SW¼, Sec. 20, T. 2 S., R. 21 W. yielded the following identifiable fossils:

Cryptograptus antennarius
Cryptograptus tricornis mut. insectiforms
Didymograptus cf. *D. euodus*
Dictyonema ritilineatum
Phyllograptus anna
Phyllograptus nobilus
Tetragraptus cf. *T. quadribrachiatus*

The age common to all of the above species and, therefore the age of the upper part of the middle shale division is lower Middle Ordovician. This agrees with the age determined by Berry (1960) and Ketner (1980), but is slightly younger than that determined by Miser and Purdue (1929) and Repetski and Ethington (1977).

A very low diversity assemblage of trace fossils was found at the tops and bottoms of some thin sandstone beds. The trace fossils that could be identified are the horizontal feeding burrows *Planolites.* Buthman (1982a) has reported the horizontal feeding burrow *Nereites* in a shale sample from the Blakely Sandstone. These two fossils suggest a *Nereites* (bathyal to abyssal) ichnofacies (Frey and Seilacher, 1980).

Womble Shale

The Womble Shale crops out across the southern part of the map area in gradational

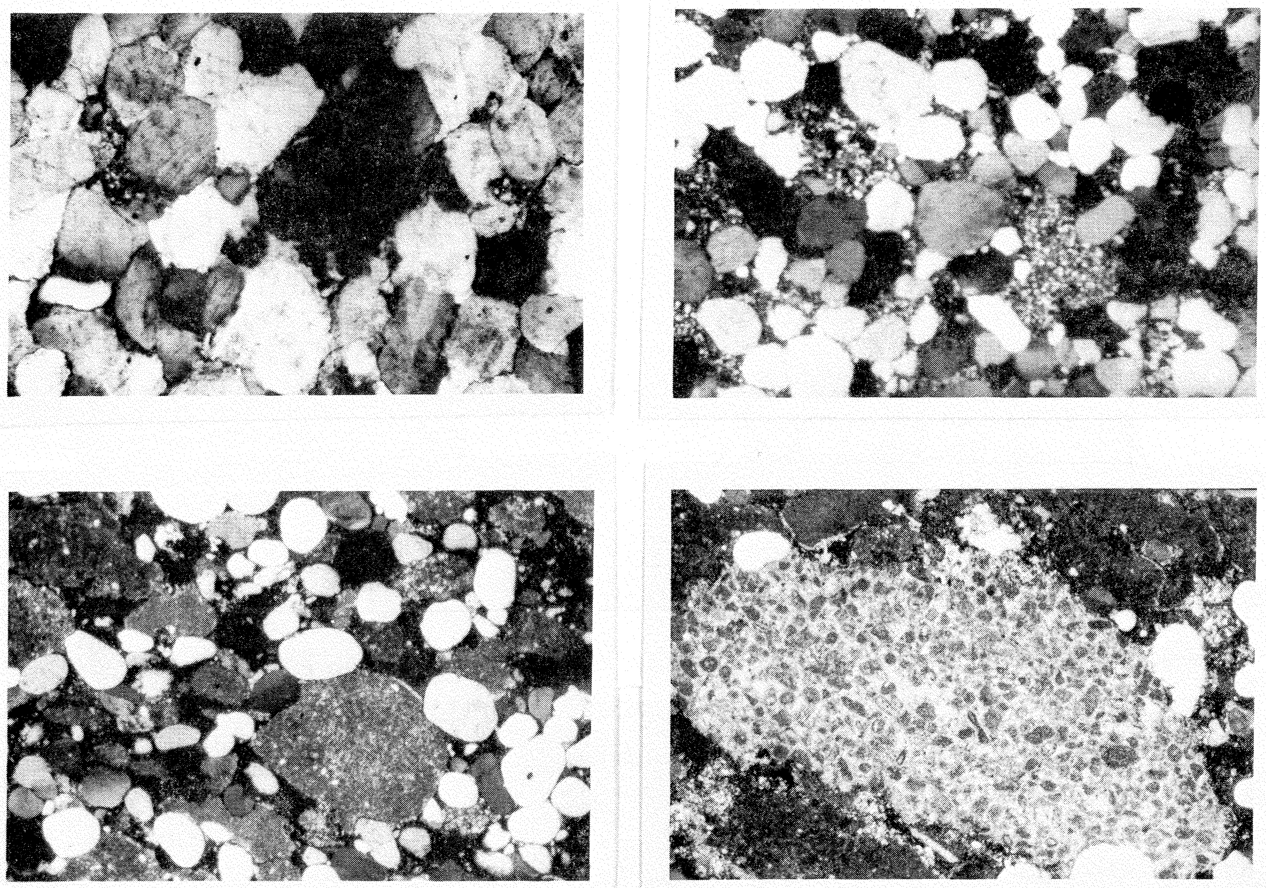

Figure 7. Photomicrographs of sandstones from the Blakely Sandstone. a) Quartz arenite with very fine to medium quartz grains, quartz overgrowths, and strong undulose extinction; sample 1835B, 40X. b) Quartz arenite with well rounded, very fine to coarse quartz grains and chert fragments; sample 18351, 20X. c) Sedarenite (litharenite) composed of well rounded quartz grains and angular to rounded micritic limestone fragments; sample 221, 20X. d) Large clast of pelletoidal limestone cemented with sparry calcite and surrounded by grains of well rounded quartz and micritic limestone fragments; 20X. All photomicrographs taken under crossed-nichols.

contact with the underlying Blakely Sandstone. The formation consists of about 3,500 feet (1067) m) of dark shale with minor amounts of interbedded sandstone, siltstone and limestone (Stone and others, 1973).

Although dark shale, typically containing thinly laminated to thinly bedded gray siltstone, is volumetrically dominant, some gray-green to black banded shale is present toward the lower parts of the formation (Miser and Purdue, 1929). Banded shales are not as abundant in the Womble as they are in the Mazarn and Blakely Formations.

Sandstone beds can be found throughout the Womble but appear to be concentrated near the base of the formation. They are typically gray, quartz-rich, very fine to medium grained, thin to medium bedded, and locally lensoidal.

Black, micritic, siliceous, thin to thick bedded, locally lenticular limestones interlayered with black, calcareous shales are concentrated in the upper 60 to 80 feet (18 to

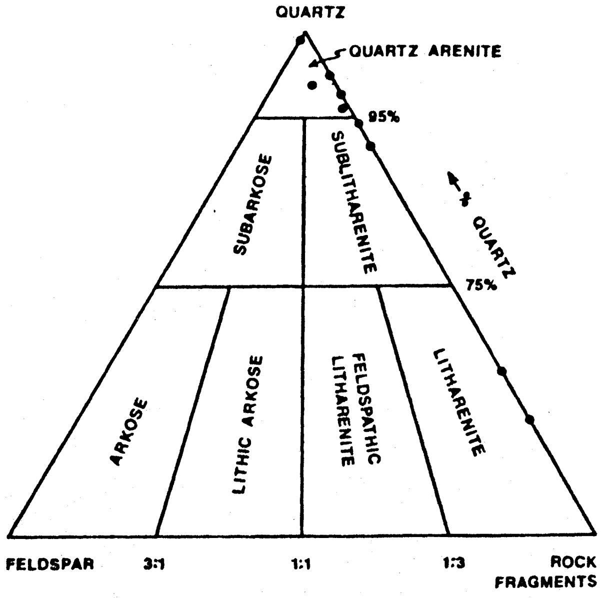

Figure 8. Ternary diagram of sandstone compositions from the Blakely Sandstone. Classification after Folk (1974).

25 m) of the Womble Shale. Carbonate beds of similar composition occur sporadically in the lower parts of the formation in which shale is the volumetrically dominant rock type.

Based on their analysis of conodont and graptolite assemblages, Repetski and Ethington (1977) assigned a Lower to Middle Ordovician age to the Womble Shale. Berry (1960) and Ketner (1980) have suggested that the age of the Womble is Middle Ordovician. The latter determination better supports our suggestion of a lower Middle Ordovician age for the Blakely Sandstone.

ENVIRONMENTS OF DEPOSITION

Two distinct environments have been suggested for the deposition of the Blakely Sandstone. Various workers (Miser and Purdue, 1929), (Brieva, 1963; Davies and Williamson, 1976 and 1977) have interpreted the formation to be a shallow water marine deposit. Specifically, Davies and Williamson suggested that the rocks were deposited as sublittoral sheet sandstones on a shallow marine shelf. In contrast, other workers (Haley and Stone, 1973; Keller and Cebull, 1973; Morris, 1974) determined the Blakely Sandstone to consist of proximal turbidites or watery slides deposited on an unstable slope in deeper water.

Davies and Williamson (1976) state that the sandstone beds do not show abundant sedimentary characteristics indicative of turbidity currents, fludized sediment flows, or grain flows citing the low percentage of graded beds and Bouma sequences and the abundance of sharp upper surfaces on individual beds. They also indicate that the Ordovician sediments of the Ouachita Mountains may have been deposited some 310 miles south of the northern paleoshoreline of the Ouachita trough. For this reason, they suggest that the Blakely Sandstone must have had a southern source and a paleoshoreline that lay south to southwest of the present Ouachita Mountains.

Morris (1974) did not cite detailed sedimentological evidence for a depositional environment but indicated that parts of the Blakely Sandstone very closely match Walker's (1967) description of proximal turbidites and, therefore, probably have a similar origin. Morris (1974) also stated that the source of the Ordovician sediments was the cration to the north of the Ouachita trough.

We have considered both depositional models and tend to favor a deep water turbidite and submarine fan model for the following reasons:

1. **Paleocurrent data.** Figure 9 is a rose diagram and geologic map showing the distribution of paleocurrent directions in the Mazarn Shale and Blakely Sandstone after the beds were rotated to a horizontal and stratigraphically upright position. The paleocurrents trend from southeast to southwest and suggest a northern source and/or southward transport for the sediments. These data favor Morris' (1974)

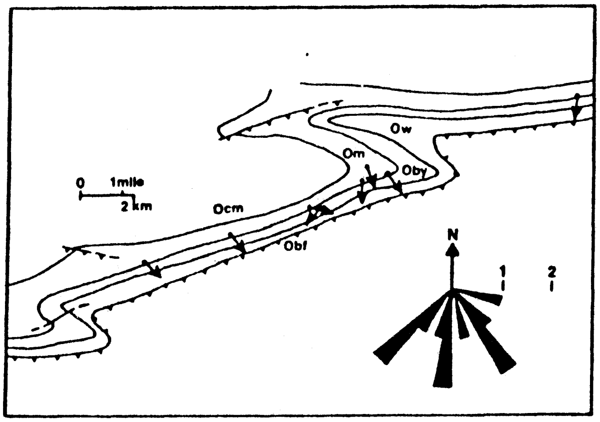

Figure 9. Areal distribution of paleocurrent directions from flute casts and forset crossbeds in the Mazarn and Blakely Formations.

northern sediment source and turbidite depositional model over those that require a southern sediment source (also see Buthman, 1982, Figs. 7 and 8).

2. **Sole marks.** Some thin-to-medium bedded sandstones in the map area contain sole marks such as flute and groove casts. Selley (1976) stated that such sole marks are characteristic of, but not restricted to, turbidite facies.

3. **Presence of possible partial Bouma sequence divisions.** Many of the thin-bedded sandstones and siltstones in the map area display parallel laminations and/or ripple crossbeds possibly representing divisions B and C of a Bouma sequence (Fig. 4) in distal turbidites. Some beds show a vertical change from parallel laminations to micro-cross laminations and rippled beds.

The thicker, massive sandstones of the Mazarn and Blakely Formations may represent division A (massive and graded sandstone beds) of a Bouma sequence, based on the occasional presence of graded subunits and load casts. These rocks may have been deposited as proximal turbidites, and the sharp upper contacts of numerous sandstone beds could be attributed to winnowing by bottom currents.

4. **Presence of possible grain-flow and debris-flow deposits.** Rocks interpreted as grain-flow and debris flow deposits (Middleton and Hampton, 1976) have been found at several locations in the map area, particularly in the upper sandstone division of the Blakely Sandstone. These rocks may represent gravity induced sediment flows down a submarine canyon or other paleoslope.

5. **Facies variations.** Figure 6 shows the inferred lateral and vertical facies variations of the major sandstone units in the upper Mazarn Shale and Blakely Sandstone. This figure is an east-west cross section looking north at an oblique angle towards the suggested paleoslope of the Ouachita trough. The facies variation depicted could be the result of coalescing and overlapping submarine fans or turbidite deposits.

6. **Trace Fossils.** The trace fossil *Nereites* found by Buthman (1982a) indicates a deep-water *Nereites* ichnofacies (Frey and Seilacher, 1980).

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REFERENCES

Arkansas Geological Commission, 1977, Geologic map of the Ouachita National Forest: Little Rock, Arkansas Geol. Commission, scale 1:250,000.

Berry, W. B. N., 1960, Graptolite faunas of the Marathon Region, West Texas: Texas Bureau of Economic Geology Pub. 6005, 179 p.

Brieva, J. A., 1963, Petrology of the Mazarn and Blakely Formations, Montgomery County, Arkansas: M. S. Thesis, University of Tulsa, 89 p.

Buthman, D. B., 1982a, Stratigraphy and structural style of the Mazarn, Blakely, and Womble Formations in the Ouachita core near Norman, Arkansas: M. S. Thesis, Southern Illinois University at Carbondale, 125 p.

————, 1982b, Stratigraphy and structural style of the Mazarn, Blakely, and Womble Formations in the Ouachita core near Norman, Arkansas, Part I: Journal of the Oklahoma City Geological Society (Shale Shaker), v. 33, No. 3, p. 20–29.

————, 1982c, Stratigraphy and structural style of the Mazarn, Blakely and Womble Formations in the Ouachita core near Norman, Arkansas, Part II: Journal of the Oklahoma City Geological Society (Shale Shaker), v. 33, No. 4, p. 31–43.

Davies, D. K., and Williamson, E. A., 1976, Shallow-marine origin for Ordovician rocks of the Ouachita Mountains, Arkansas: Geology, v. 4, p. 340–344.

————————————, 1977, Paleoenvironments and paleobathymetry of lower Paleozoic Crystal Mountains and Blakely Formations, Ouachita Mountain core, *in* Stone, C. G., ed., Symposium on the geology of the Ouachita Mountains: Little Rock, Arkansas Geological Commission, v. 1, p. 115–131.

Folk, R. L., 1974, Petrology of Sedimentary Rocks: Austin, Hemphill Publishing Company, 182 p.

Frey, T. W., and Seilacher, A., 1980, Uniformity in marine invertebrate ichnology: Lethaia, v. 13, p. 183–207.

Haley, B. R., and others, 1976, Geologic map of Arkansas: Little Rock, Arkansas Geological Commission and U. S. Geological Survey, scale 1:500,000.

Haley, B. R., and Stone, C. G., 1973, Paleozoic stratigraphy and depositional environments in the Ouachita Mountains, Arkansas: Geol. Soc. America Abs. with Program, v. 5, p. 259–260.

Keller, G. R., and Cebull, S. E., 1973, Plate tectonics and the Ouachita system in Texas, Oklahoma, and Arkansas: Geol. Soc. America Bull., v. 84, p. 1659–1666.

Ketner, K. B., 1980, Stratigraphic and tectonic parallels between Paleozoic geosynclinal siliceous sequences in northern Nevada and those of the Marathon Uplift, Texas, and Ouachita Mountains, Arkansas and Oklahoma, *in*, Fouch, T. D. and Magathan, E. R., eds. Paleozoic paleogeography of the west-central United States: Rocky Mountain Paleogeography Symposium 1: Soc. Economic Paleontologists and Mineralogists, Rocky Mountain Section, P. 363–369.

Middleton, G. V., and Hampton, M. A., 1976, Subaqueous sediment transport and deposition by sediment gravity flow, *in* Stanley, D. J., Swift, D. J. P., eds., Marine sediment transport and environmental management: Wiley-Intersci. Publ., New York, p. 197–218.

Miser, H. D., 1959, Structure and vein quartz of the Ouachita Mountains of Oklahoma and Arkansas, *in*, Cline, L. M., Hilseweck, W. J., and Feray, D. E., eds., The geology of the Ouachita Mountains: A symposium: Dallas and Ardmore Geol. Socs., p. 30–43.

Miser, H. D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap Quadrangles, Arkansas: U. S. Geological Survey Bull. 808, 195 p.

Morris, R. C., 1974, Sedimentary and tectonic history of the Ouachita Mountains, *in* Dickinson, W. R., ed., Tectonics and sedimentation: Soc. Economic Paleontologists and Mineralogists Spec. Pub. no. 22, p. 120–142.

Pitt, W. D., Cohoon, R. R., Lee, H. C., Robb, M. G., and Watson, J., 1961, Ouachita Mountain core area, Montgomery County, Arkansas: Am. Assoc. Petroleum Geologists Bull., v. 45, p. 72-94.

Repetski, J. E., and Ethington, R. L., 1977, Conodonts from graptolite facies in the Ouachita Mountains, Arkansas, *in* Stone C. G., ed., Symposium on the geology of the Ouachita Mountains, Vol. 1: Little Rock, Arkansas Geological Commission, p. 92–106.

Selley, R. C., 1976, An introduction to sedimentology: New York, Academic Press, 408 p.

Sommer, N. A., 1971, Structural analysis of the Blakely Mountain area, Arkansas: Ph.D. Dissertation, Univ. of Missouri at Columbia, 126 p.

Soustek, P. G., 1979, Structural style of the Ouachita core in a portion of the McGraw Mountain quadrangle, Arkansas: M. S. Thesis, Southern Illinois University at Carbondale, 132 p.

Stolarz, R. J., 1982, Structure and stratigraphy of the Blakely Sandstone in eastern Montgomery and western Garland Counties, Arkansas: M. S. Thesis, Southern Illinois University at Carbondale, 95 p.

Stone, C. G., Haley, B. R., and Viele, G. W., 1973, A guide book to the geology of the Ouachita Mountains, Arkansas: Little Rock, Arkansas Geological Commissio 114 p.

Stone, C. G., McFarland, J. D., and Haley, B. R., 1981, Field guide to the Paleozoic rocks of the Ouachita Mountain and Arkansas Valley Provinces, Arkansas: Little Rock, Arkansas Geological Commission, 140 p.

Walker, R. G., 1967, Turbidite sedimentary structures and their relationships to proximal and distal depositional environments: Journal Sedimentary Petrology, v. 37, p. 25–43.

THE UPPER JACKFORK SECTION, MILE POST 81, I–30

ARKADELPHIA, ARKANSAS

By

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Charles G. Stone[3]

INTRODUCTION

A readily accessible 440 foot section of the upper Jackfork Sandstone crops out along Interstate 30 near Arkadelphia, Arkansas. This section is frequently visited by geology field trip groups because it is an excellent example of a Ouachita facies turbidite sequence. Despite its renoun, it has never been described in detail.

This study presents a measured stratigraphic section of this exposure, and an interpretation and discussion of its depositional environment. The section is subdivided by the megascopic appearance of rock type, grain size, sedimentary structures, vertical distribution of bed thicknesses and cyclicity. Based on these sedimentary attributes subdivisions are then assigned to an appropriate submarine fan subenvironment.

The outcrop is on the west side of Interstate 30 between Exit 83 (Friendship) and Exit 78 (Arkadelphia, Hot Springs) at the Mile Post 81 marker point (Fig. 1). Although there is an exposure on the opposite side of the interstate, it has less section and is not described here. The outcrop, hereafter called "Mile Post 81", is part of the "Friendship Section", a term used among local geologists.

This section has received special attention in recent years by the Shell Development Company and more recently other companies in their training research programs studying deep-water clastic deposits. This led to a report in 1975 by Thomson and LeBlanc who first briefly described submarine fan channels (5 distinct) at the Mile Post 81 section.

The Jackfork Sandstone is Pennsylvanian in age (Miser, 1934 and Gordon and Stone, 1977). It overlies the Mississippian age Stanley Shale and is overlain by the Pennsylvanian age Johns Valley Shale (Fig. 2a). The Jackfork forms a portion of a thick sequence of Carboniferous age clastic sediments that was deposited as a shelf to slope complex in the Ouachita Trough and environs (Fig. 2b). Cline (1960), Walthall (1967), Morris (1971a, 1971b, 1974a, 1974b, 1977a, 1977b), Chamberlain (1971), Thomson and LeBlanc (1973), Graham et al. (1976), Lock (1979), Stone et al. (1981), and many others have established that the Ouachita facies represent deep marine continental slope rise sediments deposited largely by submarine fan processes.

GEOLOGY OF THE SECTION

The Mile Post 81 section has a true vertical thickness of 440 feet. It dips 43 degrees to the south and strikes N78E. The entire section is in the upper Jackfork Sandstone, although neither top nor bottom formation contacts are present at this exposure. Boyd Haley (personal communication) indicates that this section lies stratigraphically 700 feet above the DeGray

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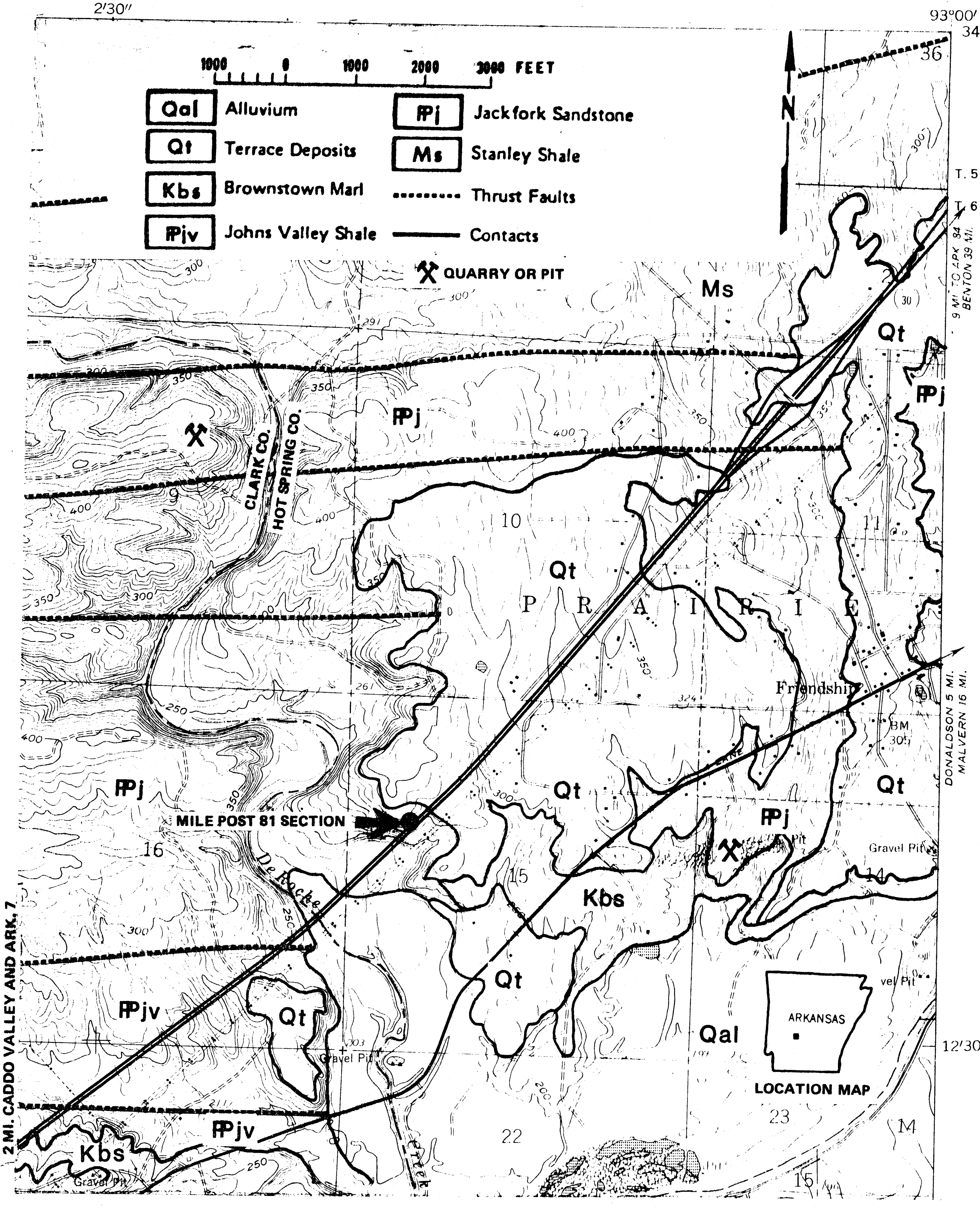

Figure 1. A geologic map of the Friendship, Arkansas area showing the location of the Mile Post 81 section. Geology slightly modified from Haley et al. (1976).

Dam Spillway section located 10 miles to the west, and 500--600 feet below the Johns Valley Shale.

A measured columnar section of the Mile Post 81 outcrop is presented in Figure 3. Two lithologies dominate the section, dark gray to blue-gray shales and buff to light brown sandstones. Measurement started at the northeastern end of the outcrop at the base of a thick massive bed of conglomeratic sandstone that contains abundant shale clasts towards the top. This begins what we have interpreted to be 9 submarine fan channels and one depositional lobe. Each fan channel is composed of at least two of the following stratigraphic subdivisions, active channel fill, partially abandoned channel fill, and abandoned channel fill. This interpretation was made following a close examination of lithology, vertical distribution of bed thicknesses, and sedimentary features. The latter allowed an evaluation of the relative energy of the depositional environment.

The top of the measured section is to the southwest and is marked by prominent reddish breccia along a fault plane. Approximately 50 ft. (15 m) of somewhat distorted sandstones and shales are present above this fault. Two intermittent springs are also present, one at 130 ft. (40 m) and a second at 315 ft. (96 m) above the base. Six reverse faults are observed but these appear to be of local extent. Two prominent joint sets are perpendicular to bedding and form a diamond pattern on exposed bedding surfaces.

RECOGNITION OF ENVIRONMENTS

The character and vertical variation of lithologies at Mile Post 81 is readily interpretable using Walker's (1978) model (Fig. 2c). This model combines the pioneer work of Bouma (1962) with recent stratigraphic studies of Mutti (1974) and Ricci Lucchi (1975) and the work on modern depositional systems such as that of Normark (1978). The importance of this model in stratigraphic studies is illustrated by the work of Wilde et al. (1978). Steeply sloping submarine canyons are incised into the slope-shelf adjacent to the area of submarine-fan deposition and are floored with slumped debris flow deposits. Sediment charged currents that spasmodically surge out of these canyons initially erode the underlying soft sediments and then, as their energy wanes, deposit a sequence of sediments whose grain size and sedimentary structures depend on where in the fan the sediments ultimately come to rest. The lateral pattern illustrated in Figure 2c migrates with time, building the fan upwards and forming a vertical succession of facies that can be used to interpret which

Figure 2. A. Carboniferous stratigraphy of the central Arkansas area. Only the upper portion of the Arkansas Novaculite is represented. No vertical scale. After Gordon and Stone (1977).

B. Idealized map showing regional depositional patterns in the Jackfork Sandstone (without significant paleogeographic reconstruction). The position marked 81 is the approximate location of the Mile Post 81 section, i. e. on the mid-fan close to the inner (upper) fan. The location marked BR is the approximate position of the Big Rock Quarry location, another frequently visited fan area. Revised by Stone et al. (1981) from M. T. Roberts, CITGO Field Guide (1980).

C. A schematic model for submarine fan deposition showing the lateral and vertical facies distribution (Walker, 1978). The probable position of the Mile Post 81 location is inside the marked box, i. e. in the area of suprafan deposition on the mid-fan area.

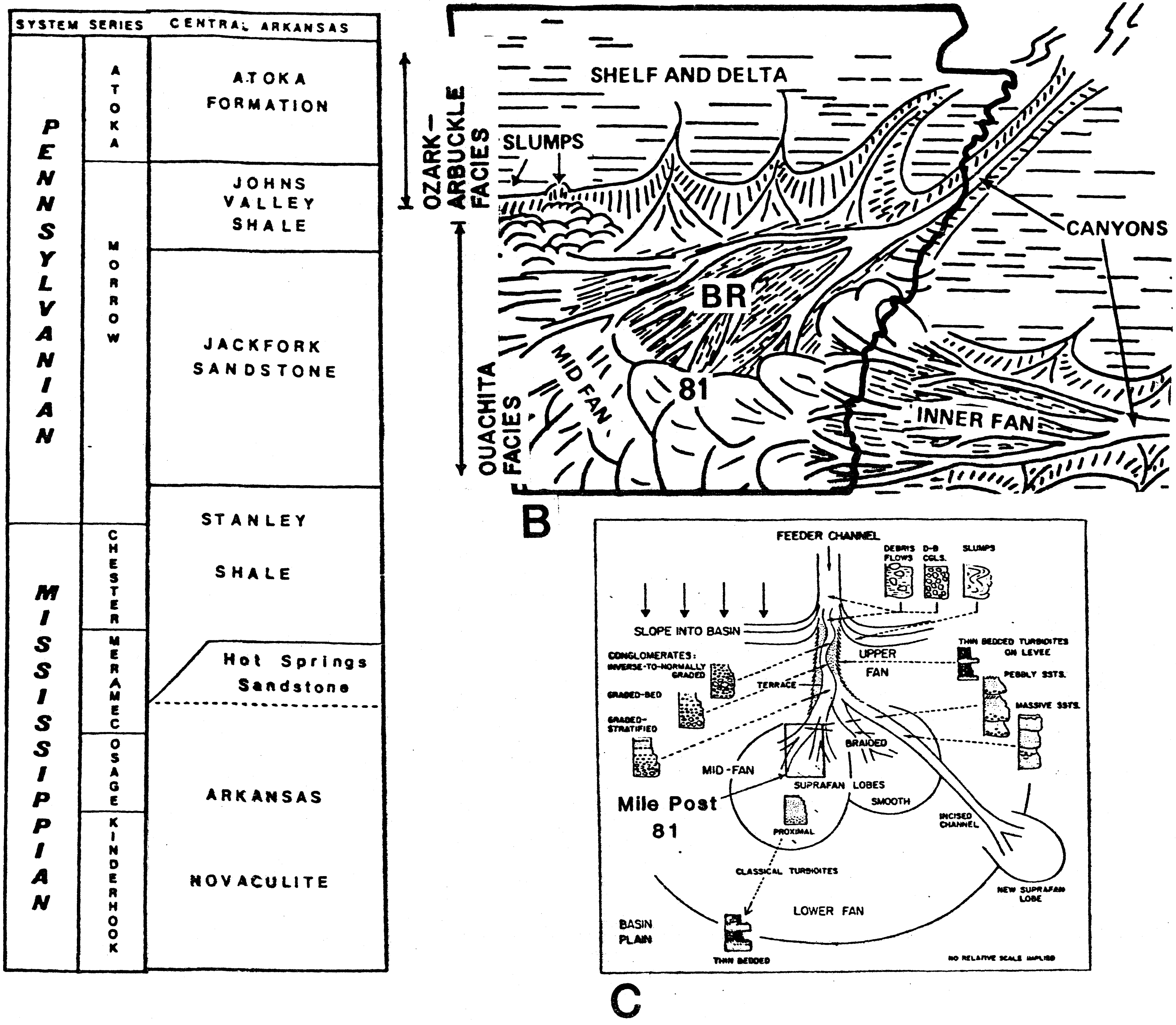SYSTEM
SERIES
CENTRAL ARKANSAS
PENNSYLVANIAN
ATOKA
MORROW
ATOKA FORMATION
JOHNS VALLEY SHALE
JACKFORK SANDSTONE
MISSISSIPPIAN
CHESTER
MERAMEC
OSAGE
KINDERHOOK
STANLEY SHALE
Hot Springs Sandstone
ARKANSAS NOVACULITE
OZARK—ARBUCKLE FACIES
OUACHITA FACIES
SHELF AND DELTA
SLUMPS
CANYONS
BR
81
MID FAN
INNER FAN
B
FEEDER CHANNEL
DEBRIS FLOWS
D-B CGLS.
SLUMPS
SLOPE INTO BASIN
THIN BEDDED TURBIDITES ON LEVEE
UPPER FAN
CONGLOMERATES: INVERSE-TO-NORMALLY GRADED
GRADED-BED
GRADED-STRATIFIED
TERRACE
PEBBLY SSTS.
MASSIVE SSTS.
BRAIDED
MID-FAN
SUPRAFAN LOBES
SMOOTH
Mile Post 81
PROXIMAL
INCISED CHANNEL
CLASSICAL TURBIDITES
NEW SUPRAFAN LOBE
LOWER FAN
BASIN PLAIN
THIN BEDDED
NO RELATIVE SCALE IMPLIED
C

Figure 2

portion of the fan occupied the outcrop location at a particular time.

Each submarine fan channel may be generally described as a fining upward sequence of interbedded sandstone and shale. Active channel fill which occupies the base of each fan channel is characterized by thick massive sandstone units containing sedimentary features indicative of high energy turbidity flow. From the base toward the top of each fan channel the massive sandstone units thin and become interbedded with siltstone and shale. These interbedded deposits, the accumulation of partially abandoned channel fill, reflect a decrease in the sand carrying capacity of the currents and an increase in the amount of silt and clay material supplied to the environment. Finally, the uppermost portion of the fan channel is the abandoned channel fill. It is recognized by thick accumulations of fissile shale with thin siltstone and iron carbonate layers. Thick turbidite sandstones are virtually absent in this portion of the fan.

Within the active channel fill individual sandstone beds with sharp upper and lower contacts are commonly stacked in thicknesses of 20–30 ft. *6–9 m) (e. g. at 55 to 75 ft., 175 to 195 ft., 265 to 275 ft.) (Figs. 3 and 4). Bedding is typically planar, but may have considerable relief (Fig. 5). This relief (e. g. at 62 ft.) is most likely due to scour, indicating that each sand bed was deposited by a single advancing current. Irregular ripples (at 265 ft. and 365 ft.) and large enigmatic "bowl structures" (at 295 ft. and 365 ft.) are also present.

High energy erosional features at lithologic contacts and or bedding planes are labeled reactivation surfaces (Fig. 3). These surfaces are evidence of a sudden change in the erosional and/or depositional power of the turbidity current. Deposition was in a sense "reactivated" from nonexistence or a slower rate by the introduction of one advancing current. This sudden surge of suspended sediment eroded a fine undisturbed substrate, leaving scour marks to be quickly filled by new material. The lateral extent of each reactivation surface, although unseen, directly corresponds with the lateral extent of each turbidity current.

Although shale beds are rather sparse in accumulations of active channel fill, some sandstone beds do contain large shale clasts. These are clasts of partially consolidated shale substrate that were ripped up, transported and eventually redeposited within the body of the current transported sediment. During transport they acquired their roundness and present shape. At the Mile Post 81 exposure (at 20, 70 and 90 ft.) many shale clasts themselves have been removed by weathering and erosion, and only molds remain (Figs. 5 and 6). They are typically located toward the top of a sandstone bed and are up to 6 inches in diameter.

Also occurring above some of these sandstones and grading upward into the shales are rather numerous intervals of chaotic sandstone and shale often with carbonized plant fragments that are considered of debris flow origin. These beds are called "blue beds" for their characteristic fresh color, but upon weathering they become light brown.

Sandstones of the active and partially abandoned channel fill are medium to fine grain, moderately sorted sublitharenites. They form approximately 58% of the Mile Post 81 section. Petrographic analysis of selected samples reveals that the sandstone composition averages 86% grains, 4% matrix, 4% quartz cement and 6% spores. Grain composition averages 92% quartz with minor amounts of both rock fragments and muscovite mica. Feldspar is noticeably absent in the thin sections examined.

The number and thickness of shale beds noticeably increases in the upper portion of

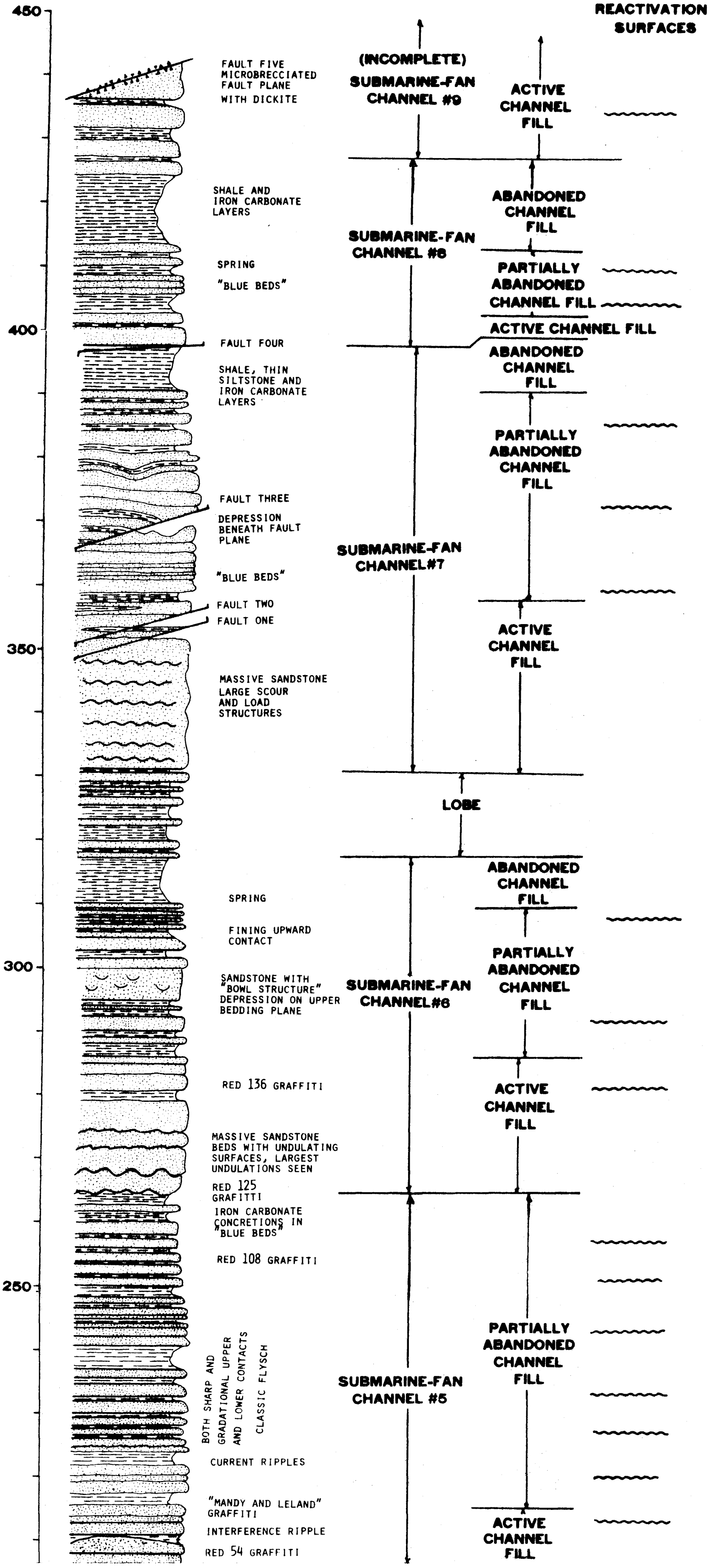

Figure 3. A detailed columnar section with interpretation of submarine-fan depositional environments of the Mile Post 81 section. The labeled subsequences represent subjective interpretations of the pattern of vertical variation in the sandstone/shale ratio and are intended as guides to interpretation, not as formal unit designations. See text for a discussion.

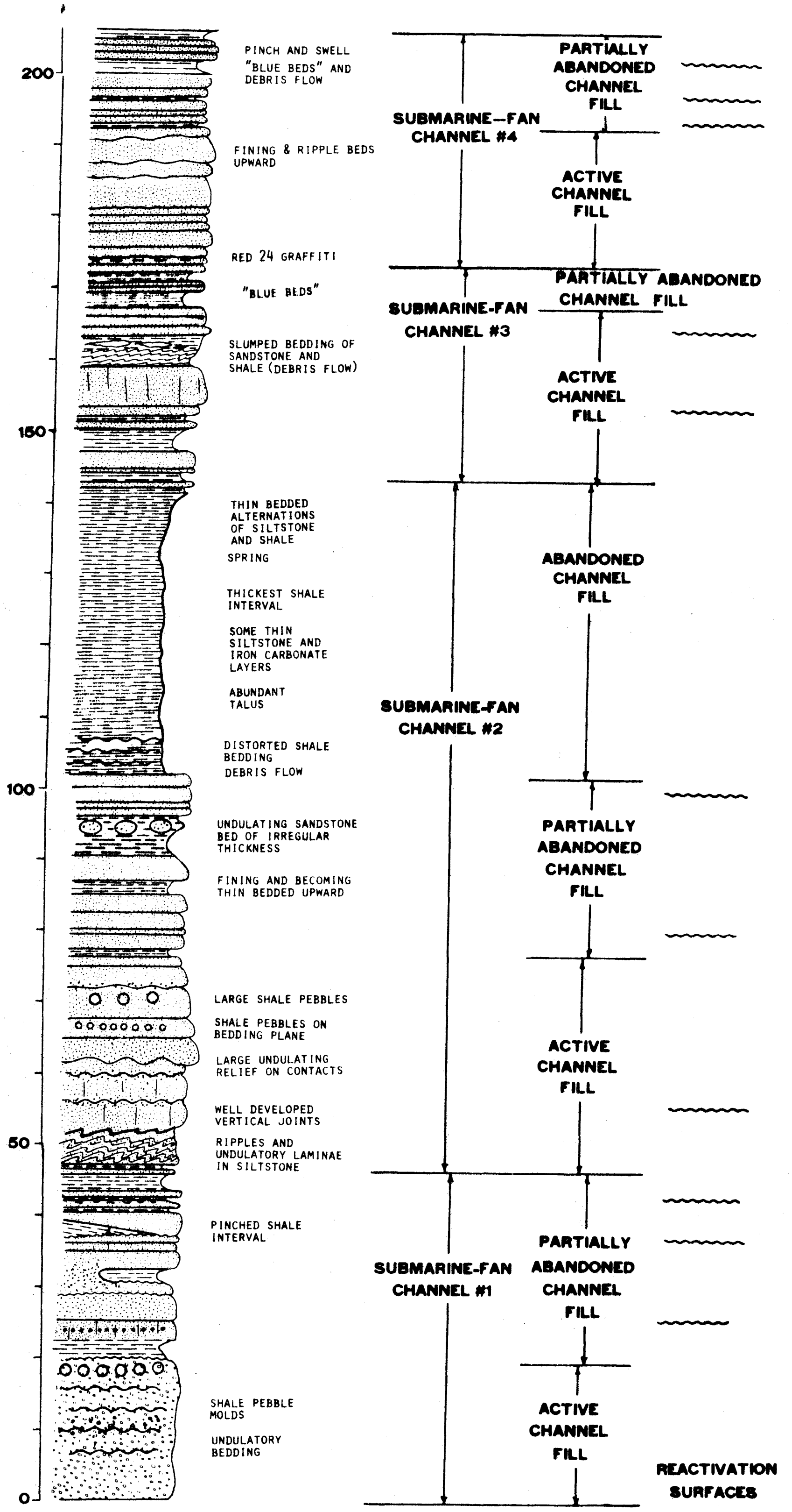

Figure 3. continued.

Figure 4. A sandstone "packet" 52 to 75 ft (16 to 23 m) above the base of the section. Note the absence of shale, the rather massive bedding, lack of internal sedimentary features, and sharp, undulatory contacts between adjacent beds.

Figure 5. Undulatory relief on the contacts between sandstone beds. In this case both the top and bottom contacts have relief of up to 4 inches (10 cm). The mold of a large shale clast is to the left, a partial mold still containing most of a shale clast is at the end of the hammer handle. Illustration of a unit 73 ft (22 m) above the base of the measured section.

Figure 6. Shale clast molds on the upper surface of a sandstone bed 67 ft (20 M) above the base of the section.

each submarine fan channel. These fine grained quartz rich shales form 42% of the measured section. They can be broadly divided into two groups; 1) thin bedded shales and 2) thick intervals of fissile shales. Differences between the two are based on sedimentary features, thickness, bedding and grain size.

Thin bedded shales occur as the upper portions of graded beds. More specifically, 1 to 2 ft. thick dark gray beds fine upwards from sandstone to silty shale to thin bedded shale. Deposits of partially abandoned channel fill are identified by these graded beds. Some exhibit partial Bouma sequences (Bouma, 1962), with A, B, C, and D divisions developed (Figs. 8 and 9). The base of the sandstones are typically in sharp contact with the subjacent

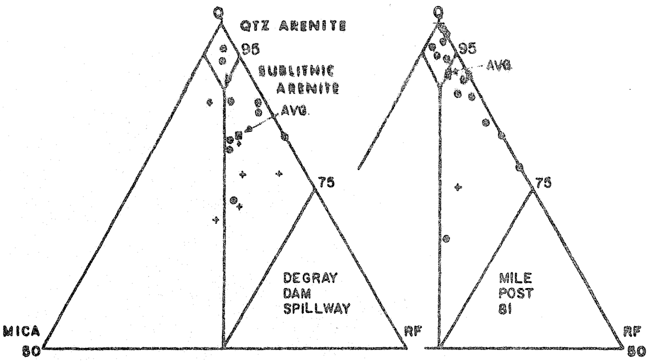

Figure 7. Ternary diagrams of the grain composition of sandstones in the upper Jackfork Sandstone at Mile Fost 81 and the nearby location at the DeGray Dam Spillway. Only the upper portion of the triangles are shown. Mica forms the left pole rather than the customary feldspar. Plus signs mark samples that have 10% matrix. Although the overall composition at the two locations is similar, the differences in average composition suggest a somewhat different source.

shales and often exhibit sole marks. These sole marks are much smaller in size however than those of the active channel. Cross laminations are also present but diminish as each bed fines upward. Current and interference ripple marks may be seen where bedding planes of silty shale are exposed (Fig. 10).

The graded beds were deposited by weaker turbidity currents or those with a lower velocity and sand carrying capacity than their active channel counterparts. Sandstones in the partially abandoned channel of each fan are therefore finer grained and thinner than those of the active channel. Waning turbidity currents with a large suspended silt and clay volume thus produced the graded bedding.

The second group; dark gray fissile shales are found in continuous thicknesses of ten feet or more (e. g. 100–140 ft.). These shales are easily recognized because they weather to splinters which form large accumulations of talus at the base of each exposure. Interbeds of slightly more resistant siltstone and siderite are common. A low energy marine environ-

Figure 8. Sandstones interbedded with silty shales at 200 ft (61 m) above the base of the measured section. Noteworthy are the sharp contacts between the sandstones, the interbedded silty shales, and the disrupted appearance of the silty shales.

Figure 9. A sandstone layer fining and becoming ripple marked upward at 188 ft (57 m) above the base of the measured section.

Figure 10. The upper surface of a sandstone bedding plane marked by large interference ripple marks. This unit is 209 ft (64 m) above the base of the section. Jointing has resulted in a rectangular edge to the bed.

ment relatively out of the path of turbidity current flow is believed to be the site of deposition for these thick shales. Very slow hemipelagic sedimentation over a long period was responsible for much of their accumulation.

Between 320 and 330 ft. an interval of coarsening upward sediment is present. Thick, dark shales are interrupted by several thickening upward sand layers. We have assumed this thickening pattern is that of an anomalous depositional lobe.

SUBMARINE FAN DEPOSITION

Based on this examination of the Jackfork Sandstone at Mile Post 81 we find that channel sedimentation has occurred in 9 cycles, and each cycle can be interpreted in terms of fan processes. Each identified submarine fan channel preserves a progression from high energy to low energy deposition and a sediment record of vertical aggradation within each channel. Thus, the relative amount of turbidity current energy has controlled not only the distribution of coarse and fine grain sediment, but the vertical succession of thick and thin beds within each channel.

Coarse grain deposition took place in the active channel over the flow paths of the turbidity currents. Thick massive sandstone was deposited with each successive current. The lateral continuity of each bed was controlled by the width and bottom topography of the channel while the volume of sand entrained by the turbidity flow was responsible for the variation of resultant layer thicknesses. Other variables such as channel slope, frequency of turbidity currents and proximity to current source also influenced the final stratigraphy produced.

Each submarine fan channel may be considered analagous to a subaerial distributary channel of a delta. As sediment fills each distributary channel, the channel gradually loses efficiency as a sediment conduit and avulsion or channel switching takes place upslope. Turbidity currents follow a more efficient flow path and the original channel is abandoned. Each submarine fan channel at Mile Post 81 has experienced this same history of vertical aggradation, avulsion and final abandonment. In the stage of final abandonment the channel is no longer in the main flow path of the turbidity currents. Locally suspended fines and pelagic material, possibly from nearby currents, accumulate without interruption for an unknown amount of time. Then, given the right conditions, the abandoned channel is reactivated or a new channel is formed and the cycle is repeated. As mentioned the new cycle is marked by a reactivation surface or a surface of erosional scour and an abrupt change from thick shale beds to coarse massive sandstones.

As seen in channels 1, 3, 4, and 5 (Fig. 3) abandoned channel fill sections are absent and the active channel massive sandstones rest directly on partially abandoned channel fill. This may be caused by either simple non-deposition or powerful erosion of the abandoned channel fill shales by the reactivating turbidity current. In either case this points out that the abandoned channel fill is not a standard component of the system and there is no ideal sedimentary cycle expected in submarine fan channel deposits. Although fining and thinning upward is expected, each channel will vary according to numerous geologic factors external to the environment of deposition. Trends in bed thickness, grain size, sedimentary structures, etc. are recognizable however, and allow us to find order and meaning in the section.

We assume that the interval between fan channels 6 and 7 is that of a depositonal lobe. Several thickening and coarsening upward sand layers within this interval do not follow the same cycle of channel deposition observed in the rest of the section. As defined by Mutti and Ghibaudo (1972) depositional lobes are broad lenticular non-channelized sandstone bodies enclosed by thinbedded shale fringe deposits. They also commonly occur as thickening upward cycles superimposed on channel deposits. Since only one such anomalous interval was observed at Mile Post 81, admittedly there may be some debate as to its origin.

The Mile Post 81 section is a vertical record of lateral channel migration across the surface of a submarine fan. Figure 2c from Walker (1978) presents a schematic diagram of the lateral distribution of environments of a submarine fan, implying no relative scale. After studying the sedimentary section at Mile Post 81 in relation to these facies association presented by Walker (1978), we believe that the 9 submarine fan channels and one depositional lobe most probably were deposited in the mid-fan, possibly suprafan lobe, environment. Both the fining and thinning upward of sand and shale beds throughout the section are consistent with Walker's mid-fan facies association. Also, the upper portion of the mid-fan is where distributaries begin forming from the upper-fan feeder channel. This is a most likely location for channel switching to take place, ultimately triggering lateral channel migration and renewed deposition over abandoned channels downslope.

SUMMARY

A 440 foot stratigraphic section of the Pennsylvanian Jackfork Sandstone near Arkadelphia, Arkansas was measured and described. This section is an excellent example of a Ouachita facies turbidite sequence. We find the section to be composed of 9 submarine fan channels and one depositional lobe. Each channel is subdivided into at least two of the following stratigraphic subdivisions: active channel fill, partially abandoned channel fill, and abandoned channel fill, and each may be simply described as a fining upward sequence of interbedded sandstone and shale. Within each channel the sedimentary units fine and thin upward indicating a progression from high energy, high sand volume to low energy, low sand volume turbidity currents moving through the area. Deposition within each channel continued until the channel could no longer serve as an efficient sediment conduit. Channel switching then took place upslope and the original channel was abandoned.

The Mile Post 81 section is a vertical record of lateral channel migration across the surface of a submarine fan. According to the lateral distribution of submarine fan environments presented by Walker (1978), we believe the section at Mile Post 81 was deposited in a suprafan lobe environment of the mid-fan. Deposition in this mid-fan environment has occurred in 9 observable cycles, each representing one submarine fan channel and each sharing common sedimentary attributes of lithology, patterns of bed thickness, and sedimentary features.

ACKNOWLEDGEMENTS

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REFERENCES

Bouma, A. H., 1962, Sedimentology of Some Flysch Deposits, Elsevier Publishing Company, 168 p.

Chamberlain, C. K., 1971, Bathmymetry and paleoecology of the Ouachita geosyncline of southeastern Oklahoma as determined from trace fossils: Am. Assoc. Petroleum Geologists Bull., v. 55, 34–50.

Cline, L. M., 1960, Stratigraphy of the late Paleozoic rocks of the Ouachita Mountains, Oklahoma: Oklahoma Geol. Survey Bull. 85, 113 p.

Gordon, M., and Stone, C. G., 1977, Correlation of the Carboniferous rocks of the Ouachita Trough with those of the adjacent foreland: *in* Stone, C. G., ed., Symposium on the Geology of the Ouachita Mountains, Volume 1, Arkansas Geol. Commission, p. 70–91.

Graham, S. A., Ingersoll, R. V., and Dickinson, W. R., 1976, Common provenance for lithic grains in Carboniferous sandstones from Ouachita Mountains and Black Warrior Basin: Jour. Sed. Petrology, v. 46, p. 620–632.

Ghibaudo, G., 1981, Deep sea fan deposits in Macigno Formation (Middle-Upper Oligocene) of the Gordana Valley, Northern Appennines, Italy – Reply: Jour. Sed. Petrology, v. 51, p. 1021–1026.

Haley, B. R., Glick, E. E., Bush, W. V., Clardy, B. F., Stone, C. G., Woodward, M. B., and Zachry, D. L., 1976, Geological Map of Arkansas: Arkansas Geol. Comm., Little Rock, Arkansas.

Lock, B. E., 1979, Outer deep-sea fan depositional lobe sequence from Jackfork Group of southern Arkansas: Abs. in Gulf Coast Assoc. of Geological Societies and SEPM Regional Meeting, San Antonio, Texas.

Miser, H. D., 1934, Carboniferous rocks of the Ouachita Mountains: Am. Assoc. Petroleum Geologists Bull., v. 18, p. 30–43.

Morris, R. C., 1971a, Stratigraphy and sedimentology of the Jackfork Group, Arkansas: Am. Assoc. Petroleum Geologists Bull., v. 55, p. 387–402.

————, 1971b, Classification and interpretation of disturbed bedding types in Jackfork flysch rocks (Upper Mississippian). Ouachita Mountains, Arkansas: Jour. Sed. Petrology, v. 41, p. 410–424.

————, 1974a, Carboniferous rocks of the Ouachita Mountains, Arkansas -- A study of facies patterns along the unstable slope and axis of a flysch trough, *in* Briggs, G., ed., Symposium on the Carboniferous Rocks of the Southeastern United States: Geological Soc. Am. Spec. Paper 148, p. 241–279.

————, 1974b, Sedimentary and tectonic history of the Ouachita Mountains, *in* Dickinson, W. R., ed., Tectonics and Sedimentation: Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 22, p. 120–142.

————, 1977a, Petrography of Stanley-Jackfork Sandstones, Ouachita Mountains, Arkansas: *in* Stone, C. G., ed., Symposium on the Geology of the Ouachita Mountains, Volume 1, Ark. Geol. Comm., p. 146–157.

————, 1977b, Flysch facies of the Ouachita Trough-- with examples from the Spillway at DeGray Dam, Arkansas, *in* Stone, C. G., ed., Symposium on the Geology of the Ouachita Mountains, Volume 1, Ark. Geol. Comm., p. 158–168.

Mutti, E., 1974, Examples of ancient deep-sea fan deposits from Circum-Mediterannean geosynclines, in Modern and Ancient Geosynclinal Sedimentation: SEPM Spec. Pub. 19, p. 92–105.

Mutti, E., and Ghibaudo, G., 1972, Un esempio di torbiditi di conoide sottomarina esterna: le Arenarie di San Salvatore (Formazione di Bobbio, Miocene) nell' Appennino di Piacenza: Mem. Acc. Sci. Torino, Classe Sci. Fis., Nat., ser. 4, n. 16, 40 p.

Normark, W. R., 1978, Fan valley channels and depositional lobes in modern submarine fans: characters for recognition of sandy turbidite environments: Am. Assoc. Petroleum Geologists Bull., v. 62, p. 912–931.

Ricci, Lucchi F., 1975, Depositional cycles in two turbidite formations of northern Appennines (Italy): Jour. Sed. Petrology, v. 45, p. 3–43.

Stone, D. M., 1981, Paleoenvironments and Secondary Porosity of the Upper Jackfork Sandstone, Central Arkansas: Unpub. M.S. Thesis, Memphis State University, 84 p.

Stone, C. G. and McFarland, J. D., III, 1981, Field guide to the Paleozoic rocks of the Ouachita Mountains and Arkansas Valley Provinces, Arkansas: Ark. Geol. Comm. Guidebook. 81–1, 140 p.

Thomson, A., and LeBlanc, R. J., 1975, Carboniferous deep-sea fan facies of Arkansas and Oklahoma: Geol. Soc. Am. Abs. with Prog., v. 7, p. 1298–1299.

Walker, R. G., 1978, Deep-water sandstone facies and ancient submarine fans: Models for exploration for stratigraphic traps: Am. Assoc. Petroleum Geologists Bull., v. 62, p. 932–966.

Walthall, B. H., 1967, Stratigraphy and structure, part of Athens Plateau, southern Ouachitas, Arkansas: Am. Assoc. Petroleum Geologists Bull., v. 51, p. 504–528.

Wilde, R., Normark, W. R., and Chase, T. E., 1978, Channel sands and petroleum potential of Monterey deep-sea fan, California: Am. Assoc. Petroleum Geologists Bull., v. 62, p. 967–983.

RECONNAISSANCE STRUCTURAL GEOLOGY IN THE WESTERN MAZARN BASIN, SOUTHERN BENTON UPLIFT, ARKANSAS

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ABSTRACT

Reconnaissance mapping of structural elements in the western Mazarn Basin between Bonnerdale and Hopper, Arkansas, confirms north and northwest verging thrust faults as the dominant structural features. Imbricate slices of Arkansas Novaculite, Stanley Shale and Jackfork Sandstone were transported northward during the thrusting phase. Relatively early syntectonic mesoscopic folds have been redeformed by later macroscopic folding. Rare southward verging disharmonic folds are probably pretectonic in origin, having formed during syndepositional slumping. Cleavage is local and poorly developed. Geometrically regular fracture sets probably formed relatively late in the structural history of the region. A genetic relationship between fracture formation and macroscopic folding appears to be mechanically feasible for the western part of the area but invalid farther to the east.

INTRODUCTION

The Mazarn Basin is an elongate, east-west trending intermontane valley bounded on all sides by the folded and thrust faulted ridges of Arkansas Novaculite (Devonian/Mississippian) that comprise the ranges of the southern Benton Uplift in Arkansas (Fig. 1, inset). Most of the valley is floored by deformed flysch-like sedimentary rocks of the Stanley Shale (Mississippian). Rocks of the Jackfork Sandstone (Pennsylvanian) and Arkansas Novaculite are exposed along Mazarn Ridge and Pigeon Roost Mountain, respectively, in the central and west-central parts of the valley. The basin has been recognized as a complex syncline (Miser and Purdue, 1929) or synclinorium (Viele, 1974).

Traditionally, the Arkansas Ouachitas have been divided into three tectonic subprovinces: the Frontal Ouachitas (Frontal Zone), Benton Uplift, and Southern Ouachitas or Athens Plateau. Haley and Stone (1982) have recently proposed new divisions (tectonic belts) based on their identification of seven distinct, internally deformed thrust plates. The thrust fault that defines the boundary between their Mt. Ida and Hopper Belts trends across the northern part of the area described in this paper, and we have informally designated it the Hopper Thrust (Fig. 1).

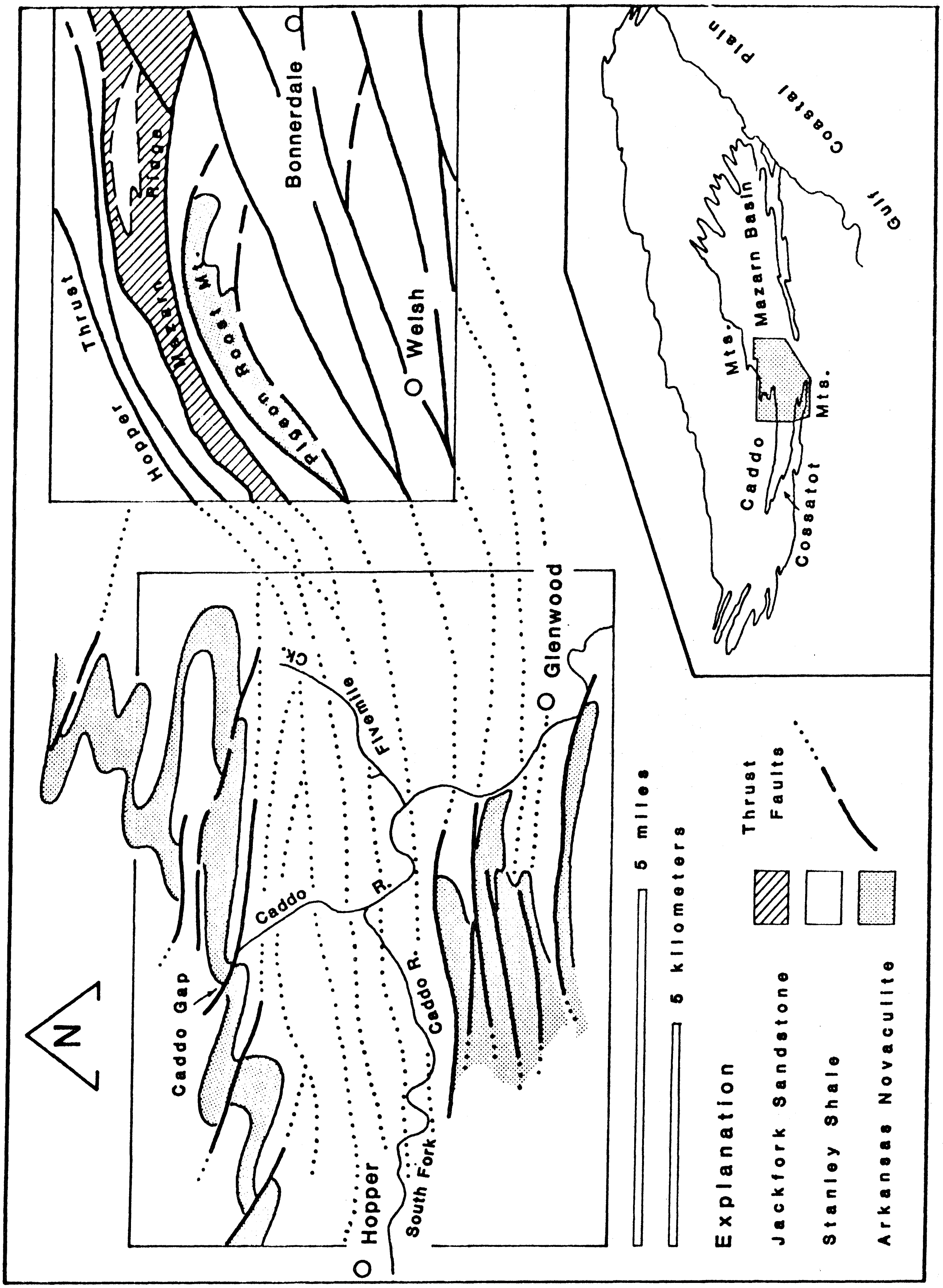

Figure 1. Generalized geologic map of part of the western Mazarn Basin showing areas mapped by Evansin (left) and Ragan (right). Linear dot pattern shows thrust faults from Haley and others (1976). Inset: location of study area in shaded pattern.

This paper presents data and preliminary structural interpretations from part of the western Mazarn Basin between the villages of Bonnerdale and Hopper, comprising parts of the Norman, Caddo Gap, Bonnerdale, Lodi, Glenwood, and Amity, Arkansas, U.S.G.S. 7.5 minute quadrangles. The data originated during separate mapping projects conducted by D. P. Evansin and V. S. Ragan in 1975 and 1976. The individual map areas are shown in Figure 1.

Observations and measurements from the South Fork Valley, Fivemile Creek and associated areas were extracted from Evansin's (1976) thesis and from his field notes. Structural data from the Pigeon Roost Mountain area were taken from the draft text and preliminary geologic map of an uncompleted thesis by Ragan. Interpretations of the structural geology have been taken in part from the above sources and are, in part, the results of additional analysis and re-evaluation of the data.

SOUTH FORK VALLEY AND ASSOCIATED AREAS (Evansin)

The South Fork of the Caddo River drains the narrow, western extremity of the Mazarn Basin between the merging ridges of the Caddo and Cossatot Ranges (Fig. 1). A second tributary, Fivemile Creek, drains an area east of the Caddo River at about the same latitude. Evansin (1976) mapped the Stanley Shale in both drainages as part of his larger study of the geology between the eastern Cossatot and southern Caddo Mountains. He referred to the areas on both sides of the Caddo River (including the Fivemile Creek drainage) as the South Fork Valley. We will use that term only for areas west of the Caddo River. Discussion in this paper will be limited to Evansin's work on the Stanley Shale in the Mazarn Basin and on local outcrops of that unit in the Cossatot Range. The results of his work in the Caddo Mountains have been reported elsewhere (Zimmerman and Evansin, 1982; Zimmerman, Roeder, Morris and Evansin, 1982).

Thrust Faults

Thrust faults are the dominant macroscopic structures in the western Mazarn Basin. Haley and others (1976) have mapped the traces of nine thrusts between Glenwood and Caddo Gap (Fig. 1), including that of the Hopper Thrust which Haley and Stone (1982) consider to be a structure of regional significance. These faults are difficult to observe in the field, particularly in areas where a single, monotonously interbedded unit such as the Stanley Shale has been imbricated. For most part, thrust faults have been mapped from air photos with subsequent field checks. Stone and Haley (personal communication) have noted concentrations of quartz, calcite and/or dickite filled fractures in close proximity to known or suspected faults.

Evansin (1976) mapped a single thrust zone in the vicinity of the Bird and Son quarry, north of Fivemile Creek (Fig. 2). A number of thrust slices ranging from about 5 to 20 feet (1.5 to 6 m) thick are exposed in the quarry. Fault planes strike WNW and dip southward between 55° and 70°. Slip surfaces are locally slickensided, indicating northward overthrusting where the sense of transport can be determined, and are associated with north verging drag folds (Evansin, 1976, Plates III and IV). Southward steepening of thrust surfaces in the quarry suggests folding of the faults on a larger scale. This sense of rotation (clockwise, looking east) is compatible with the rotation of structures noted by Zimmerman and Evansin (1982) at Caddo Gap, some 2.5 miles (4 km) to the northwest, and has long been considered an important element of structural style in much of the southern Benton Uplift.

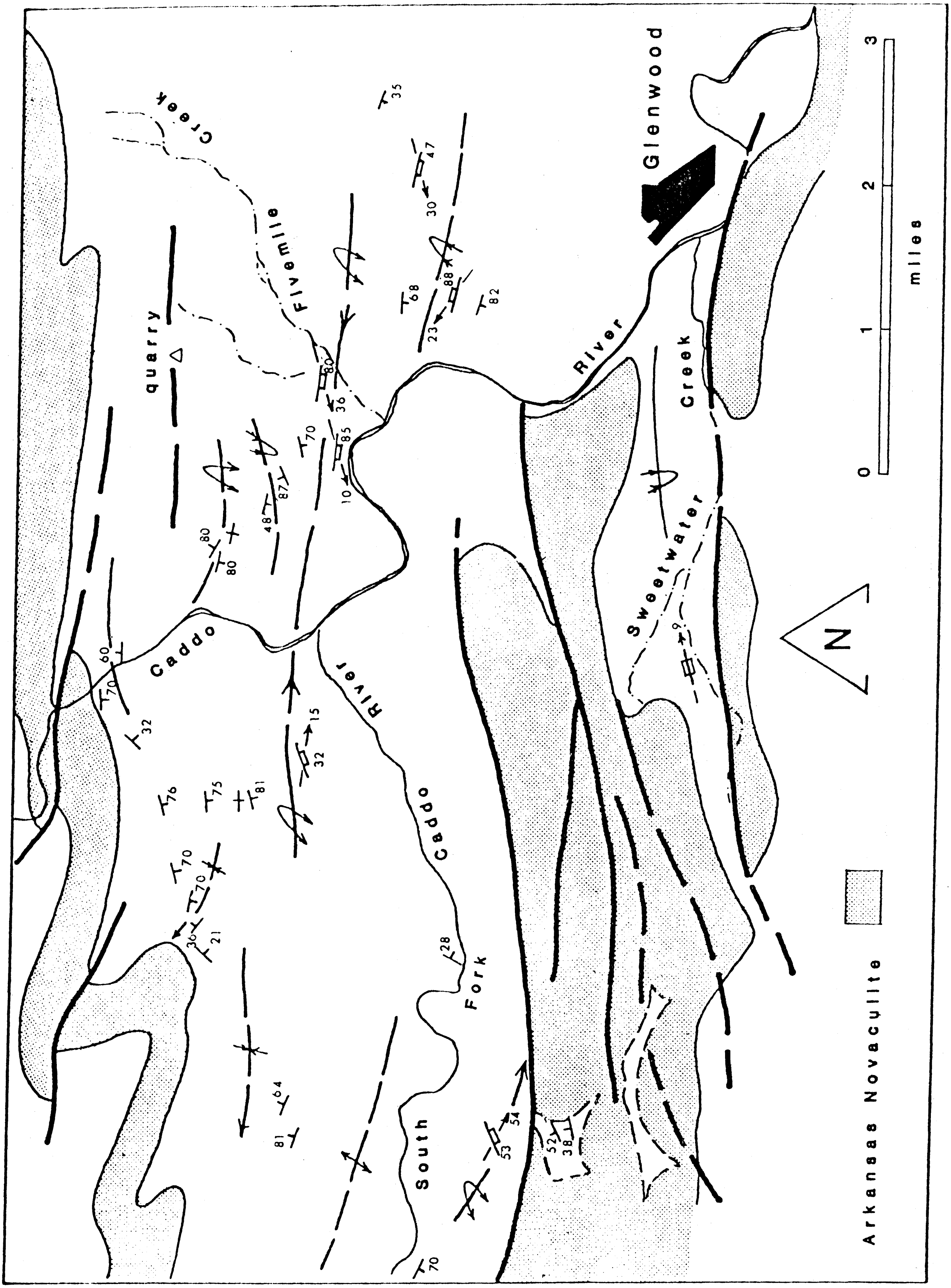

Figure 2. Generalized geologic map of the South Fork Valley, Fivemile Creek and adjacent areas in the Cossatot and Caddo Mountains; modified from Evansin (1976, Plates VII and VIII).

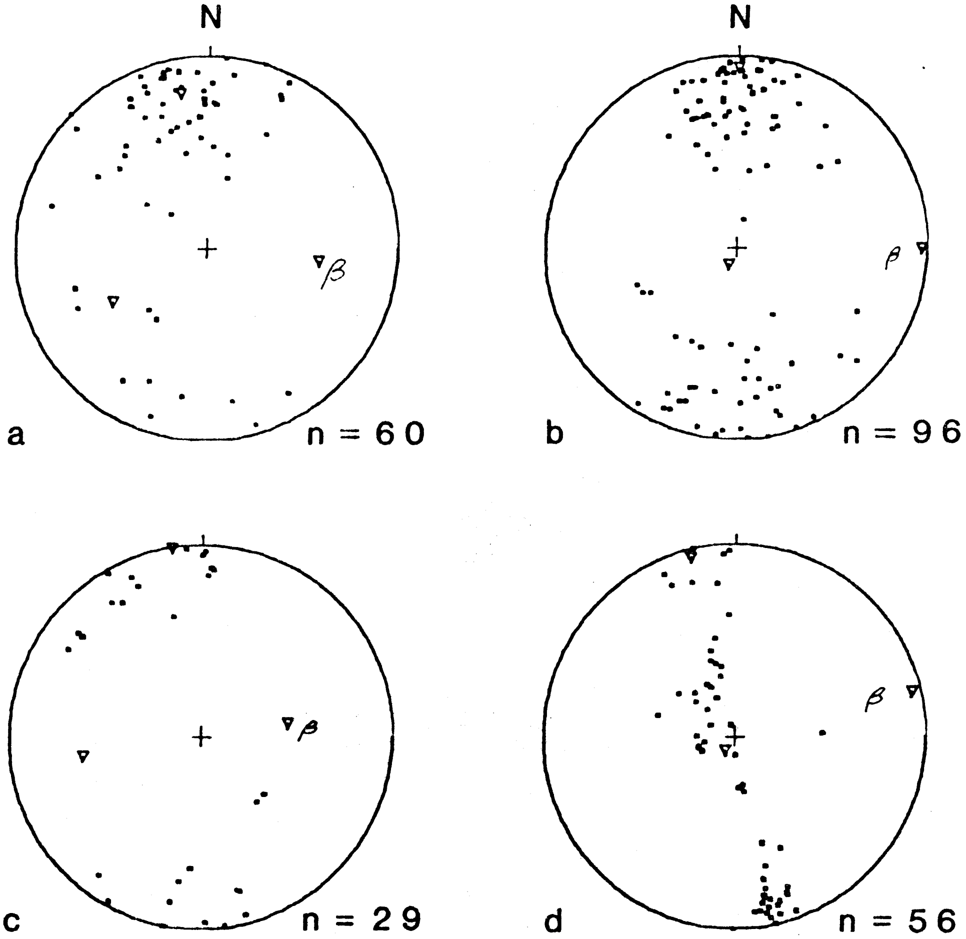

Figure 3. Lower hemisphere, equal-area projections of poles to bedding planes from: a) South Fork Valley, b) Fivemile Creek area, c) Stanley Shale outcrops in the Cossatot Range, and d) Sweetwater Creek area. n = number of poles; inverted triangles (dels) = eigenvectors; β = statistical fold (zone) axis.

There is no evidence that the high thrust density observed at the Bird and Son quarry is typical of the western Mazarn Basin as a whole, and it is probable that the quarry is located in a specific thrust zone. Nonetheless, our observations of the overall deformation between Glenwood and Caddo Gap suggest that Haley and others (1976) have not overestimated the number of faults in that area.

Bedding Attitudes

Bedding typically strikes ENE in the South Fork Valley and Fivemile Creek areas. Southward dips predominate (Fig. 2; Evansin, 1976, Plate IX) although northward dips are not uncommon. At least some of the dip reversals occur near the leading edges of thrust slices.

Patterns of bedding attitudes in the Stanley Shale are shown in Figure 3. These were measured in the South Fork Valley, the Fivemile Creek area, from outcrops of Stanley Shale in the Cossatot Mountains, and along Sweetwater Creek west of Glenwood. The overall structural pattern is similar in all four areas and indicates the presence of large-scale folding with varying amounts of eastward plunge. In three of the areas patterns are diffuse, probably

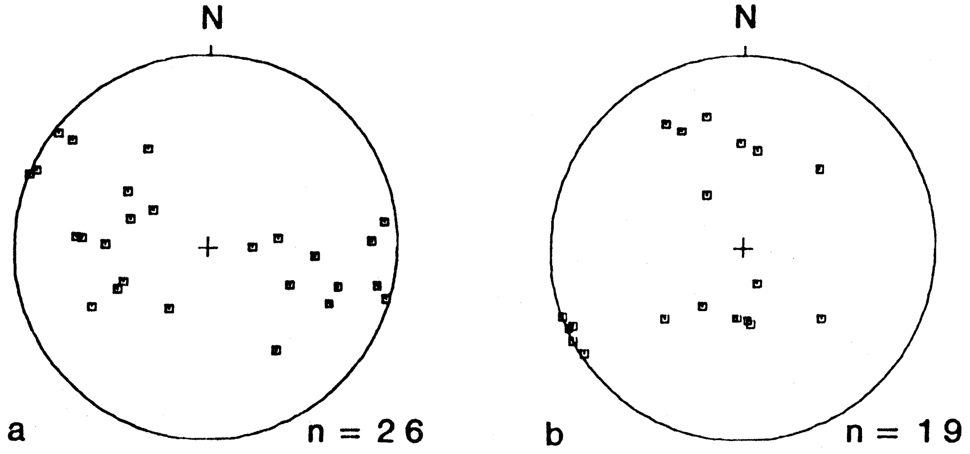

Figure 4. Lower hemisphere, equal-area projections of poles to possible cleavage planes: a) west of the Caddo River (South Fork Valley and Cossatots), and b) east of the Caddo River (Fivemile Creek area).

a result of the relative ductility of the Stanley Shale during deformation and of the fact that bedding measurements come from several distinct folds, each of which varies from others in orientation. The Sweetwater Creek data (Fig. 3d) were taken from a single complex, macroscopic, north-verging overturned syncline between thrust-bounded ridges of Arkansas Novaculite and show a more consistent pattern of bedding attitudes.

Cleavage

Cleavage in the Stanley Shale of the western Mazarn Basin is poorly defined and local. Slaty and crunulation cleavage become better developed but still unevenly distributed in pre-flysch rocks north of Caddo Gap.

Poles to surfaces that may represent incipient slaty and/or fracture cleavage are shown in Figure 4. Where present, these surfaces typically occur at high angles to bedding planes. Two entirely different patterns are obvious from measurements taken east and west of the Caddo River, suggesting that each may represent surfaces of different origin. Those west of the river are oriented nearly perpendicular to the dominant bedding strike (compare Figures 4a, 3a, and 3c) and may represent poorly developed a-c and/or [0k1] fractures (Hobbs, Means and Williams, 1976) rather than cleavage. The pattern in Figure 4b (from east of the river) is more likely to have been produced by incipient axial plane cleavage although [h01] fractures cannot be ruled out.

More data and additional observations of relationships beyween such surfaces and other, better defined structural elements must be collected from the western part of the Mazarn Basin before a complete evaluation of cleavage development and distribution can be made.

Mesoscopic Folds

While not uncommon, fewer mesoscopic folds were observed than might be expected from the overall intensity of deformation that affected the area.

Evansin (1976) mapped twenty such folds in the Stanley Shale (Figs. 2 and 5a) and divided them into three groups based on morphology and relationships to adjacent bedding. Tight folds with limbs subparallel to relatively undisturbed strata were considered

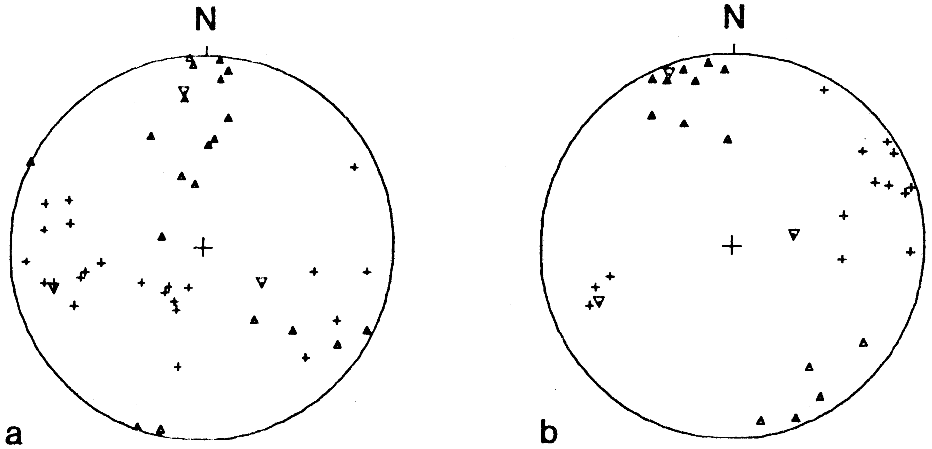

Figure 5. Lower hemisphere, equal-area projections of poles to axial surfaces (triangles) and B axes (crosses) of mesoscopic folds in the Stanley Shale. a) 20 poles to axial surfaces and 23 B axes from the South Fork Valley and Fivemile Creek areas; b) poles to axial surfaces and B axes of 14 folds from the Pigeon Roost Mountain area. Dels are eigenvectors calculated from axial surface distributions.

to be syndepositional, probably related to slumping. Folds of this type are common in the flysch facies of the Ouachita Mountains (Morris, 1974). Evansin also suggested that disharmonic folds observed in the area might be primary but could be equally well explained by ductility contrasts and variable bedding thicknesses in a tectonically deformed flysch sequence that had retained appreciable fluid pore pressure. The third and numerically dominant fold type consists of monoclinic, north verging structures produced during northward thrust transport.

Poles to axial surfaces of 20 mesoscopic folds (Fig. 5a) show a predominance of east-west strikes and southward dips although five folds strike NE–SW and four of these dip to the northwest. The β axis constructed from the pattern of axial surface attitudes plunges 22°/254° (22° to the SW) and could represent the axis of a phase of macroscopic folding that reoriented earlier mesoscopic folds. This axis is significant only if south and southwestward fold vergence is a function of tectonic rather than syndepositional processes, an eventuality that can be readily inferred but has not been firmly established from field observations.

Measured mesoscopic fold axes (B) typically plunge to the west and southwest although a few easterly plunges were recorded. Axes are distributed in a distinct girdle that further suggests the probability of later macroscopic folding (see following section).

Mesoscopic folds in the South Fork Valley and Fivemile Creek areas are typically inclined and plunging (Fig. 6) although a variety of spatial orientations is represented.

Macroscopic Folds

Evidence for macroscopic folding of the Stanley Shale can be inferred from patterns of bedding and mesoscopic fold elements.

Bedding attitudes measured away from known mesoscopic folds show both northward and southward dips in all four areas illustrated in Figure 3. β axes calculated by eigenanalysis of the data in each equal-area net probably represent axes of macroscopic folds of bedding planes. For each of the areas analyzed, the amounts and azimuths of β axis plunges are: South Fork Valley, 41°/097°; Fivemile Creek,

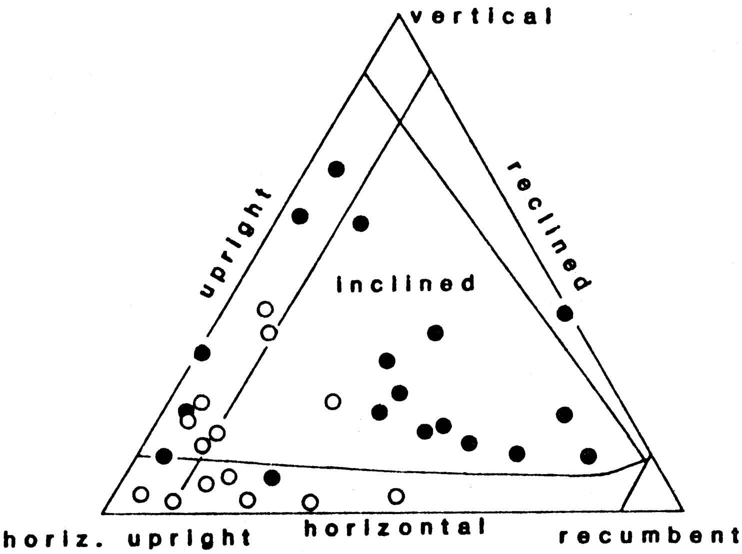

Figure 6. Rickard (1971) diagram showing spatial orientations of 18 mesoscopic folds from the western (Evansin) map area (solid circles) and 14 mesoscopic folds from the eastern (Ragan) map area (open circles).

4°/090°; Stanley Shale outcrops in the Cossatot Range, 52°/082°; and Sweetwater Creek, 5°/075°. All axes plunge in an easterly direction. It is worth noting, however, that the amount of plunge varies markedly in areas east and west of the Caddo River. West of the river, plunges are 41° and 52°, while to the east, plunges diminish to 4° and 5°. This apparently consistent variation suggests broad-scale warping of macroscopic folds about a north-south line roughly coincident with the position of the Caddo River in the central part of the western Mazarn Basin. In addition to the latitudinal plunge variation there is evidence of a longitudinal variation of macroscopic β axes which show a consistent 16° change in azimuth from north to south. This can be illustrated by direct comparison of data from the South Fork Valley (Fig. 3a) with those from the Cossatot Mountains (Fig. 3c) and by a similar comparison between data from the Fivemile Creek area (Fig. 3b) and from Sweetwater Creek (Fig. 3d). In both instances, rotation of 16° about a vertical axis brings the northern and southern macroscopic fold elements into close juxtaposition. This apparent counter-clockwise rotation of folds in the southern areas can be explained in one (or both) of two ways. The warp axis that is reflected in the plunge variation discussed above may swing from a north-south trend in the middle of the basin to a SSE orientation in the vicinity of the Cossatots. A more probable explanation is that the trend of macroscopic fold elements in the two southern areas has been controlled by thrusting in the Cossatot Range that trends ENE and brings slices of relatively rigid Arkansas Novaculite into close proximity, limiting the variation of structural orientation in the more ductile Stanley Shale.

As mentioned in the preceding section, the pattern of mesoscopic fold elements in the South Fork Valley and neighboring areas indicates subsequent macroscopic folding. Both poles to axial surfaces and B axes are distributed along nearly orthogonal girdles.

The β axis constructed from the distribution of axial surfaces plunges 22°/254° (22° WSW), an attitude inconsistent with the slight to moderate eastward plunge of macroscopic

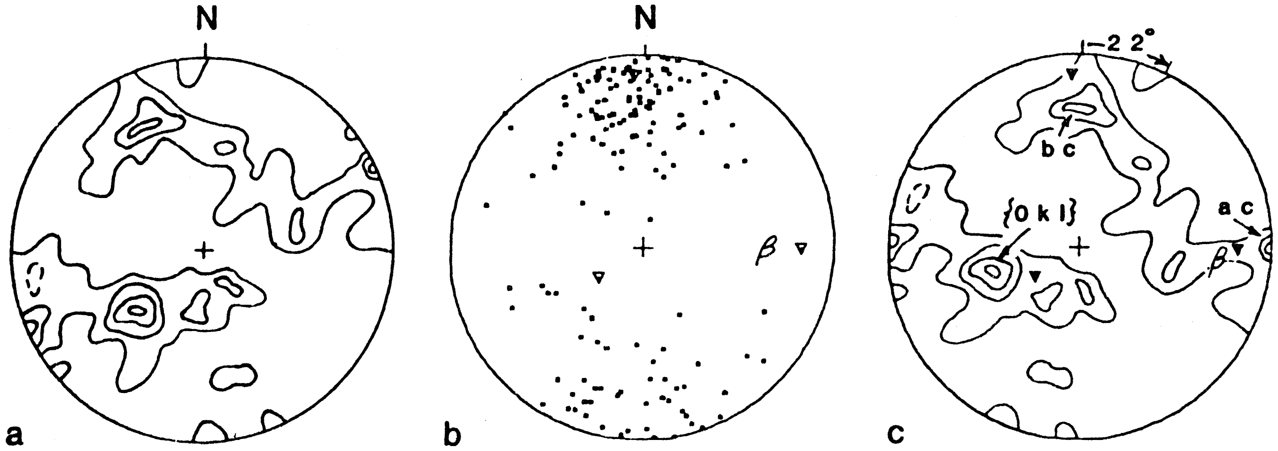

Figure 7. Lower hemisphere, equal-area projects to: a) 65 fracture surfaces in the South Fork Valley and Fivemile Creek area, contours at 1.5, 4.6, 7.7, and 10.8% per 1% area. b) 156 bedding surfaces in the Stanley Shale from the same areas. c) Close correspondence between fracture sets and mesoscopic fold elements resulting from 22° clockwise rotation of fracture data. Possible geometrical identities of fracture sets are indicated.

folding inferred from bedding orientations. It has been noted that five of the twenty mesoscopic axial surfaces plotted in Figure 5a strike northeast rather than essentially east-west. Were these not present the remaining mesoscopic folds would conform closely to the macroscopic fold pattern defined by bedding planes. There is no indication that all of the northeast trending folds are of different origin from the east-west trending structures. It is probable that the differences in macroscopic fold orientations inferred from bedding and mesoscopic fold patterns reflect original attitude variations of the latter.

Fractures

Poles to 65 well developed fracture and joint surfaces in the Stanley Shale of the South Fork Valley and Fivemile Creek areas are shown in Figure 7a. In the absence of more systematic observations of relationships between sets of fractures and other structures, conclusions about the origins of the former have been based on comparison of structural geometries from equal-area plots and should be considered speculative.

In Figure 7a, poles cluster in three fairly strong and one fairly weak point maxima. Regularity of the distribution pattern suggests that the fractures were produced during a relatively late phase of deformation and may be genetically related to the macroscopic folding discussed above. The distribution of bedding planes in the Stanley Shale provide the best indication of macroscopic fold orientation. Comparison of Figures 7a and 7b shows no close correspondence between fracture patterns and inferred fold orientation unless all fracture data are rotated about 22° clockwise about a vertical axis. With rotation (Fig. 7c) the following correspondence can be inferred: east-west trending (northeast trending before rotation), south-dipping fracture surfaces represent b-c extension fractures; north-south trending, vertical surfaces represent a-c extension fractures; and sets of east and west dipping [0k1] surfaces correspond to Stearns (1968) set 1 shear fractures, as indicated. Each fracture set is genetically linked to the formation of large folds (Stearn, 1968; Hobbs and others, 1976).

The fact that rotation is required to bring fracture and macroscopic fold patterns into reasonable juxtaposition may be due to nonparallel alinement of principal stress and principal strain directions during the folding phase, a condition that is not unusual in the deformation of anisotropic materials. In natural folds, fracture and joint orientations are related to principal stress directions (Stearns, 1968) while fold elements such as axial surfaces reflect strain directions. In their experimental study of buckle folding and fracture formation in micritic limestone at low to moderate confining pressures (equivalent to about 1 km to 8 km of overburden) Handin and others (1972) found instances in which axes of compressional and extensional stress determined from calcite twin lamellae were inclined to fold elements by amounts comparable to those suggested in Figure 7. Whether these observations can be extrapolated to folding on the scale of that found in the South Fork Valley and Fivemile Creek areas is unsubstantiated at present.

PIGEON ROOST MOUNTAIN AREA

(Ragan)

East of Fivemile Creek (Fig. 1), two northeast trending, thrust bounded ridges, Mazarn Ridge and Pigeon Roost Mountain, are surrounded by imbricate thrust slices of Stanley Shale. Rocks of the Jackfork Sandstone are exposed along Mazarn Ridge which continues to the east of the mapped area (Haley and others, 1976). Pigeon Roost Mountain consists of novaculites, cherts and shales of the Middle and Lower Divisions of the Arkansas Novaculite. Conflicting interpretations of the macroscopic structure of this area have been given by Griswold (1892) and Croneis (1930). Griswold suggested that both ridges were part of a single, northward verging, overturned anticline with the Arkansas Novaculite along Pigeon Roost Mountain at its core, the standstones that should define its southern limb having been removed by erosion. Croneis considered each ridge to be a separate overturned anticline. In his discussion of the barite deposits located at the southwestern end of Pigeon Roost Mountain, Scull (1958) accepted Griswold's interpretation. We suggest, however, that more recent mapping (Haley and others, 1976; Ragan, this paper) favor the Croneis model although a number of structural questions remain unanswered.

Thrust Faulting

As in other parts of the western Mazarn Basin thrust faults are the dominant macroscopic structures in the Pigeon Roost Mountain area. Fault traces trend northeastward in the southern part of the area but become more arcuate in the vicinity of Mazarn Ridge and Pigeon Roost Mountain (Fig. 1). Haley and others (1976) mapped over a dozen thrust traces in the area. Most of these were confirmed by Ragan and were included on her preliminary geologic map (Fig. 8) There are two major exceptions. Haley and others mapped a northeast trending fault through the interior of Mazarn Ridge, subparallel to its bounding thrusts, that was not confirmed by Ragan. They also indicate that the lobate outcrops of Arkansas Novaculite that define the southeastern edge of Pigeon Roost Mountain are fault bounded. Ragan (Fig. 8) interprets them as southeastward plunging folds.

Ragan reported the Upper Division of the Arkansas Novaculite to be missing from the southeastern flank of Pigeon Roost Mountain (Fig. 8) and suggested that if the rocks had been deposited in this area they must have been removed by erosion or by thrust faulting that brought the Stanley Shale into contact with rocks of the stratigraphically lower Middle Division. The contact is not shown as faulted in Figure 8 because Ragan originally considered

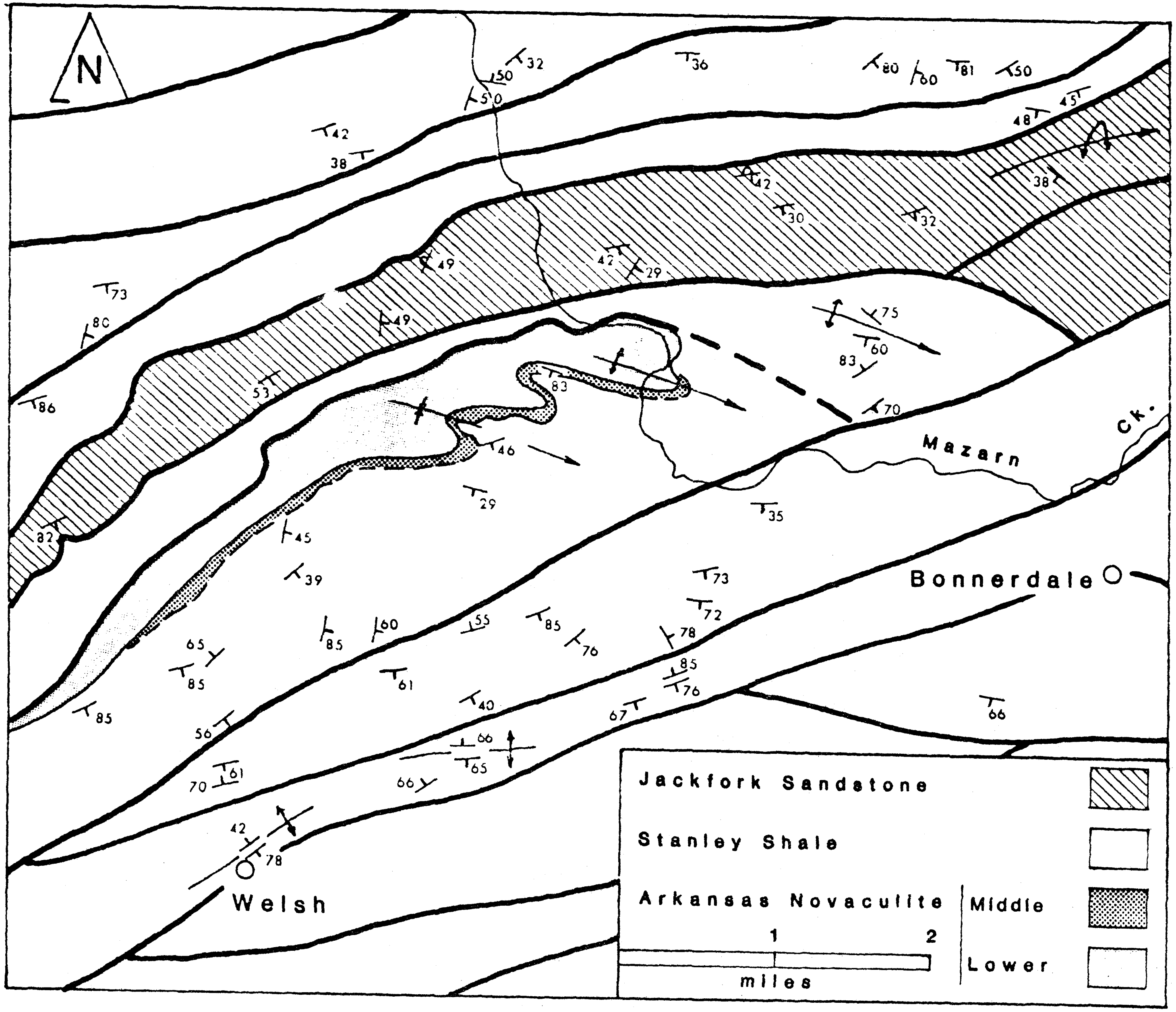

Figure 8. Generalized geologic map of the Pigeon Roost Mountain area; modified from preliminary geologic map by V. S. Ragan.

it to be gradational and favored the nondeposition hypothesis. We now consider the removal of Upper Divison and some Middle Division rocks by faulting to be a better explanation.

The differences in structural interpretations indicate that there is still a good deal to be learned about thrust faulting in the area and underscore the need for additional detailed field work in the vicinity of Pigeon Roost Mountain and Mazarn Ridge.

Bedding Attitudes

Most bedding planes in the Pigeon Roost Mountain area strike northeast and dip to the south - attitudes that are entirely consistent with northwestward thrust transport. Contoured equal area nets from Ragan's draft thesis are shown in Figure 9 together with point diagrams from the same areas plotted from bedding attitudes taken from her preliminary geologic map.

Poles to bedding planes in the Stanley Shale from north and south of the Jackfork and novaculite ridges (Fig. 9a) define a relatively diffuse pattern dominated by northeast strikes and southward dips. The pattern on Ragan's contoured diagram (Fig. 9b) is generally comparable except for a strong point maximum in the northeast quadrant that records a number of northwest striking surfaces that dip between 10° and 50° to the southwest. Ragan reported that no bedding attitudes on recognized mesoscopic folds had been plotted on the diagram. Only one bed with a comparable attitude was recorded on her preliminary map, and data from Mazarn Ridge and Pigeon Roost Mountain fail to show an equivalent point maximum. The apparent inconsistency produced by these southwestward dips cannot be resolved with the information presently available.

Bedding attitudes from Mazarn Ridge (Figs. 9c and 9d) and from Pigeon Roost Mountain (Figs. 7e and 7f) are comparable, showing a clear dominance of northeast strikes and southwestward dips. Attitudes from Mazarn Ridge are somewhat more variable, and dips are generally shallower.

Eighteen bedding measurements from Mazarn Ridge recorded on Ragan's preliminary geologic map are distributed in a pattern nearly identical to bedding in the Stanley Shale (compare Figures 9a and 9c). The eight attitudes from Pigeon Roost Mountain (Fig. 9e) are not sufficient to define a meaningful distribution.

Mesoscopic Folds

As was the case in Evansin's area farther west, fewer mesoscopic folds than might be expected were found in the Stanley Shale surrounding Pigeon Roost Mountain (Fig. 5b). The majority are northwest verging, monoclinic folds that face in the direction of tectonic transport and probably formed during thrust faulting. Ragan reported a small number of folds with identical geometry that have northwest rather than southwest dipping axial surfaces and suggest that these were originally north verging structures rotated into southward vergence during a later phase of the deformation. Other, clearly disharmonic south verging folds were probably produced by slumping soon after deposition.

Axial surfaces of mesoscopic folds in the South Fork Valley and Fivemile Creek areas strike predominantly east-west with the exception of 5 anomalous structures with northeast strikes. In the Pigeon Roost Mountain area, axial surfaces show a more regular pattern of ENE strikes consistent with the trend of related thrust faults (Figs. 1 and 5b). Measured B axes reflect a dominant northeastward plunge (the average axis plunges 6°/067°) and show less tendency to be distributed along a girdle, in contrast to B axes to the west (compare Figs. 5a and 5b). If the thrust faults

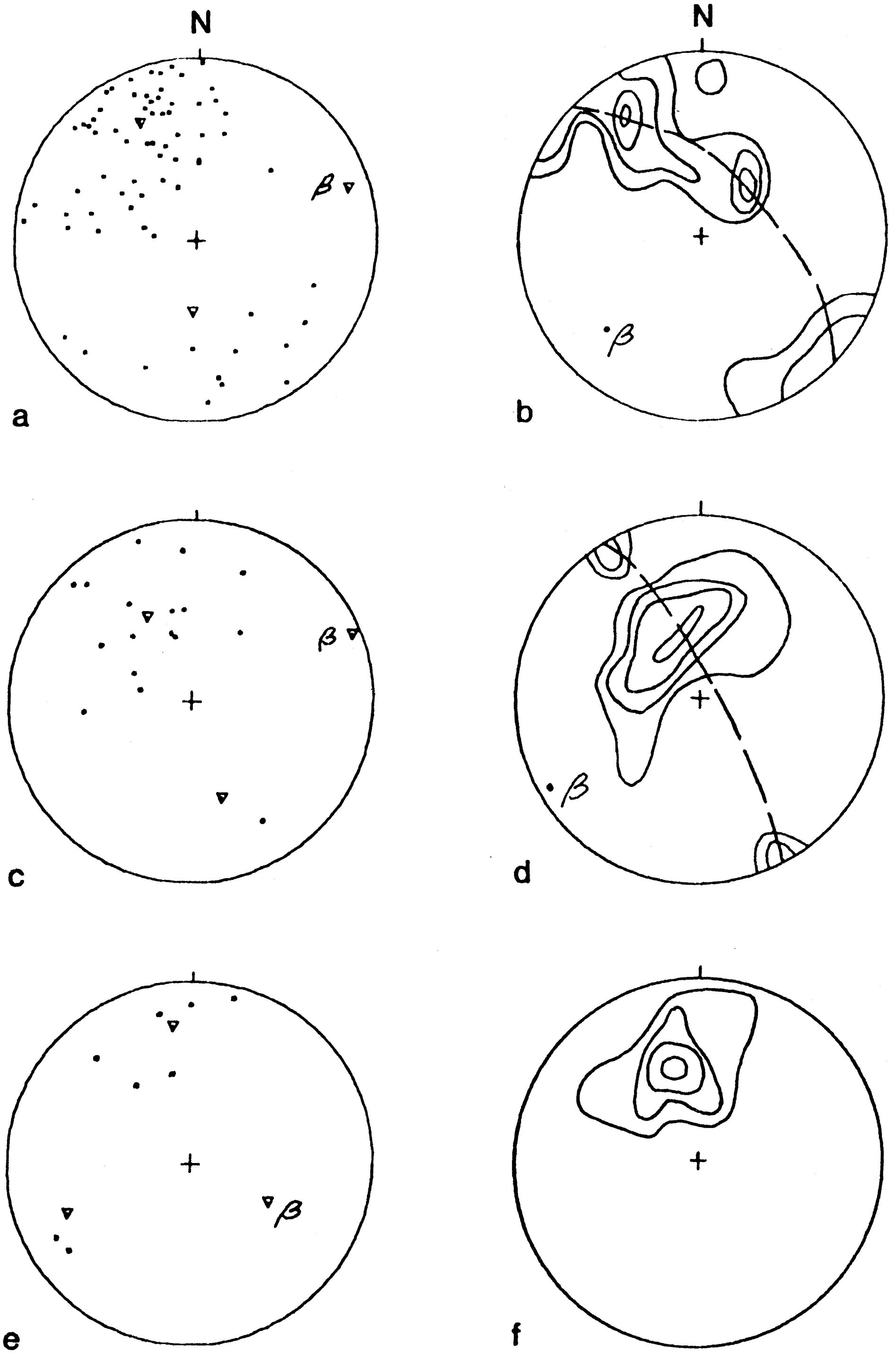

Figure 9. Lower hemisphere, equal-area projections of poles to bedding planes (So) in the Pigeon Roost Mountain area. Point diagrams constructed from data taken from preliminary geologic map and contoured diagrams from draft report by V. S. Ragan. Diagrams contoured at 2, 4, 8, and 24% per 1% area. a) 76 poles to So, Stanley Shale; b) 175 poles to So, Stanley Shale; c) 18 poles to So, Jackfork Sandstone, Mazarn Ridge; d) 103 poles to So, Jackfork Sandstone, Mazarn Ridge; e) 8 poles to So, Arkansas Novaculite, Pigeon Roost Mountain; f) 65 poles to So, Arkansas Novaculite, Pigeon Roost Mountain.

in the two areas were parallel, the mesoscopic fold patterns would be nearly identical except for some differences in the direction of average axial plunge.

Mesoscopic folds in the Pigeon Roost Mountain area tend to be slightly to moderately inclined, upright structures with much less variability in spatial orientation than comparable folds in the South Fork Valley and Fivemile Creek areas (Fig. 6).

Macroscopic Folds

The presence of macroscopic folds is indicated by overturned bedding and dip reversals at the leading edges of thrust slices (Fig. 8) and can be inferred from patterns of poles to bedding planes and the distribution of mesoscopic fold elements.

Sufficient bedding plane data are available from the Stanley Shale to calculate a macroscopic fold axis that plunges 13°/071° (Fig. 9a). Although based upon eigenanalysis of far fewer points, bedding plane attitudes from the Jackfork Sandstone on Mazarn Ridge indicate a similar direction and degree of axial plunge, 6°/066° (Fig. 9c). Ragan's contoured diagrams of bedding attitudes (Figs. 9b and 9d) confirm the northeast trend of macroscopic fold axes but indicate slight to moderate southwestward rather than northeastward plunges: β = 30°/225° for the Stanley Shale and β = 10°/239° for the Jackfork Sandstone.

Poles to axial surfaces of mesoscopic folds (Fig. 5b) also indicate a southward plunge (β = 27°/247°) while the average orientation of measured mesoscopic fold axes (β = 6°/067°) is nearly identical to the macroscopic axes calculated from map data (compare Figs. 5b, 9a and 9c).

The information contained in Figures 5 and 9 clearly establishes a northeast-southwest trend for macroscopic fold axes, a direction consistent with the orientations of most thrust fault traces in the area. Differences in the calculated azimuths and plunges of the axes (from 16° to 43°) cannot be resolved with the data presently available but in any case are not great enough to affect the basic interpretation of macroscopic folding.

Fractures

The contoured distribution of 98 poles to fracture surfaces is shown in Figure 10. Strong point maxima in the northeast and southwest quadrants of the net probably represent a conjugate set of shear fractures that dip to the northeast and southwest and whose intersection plunges norhtwestward. A third and considerably weaker point maximum indicates a set of fracture surfaces dipping to the northwest.

As in the case of fractures recorded from the South Fork Valley and Fivemile Creek areas, these fracture sets show no directional correspondence to calculated macroscopic fold elements unless rotated clockwise about a vertical axis. If rotated into juxtaposition

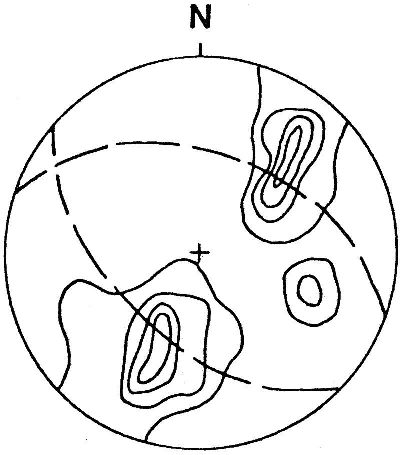

Figure 10. Lower hemisphere, equal-area projection of 98 poles to fracture surfaces, Pigeon Roost Mountain area. Contours at 2, 4, 8, 12% per 1% area. Dashed great circle arcs = average attitudes of fracture surfaces.

with inferred fold trends, the strong point maxima could denote a set of [0k1] shears and the weak maximum could represent [h01] extension fractures. In the Pigeon Roost Mountain area, however, the rotation necessary for the alinement of fold and fracture elements exceeds 40°. It would be difficult to explain so great a divergence between principal stress and strain directions in an area in which there is excellent evidence for overall colinearity between thrust transport and folding.

SUMMARY

The results of structural reconnaissance work in the western Mazarn Basin combined with evidence from regional mapping by Haley and others (1976) are summarized as follows:

1. North to northwest verging thrust faults that displace imbricate slices of Arkansas Novaculite, Stanley Shale and Jackfork Sandstone are the dominant large scale structures in the western part of the Basin.

2. Secondary structures, including mesoscopic and macroscopic folds, were produced during northward thrust transport and are best developed in Stanley Shale rocks. There is evidence of macroscopic folding along Mazarn Ridge and in the Cossatot Range (Evansin, 1976). Conflicting interpretations of the large scale structure of Pigeon Roost Mountain remain unresolved.

3. Structural styles in the western and eastern areas mapped by Evansin and Ragan, respectively, are essentially identical. Changes in the general trends of thrust faults between the two areas have reoriented secondary structures.

4. Most folds are overturned toward the north or northwest. South verging disharmonic folds are attributed to syndepositional slumping. Other folds overturned to the south may have been rotated away from their original northward vergence. There is no clear evidence that such rotation has been extensively developed in the western Mazarn Basin although it is recognized in other parts of the Benton Uplift.

5. There is ample evidence that mesoscopic folds in the Stanley Shale have been refolded during a later phase of macroscopic folding. Although there was considerable variability in the original orientations of the earlier mesoscopic structures, it is probable that both phases were essentially colinear and may have been produced sequentially during the same period of thrust transport.

6. Cleavage is poorly developed in the Stanley Shale and no data were reported from the Pigeon Roost Mountain area. There is no apparent reason for the variation in cleavage patterns east and west of the Caddo River. In some cases incipient or poorly developed fracture surfaces may have been mistaken for cleavage planes.

7. The regularity of fracture patterns from both mapped areas suggests that fracture surfaces have not been reoriented by subsequent deformation, and, therefore, developed late in the structural history of the region. It has been assumed, but not substantiated, that the fractures are associated with the relatively late macroscopic folding phase. If this was the case, non-alinement of principal stress and strain axes during the folding must be called upon to explain the lack of spatial correspondence between the fractures and macroscopic fold elements. While this may be a reasonable explanation for the structural asymmetry in the western area, it is not convincing when applied to the eastern area owing to the degree of non-alinement between stress and strain directions required. It is possible that the fracture sets are mutually unrelated and may have developed during entirely different periods in the history of the region.

8. A number of structural problems remain to be solved in the mapped areas of the western Mazarn Basin. Among others, these include the possible presence of an additional thrust fault along Mazarn Ridge, and the nature of the Arkansas Novaculite-Stanley Shale contact and the origin of the lobate outcrop pattern on the southeast flank of Pigeon Roost Mountain.

A NOTE ON DATA ANALYSIS

Most of the field data collected from the Benton Uplift and adjacent areas in recent years are scattered among a number of theses completed during mapping projects by graduate students from various universities. These reports contain a wealth of information about the spatial orientation of bedding, fold elements, fractures, cleavage, strain markers, current indicators, and other features of interest to structural geologists and sedimentologists. This information is potentially of critical importance to the development of new interpretations and syntheses of Ouachita geology but will be of practical value only if available in usable form.

A project to extract Ouachita field data from reports and to classify and store them for later use has been started at Southern Illinois University at Carbondale. Similar efforts are under way at other institutions. In our system, data are separated and classified according to type, location, stratigraphic unit, and source and are filed in computer storage. Recall of files for modification, combination with other files, plotting, and geometrical, mathematical and statistical analysis is quick and efficient.

Because of our primary interest in the structural geology of the Ouachitas, the system contains routine procedures for eigenanalysis and for plotting data and eigenfunctions on equal-area nets. In a typical case, data from a stored file are used to construct a normalized orientation tensor:

$$A = \frac{a}{N}$$

where N is the number of data, and

$$a = \begin{bmatrix} \Sigma l_i^2 & \Sigma l_i m_i & \Sigma l_i n_i \\ \Sigma m_i l_i & \Sigma m_i^2 & \Sigma m_i n_i \\ \Sigma n_i l_i & \Sigma n_i m_i & \Sigma n_i^2 \end{bmatrix}$$

is a matrix of direction cosine products summed over all measurements (Scheidegger, 1965; Woodcock, 1977), l_i, m_i, and n_i are the direction cosines defining the spatial attitude of each field measurement in a geographically based cartesian coordinate system (see Nielsen, 1983, p. 104). The eigenvalues and corresponding eigenvectors of the orientation tensor are calculated together with the equal-area net coordinates of the eigenvectors.

Eigenvectors are related to the distribution of lines in space and can be used to determine quantitative best fit planes and lines to various sorts of structural patterns on equal-area nets. This takes the guesswork out of defining π-girdles, β axes of folded surfaces, or average vectors of any sort that formerly had to be "eyeballed" into the point distribution or determined qualitatively by much more cumbersome contouring procedures. All of the equal-area point diagrams used in this paper were constructed by the series of steps outlined in the preceding paragraph. Other applications of eigenanalysis to structural and sedimentological data have been discussed by Woodcock (1977) and Nielsen (1983).

ACKNOWLEDGEMENTS

The field work discussed in this paper was made possible by the generous support of the Arkansas Geological Commission and the interest and encouragement of Norman F. Williams. Southern Illinois University at Carbondale provided additional support and computer facilities. All phases of our work in the Benton Uplift have been influenced by Boyd Haley, Charles Stone, and George Viele, who have freely lent us their time and expertise but are not responsible for our errors or omissions. John and Marie Cook and Cleve Wisenhunt of Glenwood deserve special thanks for their unfailing hospitality to us and to many of our colleagues.

REFERENCES

Croneis, C., 1930, Geology of the Arkansas Paleozoic area: Little Rock, Arkansas Geological Survey Bull. No. 3, 457 p.

Evansin, D. P., 1976, Structural style of the Ouachita core near Caddo Gap, Arkansas, M.S. Thesis, Southern Illinois University at Carbondale, 94 p.

Griswold, L. S., 1892, Whetstones and the novaculites of Arkansas: Little Rock, Arkansas Geological Survey Annual Report for 1890, v. 3, 443 p.

Haley, B. R., Glick, E. E., Bush, W. V., Clardy, B. F., and others, 1976, Geologic map of Arkansas: Little Rock, Arkansas Geological Commission and U. S. Geological Survey, scale 1:500,000.

Haley, B. R., and Stone, C. G., 1982, Structural framework of the Ouachita Mountains, Arkansas: Geological Society of America Abs. with Programs (South-Central Section), v. 14, No. 3, p. 113.

Handin, J., Friedman, M., Logan, J. M., Pattison, L. J., and Swolfs, H. S., 1972, Experimental folding of rocks under confining pressure: buckling of single-layer rock beams: *in* Heard, H. C., and others, eds., Flow and fracture of rocks (Griggs Volume): Washington, D. C., American Geophysical Union, Geophysical Monograph 16, p. 1–28.

Hobbs, B. E., Means, W. D., and Williams, P. F., 1976, An outline of structural geology: New York, John Wiley & Sons, 571 p.

Miser, H. D., and Purdue, A. H., 1929, Geology of the De Queen and Caddo Gap quadrangles, Arkansas: U. S. Geological Survey Bull. 808, 195 p.

Morris, R. C., 1974, Carboniferous rocks of the Ouachita Mountains, Arkansas: a study of facies patterns along the unstable slope and axis of a flysch trough: *in* Dickinson, W. R., ed., Geological Society of America Spec. Paper 148, p. 241–280.

Nielsen, K. C., 1983, Pebble population ellipsoid: a three-dimensional algebraic description of sedimentary fabric: Journal of Geology, v. 91, p. 103–112.

Rickard, M. J., 1971, A classification diagram for fold orientations: Geological Magazine, v. 108, p. 23–26.

Scheidegger, A. E., 1965, On the statistics of the orientation of bedding planes, grain axes, and similar sedimentological data: U. S. Geological Survey Prof. Paper 525–C, p. C164–C167.

Scull, B. J., 1958, Origin and occurrence of barite in Arkansas: Little Rock, Arkasnas Geological Commission Information Circ. 18, 101 p.

Stearns, D. W., 1968, Certain aspects of fracture in naturally deformed rocks: *in* Riecker, R. E., ed., NSF advanced science seminar in rock mechanics, Vol. 1: Bedford, Massachusetts, Air Force Cambridge Research Laboratories, p. 97–116.

Viele, G. W., 1973, Structure and tectonic history of the Ouachita Mountains, Arkansas: *in* DeJong, K. A. and Scholten, R., eds., Gravity and tectonics: New York, John Wiley & Sons, p. 361–377.

Woodcock, N. H., 1977, Specification of fabric shapes using an eigenvalue method: Geological Society of America Bull., v. 88, p. 1231–1236.

Zimmerman, J., and Evansin, D. P., 1982, Fold geometry and faulting in the Arkansas Novaculite at Caddo Gap, west Benton Uplift, Arkansas: Geological Society of America Abs. with Programs (South-Central Section), v. 14, No. 3, p. 141.

Zimmerman, J., Roeder, D. H., Morris, R. C., and Evansin, D. P., 1982, Geologic section across the Ouachita Mountains, western Arkansas: Geological Society of America Map and Chart Series, MC 28–Q, scale 1:250,000.