

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

SYMPOSIUM ON THE GEOLOGY OF THE
OUACHITA MOUNTAINS

VOLUME I

Stratigraphy, Sedimentology, Petrography,
Tectonics, and Paleontology

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T A B L E O F C O N T E N T S

Page

FOREWORD.....v
By Thomas A. Hendricks

P A R T I

DEDICATION OF SYMPOSIUM VOLUMES TO
HUGH DINSMORE MISER

A Tribute to Hugh Dinsmore Miser2
By George V. Cohee

Early Life of Hugh Dinsmore Miser4
By Lloyd G. Henbest

Comments on Some Characteristics of Hugh Dinsmore Miser6
By Thomas A. Hendricks

Selected Bibliography of Hugh Dinsmore Miser.....7

P A R T II

TECHNICAL PAPERS

Structural and Stratigraphic Continuity of the Ouachita and Appalachian Mountains.....9
By William A. Thomas

Age of Igneous and Metamorphic Activity Affecting the Ouachita Foldbelt25
By R. E. Denison, W. H. Burke, J. B. Otto, and E. A. Heatherington

Marathon Revisited41
By Philip B. King

Correlation of the Carboniferous Rocks of the Ouachita Trough with70
Those of the Adjacent Foreland
By MacKenzie Gordon, Jr. and Charles G. Stone

Conodonts from Graptolite Facies in the Ouachita Mountains, Arkansas92
and Oklahoma
By John E. Repetski and R. L. Ethington

The Occurrence and Origin of the Granite--Meta-arkose Erratics in the Ordovician	107
Blakely Sandstone, Arkansas	
By Charles G. Stone and Boyd R. Haley	
Meta-arkose Boulder from Blakely Sandstone (Ordovician) Benton, Quadrangle,	112
Arkansas	
By Kern C. Jackson	
Paleoenvironments and Paleobathymetry of Lower Paleozoic Crystal Mountain	115
and Blakely Formations, Ouachita Mountain Core	
By David K. Davies and Eddie A. Williamson	
The Arkansas Novaculite: Some Aspects of its Physical Sedimentation	132
By Donald R. Lowe	
Arkansas Novaculite Stratigraphy	139
By Mark A. Sholes	
Petrography of Stanley-Jackfork Sandstones, Ouachita Mountains, Arkansas	146
By Robert C. Morris	
Flysch Facies of the Ouachita Trough--with Examples from the Spillway at	158
DeGray Dam, Arkansas	
By Robert C. Morris	
Internal Sedimentary Structures in Sandstones of the Jackfork Group,	169
Ouachita Mountains, Oklahoma	
By Robert D. LoPiccolo	

F O R E W O R D

By Thomas A. Hendricks¹

It is my privilege to write the foreword to the volumes on the geology of the Ouachita Mountains. The high degree of interest in the geology of the Ouachitas is shown by the fact that the two Ouachita Mountains symposium sessions scheduled for the Geological Society of America South-Central Section meeting in Little Rock in 1973 were filled and additional pertinent papers were presented in other sessions. Some papers that were primarily concerned with Ouachita geology but were presented in other sessions of that meeting are included in the volumes as are some papers not presented at any of the sessions.

This is not the first symposium on the geology of the Ouachitas--nor will it be the last. The paucity of good bedrock exposures limits the detailed information available in this area of complex structure and stratigraphy. Consequently, each increment of new data takes us a step nearer to good knowledge of the geology of this fascinating area.

Pre-1900 workers on Ouachita geology carried some names famous in the profession -- R.A.F. Penrose, J. C. Branner, Joseph A. Taft, and E. O. Ulrich. Some less famous but sound workers made major contributions, such as the monumental work of L. S. Griswold on whetstones and novaculites.

After a lull of more than a decade, other prominent geologists entered the studies -- A. H. Purdue, his protege H. D. Miser, C. W. Honess, Sidney Powers, and W.A.J.M. van Waterschoot Van der Gracht, the Dutch geologist who first called attention to the flysch-like character of some of the Ouachita sediments. These stalwarts worked remarkably well with little in the way of base maps or other aids.

In 1934, controlled mosaics were prepared by Edgar Tobin Aerial Surveys for the Amerada Petroleum Corporation of an area of about 2,000 square miles in Oklahoma. This permitted mapping and other studies by Bruce H. Harlton, J. V. Howell, and their associates. When the period of exclusive use of these aerial photos expired, a new set of experienced field geologists used them as a base for extensive mapping, particularly in Oklahoma. This group included Rolf Engleman, Henry Carter Rea, Frank Notestein, Vaughn Russom, Roy P. Lehman, Paul Averitt, and myself. This phase culminated in detailed mapping and presentation of measured sedimentary direc-

tional features on the excellent topographic base of the Waldron quadrangle in western Arkansas by John A. Reinemund and Walter Danilchik and laboratory studies by August Goldstein, Jr., J. W. Bokman, C. E. Weaver, and others.

Since 1960, several theses have been prepared on the geology of the Ouachitas in Oklahoma by students from the Universities of Wisconsin and Oklahoma under the guidance of Lewis M. Cline and Kaspar Arbenz. In Arkansas much work was done by Boyd R. Haley, MacKenzie Gordon, Jr., Donald A. Brobst, Charles G. Stone, Philip J. Sterling, Drew F. Holbrook, and others for the U. S. Geological Survey and the Arkansas Geological Commission. W. E. Ham, who worked mostly west of the Ouachitas, made a major contribution to the overall knowledge by calling attention to the "starved basin" character of the older Paleozoic sediments of the Ouachitas.

Major syntheses of the geology of the Ouachita system have been published by Hugh D. Miser; August Goldstein, Jr., and T. A. Hendricks; and Peter T. Flawn, August Goldstein, Jr., Philip B. King, and C. E. Weaver. The last includes a very comprehensive bibliography.

Remote sensing from satellites and high-flying aircraft, together with geophysical studies, is likely to permit recognition of significant trends and lineaments that will add another dimension to studies of the geology of the Ouachitas, but it remains to be seen whether such information will solve more problems than it adds. The geology of the Ouachita Mountains still is one of the frontiers of the science.

It is appropriate that these volumes be dedicated to Hugh D. Miser. He established the stratigraphic nomenclature used in the Ouachita Mountains. His stratigraphic units have stood the test of their extension into parts of the area beyond those mapped by him, and have also stood the more severe test of detailed studies conducted by later workers. This is a tribute to Miser's adherence to the basic principle that stratigraphic units must be recognizable and mappable. Mr. Miser also contributed greatly to an understanding of Ouachita geology by his unstinting willingness to draw on his extensive knowledge to help less experienced geologists and students to gain the background and inspiration on which to base further studies.

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PART I
DEDICATION OF SYMPOSIUM VOLUMES TO
THE MEMORY
OF



HUGH DINSMORE MISER
(1884-1969)

The following brief eulogies to Hugh Dinsmore Miser are printed by permission from the Tulsa Geological Society in "Tulsa's Physical Environment, Digest Volume 37, 1972".

A TRIBUTE TO
HUGH DINSMORE MISER

By George V. Cohee¹

Dr. Hugh Miser, one of our most eminent geologists, died of a heart attack at his home in Washington, D. C. on August 1, 1969. Hugh Miser was born at Pea Ridge, Arkansas, December 18, 1884, and received his early education in the local schools at Pea Ridge and then attended the University of Arkansas where he received his Bachelor's and Master's degrees, the latter in 1912.

He worked for the U.S. Geological Survey first as a Geologic Aid in 1907, and as a Junior Geologist in 1910, and was appointed Associate Geologist in 1912. His entire professional career of about 60 years was with the U.S. Geological Survey, with the exception of a one-year leave of absence to serve as Acting Professor of Geology at the University of Arkansas and Acting State Geologist of Arkansas, and another year to serve as State Geologist of Tennessee. He served in many important administrative positions on the Federal Survey; and, through his effective scientific and administrative guidance, substantial contributions were made to the knowledge of the geology and economic resources of our country. From 1955 until the day before his death, he was Scientific Staff Assistant in the Office of the Director, where he reviewed and approved manuscripts and maps for publication.

Hugh Miser was internationally recognized as one of the outstanding petroleum geologists in the world and as a leading authority on the structure of rocks in the central area of North America. Because of his distinguished career as a scientist, administrator, and advisor and because of his contributions to geologic science, he was awarded a Doctor of Law degree from the University of Arkansas in 1949 and the Department of Interior's highest honor, the Distinguished Service Award, in 1955.

As Hugh Miser influenced those who were privileged to know him and work with him, he was greatly influenced by his inspiring teachers of high school and college days. In his words, his teachers were men of strong character, and their enthusiasm was kindled by their search for truth. He cherished and lived by certain philosophical do's, don'ts, and musts that were absorbed in his school days. I believe the following provided a basic philosophy for his able and productive professional career:

1. Each geologist must be honest with himself in the use of facts; he must make certain that his facts are truly facts.
2. Anything worth doing is worth doing well; do your

¹U. S. Geological Survey, Washington, D. C. 22092

best always; do not guess.

3. Do not worry about the future; do each day's work well and happily; if you do, the morrows will take care of themselves, for each in turn will become today, one at a time.
4. If something can be done today, do not postpone it until tomorrow.
5. If a job can be finished in one day, do not take two days.
6. If something can be said in two words, do not use three words.
7. A writer must stick to his subject.
8. Ability to write well requires work including concentration and logical thinking; conversely, inability to write indicates inability to think and concentrate; thus poor writing denotes inability of author to work when he attempts to write.
9. The incentive of a geologist should be love for his work and should not be salary which is in reality a means to attain ambition's goal. This last philosophy may be labeled oldfashioned by some. But do not Americans in general love to work for the satisfaction and enjoyment they receive as rewards for making and growing useful things?

I want to include one additional statement that he made many times in giving counsel to young geologists regarding their work . . . "Make it simple and make it fun."

It was always a gratifying experience to have one's geologic report reviewed by Hugh Miser because of the feeling that the report fared better for his having seen it. Also, whenever an author conferred with him on a manuscript, he inevitably came away richer by several stories from Miser's never-failing collection. I shall never forget the first geologic report that I brought to Washington for Hugh Miser to review. He, at that time, was Chief of the Fuels Branch and he made a point of personally reviewing each geologic report of the Branch. We were going over the maps, which were spread on tables in his office, when he came upon the term "iso-pachous" that had been used on one of the maps. He looked at it for a few seconds and repeated the term and finally said in his gentle way, "Why don't we use the word 'thickness,' then every one will understand what we mean?" This was a very kind way of letting me know that ordinary terms

should be used if they convey the proper meaning.

One time, when I was serving as map processor for the Fuels Branch, a map was prepared and was about ready to be printed. On this map we found that one of the words had an additional "s" through error and I went over all the pieces of copy correcting this error. When the review was completed and I was satisfied that I had removed every additional "s," the material was taken to the printing shop and the edition printed. The finished product was very attractive, and I proudly took the first copy from the press to Miser for his reference. He knew, of course, of my trying to delete all of the "s's" and, instantaneously, he said, "What's that?", and lo and behold, under a heavy blue contour line appeared one of those "s's" that I had been removing. Naturally, I was a bit overwhelmed that this had been overlooked in the intensive review given the copy. Hugh hurriedly said, "You know, we are striving to get out the perfect map, and I am sure we will have to continue working toward that end for some time to come." Those were indeed most kind words to me at that particular time. They showed the greatness of the man in not only making a point clear to me, but doing it in such a kind, gentle and impressive way.

Out in the field, one did not tell Hugh Miser there were no fossils in the rocks under observation. Invariably, he would go over the outcrop with his keen eyes and soon a few fossils would appear in his hand to prove you wrong. Other geologists, who have been in the field with him, have shared this experience, which, I might add, was always quite impressive.

During part of the time that Hugh was preparing the second geologic map of Oklahoma, I was teaching in

the Department of Geology, University of Arkansas. Frequently, Hugh and Mrs. Miser would come to Fayetteville for the weekend, and I was in very close touch with the progress on the compilation of the new map. His life was devoted to the map during the time he was working on it. It seemed to me that those were some of his most happy years because he was in the field studying the geology that he loved and was associated with his fellow geologists in Oklahoma and Arkansas. In later years, actually up until he died, he looked forward each year to an annual pilgrimage to Arkansas and Oklahoma for a month of field work, and, of course, for seeing his friends again. As we all know, he was a great field geologist.

Hugh Miser looked to the future and the well being of our country and our science, as shown in the following statement that he made more than twenty years ago.

"Geologists will always be confronted with the solution of difficult geologic problems in spite of the ever increasing progress of the science. Whatever their faith and wishes, they will never reach the utopian day when all geologic problems are solved and all geologic facts are known. Some love the science of geology for itself; others love it for the opportunity of applying the science to engineering and economic ends. Geologists who are seeking new knowledge are keyed with hope and optimism; and they are pursuing man's noblest occupation. Their future, like the present, holds work and hope, for the progress of geology and other fields of scientific endeavor means new knowledge, new things, more jobs, and advancement in human welfare."

E A R L Y L I F E O F H U G H D I N S M O R E M I S E R

By Lloyd G. Henbest¹

The work of Hugh Dinsmore Miser has an eminent place in American geology. An equally or more notable feature of his career was his character and his influence on geology and geologists. In this brief tribute, I wish to give some insight into the origins of his character and historical background for the early part of his career in Arkansas and Oklahoma.

Miser was born on a farm near Pea Ridge, Arkansas, December 18, 1884, the third child of Jordan Stanford and Eliza (Webb) Miser. His parents and grandparents were born in that region and were descended from Pennsylvania Dutch and English pioneers. The roots of his unique character and career are revealed in the people, the rural community, the geological setting, and the history into which he was born. His character personified the admirable qualities of those origins. These qualities and his sentiments for his origins were among his most distinctive and attractive traits.

Miser's first act in life was to create a rural crisis by being born during a snowstorm. Other stories that he has told about his boyhood suggest that traits that distinguished him in later life were in evidence at an early age. There was precedent within his family and relatives for his rugged intelligence, independence, and sense of values. It is also evident that he had plenty of opportunities to exercise his shrewdness or capacity for restraint in avoiding futile clashes between enlightened reason and traditional thought. At gatherings, the farmers enjoyed spinning jokes or recounting droll stories about their own or other's foibles. In young Miser they must have had a most apt apprentice because his extraordinary store of rustic anecdotes and droll manner of recounting his experiences were to become the delight of people wherever he went. In such entertainment, however, he was a true humorist. He recounted his experiences and store of folklore not as one standing apart but as one with empathy. I have known him to be angered at persons who told rustic stories snobbishly. It was for that understanding spirit of his that people of all stations of life had confidence in him. I doubt that he was ever kept out of a locality because the local residents were suspicious of strangers.

Many areas that Miser surveyed as a young man were inaccessible except on foot or on horseback. He sometimes had to make up reasonable excuses for preferring to sleep on the porches of native cabins, but he was willing to pay the prices of poor food and vermin for doing

this kind of geology which has not been surpassed. He had an extraordinary capacity for hard work and his love for geology possessed him. All his life, he had a youthful zest for seeing geology or new phenomena and expressed pleasure on seeing good work by others. It was for all those reasons that he was regarded as a natural in the field. Many colleagues learned to be cautious about drawing conclusions in his presence on the basis of insufficient field work.

Returning to the influences on Miser's early life and career, the geologic and geographic setting into which he was born were an equally germinal factor in the early history of geology in Arkansas and Oklahoma. The natural features of the upland from Fayetteville, Arkansas, northward to Springfield, Missouri attracted early settlement. The settlements and the greater accessibility of the region as compared with the rugged White River hills on the east and with the more arid plains and Indian Territory on the west, caused this region to develop rapidly and become the main communication route from St. Louis to the southwest. The Butterfield Trail (1857-1862), the stage and mail route from St. Louis to the southwest and southern California, had a station at the Elkhorn Tavern near Pea Ridge. The elongate hill and landmark called Pea Ridge was also the scene of three Civil War battles on March 7-8, 1862, that secured Missouri and northwest Arkansas for the Union. The Pea Ridge Normal School, a preparatory and vocational school, was one of the early "Academies" of the region. It was attended by Miser. Its principal, Benjamin Harvey Caldwell, is reported to have been an inspiring teacher in the school and community.

The stratigraphic section and its abundant record of ancient life, particularly that of the Mississippian and Pennsylvanian, attracted the attention of pioneer geologists and local residents from the earliest days. This sequence which has become a standard of reference, was a strong factor in the history that concerns us here and in Miser's career. The geology has inspired an exceptional number of native sons and students to become geologists. Notable among these was the geologist Curtis Fletcher Marbut (1863-1936) of Cassville, Missouri, former Director of the U. S. Soils Survey, who became known as the father of American soils science.

When Miser entered the University of Arkansas in 1903, he had not chosen a profession. He came under the influence of A. H. Purdue, Professor of Geology (1896-1912), who was, as Miser was to become, a product of the local history and natural setting. The 1858-1860 reports of the David Dale Owen Survey of Arkansas gave the first comprehensive reconnaissance and correlations of the Ozark section of Arkansas. The John Casper Branner Survey in

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the 1880's and early 1890's was notable in several ways. Its reports and work made it a model for other state surveys and a strong influence locally. Its greatest contribution, however, consisted both of the number of young employees who became distinguished geologists and of Branner's influence as a teacher and role in American science. J. A. Taff, whose place in the history of geology in Oklahoma needs no review here, started school as a civil engineer at the University of Arkansas and started his career as an employee of Branner in the Ouachita Mountains. A. H. Purdue, a later employee of Branner and student of his at Stanford, was a very energetic and able geologist. When he came to Fayetteville to teach, he began surveying the area thereabouts as an avocation. This avocation became important for Purdue and ultimately for his student Miser because it resulted in Purdue's employment by the U. S. Geological Survey to aid G. I. Adams and E. O. Ulrich complete the surveys of the Fayetteville quadrangle and of the lead and zinc deposits of north Arkansas and eventually to survey the Winslow quadrangle. Purdue's success as a geologist was comparable to the extraordinary achievements of Taff's U. S. G. S. surveys in Oklahoma which were nearing completion.

Purdue became the State Geologist of Arkansas in 1907 and contracted with the U. S. G. S. to survey the DeQueen-Caddo Gap quadrangles in the Ouachitas. He employed Rector D. Mesler and Hugh D. Miser, his students, as assistants. It is hard to imagine a more rigorous test of an undergraduate's intelligence and motivation in geology than to be transported from the undisturbed, paleontologically and lithologically differentiated section in north-west Arkansas, to the folded, poorly differentiated, enormously thick, and virtually non-fossiliferous rocks of the DeQueen area, much of which was accessible only on foot. The nature of Miser's and his mentor's success is indicated by the rapid succession of new assignments in Arkansas and Tennessee.

Taff had hoped to survey the entire Ouachita region, but as new opportunities developed elsewhere, he never returned to extend his early surveys. Miser inherited the role. Though Miser became Chief of the Fuels Branch of the U. S. G. S. in 1928 and his duties became nationwide, his love of the geology of Arkansas and his adopted state, Oklahoma, never waned.

C O M M E N T S O N S O M E
C H A R A C T E R I S T I C S O F
H U G H D I N S M O R E M I S E R

By Thomas A. Hendricks¹

Hugh Dinsmore Miser was a man of many facets. It would be presumptuous, if not impossible, for me to attempt to treat with more than a few of those facets.

He was a man of principle. Sometimes adherence to his principles led to his taking a position that was misunderstood by his peers. One such situation occurred in 1948 when he was asked by the nominating committee of the American Association of Petroleum Geologists to run for president against his long-time friend, C. W. Tomlinson. This was the first election in which two candidates were nominated in accordance with a procedure recommended by a committee chaired by another old friend -- John G. Bartram. Miser doubted that a government geologist, particularly one in an administrative position, should be president of the A. A. P. G. because of the possibility of conflict of interest arising over perfectly proper actions of the Association that might be in conflict with government regulations or actions. He also felt that an honorary membership, which he had proudly accepted, essentially eliminated a geologist from candidacy for the presidency. After discussing these questions with a few of his associates, he reached his own conclusion and, with personal reluctance, declined the invitation to be a candidate. Debates within the A. A. P. G. and affiliated societies during the subsequent year on such subjects as regulatory practices, conservation, and economic aspects of well spacing showed clearly the wisdom of Miser's decision.

Miser was thorough in everything he did and he insisted on thoroughness in the work of geologists under his supervision. During compilation of the two editions of the Geologic Map of Oklahoma he assembled maps from every possible source. If any question arose regarding the area

covered by a map, the geologic units shown, the scale, orientation, or the responsibility of authorship, he would not use that map. Every map used was also subjected to a rigorous field check. In this connection, he believed that the only place to resolve a question in geologic mapping was on the critical exposures in the field with both parties to the disagreement present and free to express their opinions.

Miser solicited and listened to criticism. For example, many petroleum geologists believed and informed him that the value of geologic mapping by geologists of the U. S. Geological Survey was seriously depreciated by the slow pace of publication. Miser acted to meet this criticism. He and members of his Fuels Branch at his request conceived the establishment of a series of Preliminary Oil and Gas Maps and Charts. Miser gave me the privilege of submitting Preliminary Map No. 1 in response to industry requests. This series served a very useful purpose at the time and it contributed to the subsequent acceleration of U. S. G. S. map publication.

Another of Hugh Miser's attributes was his loyalty, both to persons and organizations, which at times reached extreme proportions. Once a person established himself ethically or scientifically with Hugh, he would defend that person's position to the full strength of his ability. I have been defended by Hugh, and in a few instances I have had to yield to his defense of another.

Miser believed that a field geologist should see every piece of field evidence before reaching a conclusion on interpretive questions. He felt also that an interpretation was justified only if it was consistent with *all* known facts. In other words, he strove for perfection in his own work and sought to instill perfection into every geologist and into every piece of Geological Survey work that came under his official scrutiny.

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PART II
TECHNICAL PAPERS
STRUCTURAL AND STRATIGRAPHIC CONTINUITY OF THE
OUACHITA AND APPALACHIAN MOUNTAINS

By William A. Thomas¹

ABSTRACT

Regional stratigraphic and structural relations of the Ouachita and Appalachian Mountains suggest an interpretation of structural continuity of the mountain systems; the regional Ouachita-Appalachian structure appears analogous to other salients and recesses elsewhere in the Appalachians. A curve in the structural front around the Ouachitas defines a structural salient. East of the Ouachitas in the subsurface, the structural front trends southeastward and converges with Appalachian structures in central Mississippi. That southeast-trending segment of the Ouachita front is not necessarily parallel with the strike of individual structures, and it may be a line along which east-striking frontal folds flatten and end. East of the Ouachita salient, the structural system curves into a recess in Alabama.

Lower Paleozoic rocks within the Ouachita salient are characterized by black shale. On the west, north, and east, an equivalent carbonate facies rims the area of the Ouachita-shale facies. Distribution of Devonian-Lower Mississippian chert appears to be centered on the Ouachitas. Upper Paleozoic rocks of the Ouachitas comprise a thick flysch sequence, but the clastic sequence thins to the west, north, and east. Hence, in the Ouachita salient, the succession includes the lower Paleozoic black shale facies and the thick upper Paleozoic clastic sequence; but, eastward toward the Alabama recess, the structural system crosses into a lower Paleozoic carbonate facies and a thinner upper Paleozoic clastic sequence. Location and curvature of the Ouachita salient are interpreted to be related to the distribution of sedimentary facies and thickness.

INTRODUCTION

Paleozoic structures of the Appalachian Mountains plunge southwestward beneath the cover of Gulf coastal plain strata in central Alabama, and similarly Paleozoic structures of the Ouachita Mountains plunge eastward beneath the Gulf coastal plain in central Arkansas (Fig. 1). Wells drilled through the coastal plain cover are sparse, but available subsurface data show that a belt of deformed rocks extends from the Appalachians westward to the Ouachitas (Thomas, 1973). The sparse data allow for several different interpretations of geometry of the structural system (King, 1950; King, in Flawn and others, 1961, p. 97; Vernon, 1971; Thomas, 1973).

The exposed Appalachians in Alabama include a frontal belt of folded and thrust-faulted sedimentary rocks and an interior belt (Piedmont province) of metamorphic rocks (Fig. 1). Structures within the frontal belt involve a lower Paleozoic carbonate sequence and an upper Paleozoic clastic sequence (Table 1). The subsurface fold and thrust belt in western Alabama and eastern Mississippi includes at least two major structures and one apparently less extensive frontal structure. Structural strike apparently curves

gradually westward, and subsurface data from about 30 wells indicate that the structural style of the fold and thrust belt persists as far west as central Mississippi (Fig. 1; Central Mississippi deformed belt of Thomas, 1973). No subsurface data are available farther west along the projected trend, and the westward extent and limits of the belt are unknown. A stratigraphic sequence, with a few exceptions like that in the Alabama Appalachians, also extends as far west as central Mississippi (Table 1). The Talladega Slate belt of metasedimentary rocks along the northwest side of the Appalachian Piedmont may be identified as far west as eastern Mississippi, and higher grade metamorphic rocks of the Piedmont province are known as far southwest as southern Alabama (Thomas, 1973, Fig. 4).

Exposed Ouachita structures include thrust faults and folds which involve a lower Paleozoic succession of black shale, sandstone, chert, and limestone and a much thicker upper Paleozoic clastic sequence (Table 1). The Ouachita structural system gives the impression of being more complex and including more disharmonic structures than the folded and thrust-faulted belt of the Alabama Appalachians. Apparent differences in structural style possibly reflect thicker and more numerous units of incompetent rocks which alternate with relatively competent units in the Ouachita succession. Rocks of the Ouachita core zone exhibit tight folds and slaty cleavage (Miser and Purdue,

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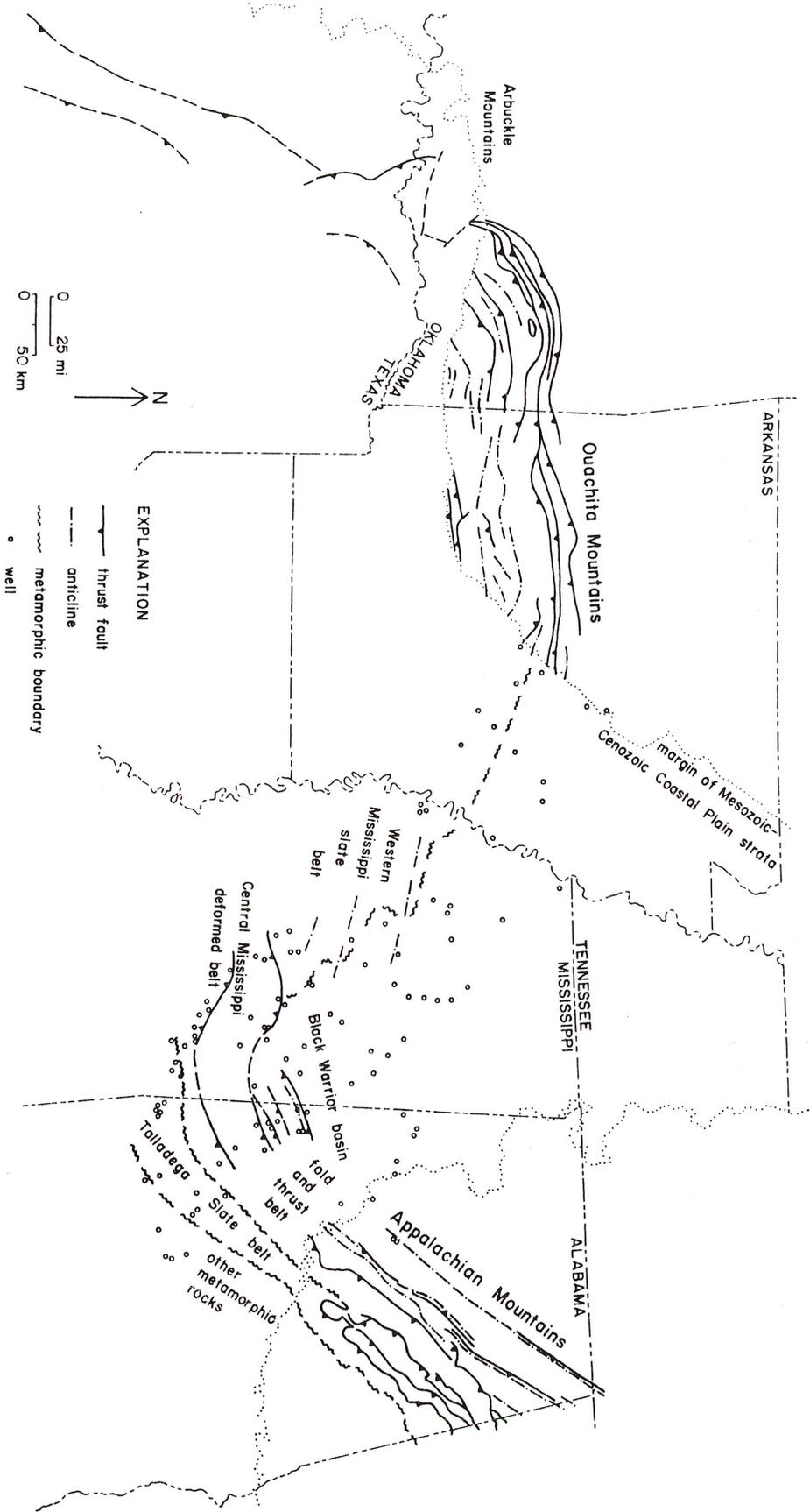


Figure 1. - Regional structural geology of Ouachita and Appalachian Mountains. Structural interpretation adapted from Flawn and others (1961), Stone (1966), and Thomas (1973). Walls on which interpretation is based are identified by Flawn and others (1961), Caplan (1964), and Thomas (1972a, 1973). Note: metamorphic boundaries may or may not be faulted.

	OUACHITA MOUNTAINS	MISSISSIPPI	APPALACHIAN MOUNTAINS, ALABAMA		
MISSISSIPPIAN - PENNSYLVANIAN	ATOKA FORMATION SHALE; SANDSTONE; COAL IN UPPER PART. 19000+ FT.	POTTSVILLE FORMATION SANDSTONE; SHALE; CONGLOMERATE; COAL. 10000 FT.	POTTSVILLE FORMATION SANDSTONE; SHALE; CONGLOMERATE; COAL. 9000 FT.		
	JOHNS VALLEY SHALE SHALE; ERRATIC BOULDERS; SANDSTONE. 1500 FT.				
	JACKFORK GROUP SANDSTONE; SHALE; SILICEOUS SHALE. 7000 FT.				
	STANLEY GROUP SHALE; SILICEOUS SHALE; SANDSTONE; TUFF IN LOWER PART. 12000 FT.	PARKWOOD FORMATION SANDSTONE; SHALE. 2000 FT.	PARKWOOD FM. SANDSTONE-SHALE FACIES ON SOUTHWEST. 2600 FT.	BANGOR LS. LIMESTONE FACIES ON NORTHEAST. 700 FT.	
	DEVONIAN	ARKANSAS NOVACULITE 950 FT.	FLOYD SHALE SHALE; SANDSTONE-LIMESTONE IN LOWER PART. 850 FT.	FLOYD SHALE 1700 FT.	HARTSELLE SANDSTONE 150 FT. PRIDE MTN. FM. SHALE; SANDSTONE. 400 FT.
			FORT PAYNE-TUSCUMBIA CHERT; CHERTY LIMESTONE. 220 FT.	TUSCUMBIA LIMESTONE CHERTY LIMESTONE. 200 FT.	FORT PAYNE CHERT CHERT; SILICEOUS LIMESTONE. 200 FT.
SILURIAN	MISSOURI MOUNTAIN SHALE 300 FT.	UN-NAMED SILICEOUS LIMESTONE; CLAYSTONE; DOLOSTONE. DARK-COLORED SHALE AND LIMESTONE ON SOUTHWEST. 700 FT.	MAURY SHALE GREEN SHALE. 10 FT.	CHATTANOOGA SHALE BLACK SHALE. 25 FT.	
	BLAYLOCK SANDSTONE 1500 FT.		-HIATUS-	FROG MOUNTAIN SANDSTONE 200 FT.	
			-HIATUS-	RED MOUNTAIN FORMATION SANDSTONE; SHALE; LIMESTONE; HEMATITE. 500 FT.	
MIDDLE & UPPER ORDOVICIAN	POLK CREEK SHALE BLACK SHALE; SANDSTONE; CHERT. 175 FT.	CHICKAMAUGA GROUP LIMESTONE; DOLOSTONE; SANDY LIMESTONE- DOLOSTONE AND SANDSTONE AT BASE. BLACK SHALE TONGUE NEAR TOP PINCHES OUT EASTWARD. SANDSTONE AT TOP ON NORTH. 3000 FT.	CHICKAMAUGA GROUP LIMESTONE; LOCAL CHERT CONGLOMERATE AT BASE. 900 FT.		
	BIGFORK CHERT CHERT; BLACK SHALE; LIMESTONE. 800 FT.				
	WOMBLE SHALE BLACK SHALE; LIMESTONE; SANDSTONE. 3500 FT.				
LOWER ORDOVICIAN	BLAKELY SANDSTONE BLACK SHALE; SANDSTONE; CHERT; BOULDERS. 400 FT.	KNOX GROUP DOLOSTONE; CHERTY DOLOSTONE.	KNOX GROUP DOLOSTONE; CHERTY DOLOSTONE. 3000 FT.		
	MAZARN SHALE BLACK SHALE; LIMESTONE; SANDSTONE. 3000 FT.				
	CRYSTAL MTN. SANDSTONE 850 FT.				
	COLLIER SHALE BLACK SHALE; LIMESTONE; CHERT. 1000+ FT.				
	-NO OLDER ROCKS EXPOSED-				

Table 1

Generalized Paleozoic stratigraphic columns (compiled and modified from Butts, 1926; Flawn and others, 1961; Sterling and others, 1966; Stone, 1966; Thomas 1972a; 1972b; Stone and others, 1973; Thomas and Drahovzal, 1973). Thicknesses are an approximate maximum for each area. Because maxima of different formations do not coincide geographically, total sedimentary thickness at any locality is less than the sum of the formation maxima for each area. Thicknesses of most units in Mississippi are from the Black Warrior basin, because formation thickness is generally undetermined within the Central Mississippi deformed belt.

1929, p. 118; Viele, 1973, p. 367). A belt containing quartz veins extends along the length of the Ouachitas (Miser, 1943, p. 94; 1959, p. 37; Engel, 1952).

In the subsurface of western Mississippi, a dark-colored shale succession is in part characterized by slaty cleavage (Fig. 1; Western Mississippi slate belt of Thomas, 1973). Quartz veins are also common in parts of the area. Identification of the subsurface Western Mississippi slate belt is based on the presence of slaty cleavage and quartz veins. However, within the area, slaty cleavage is not consistently distinct, and some rocks are not slaty. Presence of slate and quartz veins indicates an Ouachita deformational style and suggests that Ouachita structures extend into western Mississippi. The Western Mississippi slate belt apparently extends to the foreland side of Appalachian structures and, thus, is not comparable in tectonic setting to the Talladega Slate belt of the Appalachian interior (Fig. 1).

Although wells are sparse, a generalized structural front may be drawn northwestward from central Mississippi into an arc that curves around the exposed Ouachitas in Arkansas and Oklahoma (Fig. 1). The Ouachita structural system similarly may be traced southward in the subsurface of eastern Texas (Flawn, in Flawn and others, 1961, Pl. 2). The curve in outline of the Ouachita structural system defines a major structural salient convex toward the North American craton. East of the Ouachita salient, the structural system curves into a recess in Alabama. To the southwest, the structural system extends from the Ouachita salient into a major recess around the Llano uplift and farther west into the Marathon salient (Flawn, in Flawn and others, 1961, p. 166).

Enough data are available now to establish continuity of a belt of deformed rocks from the exposed Appalachians to the exposed Ouachitas; however, many details within the connecting structural system, particularly relation of the Western Mississippi slate belt to "Appalachian-style" structures, remain uncertain. Problems of specific structural interpretations and details of the subcrop map pattern evidently cannot be resolved by further review of presently available subsurface data. However, interpretation of the regional structural pattern and of the regional stratigraphy of Paleozoic rocks provides the basis for a working structural model of the "junction" of Ouachita and Appalachian structures. Some characteristics of Ouachita and Alabama Appalachian structures seem to be related to their positions within the regional salient and recess; these may be examined by analogy with exposed salients and recesses elsewhere in the Appalachians. Structures of the Ouachita salient and Alabama Appalachian recess are formed within substantially different sedimentary facies, and a genetic relationship between regional structure and stratigraphy is implied. The purpose of this paper is to review available structural and stratigraphic data from the area of the Ouachita-Appalachian junction, to compare structure and stratigraphy of other regional salients, and to develop a comprehensive Ouachita-Appalachian structural-stratigraphic model.

LOWER PALEOZOIC (CAMBRIAN-SILURIAN) STRATIGRAPHY

The lower Paleozoic succession in the Ouachita Mountains is characterized by black shale (Table 1). Conodont biostratigraphy indicates that the oldest rocks exposed in the Ouachitas are Early Ordovician (Repetski and Ethington, 1973, p. 277). Silurian rocks comprise only a small part of the total sequence, and the Silurian Blaylock Sandstone pinches out northward across the Ouachita outcrops (Sterling and others, 1966, p. 184). The Ouachita black shale facies includes distinctive units of chert, sandstone, limestone, and boulder conglomerate (Table 1). Although thickness and proportion of the different components vary, the characteristic lower Paleozoic black shale facies extends throughout the Ouachita outcrops. In contrast, in areas adjacent to the Ouachitas, the lower Paleozoic strata are mainly carbonate, and the carbonate succession generally includes quartzose sandstone and sandy carbonate (Fig. 2).

In the Arbuckle Mountains, the lower Paleozoic succession is mainly carbonate and contains quartzose sandstone units (Ham, 1959, p. 71). The Sylvan Shale is a distinctive interbed of dark-gray and greenish-gray shale in the upper part of the carbonate succession (Ham, 1959, p. 77; Frezon, 1962, p. 39). The Sylvan is interpreted to be a tongue of the Ouachita shale facies, and that shale tongue has served as a tie between the Arbuckle and Ouachita facies (Ham, 1959, p. 77).

Similarly, north of the Ouachitas in northern Arkansas, the lower Paleozoic succession is mainly carbonate and contains units of quartzose sandstone (Fig. 2). The Cason Shale within the carbonate succession of northern Arkansas is lithologically similar to the Sylvan Shale (Maher and Lantz, 1953; Frezon, 1962, p. 39), and the Cason evidently is also a tongue of the Ouachita facies. The Cason may be part of the same widespread shale unit as the Sylvan (Ham, 1959, p. 77); however, part of the Cason may be younger than the Sylvan (Wise and Caplan, 1967, Fig. 2; Craig, 1973, p. 253).

East of the Ouachitas in the subsurface of Mississippi, the lower Paleozoic succession is almost entirely carbonate, but it includes some sandy intervals (Table 1; Fig. 2). In north-central Mississippi, the Upper Ordovician includes a thin unit of black shale which pinches out eastward into the carbonate facies and has been drilled in only a few wells (Fig. 2; Thomas, 1972a, Fig. 7). The black shale is overlain by a sandstone unit which may be a distal tongue of the Sequatchie clastic wedge of the Tennessee Appalachians (Thomas, 1972a, p. 92). In central Mississippi, one well (Carter Oil Company No. 1 Denkman, Leake County) penetrated dark-gray and black shales and limestones (Thomas, 1972a, Table 5) which contain Silurian brachiopods (King, in Flawn and others, 1961, p. 355). This well marks the most southwesterly known Silurian rocks in Mississippi, and the dark-colored shales and limestones contrast with the lighter colored Silurian carbonate rocks and thin claystones farther northeast in Mississippi

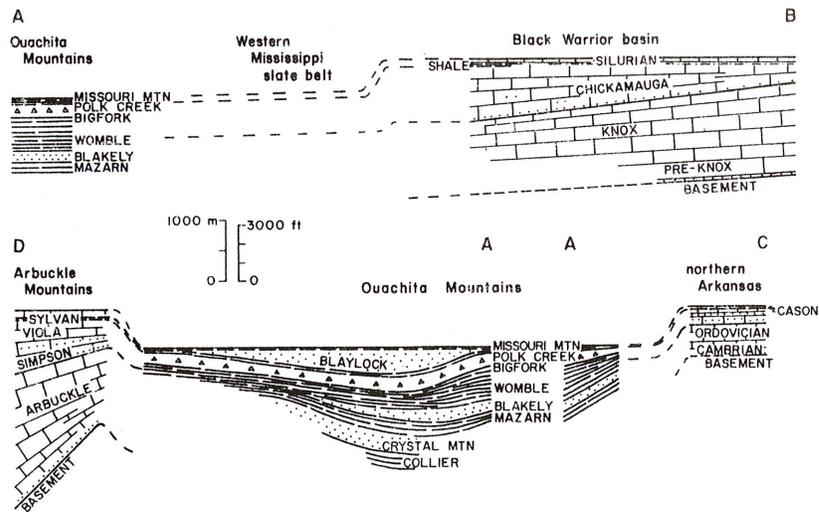
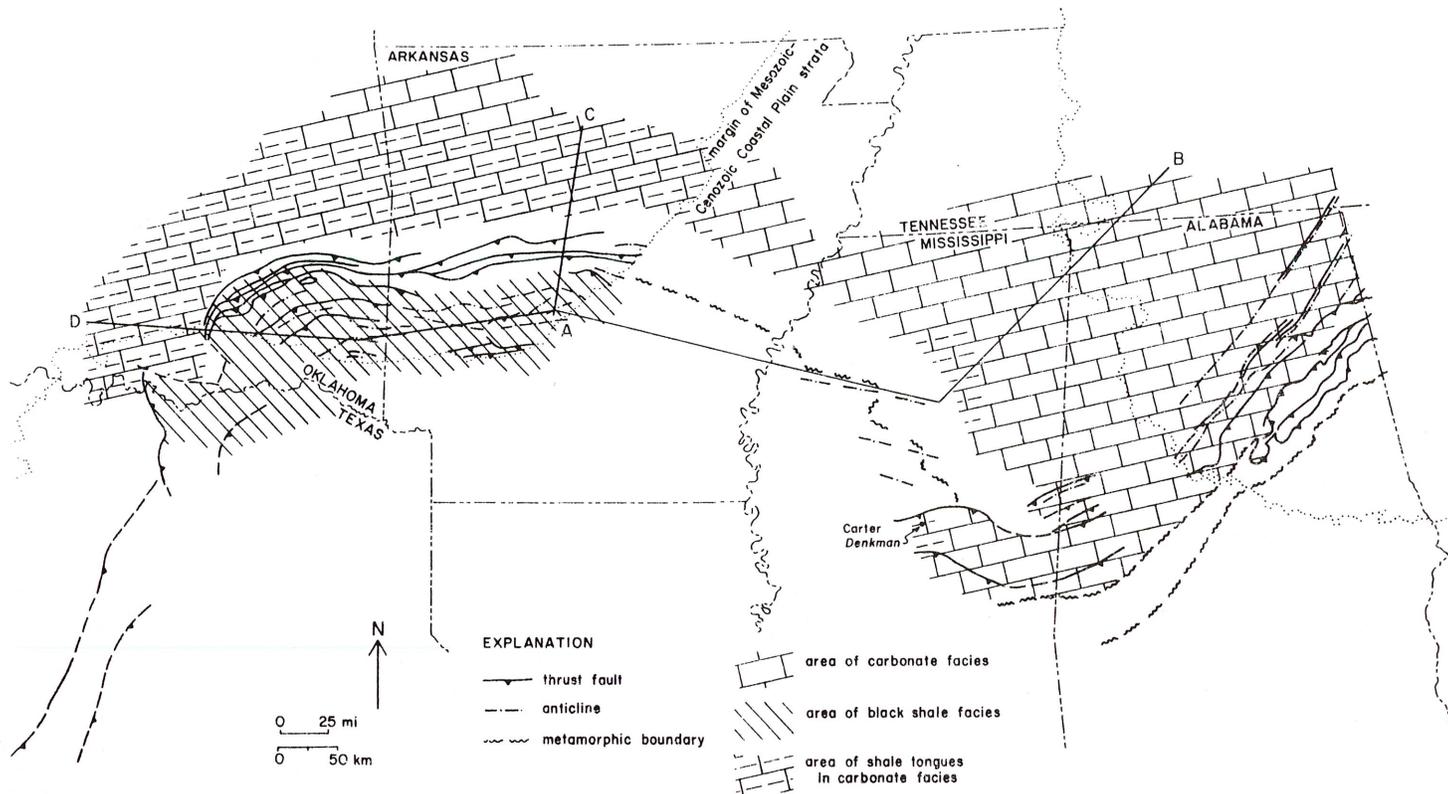


Figure 2. - Lower Paleozoic (Cambrian-Silurian) stratigraphy. Symbols on cross sections designate stratigraphic units but do not show lithologic details. Cross sections show interpreted depositional framework. Data from Miser and Purdue (1929), Wilson (1949), Maher and Lantz (1953), Huffman and others (1958), Frezon and Glick (1959), Ham (1959), Milhous (1959), Flawn and others (1961), Frezon (1962), Caplan (1964), Sterling and others (1966), Sellars (1967), Berry and Trumbly (1968), Haley and Hendricks (1968), Thomas (1972a).

(Thomas, 1972a, p. 96). In summary, the Ordovician and Silurian in the subsurface of Mississippi include two units which suggest tongues of the Ouachita black shale facies: (1) a thin tongue of black shale that pinches out eastward into the Ordovician carbonate sequence; and (2) dark-colored shale and limestone on the southwest in the Silurian (Fig. 2).

From the Arbuckle Mountains across northern Arkansas to central Mississippi, a lower Paleozoic carbonate facies rims the area of an equivalent black shale facies in the Ouachita Mountains (Fig. 2). Evidently the carbonate facies must change to black shale toward the Ouachitas. East of the Arbuckle Mountains, the Ouachita black shale facies is now in close proximity to the carbonate facies (Fig. 2), and subsurface data indicate that the Ouachita shale facies has been thrust over Arbuckle rocks (Flawn, in Flawn and others, 1961, Pl. 2). Although the original horizontal distance between the Ouachita and Arbuckle facies has been shortened by thrusting, the extent to which the black shale has been thrust over the carbonate facies is unknown. Shale tongues in the carbonate facies of the Arbuckle Mountains and northern Arkansas may be an indication of proximity to the facies boundary. Like the other shale tongues, the thin eastward-pinching black shale in the carbonate sequence of north-central Mississippi suggests a tie to the Ouachita facies and possible proximity to the facies boundary. Available data are not adequate to define the eastward extent of the Ouachita black shale facies. However, the known subsurface area of the carbonate facies and extent of the black shale tongue suggest that possibly the Ouachita black shale facies extends into western Mississippi, perhaps across the Western Mississippi slate belt (Fig. 2). Apparently the facies boundary crosses western Mississippi in a generally southeastward or southward direction. Several wells in western Mississippi have drilled dark-colored shale, the age of which is not precisely known; however, the lithology is similar to known upper Paleozoic rocks elsewhere in the region. Thus, no known lower Paleozoic shale facies has been drilled in western Mississippi, and precise definition of the eastward extent of the Ouachita shale facies awaits additional drilling.

Distributions of the carbonate and black shale facies suggest the interpretation that the facies boundary marks the steep edge of a shallow carbonate bank (see Rodgers, 1968, for discussion of carbonate bank and deep-water shale facies in the northern Appalachians). The carbonate facies and interbedded quartzose sandstone units reflect the shallow shelf environment. Shale tongues within the carbonate facies indicate transport of fine-grained clastic sediments across the bank. Apparently the carbonate bank around the Ouachita region outlined a basin in which black shale and related sediments accumulated. The inferred deep-basin setting of the black shale facies is in accord with the commonly held interpretation of a deep-water starved basin or leptogeosynclinal environment (Goldstein, in Flawn and others, 1961, p. 31; Viele, 1973, p. 363; but see summary of other interpretations by Viele, 1973, p. 363). Some sandstone interbeds within the black

shale facies suggest a supply of quartz sand from the bank. Parts of the Ouachita clastic facies (for example, the southward-thickening Blaylock Sandstone) suggest a sediment source within or south of the basin (Goldstein, in Flawn and others, 1961, p. 32). Meta-arkose and granitic boulders in the Blakely Sandstone may have been derived from scarps on the north (Stone and others, 1973, p. 37).

The structures of the Ouachita salient are within the shale facies, and the frontal structures of the Ouachitas may be approximately parallel with the carbonate bank edge. Toward the Alabama recess, the front of the structural system crosses into the carbonate facies; and, in eastern Mississippi and Alabama, Appalachian structures are in the carbonate facies (Fig. 2).

MIDDLE PALEOZOIC (DEVONIAN-LOWER MISSISSIPPIAN) STRATIGRAPHY

The Arkansas Novaculite apparently is related to other Devonian and Lower Mississippian chert units in the region around the Ouachita Mountains (Fig. 3). The novaculite has a maximum thickness of nearly 1,000 feet in the Ouachita Mountains, but it thins northward across the Ouachita outcrops (Sterling and others, 1966, p. 184). The novaculite also thins westward, but the middle division of the novaculite is continuous with the Woodford Formation of the Arbuckle Mountains (Ham, 1959, p. 75). The Woodford consists of black shale and chert beds (Ham, 1959, p. 75; Frezon, 1962, p. 33).

In northern Arkansas, chert beds equivalent to the Arkansas Novaculite are included in the Penters and Boone Formations (Sterling and others, 1966, Table 1; Gordon and Stone, 1973, p. 259). Maximum thickness of the Penters-Boone sequence is less than that of the Arkansas Novaculite; however, locally in northern Arkansas, thickness of the Penters-Boone exceeds that of the relatively thin Arkansas Novaculite of the northern Ouachitas (Fig. 3). The Penters Chert pinches out northward (Fig. 3), and in northern Arkansas the Penters contains thin interbeds of limestone (Frezon and Glick, 1959, p. 177). The Boone thickens northward and extends beyond the limit of Penters Chert (Fig. 3).

The Boone Formation of northern Arkansas consists mainly of cherty limestone and chert (Frezon and Glick, 1959, p. 179). The formation thins and changes southward into a shale facies in the subsurface north of the Ouachitas (Fig. 3; Caplan, 1957, p. 4; Frezon and Glick, 1959, p. 179). Subsurface data indicate that the Penters Chert also thins southward across the same area (Maher and Lantz, 1953; Haley and Frezon, 1965, p. 3). The Arkansas Novaculite is shaly in outcrops along the northern Ouachitas (B. R. Haley and C. G. Stone, personal communication, 1974), and evidently an intermediate shaly facies separates the thick Arkansas Novaculite of the Ouachitas from the Penters-Boone sequence of northern Arkansas. Frezon and Glick (1959, p. 179) conclude that the Boone limestone and chert facies was deposited in a shelf environment,

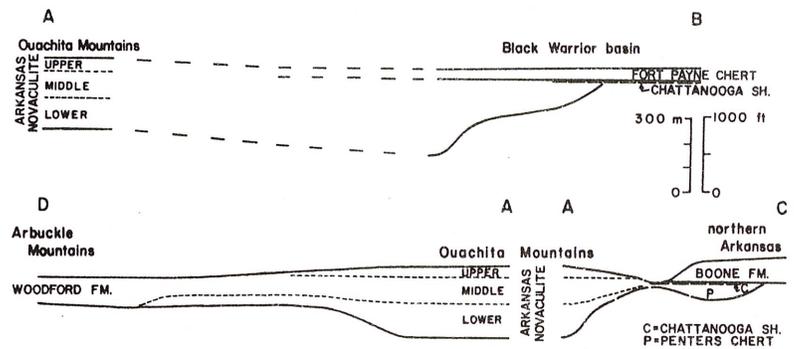
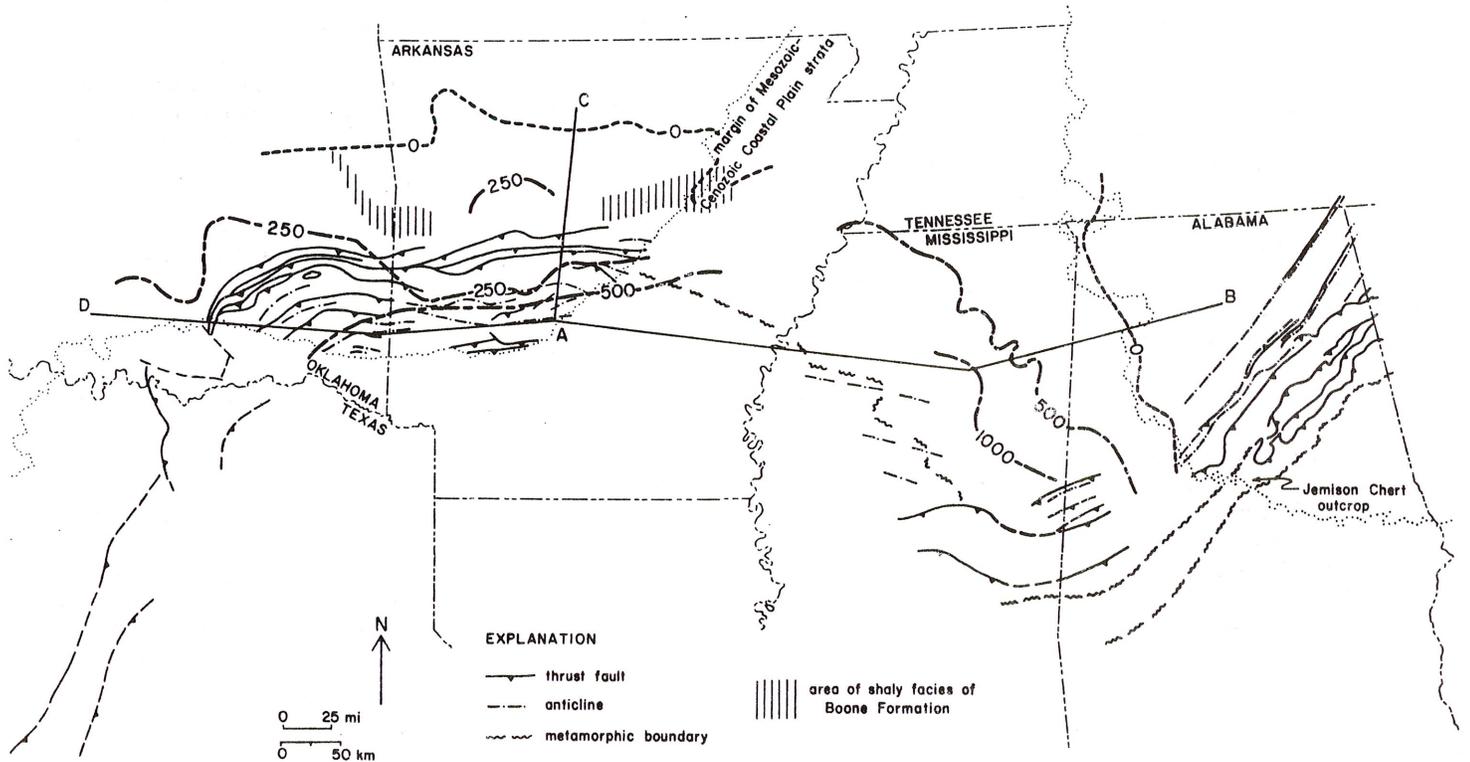


Figure 3. - Middle Paleozoic (Devonian-Lower Mississippian) stratigraphy. Isopach map of Devonian chert (contour values in feet; interval varies); Mississippian Boone and Fort Payne Formations extend north and northeast beyond pinch-out of Devonian chert. Datum of cross sections is top of Devonian. Data from Purdue and Miser (1923), Miser and Purdue (1929), Maher and Lantz (1953), Huffman and others (1958), Frezon and Glick (1959), Ham (1959), Frezon (1962), Haley and Frezon (1965), Sterling and others (1966), Sellars (1967), Wise and Caplan (1967), Thomas (1972a), Thomas and Drahozal (1973).

whereas the shaly facies was deposited in a basin south of the shelf. Geographic relation of the Arkansas Novaculite to the shaly Boone suggests that the novaculite is also a basin sediment. Park and Croneis (1969, p. 109) interpret the novaculite to have been deposited in relatively deep water; however, Goldstein and Hendricks (1953, p. 441) conclude that the novaculite is of shallow-water origin.

In the subsurface of central Mississippi the novaculitic chert sequence locally exceeds 1,000 feet in thickness (Thomas, 1972a, p. 96). The chert unit thins northeastward across eastern Mississippi but evidently is continuous northward with a succession of chert, cherty limestone, and limestone in the Devonian of western Tennessee. The pattern of northeastward thinning of the Devonian chert terminates at a southeast-trending pinch-out line in the subsurface of western Alabama (Fig. 3). Near the pinch-out edge in Alabama, the chert unit includes limestone beds which are comparable in position to similar rocks in northern Arkansas and western Tennessee.

The Lower Mississippian Fort Payne Chert of Alabama consists mainly of beds of chert and fine-grained limestone, and it includes crinoidal chert and a few interbeds of bioclastic limestone. Stratigraphic relations in northeastern Mississippi suggest possible continuity of the Fort Payne with the upper part of the subsurface novaculite sequence (Thomas, 1972a, p. 96). The Osagean age of the Fort Payne (Drahovzal, 1967, p. 14) coincides with the probable age of the upper part of the Arkansas Novaculite (Hass, 1951, p. 2540). The Fort Payne thickens northeastward in Alabama and extends far northeastward beyond the limit of Devonian chert across most of northern Alabama and Tennessee. The Fort Payne evidently is a shelf deposit which thins southwestward into the Black Warrior basin (Fig. 3).

The Lower Devonian Jemison Chert (Butts, 1926, p. 147; Carrington, 1973, p. 31) at the southwestern exposed end of the Alabama Piedmont also may be part of the chert distribution that centers on the Arkansas Novaculite. Massive chert of the Jemison is common only on the southwest where the formation is about 450 feet thick (Carrington, 1973, p. 31), but the unit can be traced northeastward as a quartz schist (Neathery, 1973, p. 52). Chert interbeds in the Devonian Frog Mountain Sandstone of the Alabama Appalachians also may be related to the Arkansas Novaculite (Thomas and Drahovzal, 1973, p. 79).

The distribution of Devonian-Lower Mississippian chert thickness defines a semicircular pattern around the Ouachita Mountains (Fig. 3). The chert thins irregularly away from the Ouachitas and evidently represents a sedimentary system centered on the Ouachitas. Devonian chert pinches out northward and eastward from the Ouachitas (Fig. 3). The thickness of the chert unit in the subsurface of Mississippi is similar to that in the Ouachitas, and possibly the thick chert unit is continuous from Mississippi to the Ouachitas (Fig. 3). Any possible intermediate shaly facies or area of thin chert (analogous to that between northern Arkansas and the Ouachitas) has not been recog-

nized in available subsurface data from Mississippi.

UPPER PALEOZOIC (UPPER MISSISSIPPIAN-LOWER PENNSYLVANIAN) STRATIGRAPHY

Upper Mississippian and Lower Pennsylvanian strata comprise a clastic sequence which thickens toward the Ouachita Mountains from the west, north, and east; and, the maximum thickness is as much as 30,000 feet in the Ouachitas (Table 1; Fig. 4). The clastic sequence is characterized by shale in the lower part and by a succession of interbedded sandstones and shales in the upper part (Table 1).

The Ouachita sequence includes dark-colored shales, several kinds of sandstones, boulder beds, and distinctive interbeds of dark-colored siliceous shale, chert, and tuff (Cline, 1960; Goldstein and Hendricks, 1962). The upper Paleozoic Ouachita sequence is interpreted to be a deep-water flysch facies (Cline, 1960, p. 100; 1966; 1970; King, in Flawn and others, 1961, p. 184; Chamberlain, 1971, p. 49).

In Oklahoma, the sequence thins westward and northward from the central Ouachitas (Fig. 4). The rate of thinning evidently has been exaggerated somewhat by thrust faulting; however, Cline (1960, p. 21) and Hammes (1965, p. 1678) suggest that the amount of overthrusting may be relatively less than is apparent because the rate of thinning may have been influenced significantly by original sedimentary convergence. Some thickening toward the Ouachitas is related to down-to-basin contemporaneous faults (Koinm and Dickey, 1967; Buchanan and Johnson, 1968; Haley and Hendricks, 1968, p. A7). Northwest of the Ouachitas, the Mississippian and Pennsylvanian include clastic and carbonate rocks which represent shallow marine and deltaic depositional environments (Laudon, 1959; Scull and others, 1959, p. 167; Visher and others, 1971). The Pennsylvanian includes prograding deltaic sandstones which were supplied from the north (Scull and others, 1959, p. 167; Visher and others, 1971, p. 1212).

In northern Arkansas, total thickness of Mississippian and Pennsylvanian strata is much less than that in the Ouachitas (Fig. 4). The Mississippian-Pennsylvanian succession includes units of shale, limestone, and sandstone (Maher and Lantz, 1953; Ogren, 1968; Glick, 1973; Zachry and Haley, 1973). These sediments reflect shallow marine and deltaic environments.

East of the Ouachita Mountains in the Black Warrior basin, the Mississippian-Pennsylvanian sequence thins northeastward from a maximum of more than 10,000 feet in east-central Mississippi (Fig. 4). A predominantly shale unit at the base of the sequence grades upward into a cyclical succession of sandstone, shale, and limestone (Thomas, 1972a, p. 98). These rocks indicate shallow marine and deltaic environments, and rock-stratigraphic relations indicate northeastward progradation of a delta complex. Upward gradation continues into a coal-bearing

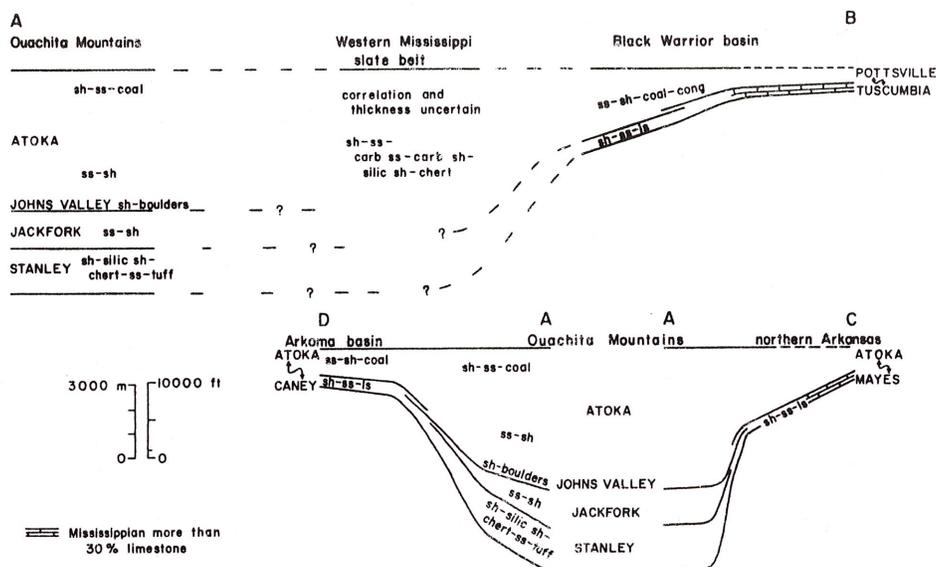
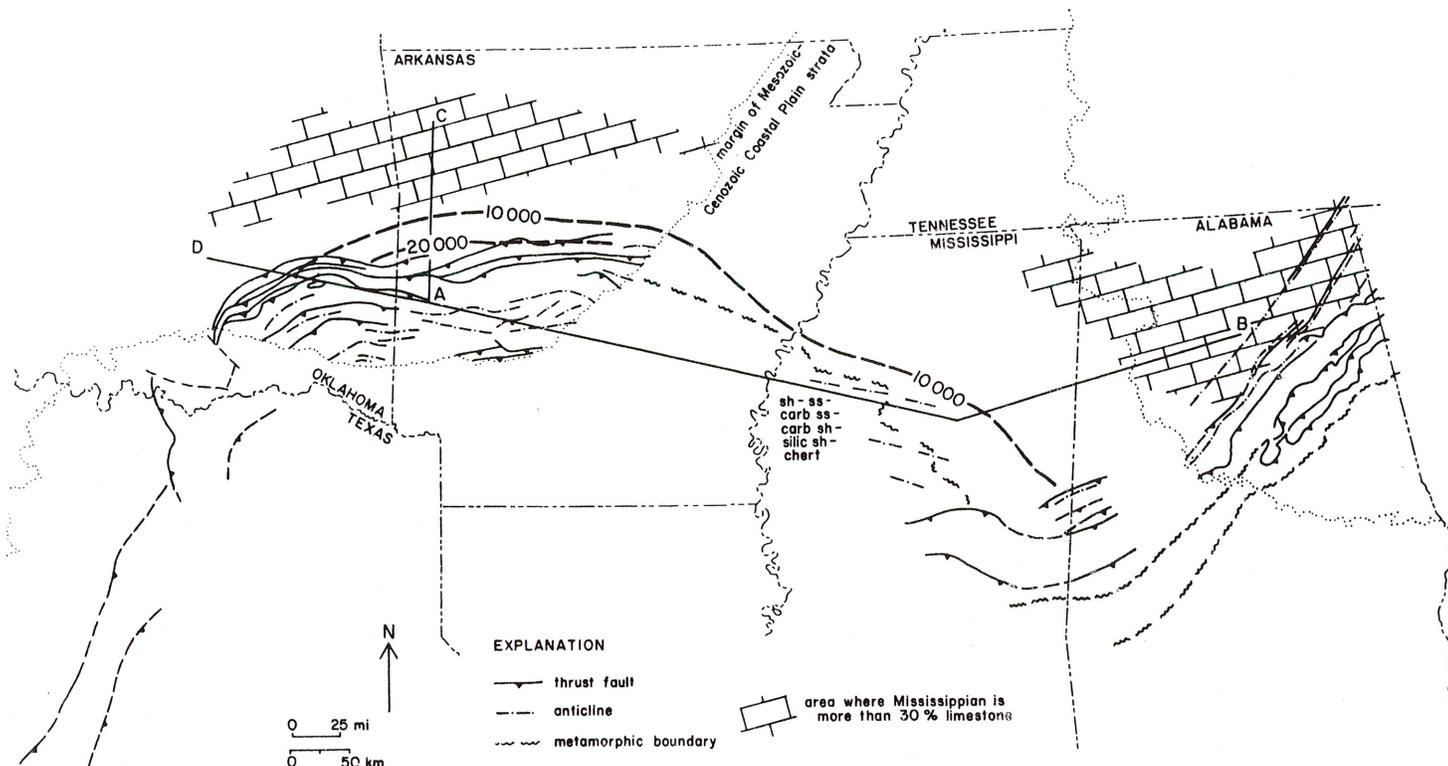


Figure 4. - Upper Paleozoic (Upper Mississippian-Lower Pennsylvanian) stratigraphy. Isopach map of Upper Mississippian-Lower Pennsylvanian (contour values in feet). Datum of cross sections is top of Atoka. Abbreviations: sh = shale, ss = sandstone, carb = carbonaceous, silic = siliceous, cong = conglomerate, ls = limestone. Data from Maher and Lantz (1953), Caplan (1957), Reinamund and Danilchik (1957), Frezon and Glick (1959), Laudon (1959), Scull and others (1959), Cline (1960), Branson (1962), Frezon (1962), Goldstein and Hendricks (1962), Haley and Frezon (1965), Stone (1966), Ogren (1968), Visher and others (1971), Thomas (1972a, 1972b).

cyclical succession dominated by carbonaceous sandstone, shale, and conglomerate beds. The lower (Mississippian) part of the clastic sequence grades northeastward into a limestone facies in the eastern part of the Black Warrior basin in Alabama (Fig. 4). The facies boundary trends southeastward and is paralleled by linear barrier sandstones and massive oolitic limestones (Thomas, 1972b, p. 103). Facies strike is nearly perpendicular to Appalachian structural strike in the frontal structures of the recess in Alabama; and, although far removed from the exposed Ouachita Mountains, the facies in Alabama are approximately concentric with the Ouachita structural front (Fig. 4). Distribution of clastic sediments indicates a sediment source southwest of the Black Warrior basin.

In the Western Mississippi slate belt, the age of the dark-colored shale is not known with certainty, but parts of the succession include components which appear similar to distinctive rocks within the Mississippian-Pennsylvanian sequences in the Ouachita Mountains and in the Black Warrior basin. The slate belt sequence includes some interbeds of carbonaceous shale and sandstone like those in the Black Warrior basin. Parts of the slate belt succession include dark-colored siliceous shale and chert similar to those of the Ouachita facies. Apparently western Mississippi is an area of facies transition from the thinner shelf sediments of the Black Warrior basin to the thicker deep-water flysch of the Ouachitas. That relation suggests correspondence between facies boundaries and thickness distribution and further indicates that western Mississippi belongs to the Ouachita stratigraphic province (Fig. 4).

The great volume of Upper Mississippian-Lower Pennsylvanian clastic sediment has been interpreted to indicate orogenic uplift of a sediment source south or southeast of the present Ouachita Mountains and within the Ouachita mobile belt (Miser, 1921; Miser and Purdue, 1929, p. 134; King, in Flawn and others, 1961, p. 184; Goldstein and Hendricks, 1962, p. 421; Johnson, 1966, p. 156; Cline, 1970, p. 100). The sediment source is described as having included basement rocks, metasedimentary rocks, and sedimentary rocks, as well as active volcanoes (which supplied tuff in addition to sediment). Petrographic data generally have been interpreted to indicate a lithologically complex sediment source at the southern margin of the Ouachita trough (Bokman, 1953, p. 168; Goldstein and Hendricks, 1962, p. 421; Hill, 1966, p. 120; Klein, 1966, p. 316; Walthall, 1967, p. 523). Other petrographic work suggests that quartz sand was introduced into the basin from a source north of the Ouachitas (Klein, 1966, p. 316; Morris, 1971, p. 398). Paleocurrent data indicate predominantly westward (axial) transport of sediment through the Ouachitas of western Arkansas and eastern Oklahoma (Briggs and Cline, 1967, p. 991; Cline, 1970, p. 93; Morris, 1971, p. 399). A comprehensive interpretation suggests that sediment was introduced into the Ouachita trough from both south and north and was transported westward along the axis of the trough (Klein, 1966, p. 323; Cline, 1970, p. 100). The orogenic sediment source south or southeast of the Ouachitas also supplied clastic sediment to the Black Warrior basin and southwestern

Alabama Appalachians. Parts of the shelf north and west of the Ouachitas received clastic sediment from northern sources (Visser and others, 1971, p. 1212). Swann (1964, p. 653) proposes that the "Michigan River" delta system prograded through the Illinois basin and transported sand to the northeast edge of the Ouachita trough as well as to the western part of the Black Warrior basin. Fault scarps along the northern margin of the Ouachita trough are proposed as the source of erratic boulders in the Johns Valley Shale (Shideler, 1970, p. 803).

SUMMARY OF STRATIGRAPHY WITHIN THE OUACHITA SALIENT

Available data indicate a regional geographic coincidence of the Ouachita structural salient with the extent of the Ouachita sedimentary facies. The Ouachita salient structures are within the area of the lower Paleozoic Ouachita black shale facies. Eastward from the Ouachita salient, the belt of deformed rocks crosses into the carbonate facies in the Alabama recess (Fig. 2). Distribution of Devonian-Lower Mississippian chert defines a generally semicircular pattern that apparently is centered on the Ouachita salient (Fig. 3). Thickness of the upper Paleozoic clastic sequence decreases westward, northward, and eastward from the Ouachitas, and isopach lines appear to be roughly semicircular around the Ouachita salient (Fig. 4). Eastward toward the Alabama structural recess, the structural front intersects isopach strike, and the belt of deformed rocks crosses into a thinner upper Paleozoic clastic sequence. The Mississippian part of the clastic facies grades northeastward into a carbonate facies along the frontal part of the Alabama Appalachian recess; the Mississippian of northern Arkansas includes a similar carbonate facies.

In northwestern Mississippi, the boundary between the Western Mississippi slate belt and undeformed rocks of the Black Warrior basin trends generally southeastward, but subsurface data are too sparse to define precisely the position and shape of the boundary (Thomas, 1973, Fig. 8). Between the area of slaty rocks and the area of undeformed rocks of the Black Warrior basin, several wells have cored beds which have dips of 15 degrees or more (Thomas, 1973, p. 385), but the relation of strike of individual structures to strike of the structural front is unknown. Some rocks of the Western Mississippi slate belt suggest affinities with the upper Paleozoic Ouachita facies, whereas other slate belt rocks are similar to upper Paleozoic rocks of the Black Warrior basin. Northeast of the slate belt, the succession is typical of that in the Black Warrior basin. Thus, the front of the Western Mississippi slate belt may coincide approximately with the limits of the thick Ouachita clastic facies.

STRUCTURE AND STRATIGRAPHY OF OTHER APPALACHIAN SALIENTS

Interpretation of structures between the Ouachita salient

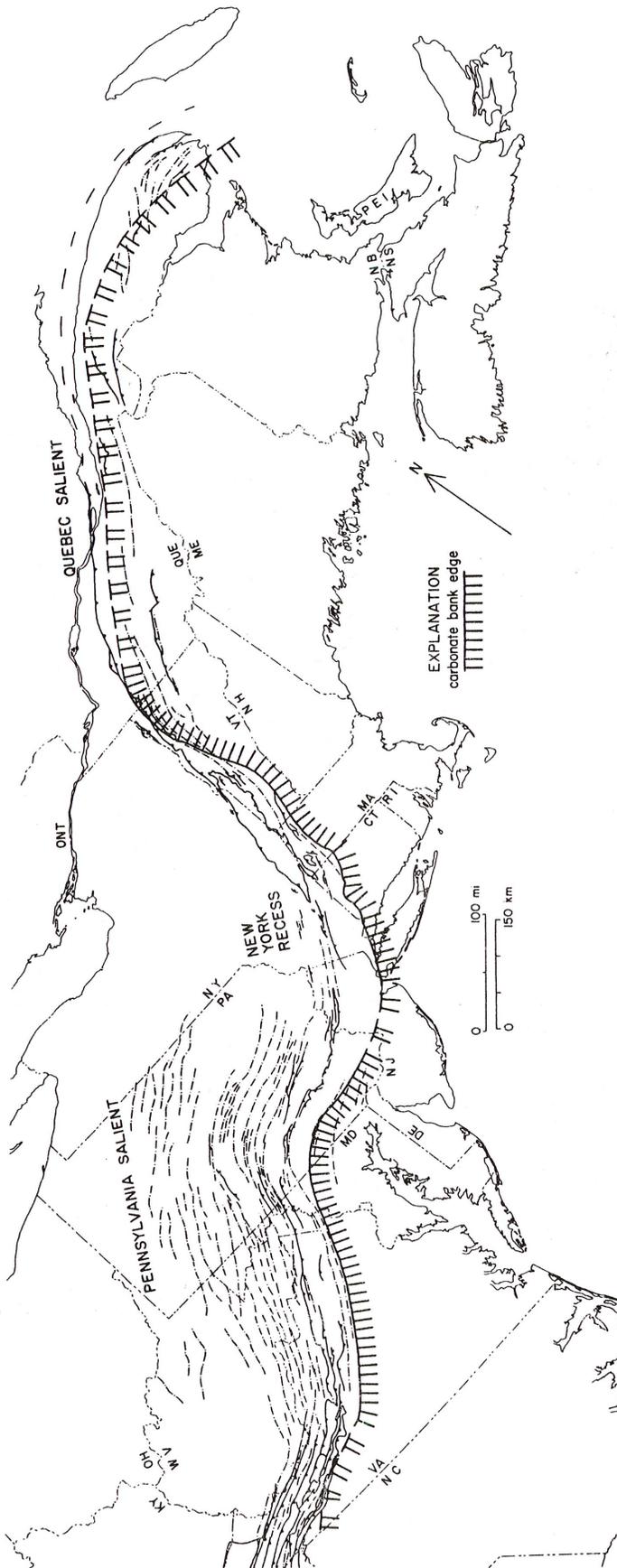
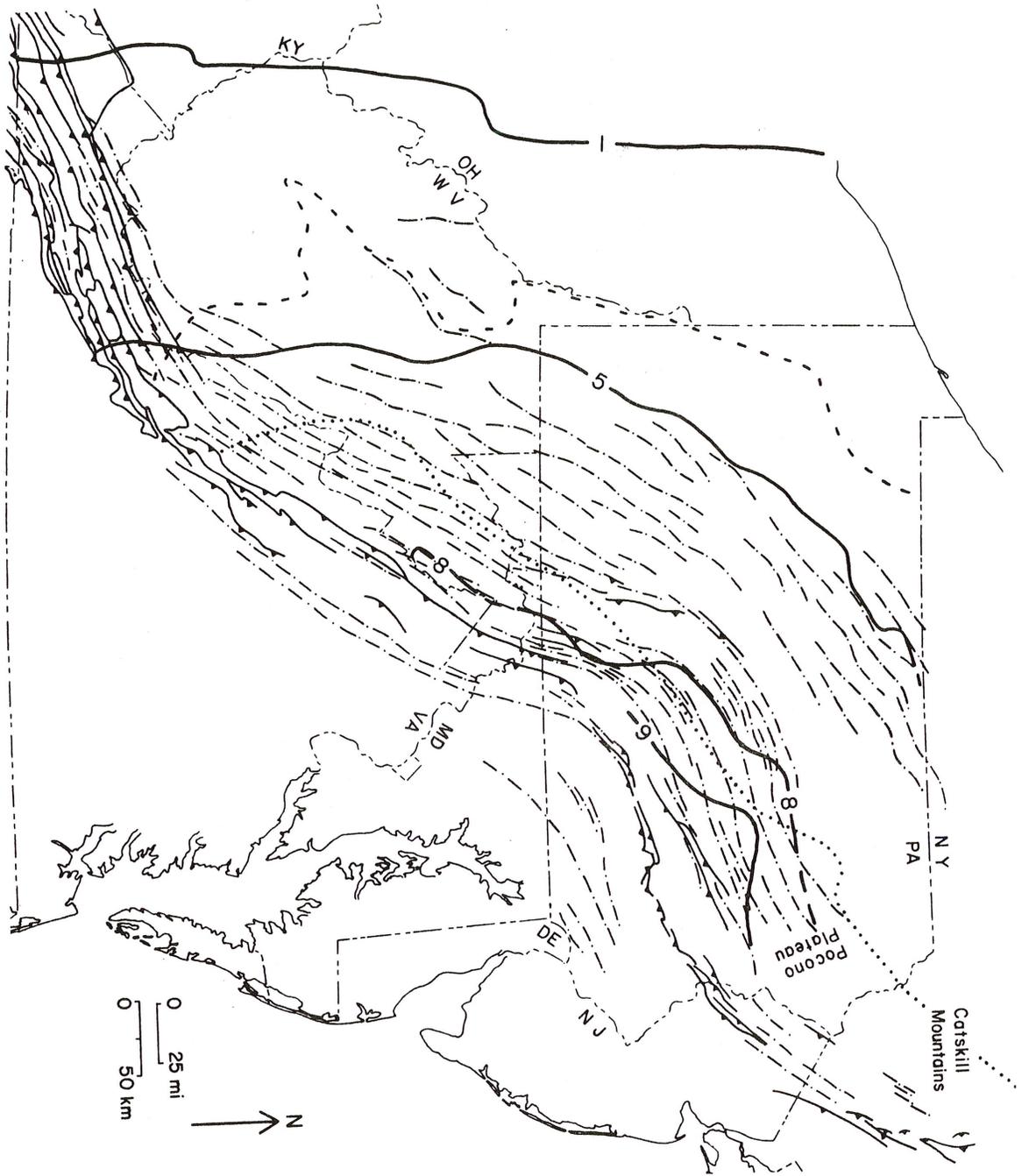


Figure 5. - Outline of Cambrian-Ordovician carbonate bank in northern Appalachians (after Rodgers, 1968).

Figure 6. . Distribution of Devonian clastic wedge in Pennsylvania salient (after Oliver and others, 1967; Cotton, 1970). Isopach map of Upper Devonian rocks (contour values in thousands of feet; interval varies). Explanation: short-dashed line = western limit of red beds; dotted line = western boundary of area where sandstone-shale ratio exceeds 1:4.



and Alabama recess may be guided by analogy with other salients and recesses in the Appalachians. The implied relations between Ouachita structure and stratigraphy also may be evaluated in other salients.

The structural configuration of the northern Appalachians may be described in terms of the Quebec and Pennsylvania salients and the New York recess (Fig. 5). In the northern Appalachians, the Cambrian and Ordovician include a carbonate facies on the northwest and a black shale facies on the southeast (Fig. 5). Rodgers (1968, p. 143) interprets the black shale to have been deposited in deep water east of a shallow carbonate bank; the facies boundary is interpreted to be the steep edge of the bank. In northern Vermont, the facies boundary crosses structural strike diagonally (Rodgers, 1968, p. 144). On the north in the Quebec salient, the deformed belt is within the black shale facies; but, southward toward the New York recess, the structural system crosses into the carbonate facies (Fig. 5). Similarly, a curve in the carbonate bank edge is concentric with the Pennsylvania salient (Rodgers, 1968, Fig. 10-3); however, that shale facies does not extend to the frontal structures of the salient (Fig. 5). In Quebec, the edge of the carbonate bank is obscured beneath the overthrust shale facies southeast of Logan's Line; and, outlying masses of the shale facies (for example, Taconic slate mass) within the area of the carbonate facies are interpreted to be allochthonous (Rodgers, 1968, p. 146).

The area of maximum thickness of Devonian clastic rocks of and related to the Catskill delta is shown to be within the Pennsylvania salient (Oliver and others, 1967, Fig. 9; Colton, 1970, Fig. 22). The Devonian clastic succession thins along strike from the Pennsylvania salient southwestward into the Virginia structural recess (Fig. 6). Isopach lines intersect structural strike at a large acute angle in the Virginia recess (Fig. 6), and the arcuate isopach lines are generally concentric around the center of the Pennsylvania salient. Boundaries of various facies components have been shown to be approximately parallel with isopach lines and, thus, also concentric with the structural salient (Fig. 6). At the northern limit of the salient, distributions of thickness and facies are obscured by erosion; however, there is a hint of an eastward curve of isopach lines and of northward thinning (Fig. 6). Possibly original isopach strike did not parallel structural strike in the New York recess.

COMPARISON OF OUACHITA AND NORTHERN APPALACHIAN SALIENTS

Regional structural salients coincide geographically with curves in the carbonate-black shale facies boundary and/or with areas of thick clastic wedges (Figs. 2, 4, 5, 6). Particularly in the Quebec and Ouachita salients, the curvature of the structural salient seems to coincide approximately with the facies boundary between black shales and carbonates (Figs. 2, 5). Similarly, the greatest thicknesses of clastic sediments in Pennsylvania and in the Ouachitas seem to be concentrated near the center of curvature of the structural salients. (Figs. 4, 6).

Salients are curves in the structural system convex toward the craton, and the belt of deformed rocks within salients extends farther toward the continental interior than that in the adjacent recesses. Part of that greater extent reflects a curve in structural strike. However, in the Pennsylvania salient, the deformed belt includes frontal folds that end along strike toward the recesses (Figs. 5, 6). Part of the greater extent of the Pennsylvania salient toward the craton reflects the greater width of the deformed belt in the salient. In the northeastern part of the Pennsylvania salient, the structural front trends southeastward and is not parallel with strike of individual structures (Fig. 6). Rather, that structural front is a line along which northeast-trending folds flatten along strike and end beneath undeformed rocks in the Pocono Plateau (Wood and Bergin, 1970, p. 147).

A similar interpretation of structural boundaries may be suggested for the frontal structures of the Ouachitas. In the subsurface of northwestern Mississippi, a southeast-trending structural front is defined between Ouachita structures on the southwest (Western Mississippi slate belt) and undeformed strata farther east (Fig. 1). Possibly that front is not parallel with structural strike but rather is a line marking the ends of east-striking folds that flatten eastward beneath undeformed rocks in the Black Warrior basin. The structures of the exposed Ouachitas may be analogous to part of the frontal structures of the Pennsylvania salient. And by analogy with the New York recess, structures within the Alabama recess perhaps project along strike westward into the Ouachita salient far south of the exposed Ouachita structural front. Identity of Appalachian structures may be lost where they cross westward into the lower Paleozoic black shale facies in the salient. The limits of Ouachita frontal folds and the structural front of the salient thus may coincide approximately with sedimentary facies and thickness outlines.

CONCLUSION: PROPOSED STRUCTURAL-STRATIGRAPHIC MODEL

Structural and stratigraphic relations suggest a theoretical model for evolution of the Ouachita salient and adjacent Alabama recess (Fig. 7). Lower Paleozoic facies relations may be interpreted in the framework of a carbonate bank (similar to the northern Appalachian carbonate bank of Rodgers, 1968). The shallow-water shelf extends southward to a steep bank edge, and the deep-water Ouachita black shale facies occupies a semicircular reentrant in the bank margin (Fig. 7 - panel 1). A limited amount of clastic sediment was supplied by sources within and/or south of the basin; erratic boulders were supplied from steep scarps within and around the basin.

The area of thick upper Paleozoic deep-water flysch is bordered on the west, north, and east by thinner shallow marine and deltaic sediments which occupy the area of the earlier carbonate bank (Fig. 7 - panel 2). Ouachita orogenesis on the south provided clastic sediment to the deep Ouachita trough as well as to the shallow marine shelf on

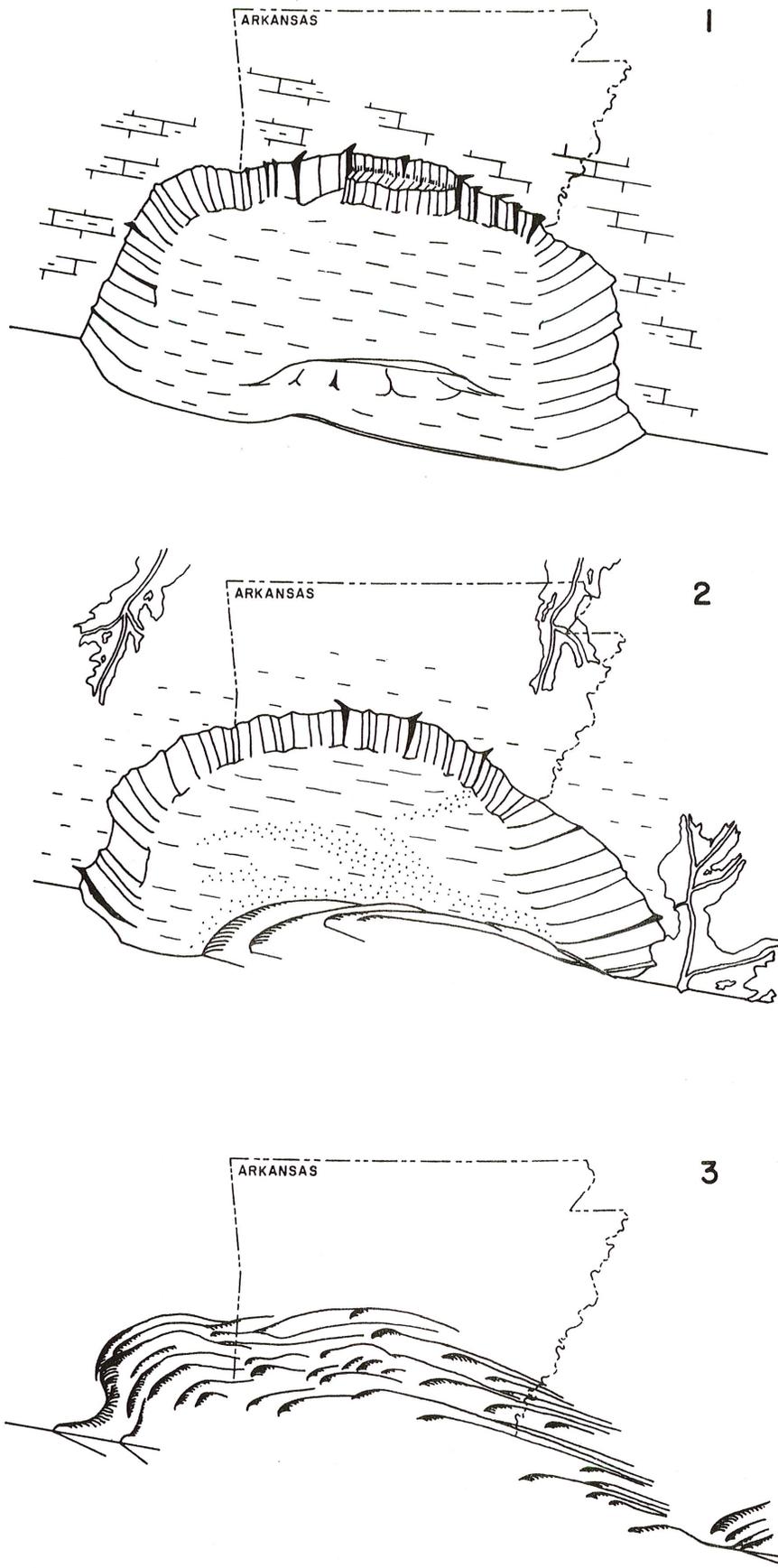


Figure 7. Sequential diagram of Ouachita-Appalachian structural-stratigraphic model.
Panel 1 - Lower Paleozoic carbonate bank and deep basin.
Panel 2 - Upper Paleozoic shelf and flysch basin.
Panel 3 - Structural configuration of the Ouachita Salient.

the east in the Black Warrior basin. Scarps within or marginal to the basin supplied erratic boulders. Other clastic sediments were supplied to parts of the shelf and the northern part of the Ouachita trough from the craton.

The structural configuration of the Ouachita system may be interpreted within the stratigraphic framework. The thrust faults and folds of the Ouachita Mountains have formed within the lower Paleozoic black shale facies and the thick upper Paleozoic clastic facies (Fig. 7 - panel 3). Possibly the frontal structures flatten toward the east into the carbonate facies and the thinner upper Paleozoic succession of the Black Warrior basin, and only the interior structures of the salient extend far across the facies boundaries into the Alabama recess.

Distribution of the major sedimentary facies was controlled by the shape and position of the scalloped edge of the shallow shelf (which possibly reflects the approximate shape of the margin of continental crust as proposed by Rodgers, 1968, p. 148). The ultimate structural configuration of the Ouachita salient appears to be related to the distribution of clastic facies and of maximum sedimentary

thicknesses. Possibly the greater thickness of incompetent rocks permitted deformation to expand farther toward the craton in the salient, and tectonic transport apparently is greatest in the salient. Rodgers (1968, p. 144) concludes that the contrast in competence between the carbonate and black shale sequences served to localize thrusting along the facies boundary. King (in Flawn and others, 1961, p. 184) notes that Mississippian-Pennsylvanian clastic units have maximum volumes in the Ouachita and Marathon salients. The association of the structural salient with the black shale facies and thick clastic sequence suggests a genetic relationship between these major regional structural and stratigraphic elements.

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AGE OF IGNEOUS AND METAMORPHIC ACTIVITY AFFECTING THE OUACHITA FOLDBELT

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ABSTRACT

Thirty-eight isotopic ages have been determined on a variety of igneous and metamorphic rocks associated with the Ouachita foldbelt in Arkansas, Oklahoma, and Texas. These new ages have been integrated with previously published ages to determine the timing of igneous and metamorphic activity associated with the belt. The major time of movement, based on K/Ar ages from the interior zone, is Pennsylvanian and Early Permian. An older period of metamorphism and movement is suggested by apparent Devonian K/Ar ages. Post-tectonic intrusive igneous rocks were emplaced during two discrete periods. Diabasic intrusions yield apparent ages of Triassic. A petrographically more diverse suite of subsilicic igneous rock is Cretaceous in age.

INTRODUCTION

The interpretation of isotopic ages from the Ouachita system can hopefully give insight into the timing for the metamorphic recrystallization and movement of these rocks. The pattern of isotopic ages within the low rank metamorphic rocks from the Marathon-Ouachita fold system is complex. The acceptance of the isotopic ages at face value requires a model more complex than can be firmly supported by other data. However, the distribution of isotopic ages and their consistency suggest that periods of heretofore unrecognized regional metamorphism may have significantly affected the Ouachita rocks. The object here is to present and interpret the results of our petrography and isotopic dating and to integrate these with previously published work. We have not commented on the broader implications of these results in terms of a model for Ouachita deformation. This we leave to other workers.

The data have mostly been obtained from the interior zone of the Ouachita system. Flawn (in Flawn and others, 1961, p. 164) described these rocks as: "sedimentary rocks showing weak to low grade metamorphism with a high shearing component and metamorphic structures associated with high shear (foliation, slaty cleavage, fracture cleavage, wrinkling, rucking, microimbricate structures, flowage around augen) . . ." Numerous isotopic ages have been determined on the sericite-muscovite developed during this metamorphism-shearing.

Ideally the micas are formed at the peak of metamorphic

activity. The evidence from the outcrop suggests that the peak of the metamorphism and shearing were in part coincident. One example of such evidence is the relatively more complete recrystallization of the sedimentary rock at the fold axes. The retention of radiogenic daughter products begins after the rock has cooled below the threshold of rapid diffusion of daughter isotopes. In igneous rocks this usually begins immediately after crystallization as can be shown by the comparison of K/Ar ages of highly retentive (hornblende) and less retentive minerals (micas). In low rank metamorphic suites, such as we are dealing with here, there is probably a comparatively slow cooling. Thus, apparent ages from minerals would not give the "true" time of maximum metamorphism but some later time when the rock cools sufficiently to retain argon. There are no retentive-unretentive mineral pairs to compare in the same rocks from the interior zone of the Ouachitas; the results are almost exclusively K/Ar ages on muscovite-sericite.

The evaluation of the seriousness of this effect has not been fully satisfactory. Our evidence indicates that there are additional factors which complicate the simple cooling picture. Late faulting appears to be the most likely serious complication.

The interpretation of ages also hinges on the degree of metamorphism. In general, coarser, more fully recrystallized, samples give the best ages, as might be expected. Mica separates containing a high percent potassium are the best for dating. However, in this study we have had to use whole rocks containing muscovite, hoping for the best because of the limited amount of sample available from wells.

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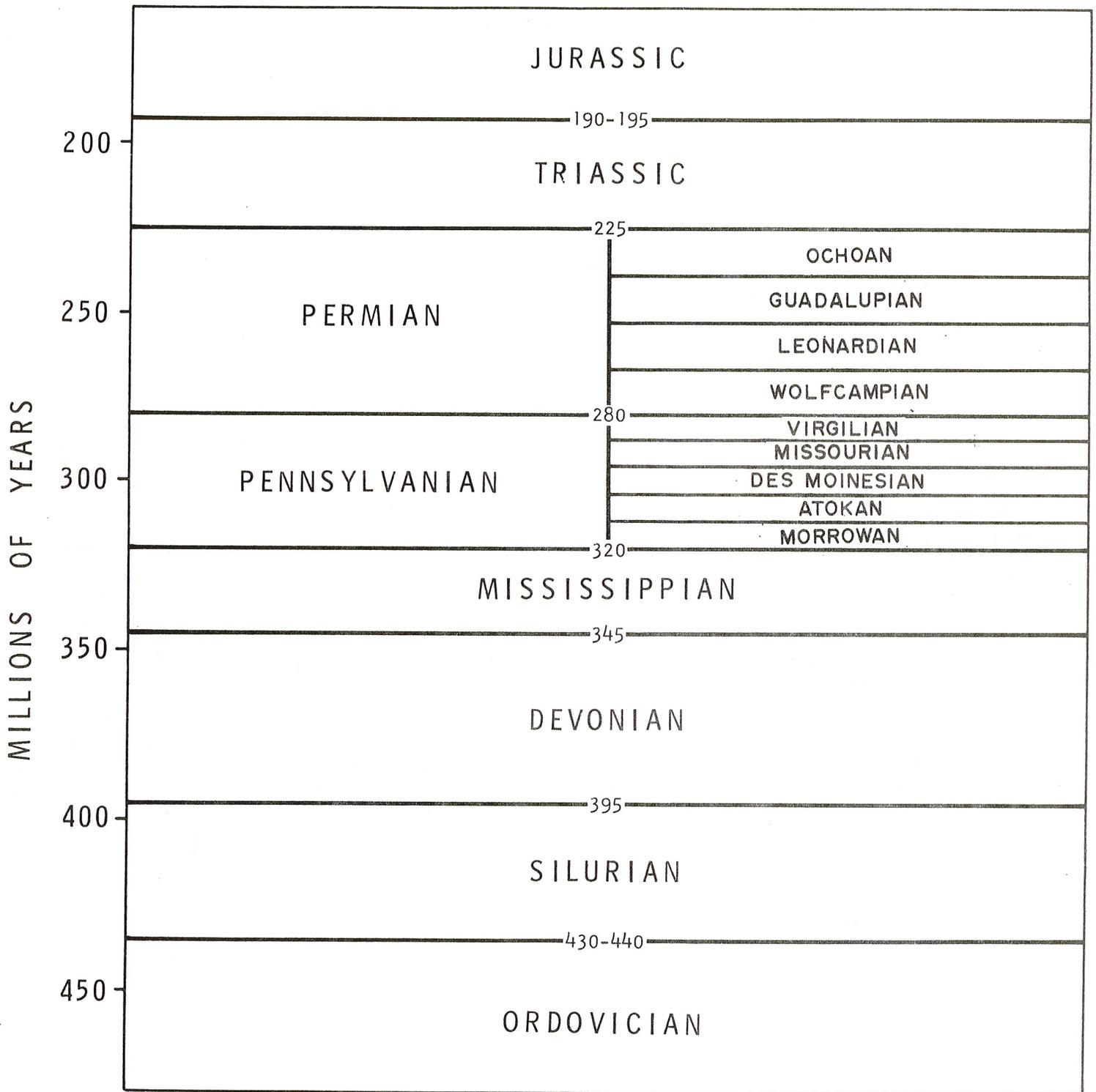


Figure 1. Part of Phanerozoic time scale taken mostly from the Geologic Names Committee, U. S. Geological Survey, 1972. Divisions of the Pennsylvanian and Permian are arbitrarily evenly divided.

AGES OF UPLIFT

Virtually all isotopic ages from the interior zone are Paleozoic. Therefore, it is useful to define the time scale during this period. Figure 1 shows the time scale accepted by the Geologic Names Committee of the U. S. Geological Survey in 1972.

Because the metamorphism is tied closely to times of shearing, it is important to review the periods during which there is evidence for strong movements that affect sedimentation and structure in the Ouachita system. This information is taken from the discussion of King (in Flawn and others, 1961, pp. 175-190).

Atokan and younger - Deformation presumably began in the Ouachita Mountains in Atokan time and continued at least through Des Moinesian time, the youngest rocks deformed. In the Marathon area deformation began in Missourian time and continued into Wolfcampian time. Movement thus began later and continued later than is demonstrable in the Ouachita Mountains.

The times of movement affecting the Ouachita system therefore fall into two categories: those during which sedimentation was affected and those during which the actual rocks in the Ouachita system were deformed. It appears from the isotopic evidence that both these types are represented.

Early Middle Ordovician - A minor pulse is suggested by the Blakely Sandstone in the Ouachita Mountains and the Rodriguez Tank Sandstone in West Texas.

Early Silurian - The wedge of Blaylock Sandstone in the Ouachita Mountains indicates a local but significant uplift and exposure of granitic rocks in the source area to the south.

Earlier Mississippian (post-Novaculite) through Morrowan - The beginning of very rapid flysch sedimentation through Atokan time indicates very unstable conditions and major uplift in the source areas.

IGNEOUS AND METAMORPHIC BOULDERS

Introduction

Several units within the Ouachita Mountains and Marathon Basin contain igneous and/or metamorphic boulders. These have an important bearing on the source area for various units deposited in the Ouachita-Marathon trough.

Marathon Region

Dagger Flat Sandstone

The Dagger Flat Sandstone of the Marathon Basin is of Late Cambrian age (Wilson, 1954). McBride (1969) has summarized the petrography and current studies of Anan (1964). The Dagger Flat contains a minor fraction of

granitic, volcanic, and metamorphic rock clasts. The source area is suggested to be to the northwest. Thus, the source is to the shelf side and may have been covered by the allochthonous thrust sheets containing the Dagger Flat.

Marathon Limestone

Young (1970) has provided an excellent description of the Marathon Limestone. He notes several conglomeratic and sandstone beds within the dominantly limestone sequence. The evidence suggests a metamorphic and volcanic source for many of the clasts. Young cites "floods" of biotite and plagioclase as indicating a volcanic source but it could also be ascribed, perhaps more appropriately, to a granodioritic igneous source. As in the Dagger Flat Sandstone, the source area for the Marathon Limestone boulder material is from the northwest.

Woods Hollow Formation

The Woods Hollow Formation in the Marathon Basin contains rare clasts of schist and igneous rocks (McBride, 1969). None of these clasts have been isotopically dated and samples loaned to us by McBride are not suitable for dating due to alteration. McBride describes five pebbles of granite, microsyenite and porphyritic amygdaloidal trachyte. The five samples we have examined are all calcitized and altered trachytic to dioritic rocks containing quartz where present as either a late intergranular mineral or in amygdules. These igneous pebbles are petrographically quite distinctive and do not come from any known basement rock province in Texas or Mexico.

Thus in mid-Ordovician the Marathon depositional basin had a minor source of igneous and metamorphic material and that source is unlike any known in the basement of Texas or Mexico.

Haymond Formation

The most famous of the crystalline boulder occurrences is in the Haymond Formation (Atokan) of the Marathon Basin. McBride (1966) has provided an excellent description of the Haymond. He ascribed the boulders to a submarine slide or wildflysch. Denison and others (1969) have suggested that the boulders were derived from a source area that was composed of petrologically related granitic and volcanic rocks. The boulders yield latest Silurian and Devonian ages. Further these rocks were formed by the partial melting of clastic sediment in a geosyncline during Silurian-Devonian time. The geosyncline was located to the southeast of the Marathon depositional basin.

Ouachita Region

Collier "Shale"

Honess (1923, p. 45) interprets the occurrence of "granitic gravel and arkosic fragments" in certain layers of the Collier marbles as indicating the presence of an igneous source area "at some not distant locality" during Early

Ordovician. These granitic fragments have received no subsequent work.

Blakely Sandstone

The Blakely Sandstone contains igneous and metamorphic cobbles and boulders in Arkansas. This unit is Middle Ordovician in age. Boulders collected from several localities in the eastern Benton uplift showed a fairly uniform petrographic character.

Several of the boulders are granite or closely related rocks of uncertain origin. These were collected as discrete boulders weathered loose on the surface and no relationship with the enclosing rock could be seen. These boulders were quite weathered but were fairly fresh when seen in thin section.

The texture and mineralogy of the original granite is quite uniform even though the boulders were collected at three localities. They are highly pure quartz-feldspar rocks containing only a minor amount of accessory minerals. These accessory minerals include zircon as rather large abundant crystals, iron oxides (now hematite) and sphene. Iron-stained chloritic minerals are found as replacement of original feric minerals and as thin veinlets. The quartz is in large single crystals, most mildly strained. The feldspar has the appearance of a mesoperthite with some late sodic plagioclase found between larger grains. However, the "perthite" has been albitized in samples examined from the three locations. Quartz veins are common in some samples.

The texture in the granite samples is a simple hypidiomorphic type with straight, smooth grain boundaries. The granites are quite coarse with some feldspar crystals exceeding 1 cm and most quartz being in the range 4 to 7 mm. There are several samples which show a modified texture. In these the rock appears to have been brecciated or shattered and the zones between the resulting pieces have been filled by a mosaic of quartz-feldspar. This mosaic contains minor iron sulfides (now hematite), zircon, and chloritic micas. This matrix is of unknown origin; it may be the fine brecciated granitic material recrystallized to a granoblastic mosaic during the low rank metamorphism in late Paleozoic. Alternately, the mosaic could be igneous in origin, an aplitic intrusion associated with brecciation. The former explanation is regarded as more likely although not on particularly firm textural evidence.

We performed three isotopic determinations on the feldspars from the granites showing no brecciation. Samples were chosen from two of the localities. The results (Table 11) are disappointing. The ages range from 283 + or - 25 to 489 + or - 55 million years assuming an initial Sr 87/86 ratio of 0.706. The age of sedimentation for the Blakely Sandstone should be in the range 460-480 million years. Only one apparent age is older than this and there is no real reason to place any higher reliance on this determination.

There are at least three factors which play an important role in the low ages. First, is the rather weathered condition

of the rocks. Second, is the late Paleozoic metamorphism which may have disturbed the equilibrium of the parent-daughter relationship. The third has to do with the albitization of the feldspars. Because the feldspars from three separate localities have the same characteristics, the albitization either occurred in the source area or was caused by a pervasive effect after deposition.

It should be noted that the potash feldspars in the granitic rocks from the Haymond Formation also have been albitized. Denison and others (1969) concluded that this happened in the source area. In the Haymond there is no late Paleozoic metamorphism to confuse the issue and this can be eliminated as a potential explanation.

Aside from the albitization of the potash feldspar and highly siliceous character of the granitic rocks, the boulders from the Haymond and Blakely share no common petrographic characteristics. The granitic boulders in the Blakely are most typical of shallow intrusions. These type granites are common throughout the south-central United States. The closest of these is in northeastern Oklahoma (~1250 million years old; Denison and others, 1969b) and in south-central Oklahoma (~500 million years old; Ham and others, 1964). Both granites were covered during mid-Ordovician everywhere the relationships are known.

Some conclusions can be drawn from this petrographic and isotopic data. In mid-Ordovician time a granitic landmass rather suddenly was available for rapid erosion. The isotopic age of the landmass is unknown. The landmass was most influential on deposition of the Arkansas rocks but not important for rocks now in Oklahoma. The landmass contained rather coarse highly silicic rocks typical of shallow intrusions. These rather distinctive rocks are not recorded as boulders in any other unit.

Johns Valley Shale

There are no reports of igneous or metamorphic boulders in the conglomeratic beds in the Johns Valley Shale of Late Mississippian-Early Pennsylvanian age (see Shideler, 1970, for a recent discussion and description of these beds).

IMPORTANT RECENT WELLS IN THE OUACHITA BELT

There are four exceptionally important wells that have been drilled into the Ouachita belt since the comprehensive work of Flawn and others in 1961. Two were drilled by Shell Oil Company on the Devils River Uplift in southwest Texas. Another was drilled by Shell on a previously unrecognized regional structure in northeast Texas. The fourth important well was drilled near or at the axis of the Broken Bow anticlinorium in southeastern Oklahoma (Fig. 2). Isotopic ages which have a direct bearing on determining the age of movement of the Ouachita belt have been determined on samples from each of these wells.

Each of these wells will be briefly described here with a

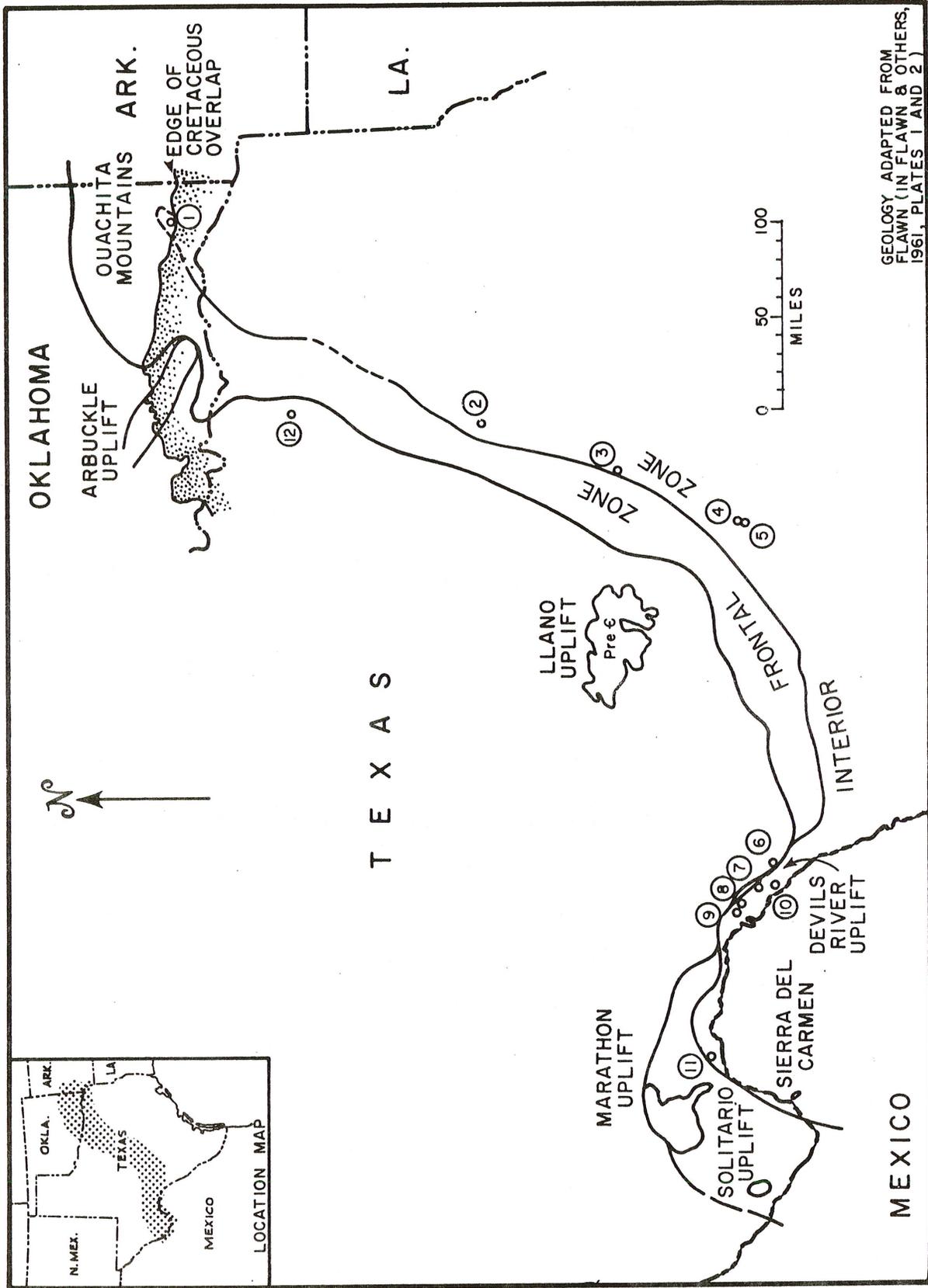


Figure 2. Location of wells from which isotopic ages have been determined in relation to major geologic structures. Sample numbers are keyed to Table 1.

discussion of the isotopic ages we and others have determined.

Shell No. 1 Gillis

The Shell No. 1 Gillis was drilled to a depth of 10,210 feet on the Devils River Uplift in Val Verde County, Texas. In our interpretation the rocks below the Cretaceous in the well can be broken into five essential units with these very approximate depths.

0 to 2,500 feet Cretaceous

Unit 1 ~ 2,500 to ~ 4,900 feet	Sheared graphitic phyllites
Unit 2 ~ 4,900 to ~ 6,500 feet	Carbonaceous micaceous sheared marbles
Unit 3 ~ 6,500 to ~ 9,500 feet	Relatively undisturbed dolomitic carbonates
Unit 4 ~ 9,500 to ~ 10,100 feet	Calcareous quartzite locally schistose
Unit 5 ~ 10,100 to ~ 10,210 feet	T. D. Greenschist

The cuttings from the Gillis well are small in size and it is difficult to determine some of the critical textural parameters. However, some basic compositional and textural evidence is clear.

Unit 1 is composed of graphitic phyllite and sheared quartzite. The sequence was deposited as a carbonaceous shale with sandy and lesser calcareous intervals, and was later metamorphosed to lower greenschist facies during a strong shearing episode.

Unit 2 is composed of sheared micaceous marbles. Some show irregular distribution of quartz both as detrital grains and veinlets. The texture is one of complete recrystallization, but is erratic. Larger calcite crystals are prominently twinned. The sequence was evidently deposited as a limestone with argillaceous intervals, which was later sheared and metamorphosed.

Unit 3 is a rather pale pure dolomitic carbonate containing irregular amounts of quartz detritus. Shearing is essentially lacking, although coarse twinned calcite is widely present though not abundant. Demonstrable metamorphic effects are difficult to define. However, an insoluble residue from the interval 9450 to 9460 feet (the only one checked) contained talc, interpreted as metamorphic in origin. The unit was evidently deposited as a pure dolomitic carbonate and escaped or did not respond to the effects of major shearing seen above but was mildly metamorphosed.

Unit 4 is a calcareous quartzite. The grain size shows considerable variation; both coarse and fine-grained detritus are present. Both plagioclase and microcline are common as detrital grains. Carbonate is found as evidently detrital grains and films and veinlets around the sand grains. Various chips show the development of schistosity and various metamorphic minerals. Olive-green biotite is well developed in some of the finer grained cutting chips. Opaque minerals are common. The rock was deposited as a calcareous felds-

pathic sandstone later metamorphosed to lower greenschist facies.

Unit 5 is a greenschist characterized by the assemblage actinolite-chlorite-epidote-albite and lesser quartz, calcite, biotite and opaque minerals. The albite contains considerable sericitic alteration. The cutting chips are too small to define the schistose character if present. The rock is interpreted on the basis of bulk composition and relict textural evidence to be derived from a parent of basaltic composition.

Shell Development determined ages (Nicholas and Rosendal, 1975) 290 + or - 11 and 300 + or - 11 m. y. on phyllites from the interval 3000 to 3200 feet and an age of 274 + or - 10 m. y. from a sample near 10,000 feet. We have determined ages (Table 1) of 324 + or - 10 m. y. on a phyllite from 3700-3710 feet and another of 295 + or - 10 m. y. on a phyllite from 4700-4720 feet. In addition an age was determined on the greenschist from 10,150-10,160 feet of 254 + or - 5 m. y.

The interpretation of results from the well is far from unequivocal. The problem is what to do with the carbonate sequence below ~4900 feet and particularly below ~6500 feet. Is it the equivalent of foreland Ellenburger Group or is it the equivalent of the Marathon Limestone of the Marathon Basin? The direct evidence is lacking. However, because of the rather sharp break between the strongly sheared carbonaceous (possible graphitic) micaceous carbonates and the relatively undisturbed condition of the dolomitic carbonates below ~6500 feet, we interpret this as the thrust faulted contact between foreland and Ouachita facies. Another explanation is that the sharp break is simply caused by a difference in structural behavior of the relatively stiff underlying dolomite and the less competent impure carbonates.

If there is a thrust at ~6500 feet, then what is the age of the biotite quartzite below the carbonate - is it metamorphosed basal Paleozoic sandstone (Reagan equivalent) or is it Precambrian? Here the interpretation is less important to understanding the sequence in the well. The clear relict detrital texture in the quartzite as opposed to the virtually complete recrystallization of the greenschist suggests it is a metabasal sandstone. The isotopic ages indicate a strong enough metamorphic pulse in late Paleozoic to reset the isotopic ages. The greenschist is interpreted as being of Precambrian age but as having had the isotopic clock reset in late Paleozoic time.

Shell No. 1 Stewart

The Shell No. 1 Stewart in Val Verde County, Texas was drilled to a depth of 9693 feet on the Devils River Uplift. Nicholas and Rosendal (1975) have described this well and presented the results of numerous isotopic age determinations.

They report that the well drilled about 1500 feet of limestones and dolomites of the Ellenburger Group beneath

TABLE I. POTASSIUM-ARGON AGES FROM ROCKS ALONG THE OUACHITA BELT

Map #	Sample Type	Sample Wt. Gm.	% K	Ar ⁴⁰ : x10 ⁻⁹ moles	% Ar ⁴⁰	Age m. y.	Sample, Depth	Rock Type	Well, Location, Remarks
1	M	0.036 0.037	5.83	0.123 0.123	52 53	307 ± 6 296 ± 6	core 6819-20'	schist	Viersen and Cochran No. 1 Weyerhaeuser, NWNE SE, NW 25-5S-23E, McCurtain County, Oklahoma, drilled in the core of the Broken Bow Anticline.
	M	1.91 1.57	6.27	6.05 5.02	3 2	265 ± 5 268 ± 5	cuttings 9140-50'	schist	
	M	0.382 0.372	6.94	1.44 1.39	11 4	284 ± 6 282 ± 6	cuttings 9800-10'	schist	
	M	0.514 0.502	5.17	1.38 1.38	8 12	272 ± 5 277 ± 6	cuttings 10,019'	schist	
2	W	0.167 0.177	2.71	0.245 0.241	17 31	282 ± 6 273 ± 5	cuttings 13,180- 13,230'	phyllite	Shell No. 1 Barrett, Hill County, Texas, drilled on the Waco uplift.
	T	0.133 0.129	2.47	0.193 0.180	12 17	304 ± 6 294 ± 6	core 13,911'	marble	
	B	1.21 1.30	7.41	5.85 6.28	6 6	337 ± 7 336 ± 6	core 20,324'	granitic rock	
3	M	2.42 2.24	4.84	7.14 6.49	7 6	315 ± 6 310 ± 6	core 3759-90'	phyllite	Davis No. 1 Coffee, Milam County, Texas
4	M	2.57 2.54	3.06	4.12 3.94	12 17	273 ± 5 266 ± 5	core 4717-20'	phyllite	Magnolia No. 40 Mercer, Caldwell County, Texas
5	M	.274 .288	3.71	0.505 0.540	6 14	260 ± 5 265 ±	core 7805'	phyllite	United North and South No. 1 Kelly, Caldwell County, Texas
7	W	1.15 1.10	3.32	2.88 2.77	5 4	383 ± 8 385 ± 8	core 4390'	meta- rhyolite	Shell No. 1 Stewart, Val Verde County, Texas, drilled on the Devils River Uplift
	W	2.10 2.10	1.46	2.24 2.26	8 6	372 ± 7 373 ± 7	core 6001'	meta- dacite	
8	M	0.203 0.204	1.78	0.229 0.225	43 21	327 ± 7 320 ± 6	cuttings 3700-10'	phyllite	Shell No. 1 Gillis, Val Verde County, Texas, drilled on the Devils River Uplift
	M	0.207 0.205	1.65	0.188 0.199	23 34	287 ± 6 304 ± 6	cuttings 4700-20'	phyllite	
	W	0.245 0.205	1.44	0.171 0.144	24 22	254 ± 5 254 ± 5	cuttings 10,150-60'	greenschist	
9	W	3.00 3.26	0.074	0.104 0.115	74 77	245 ± 14 251 ± 17	core 2400-08'	greenschist	Husky No. 1 Rose-Robertson, Val Verde County, Texas, drilled on the Devils River Uplift
10	W	2.37 2.87	1.102	1.33 1.61	20 28	266 ± 5 267 ± 5	cuttings 4100-4200'	schist	Werblow No. 1 Newton, Val Verde County, Texas, drilled on the Devils River Uplift
11	W	2.25 2.52	1.74	2.36 2.80	19 12	311 ± 6 329 ± 7	cuttings 1848-3007'	schist	Plumer and Schwab No. 1 Roark, Brewster County, Texas
12	W	1.00 1.50	0.548	0.416 0.634	33 21	384 ± 8 389 ± 8	core 10,802-03'	diabase	Humble No. 1 Miller, Collin County, Texas, intrudes equivalents of Arbuckle Group

about 2000 feet of Cretaceous rocks. An additional 500 feet of dolomitic quartzite and sandy dolomite were drilled before metasedimentary and metavolcanic rocks were entered at about 4000 feet.

These metamorphic rocks were drilled to a depth of about 6800 feet where more massive meta-igneous rocks were encountered. At 9,070 feet a epidote-tremolite schist was encountered and drilled to total depth, 9693 feet.

The results of potassium-argon isotopic dating reported by Nicholas and Rozendal are disappointingly erratic. These ages should be the time of metamorphism under favorable conditions. But the ages scatter from 256 to 431 m. y. and show no relationship to depth. Six of the fourteen ages do fall in the range 256-287 m. y. and these six agree within limits of error at 265 m. y. This age is similar to other ages determined on interior zone rocks along the Ouachita System. Our ages (Table 1) determined on concentrations of micas from metarhyolites at 4390 and 6001 feet were 384 + or - 8 m. y. and 373 + or - 8 m. y. were in good agreement even though the potassium contents varied greatly. The older ages may represent isotopic systems set in the Precambrian and only partially reset in the late Paleozoic. However, there appears to be increasing evidence to indicate a widespread Devonian period of movement and metamorphism and these ages may well represent this time.

Nicholas and Rozendal reported Rb/Sr ages of 481 + or - 20 and 529 + or - 31 m. y. on isochrons from the upper metavolcanic and metasedimentary sequence. They interpret these as possibly equivalent to the Cambrian Carlton Rhyolite Group of southern Oklahoma. The lower more massive meta-igneous rocks yielded isochron ages of 1121 + or - 244 and 1246 + or - 270 m. y. indicating a definite Precambrian age for this unit.

We have determined three ages on three metarhyolites (Table 11) drilled on the Devils River Uplift. Two are from the Shell No. 1 Stewart and the other is from the Havoline No. 1 Weatherby in Kinney County (see Flawn and others, 1961, p. 284). Our isochron yielded an age of 699 + or - 26 m. y. with an initial Sr 87/86 ratio of 0.7072 + or - .0007. On the basis of this evidence we interpret the rhyolites as being extruded about 700 m. y. ago over an area of older, more massive igneous rocks.

The age of metamorphism is more difficult to define. Flawn (in Flawn and others, 1961, p. 144-145) notes that the foreland Paleozoic carbonates on the southern end of the Devils River Uplift show slight metamorphic effects. The effects seen in the basement rocks in the Stewart well are not mild. These are highly competent igneous rocks that have undergone extensive shearing, metamorphism, and plastic deformation. On this basis, it is concluded that they were initially metamorphosed before the transgression of the sea in early Paleozoic time. The apparent K/Ar ages in this interpretation would be the result of the resetting of the isotopic clock during one or more periods of thermal activity during Paleozoic time.

Shell No. 1 Barrett

The Shell No. 1 Barrett was drilled on an exceptionally large anticlinal feature lying about 25 miles southeast of the frontal edge of the Ouachitas. Rozendal and Erskine (1971) have described and interpreted the results of the drilling of the 20,310 foot test. Briefly summarized, the well drilled into low rank metamorphic rock of the interior zone beneath Cretaceous rock at 3,904 feet. At 13,595 feet, after drilling over 9,800 feet of slates and phyllites of the interior zone, the well entered a sequence of marbles. Underlying the marble at 19,780 feet is a biotite schist 125 feet thick

TABLE II. RUBIDIUM-STRONTIUM AGES FROM THE OUACHITA BELT

Map #	Sr ppm	Rb ppm	Sr ⁸⁷ /Sr ⁸⁶	Rb ⁸⁷ /Sr ⁸⁶	Age m.y.	Location, Rock Type, Remarks
6W	25.8	142.6	.8825	15.98	740 ± 30 ^{**}	Havoline No. 1 Weatherby, Kinney County, Texas, a core of metarhyolite taken at 4400' ±.
7W	151.1	147.0	.7360	2.81	692 ± 35 ^{**}	Shell No. 1 Stewart, Val Verde County, Texas, the older age is from a core of metadacite at 6001'; the younger age is from a core of metarhyolite at 4390'.
	336.1	84.1	.7148	0.72	712 ± 95 ^{**}	
F	8.56	20.06	.7551	6.76	489 ± 55	Granite boulders in Blakely sandstone NE NE NE SE 11-1N-15W, Saline County, Arkansas
F	3.63	20.46	.7743	16.28	283 ± 25	
F	4.57	15.74	.7696	9.93	432 ± 75	
F	7.19	23.42	.7548	9.40	351 ± 47	Granite boulder in Blakely sandstone SE NE SW 17-1N-14W, Pulaski County, Arkansas

W = Whole Rock
F = Feldspar

$$\lambda_{\beta} = 1.47 \times 10^{-11} \text{ yr}^{-1}$$

$$Rb^{87} = 0.283 \text{ gm/gm Rb.}$$

*Calculated initial ratio = .7072 feldspar ages have an assumed ratio of .7060.

which in turn was underlain by a pegmatitic granitoid rock to total depth.

Isotopic ages were reported on all pre-Mesozoic units. They ranged from 257 + or - 5 to 380 + or - 20 m. y. and showed no relationship to lithology or depth. Rozendal and Erskine concluded that the interior zone had been thrust over the marbles. The marbles contained inarticulate brachiopods and crinoid columnals which indicate a Paleozoic age. They interpreted the marbles as shelf carbonates of the foreland facies, possibly Cambrian-Ordovician in age.

We have also examined samples from the well and have reached the same conclusion. The marble sequence is highly disturbed near the top with plastic isoclinal folds but shows considerably less deformation at the base. The well-developed schist at the base of the marble seems to record a metamorphic event absent in the lower marbles. As a consequence, the rocks below the marble are firmly interpreted as Precambrian in age. Our isotopic ages are only a bit more straightforward than those determined by Shell Development (Table 1). What is clear is that the system has been intensely disturbed. The age determined near the base of the interior zone sequence was 277 + or - 10 m. y. Just below the interior zone phyllites, the talcose white mica in an isoclinally folded marble yielded an age of 298 + or - 10 m. y. The biotite near the bottom of the well yielded an age of 336 + or - 7 m. y.

Thus, while our ages show less scatter, they are fewer in number. The two ages on the marble and phyllite are regarded a shearing-metamorphism age. The biotite age on the basement is interpreted a partial resetting of an originally Precambrian, most likely ~ 1000 m. y., age.

The Barrett is exceptionally important because it demonstrates that: (1) the foreland carbonates extend at least 25 miles behind the frontal edge of the Ouachita system, and (2) the basement and overlying competent carbonates of the foreland are intimately involved in deformation.

Viersen and Cochran No. 1 Weyerhaeuser

The Viersen and Cochran No. 1 Weyerhaeuser was drilled in the core area of the Ouachita Mountains. The well in the NW NE SE NW of 25-5S-23E, McCurtain County, Oklahoma, was spudded in "Lukfata Sandstone"³ according to the map of Pitt (1955) and drilled to a depth of 10,010 feet. The Lukfata is the oldest unit in the Oklahoma Ouachitas and is of probable Late Cambrian or

³ The "Lukfata Sandstone" was believed to be overturned Crystal Mountain Sandstone and possibly Mazarn Shale based on limited lithologic data (Charles Stone, Arkansas Geological Commission, personal communication, 1970). John Repetski and Ray Ethington, University of Missouri (see paper in this volume) have now confirmed these correlations by conodont determinations.

Early Ordovician age. Ham (1959) estimated that the depth of true basement rocks (Precambrian) was less than 1000 feet stratigraphically below the "Lukfata."

It appears on the basis of this well that, in this very highly disturbed area, true Precambrian crystalline basement lies at a depth greater than 10,000 feet, quite possibly considerably greater. Over 140 thin sections have been examined from the well. Almost all of the thin sections were provided by the Oklahoma Geological Survey. On the basis of a reconnaissance examination the rocks in the well have been separated into four rather gross units on the basis of bulk composition.

Unit A	0-1800 feet	Composed of graphitic marbles with sandy and slaty-phyllitic intervals.
Unit B	1800-6900 feet	Composed mostly of calcareous quartz-albite-mica schists.
Unit C	6900-8800 feet	Composed mostly of rutiliferous graphitic phyllites and fine schists.
Unit D	8800-10,019 feet	Composed of mostly quartz-feldspathic schists with graphitic phyllite intervals.

The mineral assemblage changes with depth but the grade of metamorphism shows no significant differences. The rocks are all greenschist facies with the mineral assemblage depending on bulk-starting composition of the original sedimentary rocks.

For example while chlorite is the characteristic colored mica, biotite occurs in Units A and D but only in restricted intervals. The basic metamorphic assemblage found in all intervals is quartz-albite-muscovite (chlorite-rutile-opaque minerals-sphene-carbonate) with gross variation in the relative amounts of these minerals representing the difference in the recognized units.

Unit A is the most calcareous and contains many marble intervals. These are characterized by dolomite and calcite. Graphite occurs in disseminated specks in the carbonate and as intercrystalline masses. Many intervals show extensive recrystallization and twinning of the calcite. Graphitic phyllites and fine schists, many calcareous, are interbedded with the marbles. Opaque minerals include graphite and iron sulfides. Sphene and rutile are common, the latter nearly always present. Biotite and chlorite are erratic in distribution and generally very sparse. Shearing is best developed in the less competent, impure intervals. It was carbonaceous and probably dark in color. The unit was later metamorphosed to lower greenschist facies during a shearing episode.

Unit B is basically very similar in mineral assemblage to Unit A but contains considerably less carbonate. The

basic assemblage is quartz-albite-carbonate with lesser and erratically distributed muscovite, chlorite and phlogopite. Rutile and opaque minerals are never absent and locally are quite abundant. Rutiliferous graphitic phyllites and fine calcareous quartzo-feldspathic schists are the most common rock types. A core taken at 6818-6820 feet shows a highly deformed graphitic phyllite. Iron sulfides are in large porphyroblasts as well as in anhedral masses parallel to foliation. The mica-poor intervals are crystalloblastic mosaics of quartz and lesser albite containing discrete calcite masses and euhedral dolomite rhombs. Dimensional orientation in the quartzo-feldspathic intervals does not appear pronounced. The unit was deposited as a sandy and calcareous shale and later metamorphosed to lower greenschist facies.

Unit C is composed mostly of rutiliferous graphitic phyllites and fine schists. The typical assemblage is quartz-albite-muscovite-chlorite with lesser opaque minerals, rutile and sphene. Carbonate is erratic in distribution and lacking in many intervals. The phyllites show crinkled and disturbed foliation. The quartzo-feldspathic intervals are somewhat coarser and have well crystallized muscovite and chlorite. Detrital potash feldspar, mostly perthite, is in scattered sparse grains. The interval was deposited as alternating sandstones and shales later metamorphosed to lower greenschist facies.

Unit D is composed mostly of quartz-feldspathic schists with some intervals of graphitic phyllite. The schists have a relict detrital texture. These intervals contain large (to 3.5 mm) relict sand grains set in a crystalloblastic matrix characterized by quartz-albite-muscovite. The large relict sand grains are mostly single quartz but include plagioclase, perthite and microcline. The large quartz clasts are generally only mildly strained. Chlorite and reddish brown biotite are present but the former is much more widespread and abundant. Opaque minerals are never lacking and locally abundant. Rutile and sphene are widespread. The rock was deposited as a poorly sorted arkose, later metamorphosed to lower greenschist facies.

There is not a wide variety of trace accessory minerals in any unit. Detrital zircons are scattered through the entire sequence. Sparse, pale tourmaline is a metamorphic mineral in some intervals. In addition there were some metamorphic and detrital minerals that could not be identified. These did not characterize any interval and are sparse. The graphitic portions of the rock are less fully recrystallized which suggests that graphite or the carbonaceous material from which it was derived had a suppressive effect on crystal growth.

Two important questions concern this sequence of rocks: What is the age of deposition and from what kind of source area were they derived?

In a normal stratigraphic sequence these rocks should be older than the "Lukfata Sandstone". But this is in a highly disturbed area and large recumbent folds are common as well as thrust faults of considerable magnitude. There is no unequivocal reason for believing the rocks in the well are

older than "Lukfata".

What if the well was drilled on a large recumbent fold? In this case Unit A would be "Lukfata" and Collier, and in reverse stratigraphic order Unit B would have to be Crystal Mountain Sandstone. It becomes critical to the interpretation as to whether we accept the approximate thickness of 500 feet suggested by Honess (1923) or that of 5 to 100 feet estimated by Pitt (1955). If Pitt is correct, it is doubtful that the interval could be identified in the well without considerably more careful sampling and petrography in the critical interval. Let us accept Pitt's estimate simply because no Crystal Mountain can be recognized in the well at the degree of detail used. The Unit B becomes Mazarn "Shale". There appears to be more carbonate in Unit B than is reported by workers on the outcrop for Mazarn. Unit C is the speculative correlative of the Womble which Honess (1923) called a "schistose sandstone". This is probably not a bad match. Unit D, however, is definitely dissimilar to the Bigfork Chert, which is composed of chert, limestone and shale. Unit D does have considerable similarity to the Blaylock "Sandstone" which overlies the Bigfork. The Blaylock in McCurtain County according to Honess (1923) is about 800 feet thick and is reportedly feldspathic. The microscopic description by Honess of the Blaylock does not fit too well with Unit D.

It is concluded that while there are some positive correlations for a recumbent fold at the site of the Weyerhaeuser well, it is not a close analogy. One might develop a sequence through faulting which would make a near perfect match but for which there is no positive evidence.

The alternative is to call upon a normal stratigraphic sequence, getting older with depth or some modification thereof. The modification would be caused by structure and would require, on a gross scale, the absence of unknown units by faulting. Each of the four basic units is distinctive and is not repeated.

If inclined to accept this latter interpretation, then one might entertain the idea that the well at total depth was in a stratigraphic unit originally deposited near the base of the normal sedimentary sequence. This is prompted by the highly feldspathic character of the unit which suggests it was derived from a nearby granitic source. Perhaps it was deposited as a basal transgressive sandstone.

Much of the coarse detrital fraction of the rocks drilled in the well suggests a granitic source. Detrital sodic plagioclase is widely distributed and large single quartz crystals as sand grains suggest a coarse granitic source. The rather limited suite of stable heavy minerals supports this. Only in Unit D is the evidence for a coarse granitic source overwhelming. Here the abundance of detrital plagioclase, microcline and perthite together with large single quartz virtually eliminates any other type of source. It must have been geographically close to the depositional site.

The source indicated for Unit D is similar to that dictated for the Blakely in Arkansas by granitic boulders, dis-

cussed earlier. Ages were determined on four intervals in the well. The ages fall into a range from 307 + or - 6 to 265 + or - 5 m. y. (Table 1). There is no direct relationship between the apparent age and depth or potassium content. The oldest age was determined on mica separated from a core taken at the base of Unit B. The other ages were determined on micas separated from quartzo-feldspathic schist cuttings taken in Unit D. The apparent ages from Unit D are latest Pennsylvanian to Early Permian. The ages are well within the range of those determined on other rocks from the interior zone. The ages from Unit D are younger than any yet determined from outcrop samples in the Broken Bow anticlinorium.

The rocks are very highly disturbed and meet most of the descriptive criteria listed by Flawn for interior zone rocks. The metamorphic grade and mineral assemblages are also typical of the interior zone. It is concluded that the well penetrated rocks from surface to total depth that are equivalent to the interior zone of the Ouachita fold belt.

OTHER AGES FROM WELLS

We have determined ages on several older wells penetrating interior zone rock along the length of the Ouachita system. These will be briefly described with the appropriate page in Flawn and others (1961) where the wells are discussed and locations described. The analytical data are reported in Table 1.

Coffee No. 1 Davis, Milam County, Texas (Flawn and others, 1961, p. 297). A mica concentrate from a core at 3759-3790 feet yielded a K/Ar age of 313 + or - 7 m. y. average of three agreeing measurements. This age is well within the range of determination from the interior zone.

Werblow No. 1 Newton, Val Verde County, Texas (Flawn and others, 1961, p. 331). Cuttings from calcitic schists from 4100-4200 feet were used to determine a whole rock K/Ar age. The ages do not agree within error although there are four determinations, but the average of 268 + or - 12 m. y. is certainly close to the average isotopic age on the interior zone rocks on the Devils River Uplift.

Plumer and Schwab No. 1 Roark, Brewster County, Texas (Flawn and others, 1961, p. 235-236). The graphitic slate penetrated from 1848-3007 feet in this well yielded an average K/Ar age of 320 + or - 13 m. y. This is near the maximum peak of ages of younger Ouachita metamorphism.

United North and South No. 1 Kelly, Caldwell County, Texas (Flawn and others, 1961, p. 240). A core of fine sericite schist taken at 7805 feet yielded an average K/Ar age of 263 + or - 7 m. y. Another close well, the Magnolia No. 1 Mercer, had interior zone rock of the same age.

Magnolia No. 40 Mercer, Caldwell County, Texas (Flawn and others, 1961, p. 239-240). A core of fine sericite schist taken at 4717-20 feet yielded an average K/Ar age of 270 + or - 6 m. y. The agreement with this and another age in

Caldwell County is interesting because they lie directly opposite the buttress of the Llano Uplift.

The Husky No. 1 Rose-Robertson, Val Verde County, Texas (Flawn and others, 1961, p. 324). A core of actinolite-epidote-chlorite schist was probably from the interval 2400-2408 feet. The whole rock from this greenschist yielded an average K/Ar age of 248 + or - 15 m. y. The very low potassium content (<0.1% K) suggests that if the rock were derived from a basaltic rock as it appears, then it may be of "oceanic" origin.

The Humble No. 1 Miller, Collin County, Texas (Flawn and others, 1961, p. 242). A diabase within the Ellenburger was cored in this well. We determined an age of 387 + or - 10 m. y. on a core sample at 10,802-03 feet which substantiates the interpretation that the igneous rock is intrusive into the Ellenburger.

SOUTH INTO MEXICO

The extension of the Ouachita system into Mexico has been of great interest. Unfortunately outcrops and well control are very sparse southward. Nonetheless, there are occurrences of rock bearing an astonishing resemblance to typical interior zone schist in northeastern Mexico. These rocks have yielded ages identical to those determined in Texas and Oklahoma.

The graphitic schists occurring at Sierra del Carmen in Coahuila were interpreted as equivalent to the interior zone (Flawn and Maxwell, 1958; Flawn, in Flawn and others, 1961). Denison and others (1969) determined ages 263 + or - 5 m. y. by K/Ar and 275 + or - 20 m. y. by Rb/Sr on samples collected from this outcrop. Thus, the outcrop closest to the known Ouachita rocks yields ages in the proper range.

About 660 km (410 miles) southeast of Sierra del Carmen lies a small outcrop of graphitic schists near Aramberri, Nuevo Leon. These outcrops have been described by Mixon (1963). Denison and others (1971) determined ages of 270 + or - 5 and 294 + or - 6 m. y. on mica separates from these schists.

These schists at Aramberri are identical to more extensive outcrops in several canyons outside Ciudad Victoria, Tamaulipas, some 65-75 km (40-45 miles) to the southeast. Carrillo (1961) has provided the definitive description of these exceptionally important rocks.

Three types of rock are found unconformably beneath folded Jurassic and Cretaceous rocks. There are two types of crystalline rocks, the Granjeno Schist and the Novillo Gneiss. In addition unmetamorphosed sedimentary rocks that are definitely as old as Silurian and possibly as old as Cambrian-Ordovician have been identified. The Granjeno is a graphitic schist containing a typical greenschist facies assemblage. The Novillo is a high grade quartzo-feldspathic banded gneiss.

Denison and others (1971) determined an age of about 900 m. y. for the age of metamorphism of the Novillo Gneiss. They determined ages of close to 300 m. y. on the micas in the Granjeno Schist for the age of metamorphism. In order to get the unmetamorphosed pre-Pennsylvanian rocks in juxtaposition with the Granjeno Schist, Denison and others interpreted the schist as an allochthonous thrust in order to account for the differences in isotopic age and grade of metamorphism.

The Granjeno and the schists at Aramberri have the same metamorphic age range as interior zone rocks. They bear an astonishing petrographic resemblance to the interior zone schists. The problem is that these exposures lie 550 km (340 miles) south of known interior zone rocks in Maverick and Zavala Counties, Texas, and, as noted, even farther from the schists at Sierra del Carmen. Such great distances cause justified trepidation in making an unequivocal correlation. However, it should be recognized that the distance between these Texas wells and the Broken Bow uplift to the northeast is about 750 km (460 miles). Viewed in this context it does not seem unreasonable that these outcrops in northeastern Mexico are indeed equivalent to the interior zone of the Ouachita system.

METAMORPHISM IN OUACHITA ROCKS

The metamorphic character of the early Paleozoic rocks in the Ouachita Mountains has been recognized for many years (Purdue, 1909; Honess, 1923; Miser, 1943; Goldstein and Reno, 1952, for just a few of the many papers which describe the metamorphism). However, the pervasive metamorphic effects are not widely appreciated, perhaps because the stratigraphic units have sedimentary names. For example, the Ordovician Collier is called a shale (though it contains much marble and some metasandstones) when it is actually a slate of phyllite. On the outcrop, axial plane cleavage and isoclinal folding are well displayed and clearly indicate the low rank metamorphism that took place during shearing. However, metamorphic minerals in hand specimen are difficult to identify, other than the development of silky metamorphic sheen on cleavage surfaces.

Goldstein and Reno (1952) and Goldstein (in Flawn and others, 1961) describe the metamorphic assemblages found by thin section examination of the Ouachita rocks. This is not to say this had not been done before; for example, Honess (1923) reports chlorite and sericite as typical crystalloblastic minerals in the older units.

Actually in the more fully recrystallized rocks seen in the outcrop and in the No. 1 Weyerhaeuser well, muscovite, chlorite, biotite, phlogopite, rutile, sphene, albite, tourmaline, various iron sulfides, graphite, and epidote are found as metamorphic minerals. Quartz and carbonates are found as completely crystallized depositional minerals. How can rocks bearing such assemblages be called shale, sandstone, and limestone? One reason is that the rocks are very commonly incompletely recrystallized with relict depositional fabric still clearly defined both in hand specimen and thin section. The second reason is perhaps that we are

conditioned to hearing them called shale, sandstone, and limestone and resist the somewhat foreign metamorphic terms. Nonetheless, the older units beneath the Bigfork Chert are more accurately referred to as slate, phyllite, marble, and metasandstone.

Although it is the older rocks in the Broken Bow-Benton Uplift that show the recrystallization most clearly, younger rocks show these effects to a lesser degree. Goldstein (in Flawn and others, 1961, pp. 33-42) describes recrystallized clay matrix in rocks as young as Atoka near the central anticlinorium but no recrystallization away from the older rocks.

Miser (1943) pointed out that the areas of most intense metamorphism coincide with the concentration of quartz veins. These veins are as large as 100 feet in width. In addition to quartz, the veins commonly contain calcite, dickite, rectorite, chlorite, adularia, and orthoclase as relatively minor associated minerals.

Bass and Ferrara (1969) determined ages on adularia from quartz veins at the Hamilton Hill locality in Garland County, Arkansas. Their results by K/Ar were 190 + or - 5, 205 + or - 6 and 214 + or - 6 m. y., apparent Triassic ages. They ascribed the low ages to argon loss which appears quite reasonable. The Rb/Sr data on the 205 m. y. old sample yields an age between 263 to 271 m. y. (reduced to 47 b. y. half life of Rb 87 used here). The difference in age is dependent on whether an initial Sr 87/86 ratio of 0.704 or 0.709 (for the younger age) is chosen. This would seem to be the initial Sr. 87/86 ratio range that would yield the maximum age, even an initial of 0.700 would yield an age of less than 280 m. y. Bass and Ferrara conclude that the data indicate that the quartz veins were implaced well after the mid-Pennsylvanian time postulated by previous workers. Using our timescale and recalculated ages, it would indicate an Early Permian time of crystallization, if the initial ratio range used is reasonable.

AGES FROM THE BROKEN BOW UPLIFT

We have determined ages on nine outcrop samples near the core of the Broken Bow Anticline. The results are very consistent but raise more questions than they resolve.

We have seen that isotopic ages from the Viersen and Cochran Weyerhaeuser well drilled in the core area yielded reasonably consistent ages from 267 to 303 m. y. Our ages from the Collier Slate from outcrops near the well are 301 and 318 m. y. These are certainly well within the range of ages from along the belt.

However, our results from the overlying Mazarn Slate fall into two sets. Two of the seven fall into the upper range of Collier-interior zone ages, 313 and 324 m. y. The remaining five ages have a range of 358 to 378 m. y. - distinctly older than the upper limit of interior zone ages.

All these ages were determined on micaceous concentrates. Some are more pure sericite-fine muscovite than others

and there appears to be a direct but imprecise relationship between the age and the percentage of potassium in the dated separation (Table III). Certainly all the ages on samples with more than 5% K are in the interior zone range. And all but one of the samples below 4% K have an age above 350 m. y. It should be noted that the one sample below 4% K that yielded a young age is a Collier Slate. All the samples yielding high ages are Mazarn.

The imprecise relationship between age and potassium content suggests that the micas giving high ages are less pure or less metamorphosed than the high potassium samples. X-ray data were gathered on the mica separates using the techniques of Weaver (in Flawn and others, p. 147-162). None of our sharpness ratios were over 5 which is Weaver's cutoff for incipiently metamorphosed rocks, but this is probably due to our machine settings and sample handling. In thin section the rocks are very clearly metamorphic.

The results of this work showed that there was no essential difference in the metamorphic recrystallization of the samples in relationship to their apparent isotopic age. The petrography also shows no essential difference in degree of metamorphism or mineral assemblage. Sericite and chlorite are present in all samples.

These results tend to discount the possibility that the older ages are caused by inherited detrital micas. This leads to the conclusion that the older ages are genuine and reflect the time of metamorphism of these samples.

Regionally along the Ouachita belt there is support for this as being a time of metamorphism and shearing. The highest ages from the metamorphic rocks along the Waco Uplift and Devils River Uplift are in this range. The age of metamorphism found in the Haymond boulders is within the limits of error of the upper range of these ages.

What then is the explanation of how these rocks apparently metamorphosed during Devonian are in close geographic proximity to low rank metamorphics yielding considerably younger ages? These ages also suggest that the metamorphism was taking place while the Arkansas Novaculite was being deposited only a few thousand feet above the metamorphism-shearing zone!

If the peaks of ages are accepted as representing periods of metamorphic activity, then the explanation must be tectonic. This requires fault slices that bring blocks of rock yielding younger ages, older ages, and Arkansas Novaculite into juxtaposition. The field evidence for this is not avail-

TABLE III. POTASSIUM-ARGON AGES FROM ROCKS ALONG THE OUACHITA BELT (CONTINUED)

Sample Type	Sample wt. gms.	% K	$\frac{Ar^{40}}{K} \times 10^{-5}$ moles	% $\frac{Ar^{40}}{Ar^{40}}$	Age m. y.	Rock Unit and Type, Location and Remarks
M	0.273 0.278	5.84	0.981 0.995	3 10	318 ± 6 317 ± 6	Collier phyllite, SW SE 17-5S-24E, McCurtain County, Oklahoma
M	2.50 2.50	3.12	4.52 4.53	14 16	301 ± 6 301 ± 6	Collier phyllite, SE SE SW-27-5S-23E, McCurtain County, Oklahoma
M	0.069 0.065	5.77	0.241 0.228	19 39	311 ± 6 315 ± 6	Mazarn phyllite, C NW 35-4S-24E, McCurtain County, Oklahoma
M	0.276 0.278	2.71	0.522 0.531	12 13	356 ± 7 360 ± 7	Mazarn phyllite, NE NW SW 22-5S-24E, McCurtain County, Oklahoma
M	0.490 0.483	2.67	0.947 0.932	15 14	369 ± 7 369 ± 7	Mazarn phyllite, NE NW SW 22-5S-24E, McCurtain County, Oklahoma
M	0.282 0.271	2.72	0.554 0.533	38 45	368 ± 7 368 ± 7	Mazarn schistose metasandstone, SW SW NW 23-5S-23E, McCurtain County, Oklahoma
M	2.20 2.00	3.60	5.90 5.37	13 8	378 ± 8 378 ± 8	Mazarn phyllite, same location as above
M	0.070 0.074	5.43	0.238 0.254	24 38	322 ± 6 325 ± 7	Mazarn phyllite, NE SW 20-5S-23E, McCurtain County, Oklahoma
W	0.756 0.816	0.367	0.0954 0.102	57 29	185 ± 6 183 ± 4	Diabase from 4245-48', Lion No. 1 Reap, SW SE NE 25-11S-9W, Cleveland County, Arkansas
W	0.760 0.754	0.768	0.217 0.213	22 47	198 ± 4 196 ± 4	Diabase from 4683-90', Barnsdall and Ohio No. 1 Schulz, NE NE 34-13S-26W, Hempstead County, Arkansas
W	0.840 0.837	0.747	0.213 0.211	25 37	182 ± 4 181 ± 4	Diabase from 3155-3255', Ryan Consolidated No. 1 Laws, 34-12S-23W, Nevada County, Arkansas
W	0.865 0.783	0.699	0.145 0.129	50 51	131 ± 3 129 ± 3	Diabase from 9967-73', Atlantic No. 1 Montana Realty, 17-17S-28W, Miller County, Arkansas

M = Muscovite
W = Whole Rock
T = Talc
B = Biotite

Ar^{40} is radiogenic argon.
 $K^{40}/K = 1.22 \times 10^{-4}$ g./g.
 $\lambda_{\beta} = 4.72 \times 10^{-10}$ yr⁻¹.
 $\lambda_{\epsilon} = 5.85 \times 10^{-11}$ yr⁻¹.

able.

For example, in the highly deformed, isoclinally folded anticlinal core of the Broken Bow Uplift, Honess (1923) shows few faults and Pitt (1955) shows none at all! This is caused by very poor exposures. It is exceptionally difficult to believe that there are not numerous faults in this highly disturbed area, but they have not been recognized and it will not be an easy matter to do so.

What then can be concluded from the ages determined in the Broken Bow Uplift area? The younger suite of ages indicates a time of metamorphism ranging from near the Mississippian-Pennsylvanian boundary through Early Permian, say Wolfcampian. The tailing of ages may represent the later cooling through a critical temperature range or may actually represent the final stages of metamorphism-shearing. The older cluster of ages is Devonian and, if accepted, requires considerable reevaluation of the available mapping in the core of the uplift. Although we are not confident that this is actually the case, we do feel the evidence is strong enough to merit consideration.

YOUNGER IGNEOUS INTRUSIONS INTO THE OUACHITA SYSTEM

A wide variety of igneous intrusions cut the deformed rocks of the Ouachita system. These have been described in a number of papers, the more important of which include Williams (1891), Miser and Ross (1922), Lonsdale (1927), Moody (1949), Kidwell (1949), and Spencer (1969). Kidwell reports the composition of intrusions fall into two petrographic provinces: (1) a diabase-diorite and (2) a highly diverse alkalic and generally subsilicic suite. The age of the alkalic suite has generally been regarded as Cretaceous based on firm geologic evidence in Arkansas (Ross and others, 1929). The age of the diabase-diorite suite was shown by Kidwell (p. 228) and Moody (p. 1417) to be post-Eagle Mills (Late Triassic) and pre-Cotton Valley (Late Jurassic).

The work of Zartman and others (1967) and Baldwin and Adams (1971) has shown the alkalic suite in Arkansas was intruded during an interval 90-100 m. y. ago (mid-Cretaceous). The alkalic rock in the Uvalde area of Texas has a somewhat younger range from 63-86 m. y. (Baldwin and Adams, 1971, Burke and others, 1969). These ages range from mid-Late Cretaceous to earliest Paleocene.

The diabase-diorite petrographic suite has yielded a fairly narrow range of isotopic ages with some exceptions. Baldwin and Adams report ages of 196 ± 8 m. y. on a diorite from the Carter No. C-1 Hope in Morehouse Parish, Louisiana, and 197 ± 8 m. y. from a quartz diabase from the Barnsdall No. 1 Schultz in Hempstead County, Arkansas. We have obtained an identical age from the Hempstead County well (see Table III) as well as two lower ages (181-185 m. y.) on diabases from wells in Nevada and Cleveland Counties, Arkansas. A significantly lower age of 130 ± 3 m. y. was obtained on a diabase from the Atlantic No. 1 Montana Realty in Miller County,

Arkansas. The meaning of this lower age is not known; the rock is somewhat altered but not significantly more than other diabases from which higher ages were obtained.

These ages are Late Triassic to Early Jurassic and are recorded through much of the Gulf Coastal area eastward through Mississippi (Denison and Muehlberger, 1963) and into Florida (Milton and Grasty, 1969). This appears to be a more widespread period of igneous activity than that in the Cretaceous although less visible because rocks of this older age are not known to crop out west of the Appalachians.

IGNEOUS ROCKS OF UNKNOWN AGE

There are minor occurrences of igneous rocks of unknown age found on the outcrop and in the subsurface.

Honess (1923, p. 210-212) describes a diorite sill intruded into Womble "schistose sandstone" in McCurtain County. He noted that the rock is intensely fractured and was, therefore, younger than Womble (mid-Ordovician) and older than the folding (Pennsylvanian). Miser (1954) indicated the age as Ordovician(?) on the state geologic map.

Wilson (1954, p. 2465) describes a pre-Cretaceous sill of gabbro intruded into the Marathon Limestone (Early Ordovician).

Flawn (in Flawn and others, 1961, p. 109-115) describes several occurrences of igneous rocks within the subsurface Ouachita belt that may be either Paleozoic or Precambrian.

Miser (1943) believed that the numerous quartz veins in Arkansas and Oklahoma were evidence of widespread igneous activity. This is not supported by the occurrence of igneous bodies of any significance. It is clear that the metamorphism seen particularly well in the central anticlinorium of the Ouachitas is caused by a significantly higher than normal thermal gradient during shearing. Certainly this indicates a favorable situation for the generation of magmas. If large amounts of magma are generated, the resultant plutons have not been found either on the outcrop or in the subsurface.

HATTON TUFF

The Hatton Tuff and Chickasaw Creek Tuff beds occur at the base and top of the Stanley Shale respectively. The shale is thought to be of Mississippian and/or Pennsylvanian age. Honess (1923, p. 179-188) showed that several samples of tuff were demonstrably ash falls. The composition of the tuffs is highly siliceous containing considerably more plagioclase than potash feldspar based on normative calculation as well as thin section examination (Honess, p. 188). Thin sections examined in our studies clearly show an abundance of devitrified pumiceous fragments.

Mose (1969) determined an isochron age on samples

from the Hatton Tuff of 293 ± 15 m. y. (reduced to the decay constant used here). This would suggest a mid- to late Pennsylvanian age and is probably too young. A mineral isochron from one of the samples yields an age of 322 ± 26 m. y. This age, although the same within error as the whole rock isochron, gives one more latitude and certainly is in the range expected for rocks of mid-Carboniferous age. Mose speculated that the tuff may have taken "several tens of millions of years" to equilibrate with the seawater into which it fell. He, however, rejects this as a possible explanation. The high initial Sr 87/86 ratio of 0.711 is certainly higher than is known from Mississippian and Pennsylvanian seawater (Peterman and others, 1970). The high initial ratio suggests that the magma which gave rise to the Hatton Tuff was derived from or contaminated by continental material. The Haymond boulders also show a high initial ratio (0.712 found by Denison and others, 1969). Mose also determined the Rb/Sr age on four samples of the Chickasaw Creek Tuff lying at top of the Stanley but they did not form an isochron.

The occurrence of the Hatton Tuff in the Stanley Shale demonstrates that during this time period explosive igneous activity affected a wide area. As Honess put it (p. 182), "a Carboniferous Krakatoa if you please". Further isotopic evidence suggests that this magma was derived from the partial melting of continental rocks or strongly contaminated by them.

SUMMARY

It is worthwhile to review our conclusions and supporting evidence concerning the provenance and metamorphic activity that have affected the Ouachita system.

Pre-Devonian

In Cambrian through Ordovician in the Marathon Basin igneous and metamorphic rocks contributed a small but rather consistent part of the detritus in the clastic rocks. This appears to be derived from the shelf side (northwest) based on current direction studies.

During the same period in the Ouachita Mountains and extending into the Silurian the units in the Ouachita Mountains show that granitic basement rocks were exposed in the source area. It is not clear from available data where this source area was in relationship to the depositional basin. The metamorphism and highly sheared state of these older rocks will make the determination of source direction from depositional structure studies difficult. The thickness of the Blaylock Sandstone shows that the source area was considerably closer to the Arkansas portion of the depositional basin and the unit thins markedly into Oklahoma.

Devonian

The first evidence for igneous and metamorphic activity occurs during the Devonian. None of the evidence is parti-

cularly strong when taken individually, but collectively it appears to be geographically widely distributed.

In the Marathon Basin sequence the igneous and metamorphic boulders in the Pennsylvanian Haymond Formation yield Devonian ages. The isotopic and petrographic evidence suggests these rocks were being formed in Silurian-Devonian time by partial melting of a geosynclinal sequence which later gave rise to the source of some of the Haymond detritus.

Along the Devils River Uplift the oldest metamorphic ages on metarhyolites, formed in Precambrian time, are Devonian.

A single diabase intruded into foreland Ordovician carbonates in Collin County, Texas, yields an apparent Devonian age.

The older suite of ages from the Oklahoma Ouachitas are Devonian. These were determined on low rank metamorphic slates and phyllites deposited in Ordovician time.

Pennsylvanian-Permian

The period beginning about 320 m. y. ago (near the Pennsylvanian-Mississippian boundary) and continuing for about 70 m. y. into the mid-Permian was the major time of tectonic movement and also the latest period of metamorphic activity.

The activity is distributed from Nuevo Leon and Tamaulipas in Mexico through the outcrop of the Ouachita Mountains. Isotopic ages along the buried front show that this period of movement-metamorphism affected the entire belt and undoubtedly reset ages of many rocks that are interpreted here as having been originally metamorphosed in Devonian time.

The isotopic evidence strongly indicates that the last movements all along the Ouachita belt occurred in Permian time. It is known that the last demonstrable movement occurred in the Marathon area during Early Permian time, but only into the mid-Pennsylvanian in the Ouachita area. We believe that two findings of our work strongly support, if not prove, the extension of these movements into the Permian.

First is the very wide distribution of apparent Permian ages from northern Mexico to southeastern Oklahoma determined on interior zone metamorphic rocks. This shows, in the most conservative interpretation, that the area was hot enough so that the K/Ar isotopic clock was not set until this time. The second and most critical observation is that low rank metamorphic rocks yielding apparent ages of Permian and Devonian are in close geographic proximity in the core area of the Oklahoma Ouachitas. Surely the heat that caused the setting of the Permian ages would have affected these rocks in juxtaposition yielding Devonian ages if they had been in close proximity during the setting of the clocks. Our interpreta-

tion based on this evidence is that the rocks yielding Permian age were brought into contact with those yielding Devonian ages by tectonic transport *after* the setting of the Permian clocks.

This history of the movement and metamorphism of the Ouachita belt as suggested by isotopic ages is more complex than was previously thought. The validity of our suggested sequence of events must wait further substantiation. However we believe that the evidence, while not conclusive, is strong enough to merit careful consideration.

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MARATHON REVISITED¹

By Philip B. King²

ABSTRACT

Known history of the Marathon segment of the Ouachita orogenic belt begins with a leptogeosynclinal phase from Cambrian to early Mississippian time, when modest thickness of 2,500 feet of sediments accumulated in the geosynclinal trough under stable conditions.

These are succeeded abruptly by a flysch sequence of late Mississippian through Atokan age (Tesnus, Dimple, and Haymond Formations), which expresses quickening tectonic activity. The flysch sequence attains a thickness of 15,000 feet within the trough, but nearly wedges out at its northwestern edge. An exaggerated phase is the wildflysch of the Haymond, whose small to giant clasts were derived partly from uplifts within the trough, and partly from the foreland and backland.

Succeeding the flysch are shallow-water shales and carbonates 2,000 feet thick (Gaptank and Neal Ranch Formations, Des Moinesian into Wolfcampian), with roundstone conglomerates in the lower part indicative of major orogeny within the geosyncline; probably most of the complex structures visible in the geosynclinal rocks of the Marathon Region result from this orogeny.

In late Pennsylvanian time a new flysch trough grew in the Val Verde basin in front of the orogenic belt, whose deposits are known mostly from subsurface, but which are exposed in a small area in the northwestern part of the Marathon Region; this flysch was deposited on cratonic rather than geosynclinal rocks. Flysch accumulation closed with a late Virgilian-early Wolfcampian orogeny, during which the already deformed rocks of the geosyncline were sheared off and carried many miles northwestward along the Dugout Creek thrust, whose nearly flat surface contrasts with the décollement thrusts along the orogenic front in the Southern Appalachian and Ouachita Mountains. Succeeding Permian strata of the Glass Mountains lie with right-angled unconformity on rocks deformed by this second orogeny.

INTRODUCTION

In this article I will review the pre-Permian sedimentary and tectonic evolution of the Marathon Region -- the far-flung western outpost of the Ouachita orogenic belt in western Texas. In doing so, I return to the subject of my first Geological Survey project of 43 years ago; it is an alluring subject, and it is difficult to let it go. After my report was published (King, 1937) many other geologists have worked there and an extensive literature has grown up, not all of it in agreement. Also, my own ideas on the region have evolved through the years, partly because of increase in knowledge of the region, partly because of changes in general tectonic and sedimentologic concepts; thus, interpretations which I published earlier may not correspond to those I hold today. But after all, even in the classic region of the European Alps the tectonic and sedimentologic concepts are quite different now from what they were 43 years ago, we geologists in the outlands may therefore be pardoned for also changing our views.

Much of the recent work in the Marathon Region has dealt with the sedimentology, and the tectonics has been treated only by implication; my own interest is primarily in tectonics. In general, the concepts which I have evolved correspond closely to those of Earle McBride of the University of Texas at Austin and his collaborators (for summaries see Thomson, ed., 1964; McBride, ed., 1969; McBride,

1970b), but I will present some emendations and additions of my own. Through the years, various other interpretations of the sedimentology have been made, most recently by Flores and his associates (Flores and Ferm, 1970; Flores, 1972; and later abstracts), and by Folk (1973). These deserve more respectful attention than can be given here. I will leave the details of the sedimentological arguments to others.

GENERAL FEATURES OF MARATHON REGION

The Marathon Region is doubtless familiar to many geologists in the United States but some recapitulation is worthwhile. In Figure 1, observe the Midcontinent Paleozoic, with older rocks emerging in the Ozark, Arbuckle, Wichita, and Llano uplifts. Observe the Mesozoic and Cenozoic cover rocks to the southeast, south, and west in the Gulf Coastal Plain and High Plains. The quite different Paleozoic rocks of the Ouachita orogenic belt emerge north of the Coastal Plain border in the Ouachita Mountains of Arkansas and Oklahoma, and again in smaller areas 500 miles away in western Texas, in the Marathon Region and The Solitario. Despite the wide separation, the rocks and structures of the Ouachita belt in the two regions are surprisingly alike; evidently they are connected beneath the cover rocks, around the southeastern side of the Llano uplift.

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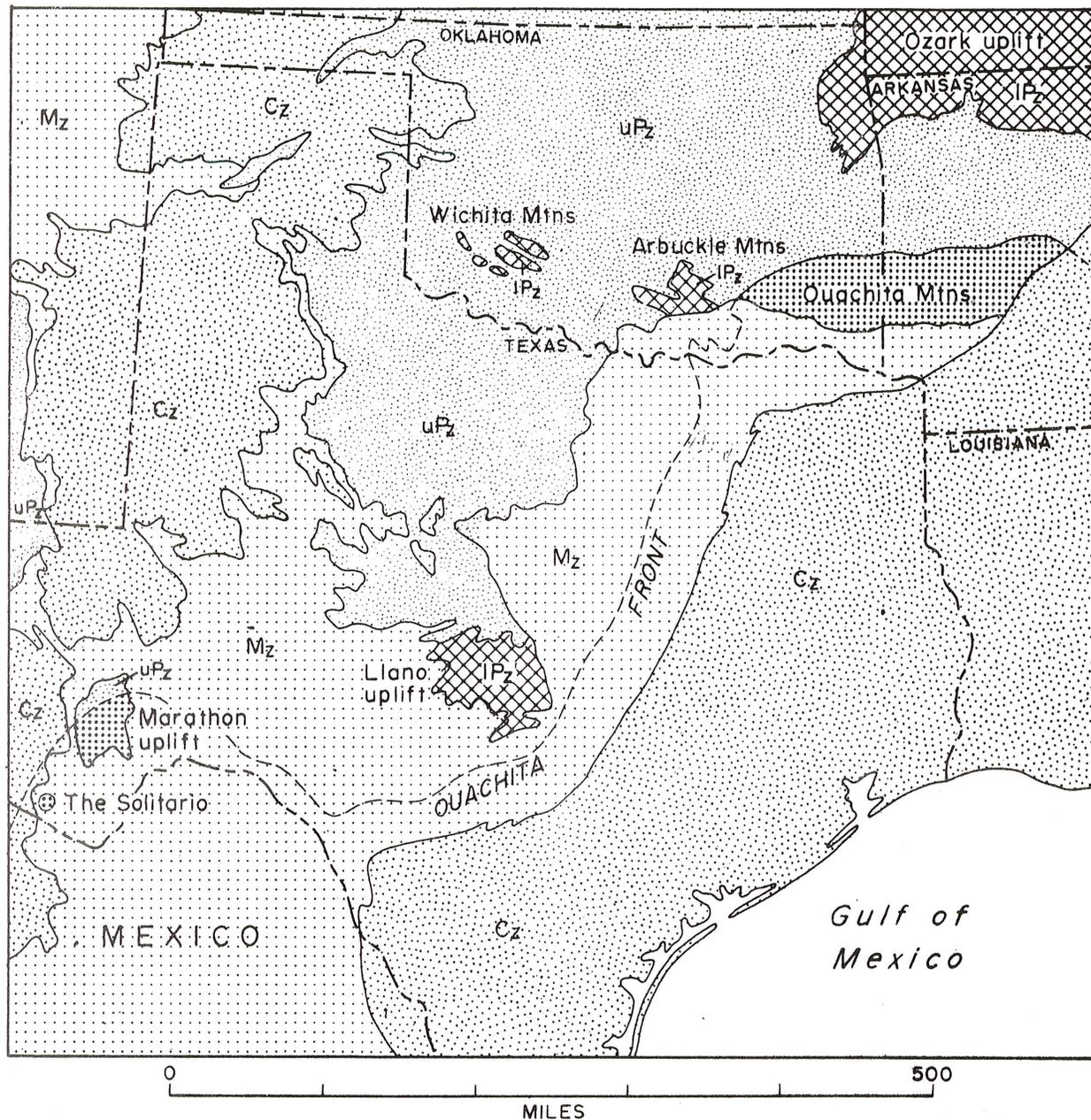
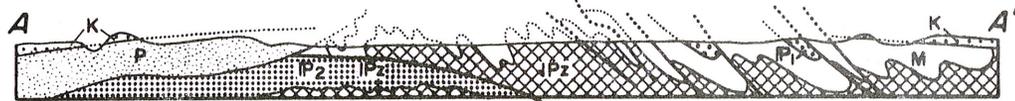
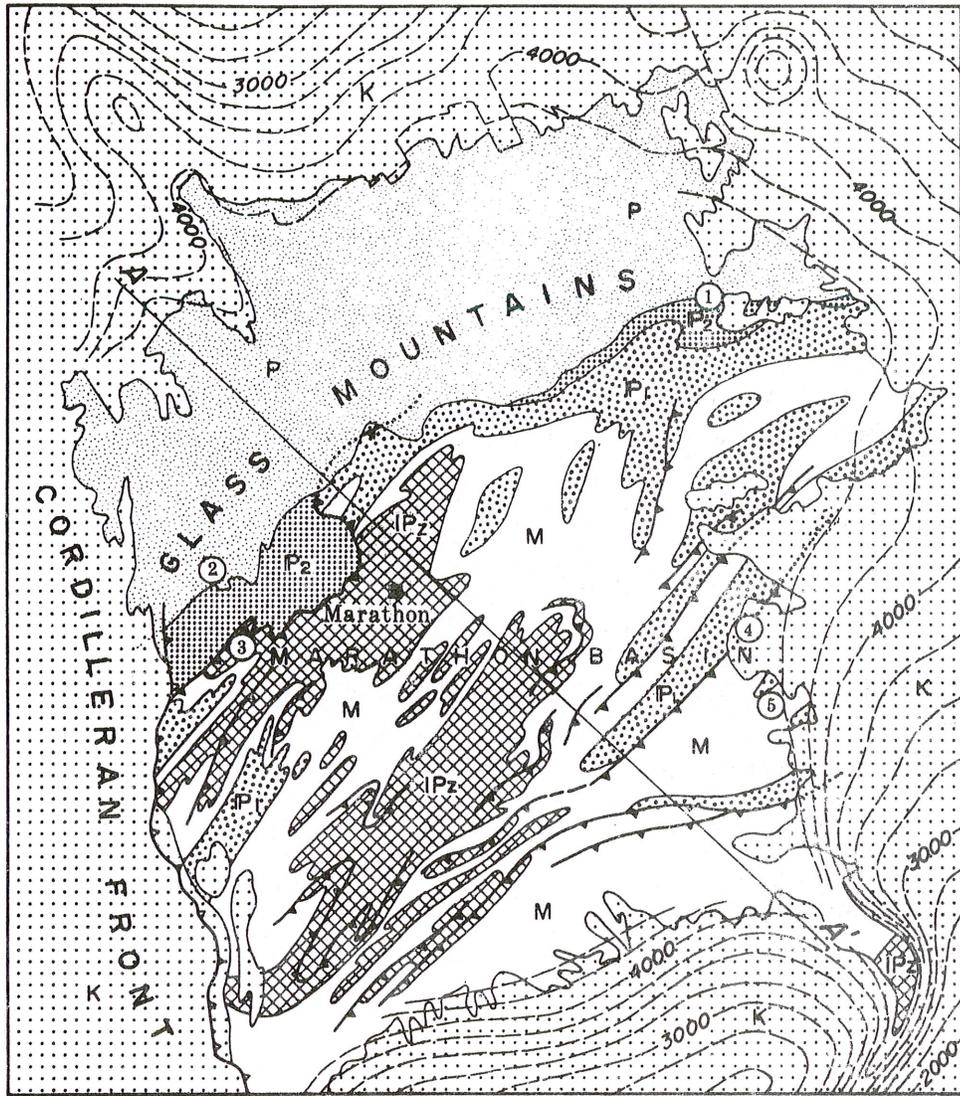


Figure 1. -- Map of part of the south-central United States, showing location of Ouachita Mountains and Marathon Region.



Cratonic sequence

Geosynclinal sequence

25 Miles



Figure 2. -- Generalized geologic map and section of the Marathon Region, west Texas. Numbers indicate localities mentioned in text; 1 - Gap Tank; 2 - Dugout Mountain; 3 - Payne Hills; 4 - Housstop Mountain; 5 - Tesnus.



Figure 3. -- Angular unconformity between Paleozoic and cover rocks of Marathon Region; Housatop Mountain 18 miles east of Marathon, on eastern rim of Marathon Basin. Tilted Mississippian sandstones and shales (Tesusus Formation) overlain by flat-lying Lower Cretaceous carbonates (Glen Rose Formation).

Figure 2 shows the Marathon Region in more detail, a grossly circular domical area 40 miles across, surrounded by cover rocks -- Cretaceous on all sides, Permian in the Glass Mountains on the north. The dome is nested against the front of the Cordillera, whose first structures appear along its western side. From its crest the cover rocks have been stripped, revealing the Paleozoic, the pre-Permian part of which has been worn down to a plain termed the Marathon Basin -- a basin in a physiographic rather than a structural sense. On the infacing escarpments around the edges of the basin a great angular unconformity is visible between the cover rocks and the Paleozoic (Fig. 3), and in the basin itself little "Appalachian ridges" formed of the stronger Paleozoic components trend northeastward (Fig. 4).

On the map (Fig. 2) observe the complex areal patterns of the pre-Permian rocks in the Marathon Basin, reflecting strong and complex folding and faulting. Pre-Carboniferous strata emerge in two large anticlinoria in the western part of the basin (the Marathon and Dagger Flat anticlinoria) with Cambrian at the base, followed by Ordovician and Devonian. The two anticlinoria are surrounded by more

extensive Carboniferous, mainly a flysch sequence -- Mississippian below, Pennsylvanian above, Upper Pennsylvanian occurs only in the north at the foot of the Glass Mountains, in one area to the northeast, another to the northwest, whose significance we will discuss later.³

³Consistent nomenclature is desirable, especially in this condensed account. I will accordingly use terms as follows: The *Ouachita orogenic belt* is the whole belt of similar rocks and structures that has a known extent from Mississippi to western Texas. The *Ouachita Mountains* are the exposure of the orogenic belt in Arkansas and Oklahoma. The *Marathon Region* is a structural dome in western Texas which includes an exposure of the orogenic belt in the *Marathon Basin* and the Permian rocks in the *Glass Mountains*, as well as a fringe of surrounding tilted cover rocks. The rocks of the orogenic belt in the Marathon Basin are the *Marathon geosynclinal rocks*. The area north of the orogenic belt is the *craton*, whose Paleozoic rocks are a *cratonic sequence*. During late Paleozoic orogeny the craton next to the orogenic belt was a *fore-deep* that received thick deposits; a small part of the foredeep is exposed in the extreme northwestern part of the Marathon Basin.



Figure 4. -- Miniature "Appalachian ridges" of resistant Caballos Novaculite projecting from the lowlands of the Marathon Basin. View westward from summit of Horse Mountain, 14 miles south-southeast of Marathon. Creaceous rocks on western rim of Marathon Basin on skyline.

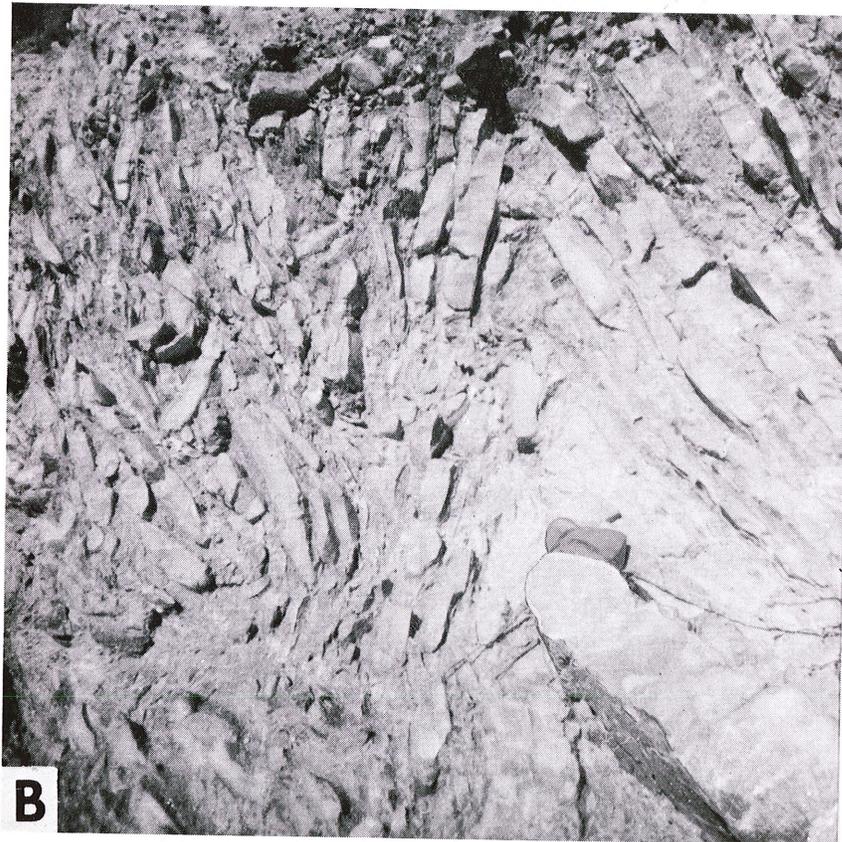
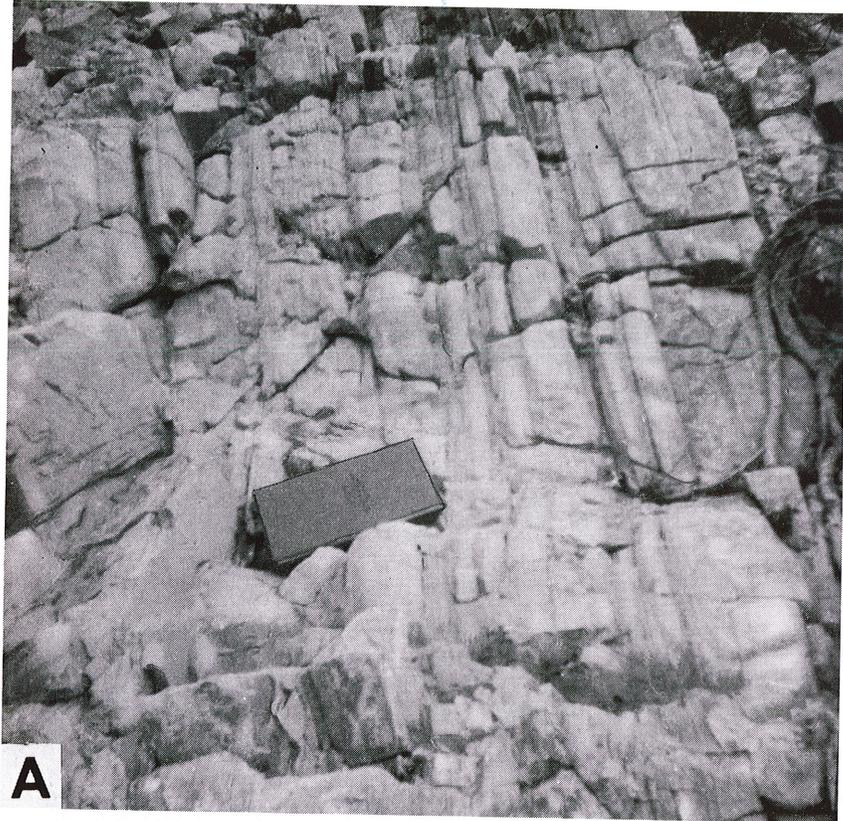


Figure 5, A and B. -- Upper part of Marathon Limestone (Lower Ordovician); Alsate Creek, 6 miles southwest of Marathon. Flaggy and shaly limestone with a fauna mainly of graptolites. The rock is incompetent, hence is contorted in many places, as in B.



Figure 6. -- Upper part of Fort Peña Formation; Alsate Creek, 6 miles southwest of Marathon. Thin-bedded cherty shales and limestones.



Figure 7. -- Upper part of Maravillas Chert; Rock House Gap, 13 miles southwest of Marathon. Black layered chert, with a few thin limestone intercalations.



Figure 8.-- Upper part of pre-Carboniferous sequence; view south from East Bourland Mountain toward Simpson Springs Mountain, 8 miles south of Marathon. In lower right is Woods Hollow Shale (W), Middle Ordovician; followed by Maravillas Chert (M), Upper Ordovician; and by Caballos Novaculite (C), Devonian and lower Mississippian, with two strong novaculite ledges. Tesnus Formation (T), upper Mississippian, forms the synclinal valley beyond, between the two mountains.

PRE-CARBONIFEROUS

The lowest exposed unit of the sequence is the Dagger Flat Sandstone, with Upper Cambrian fossils (Wilson, 1954, p. 254), hence older than the lowest visible rocks in the Ouachita Mountains, of earliest Ordovician age.

Next above it is the Lower Ordovician Marathon Limestone (Fig. 5) -- actually not a very good limestone, as it is shaly and flaggy, with many partings of shale and sandstone, and containing few fossils other than graptolites (Berry, 1960, p. 8-20; Young, 1970, p. 2304-2312). Being weak and incompetent, the limestone is frequently contorted (Fig. 5B). Although it differs somewhat from the Lower Ordovician Mazarn Shale of the Ouachita Mountains, both are of graptolite shale facies and are of broadly the same age (see Table 1).

Higher up is the Middle Ordovician Fort Peña Formation, very cherty throughout, both in its limy and shaly layers (Fig. 6), again with few fossils other than graptolites. The Fort Peña is followed by the Woods Hollow Shale (McBride, 1969), broadly equivalent to the Womble Shale of the Ouachita Mountains, and this by the Upper Ordovician Maravillas Chert (Fig. 7). The Maravillas contains some interbedded limestone below, but passes up into a nearly continuous body of bedded chert (McBride, 1970a); it is

lithically nearly identical with the Bigfork Chert of the Ouachita Mountains, and is broadly of the same age.

All the upper part of the pre-Carboniferous sequence can be viewed in single exposures on many of the ridges of the Marathon Basin, as it is rather thin. Figure 8 shows such an exposure on the flank of a minor anticlinal ridge near the center of the basin. At the base is the slope-making Woods Hollow Shale, followed by dark ledges of the Maravillas Chert. Above this is the Caballos Novaculite, with two prominent novaculite ledges, separated and overlain by weaker beds of chert; it is obviously equivalent to the Arkansas Novaculite of the Ouachita Mountains, and like it is of Devonian and early Mississippian age.⁴ The Tesnus Formation, or higher Mississippian, lies in the synclinal valley beyond.

⁴The Caballos Novaculite is strikingly like the Arkansas Novaculite, even in its subdivision into two strong novaculite units, with associated bedded chert members. Available paleontological data seemingly conflict with this comparison; conodonts from the upper chert member at Marathon are ascribed to the Upper Devonian (Graves, 1952), whereas conodonts indicate that the Devonian-Mississippian boundary in Arkansas is within the middle chert member (Hass, 1951). Nevertheless, I prefer to accept the lithologic comparison until more comprehensive paleontological data are available, especially for the Marathon sequence.

TABLE 1. -- APPROXIMATE CORRELATION OF PRE-CARBONIFEROUS ROCKS IN
MARATHON REGION WITH THOSE OF THE OUACHITA MOUNTAINS.

(Asterisks* mark the occurrence of exotic bouldery debris)

SYSTEM AND SERIES	MARATHON REGION	OUACHITA MOUNTAINS
Lower Mississippian	Caballos Novaculite	Arkansas Novaculite
Devonian		
Silurian	H I A T U S	Missouri Mountain Shale Blaylock Sandstone
Upper Ordovician	Maravillas Chert*	Polk Creek Shale Bigfork Chert
Middle Ordovician	Woods Hollow Shale*	Womble Shale
	Fort Peña Formation	Blakely Sandstone*
	Alsate Shale	Mazarn Shale
Lower Ordovician	Marathon Limestone*	Crystal Mountain Sandstone Collier Shale
Upper Cambrian	Dagger Flat Sandstone	NOT EXPOSED

The sequence shown in Figure 8 is no more than 600 feet thick, and clearly very condensed. Note the apparent absence of any Silurian -- nothing like the Blaylock Sandstone or Missouri Mountain Shale of the Ouachita Mountains. McBride and his colleagues have proposed that the lower part of the Caballos is Silurian (McBride and Thomson, 1970, p. 37; McBride, 1970a, p. 1723), but this seems to be wishful thinking and lacks any tangible evidence. Certainly there was no unconformity between the Maravillas and Caballos in the usual sense -- no emergence, erosion, or truncation of beds. But I prefer to consider the interval as a long period of failure of deposition. (Actually, a nearly identical condition occurs in the northern belts of the Ouachita Mountains, where the Blaylock is missing entirely, and the Missouri Mountain is virtually so). Other failures of deposition probably occurred during the Ordovician; a continuous sequence has been claimed there also

(Berry, 1960, p. 1), but graptolite zones known elsewhere in North America are missing (Riva, 1972, p. 6-7).

Ideas (including my own) have fluctuated as to the environment in which these pre-Carboniferous deposits accumulated. Did they form in shallow water, or even on tidal flats? Or are they basinal deposits, formed in water of much depth? These questions are not yet fully resolved, but my present inclination is to think that they formed in water of much depth. Resolution of the problem depends on sedimentological evidence and argument which I will leave to the specialists, but there is one aspect that I will discuss here -- the occurrence of exotic bouldery debris in several of the Ordovician formations.

The lowest occurrence of debris is in the Lower Ordovician Marathon Limestone, principally in what I originally



Figure 9, A and B. -- Monument Spring Dolomite Member of Marathon Limestone; near Fort Peña Colorado and Alsate Creek, 5 to 6 miles southwest of Marathon. Shelf carbonate with shelly fossils, lying in the prevailing graptolite-bearing shaly strata of the main body of the formation. A-- a large slab; B-- a smaller block.



Figure 10, A and B. -- Large boulders in lower part of Maravillas Chert; Monument Spring and Rock House Gap, 12 and 13 miles southwest of Marathon. Exotic clasts of shelf carbonate and sandy carbonate, embedded in pebbly limestone.

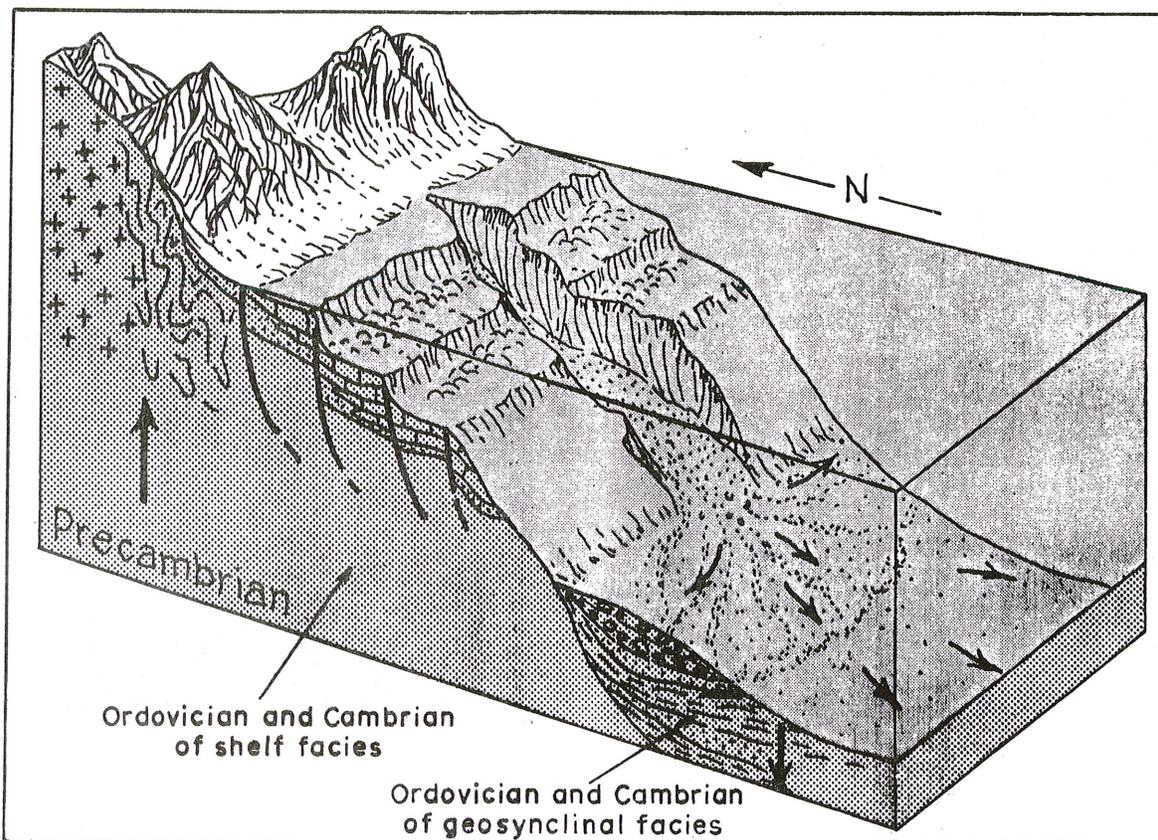


Figure 11. -- Block diagram illustrating probable origin of exotic bouldery debris in Ordovician geosynclinal deposits along the front of the Northern Appalachians in Quebec. Modified from Hubert and others (1970, Fig. 12, p. 124).

called the Monument Spring Dolomite Member.⁵ The member is a chain of closely spaced blocks and slabs of shelf carbonate similar to the Ellenburger (=El Paso) Limestone of the cratonic sequence, containing varied shelly fossils of Beekmantown type (Berry, 1960, p. 17), enclosed in the usual Marathon graptolitic shaly limestones (Fig. 9). The member is prominent in the northwestern or Marathon anticlinorium, and fades out across the southeastern or Dagger Flat anticlinorium. When first discovered, it was thought to be an autochthonous carbonate bed that had been broken during deformation, but the material is consistently fragmental, and is mingled with clasts of other lithologies, including sandstone of probable Cambrian age and volcanic rocks (Young, 1970, p. 2306-2207). The fragments are allochthonous slide blocks that have moved into a foreign environment, where they lie in strata of another facies, although of about the same age.

Exotic debris also occurs in the Woods Hollow Shale, and consists of rounded blocks as much as 10 feet in diameter of shelf carbonate, lying in graptolite shale of Middle Ordovician age. These blocks, again, contain shelly fossils, but these are all much older than their matrix, being late

Cambrian (Franconian) to Early Ordovician (Tremadocian) (Wilson, 1954a, 258-263) -- zones which are poorly represented, if at all, in nearby autochthonous cratonic sequences. Similar boulders occur in the lower part of the Maravillas Chert in the northwestern part of the Marathon Basin (Fig. 10); no fossils have been collected from them, but many of them are lithically like the Ellenburger (=El Paso) Limestone and Simpson Group of the cratonic sequence (Wilson, 1954b, p. 2469).

This exotic bouldery debris resembles that which occurs in Ordovician geosynclinal rocks all along the front of the Northern Appalachians in Canada and the United States (Zen, 1972, p. 11-13), of which there are many fine examples near Quebec City (Bailey, Collet, and Field, 1928; Osborne, 1956, p. 183-188). Figure 11 shows an interpretation of their probable origin (Hubert and others, 1970, Fig. 12, p. 124) -- a cratonic area of crystalline rocks is bordered by an unstable carbonate shelf, from which blocks slumped and slid from time to time into the adjacent geosynclinal trough, where they were embedded in sediments of a different facies. A similar explanation applies very well to the exotic bouldery debris in the Marathon Ordovician; if so, a shallow-water origin for the shelf carbonates and a deep-water origin for the geosynclinal deposits is required.

The whole thickness of exposed pre-Carboniferous

⁵Young (1970, p. 2306) assigns some of the material, probably originally mapped by me as Monument Spring, to a lower stratigraphic level (his "boulder-bed no. 1").

rocks in the Marathon geosynclinal sequence is no more than 2,500 feet, whereas the equivalent cratonic rocks to the north (beneath the West Texas Permian basin) are more than twice that in many places (Barton, 1945, p. 1339-1343). I interpret this as caused by differences in the availability of sediment, rather than differences in the amount of subsidence (as inferred by Barton). The cratonic sequence formed in relatively shallow water on carbonate banks; the Marathon sequence formed in deeper water in a leptogeosyncline or starved basin. The boundary between the two sequences is no longer visible, being either concealed by younger rocks or by the frontal thrust sheets of the orogenic belt. During the Ordovician one can infer that the boundary was an abrupt shelf break. Less can be inferred regarding the nature of the boundary during Caballos deposition; however, an approach to some sort of depositional boundary is suggested in the northwestern part of the Marathon Basin, where the novaculite layers nearly disappear and the formation is dominantly bedded chert.

CARBONIFEROUS

Mississippian-lower Pennsylvanian flysch sequence -

In both the Marathon Region and the Ouachita Mountains there is a great contrast between the novaculite formations and those above, recording an abrupt change in style of sedimentation -- from slow deposition of non-clastic siliceous sediments to rapid deposition of clastic flysch (McBride, 1970b). This is an early Mississippian event; fossils from some distance below and above are Kinderhookian and Meramecian, but paleontological control near the contact is sparse; whether there was a pause in deposition between the two events is uncertain. The succeeding flysch accumulated to great thickness; in the eastern part of the Marathon Basin the known maximum approaches 15,000 feet (right-hand column, Fig. 12) -- a healthy enough figure, but small compared with the known maximum in the Ouachita Mountains.

In the Marathon Basin the components are the Tesnus Formation, late Mississippian and early Pennsylvanian, approximately equivalent to the Stanley and Jackfork (see Table 2); the Dimple Limestone, approximately Morrowan; and the Haymond Formation, approximately Atokan. Prominent boulder beds occur in the Haymond, or at a level somewhat younger than the Johns Valley boulder beds of

TABLE 2. -- APPROXIMATE CORRELATION OF CARBONIFEROUS STRATA IN MARATHON REGION WITH THOSE OF THE OUACHITA MOUNTAINS.

(Asterisks* indicate the occurrence of exotic bouldery debris)

SERIES	MARATHON REGION	OUACHITA MOUNTAINS
Wolfcampian	Neal Ranch Formation	A B S E N T
Virgilian	.	.
Missourian	Gaptank Formation	Hartshorne Sandstone and higher Formations in Arkoma Basin.
Des Moinesian	.	.
Atokan	Haymond Formation *	Atoka Formation
Morrowan	Dimple Limestone	Johns Valley Shale *, Wapanucka Limestone, and related units
		Jackfork Sandstone
Chesterian	Tesnus Formation	Stanley Shale
Meramecian	.	.
Osageian	P R E S E N T ?	P R E S E N T ?
Kinderhookian	.	.
Devonian	Caballos Novaculite	Arkansas Novaculite

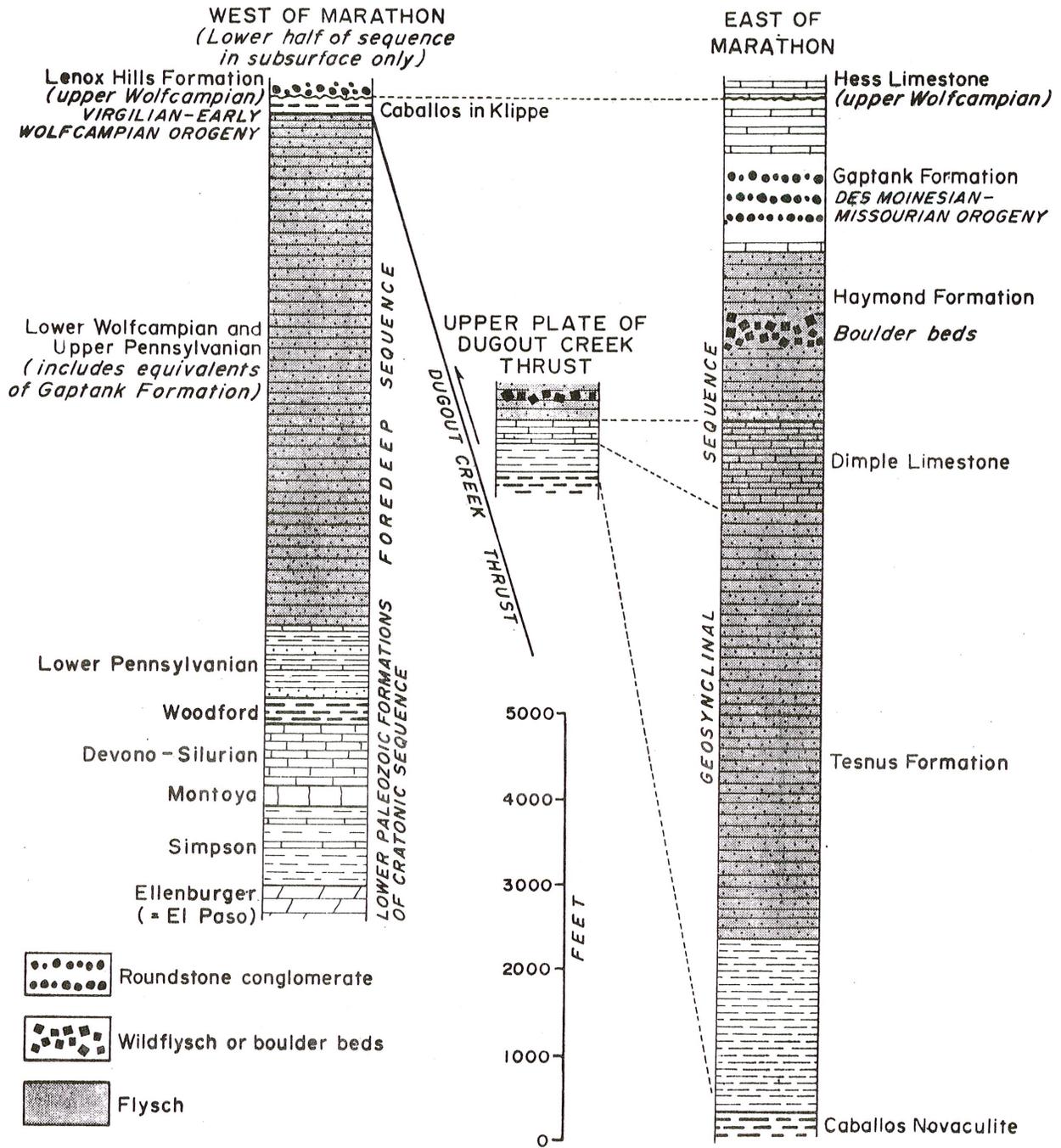


Figure 12. -- Columnar sections of Carboniferous rocks exposed in different parts of the Marathon Basin. *Right column*-- Sequence east and southeast of Marathon, where maximum exposed thickness is displayed. *Middle column*-- Thin sequence in upper plate of Dugout Creek thrust sheet west of Marathon. *Left column*-- Sequence in lower plate of Dugout Creek thrust sheet west of Marathon. In left column, thicknesses are approximate, due to complex structure; lower part of sequence is based on record of the Slick-Urschel, No. 1 Mary Decie-Sinclair well.

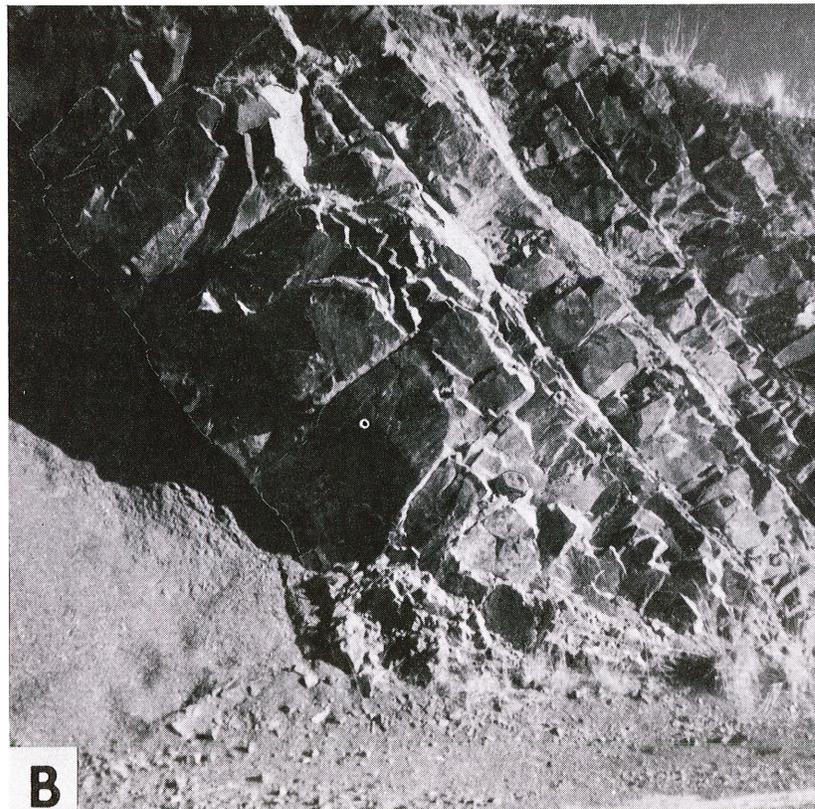


Figure 13, A and B. -- Upper part of Tesnus Formation, showing characteristic thick units of sandstone and shale. A-- On San Francisco Creek, 16 miles southeast of Marathon. B-- Cut on U. S. Highway 90 east of Lemons Gap, 19 miles east of Marathon, shale unit in lower left.



Figure 14. -- Dimple Limestone east of Haymond, 15 miles east-southeast of Marathon. Inverted beds of resedimented carbonate, with convolute layering emphasized by chert bands.



Figure 15. -- Lower part of Haymond Formation; cut on U. S. Highway 90 east of Lemons Gap, 18 miles east of Marathon. Shows characteristic thin alternation of sandstone and shale; bending of strata at top of cut results from soil creep.

the Ouachita Mountains. The sequence thins dramatically in the northwest part of the basin (middle column, Fig. 12), indicating an approach to the edge of the geosynclinal trough.

Parts of the Tesnus Formation⁶ are lithically much like the Stanley -- dark shale below, interbedded sandstone and shale above. The thicker sandstone beds resemble the "dirtier" sandstone beds in the Jackfork, but the Tesnus contains only a few quartz sandstone beds like those which are common in the Jackfork. The sandstone and shale

layers are typically grouped in thick units (Fig. 13), which is a good field guide for distinguishing the Tesnus from the other flysch units.

The Dimple Limestone marks a break in the siliciclastic flysch sedimentation, but much of its carbonate is resedimented, and is a flysch or turbidite like the rest (Thomson and Thomasson, 1969). A characteristic feature of many layers in the eastern part of the basin is convolute bedding, caused by the passage of turbid flows (Fig. 14).⁷ However, the Dimple Limestone in its extreme northwestern exposures is of different aspect, and is interpreted as a shallow-water shelf facies (Thomson and Thomasson, 1969, p. 58-60).

The Haymond Formation is again a sequence of interbedded sandstone and shale, but large parts of it differ from the Tesnus in that the alternating layers are thinly spaced (Fig. 15). Thicker sandstone beds of another facies domi-

⁶ Named for Tesnus Station on the Sunset Route of the Southern Pacific railroad in the eastern part of the Marathon Basin. Long after I was in the region I learned that the name "Tesusus" is simply the word "Sunset" spelled backwards. Geologists owe a debt of gratitude to the unknown namer of way points on the railroad, for providing so elegant a name for this distinctive formation at a suitable type locality. We are fortunate that "Tesusus" was not applied to some geologically uninteresting spot, and that the creator of names gave up on poets before he reached here; way points farther east include "Dryden," "Tennyson" and "Longfellow!"

⁷ One of these layers was figured by King (1930, pl. 2A) as "dome-like structures outlined by chert bands," but the layer is evidently inverted.



Figure 16, A and B. -- Boulder beds of Haymond Formation at foot of Housetop Mountain, 19 miles east-southeast of Marathon. A-- General view, looking north. The small knobs in the foreground are giant exotic clasts or slabs, primarily of fossiliferous lower Pennsylvanian shelf limestone; lower part of Haymond Formation and Dimple Limestone in middle distance; scarps of Lower Cretaceous on skyline. B-- Nearer view of one of the limestone clasts, with mudstone matrix of boulder bed in foreground.

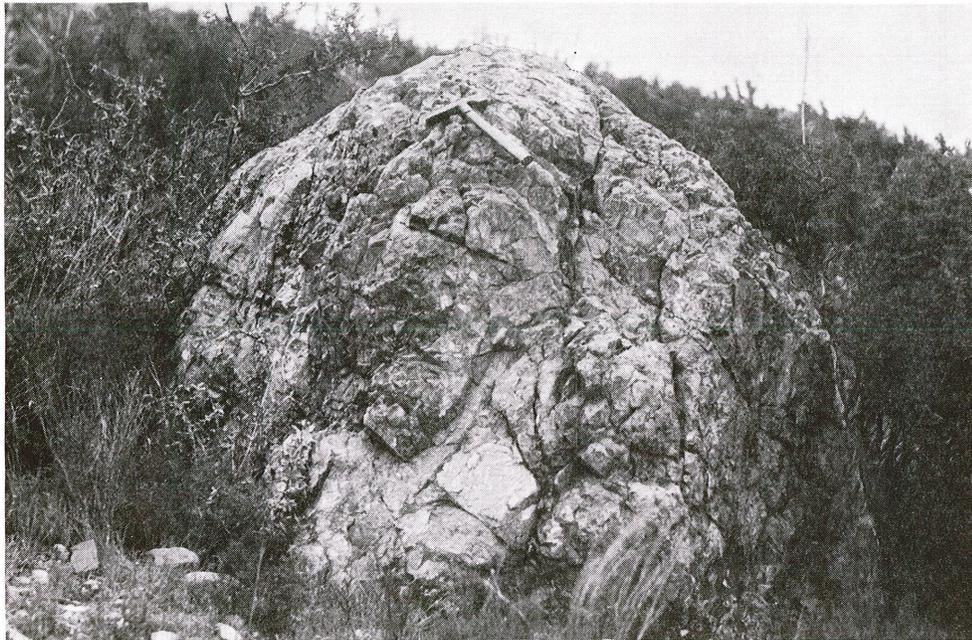


Figure 17. -- Large rounded boulder of brecciated Caballos Novaculite in Haymond boulder bed at a nearby locality.

nate the upper half of the formation in the northeastern part of the Marathon Basin southeast of Gap Tank (King, 1930, p. 41; McBride, 1966, p. 22; Flores, 1972). Flores (1972, p. 3418-3424) proposes that these are delta-front and delta-plain deposits; this is possible, as by late Haymond time the flysch trough had been nearly filled, especially along its northern margin.

Of all the rock types in the Haymond Formation, the boulder beds (wildflysch) are the most famous. Their main occurrence is in a belt 8 miles long at the foot of Housetop Mountain in the eastern part of the Marathon Basin (Fig. 16), where they occur through a sequence as much as 900 feet thick. A thinner, somewhat different boulder bed at about the same stratigraphic level crops out for 4 miles in the northeastern part of the Marathon Basin, southeast of Gap Tank. Another, more problematical occurrence of bouldery debris is in the northwestern part of the basin, in the Payne Hills on the upper plate of the Dugout Creek thrust; here, blocks of chert and carbonate rocks are scattered over the surface of the Tesnus and Dimple Formations (King, 1937, Pl. 16, in which they were shown as tectonic klippen). McBride (1973) interprets them as being embedded in these formations, but I question whether they might not be residuals from a formerly overlying Haymond Formation, now eroded away. In other parts of the Marathon Basin the Haymond boulder beds have been lost by erosion or concealed by Cretaceous cover rocks, so that very little remains of their original extent -- as compared with the occurrence of the Johns Valley boulder beds in the Ouachita Mountains for 100 miles along the strike and 25 miles across it.

The most prominent clasts in the boulder beds of the Housetop Mountain area are huge blocks or slabs of carbon-

ate rock, some more than 100 feet across. One of these is recognizably from the Dimple Limestone, but the remainder are a shelf carbonate containing shelly fossils of early Pennsylvanian age (Fig. 16 and Table 3), or approximately equivalent to the Dimple but of different facies; nothing like them is known in place in the Marathon Basin or nearby regions. Numerous smaller boulders are identifiable with units at lower levels of the geosynclinal sequence itself -- from the Tesnus Formation, the Caballos Novaculite (Fig. 17), and the Maravillas Chert. In addition, there are fairly

TABLE 3. -- FORMATIONS REPRESENTED IN LARGE HAYMOND BOULDERS OF HOusetop MOUNTAIN AREA (as recorded by King, 1937, Pl. 10).

FORMATION	DIAMETER OF BOULDERS			TOTAL
	3-10 ft..	10-50 ft..	50+ ft..	
Fossiliferous Pennsylvanian limestone	24	24	7	55
Dimple Limestone	1	1	1	3
Tesnus Formation	5	6	---	11
Caballos Novaculite	73	15	---	88
Maravillas Chert	1	---	---	1

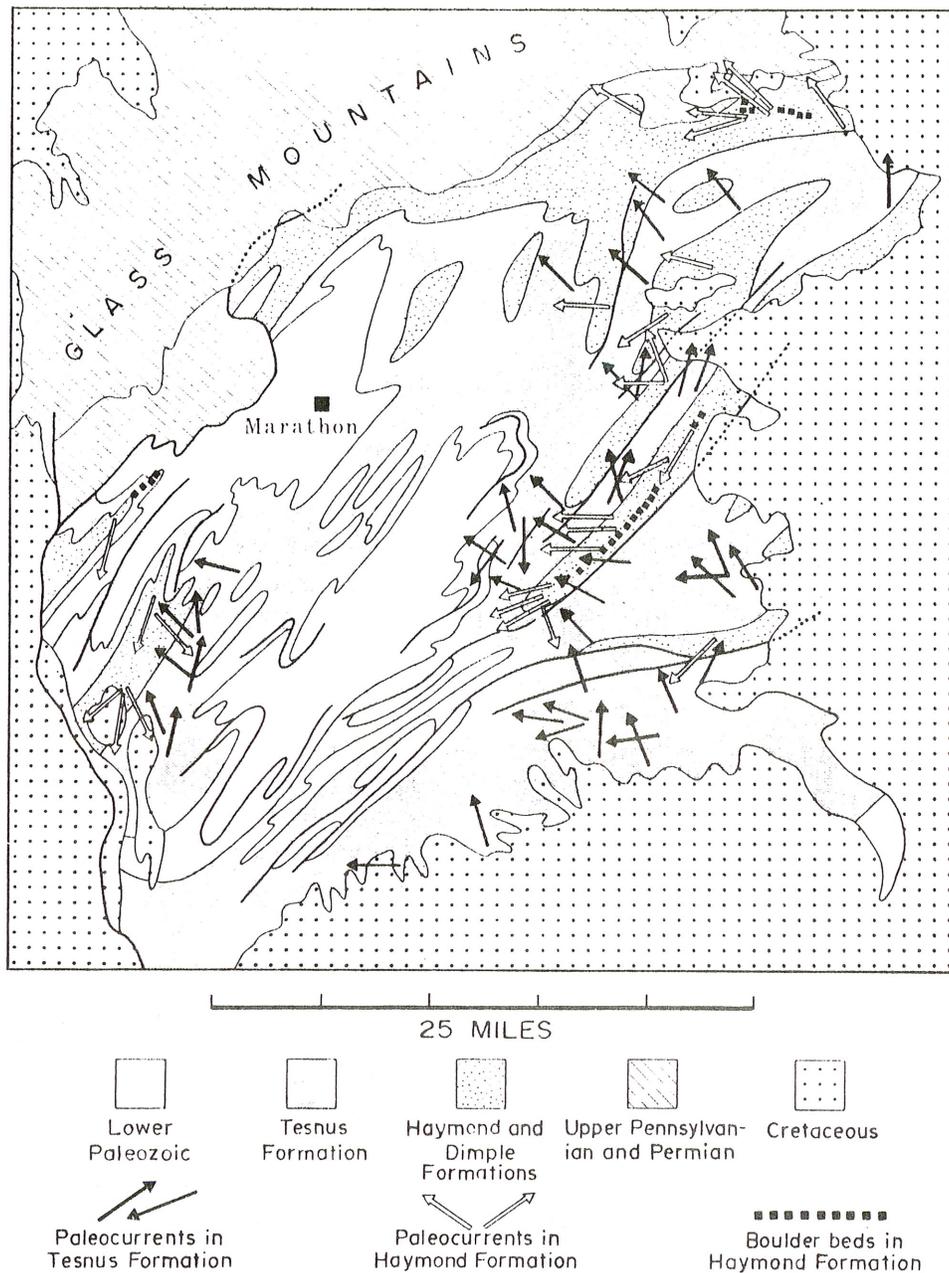


Figure 18. -- Map showing Carboniferous rocks of Marathon Basin, and paleocurrents in Tesnus and Haymond Formations, as observed by Johnson (1962, p. 782), and by McBride (1966, p. 55). Also shown are known occurrences of boulder beds of Haymond Formation.

common well-rounded pebbles and cobbles of granite gneiss and rhyolite that have yielded middle Paleozoic Rb/Sr dates of 370 to 410 m. y. (Denison and others, 1969).

What was the provenance of the clastic detritus in the flysch, and specifically of the clasts in the wildflysch? The results of paleocurrent studies of the Tesnus and Haymond Formations are shown in Figure 18. These studies indicate dominant directions of current flow toward the northwest, with a minor tendency toward southwestward turning down the axis of the trough (Johnson, 1962, p. 789-791; McBride, 1966, p. 54-56). So far as the main body of the flysch goes, this seems to be decisive for a southeastward source. Moreover, the mineralogy of the sandstone indicates derivation from crystalline and metamorphic rocks, but these could not have been exposed in the craton to the north, as it was blanketed by lower Paleozoic shelf sediments.

The paleocurrents are also seemingly conclusive indicators of the source of the small to large clasts in the Haymond boulder beds, but are they? The cobbles of gneiss and rhyolite most likely came from crystalline lands southeast of the geosyncline, and the boulders of geosynclinal rocks from structural ridges within the geosyncline itself. But the giant blocks and slabs of carbonate rocks are of shelf or northwestern provenance; even the slab of Dimple Limestone is lithically most like its shelf facies in the northwestern part of the Marathon Basin. How could these shelf rocks have traveled across the floor of the geosyncline to their present positions, against the paleocurrents? There is no ready answer to the question from the scanty available exposures of the boulder bed. (The provenance of the Haymond boulders is thus more complex than that of the Johns Valley boulders of the Ouachita Mountains, all of which were derived from the edge of the craton to the north.)



Figure 19, A and B. -- First (lowest) conglomerate bed of Gaptank Formation; south of Gap Tank and 20 miles northeast of Marathon. Round-stone clasts are principally from Dimple Limestone, but some are from *Chaetetes* Limestone Member at base of Gaptank Formation.

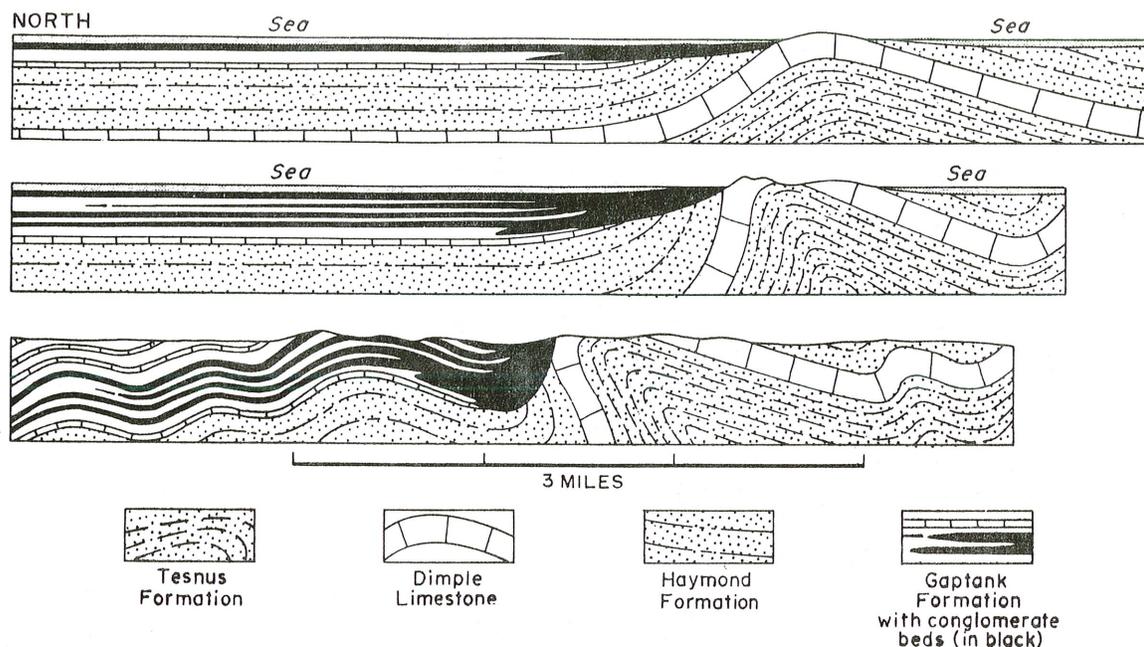


Figure 20. -- Sections showing probable sequential development of conglomerate deposits of Gaptank Formation during the Des Moinesian-Missourian orogeny. Section extends from the approximate present site of the Dimple Hills (right) to that of Gap Tank (left). After King (1930, Fig. 39, p. 112).

Upper Pennsylvanian. -- In sequence above the Haymond Formation near Gap Tank⁸ in the northeastern part of the Marathon Basin (Fig. 2) is the upper Pennsylvanian Gaptank Formation, of Des Moinesian to Virgilian age, a different sort of deposit from the flysch below -- interlensing limestones and shales with layers of roundstone conglomerate in the lower part, with a rather modest thickness of about 1,800 feet. The roundstone conglomerates differ from the earlier boulder beds of the Haymond Formation; they are formed of stream-rounded or beach-rounded cobbles, largely Dimple Limestone, but including significant numbers from the basal, or *Chaetetes* Limestone Member of the Gaptank itself (Fig. 19). The layers wedge out northward, even within their limited area of exposure, and were undoubtedly derived from areas a few miles to the south that were in process of deformation (Fig. 20).⁹

The Gaptank Formation is only broadly folded, in contrast to the steeply tilted older Carboniferous formations

⁸The name Gap Tank was unfortunately omitted from the recently published (1968) Marathon Gap 7 1/2-minute topographic map. The tank is the artificial pond about a mile southwest of Marathon Gap, east of the Marathon-Fort Stockton highway near elevation point 4485.

⁹I cannot accept Ross's proposals (1967, p. 371-372) to redefine the Gaptank Formation to exclude the *Chaetetes* Limestone, and to place a major structural unconformity beneath the first conglomerate layer above it. The *Chaetetes* Limestone would be an unnatural appendage to the overwhelmingly siliciclastic Haymond Formation, and formed under quite different conditions. Apparent unconformities are a "dime-a-dozen" in the uppermost Pennsylvanian and lowermost Permian of the Marathon Region, in which lenticular deposition, channeling, soft-sediment deformation, and disharmonic folding are the rule; the seeker after an unconformity can thus provide himself with one at almost any level that suits his fancy. Few of the apparent unconformities in this part of the sequence are genuine, and the only unconformity of regional extent is that beneath the upper Wolfcampian formations (Lenox Hills Formation and Hess Limestone).

to the south, but it is followed on the north with slight angular discordance by the Hess Limestone of late Wolfcampian age. The Gaptank is synorogenic to post-orogenic in its type area, and was laid down in shoal water after the preceding flysch trough had been filled up, and had been closed up by deformation (McBride, 1964, p. 44; Ross, 1967, p. 379-382). One can reasonably infer that early Gaptank time was one of general deformation of the geosynclinal rocks of the Marathon Basin, during which most of the complex structures visible there were created -- a major orogeny of Des Moinesian to Missourian age.

So far as the northeastern part of the Marathon Region goes, this is the virtual end of the tectonic story, but a further chapter is expressed in the upper Pennsylvanian rocks of the northwestern part of the Marathon Basin, west of the town of Marathon (Fig. 2).

The relation of the upper Pennsylvanian rocks in the northwestern part of the Marathon Basin to those of the type Gaptank Formation has long puzzled geologists. They contain fossiliferous layers that are of about the same age as different parts of the Gaptank (Des Moinesian to Virgilian) (King, 1937, p. 80-82), and in places fusulinids of probable early Wolfcampian age have been reported (Adams and Frenzel, 1952, p. 25-28).¹⁰ The rocks are more clastic and

¹⁰The geologist whose discussion involves the Wolfcampian Series is plagued by semantic problems. The original Wolfcamp Formation of Udden (1917, p. 41-42) and Böse (1917, p. 15-17), which became the basis for the Wolfcamp Series (Adams and others, 1939, p. 1674-1675), was the 600 feet or so of strata exposed in the Wolf Camp Hills in the eastern Glass Mountains, but the series has undergone confusing redefinitions. Most geologists now equate it with the zone of *Pseudoschwagerina* and related fusulinids which extends both higher and lower, so that various thicknesses of strata have been added to the base and top. For redefinitions of the outcrop sequence in the Glass Mountains, see Ross (1963, p. 20-30) and Cooper and Grant (1972, p. 30-38). In addition, great thicknesses of strata known only in subsurface and probably older than the original Wolfcamp have been ascribed to the Wolfcampian (Oriol and others, 1967, p. 30-32).

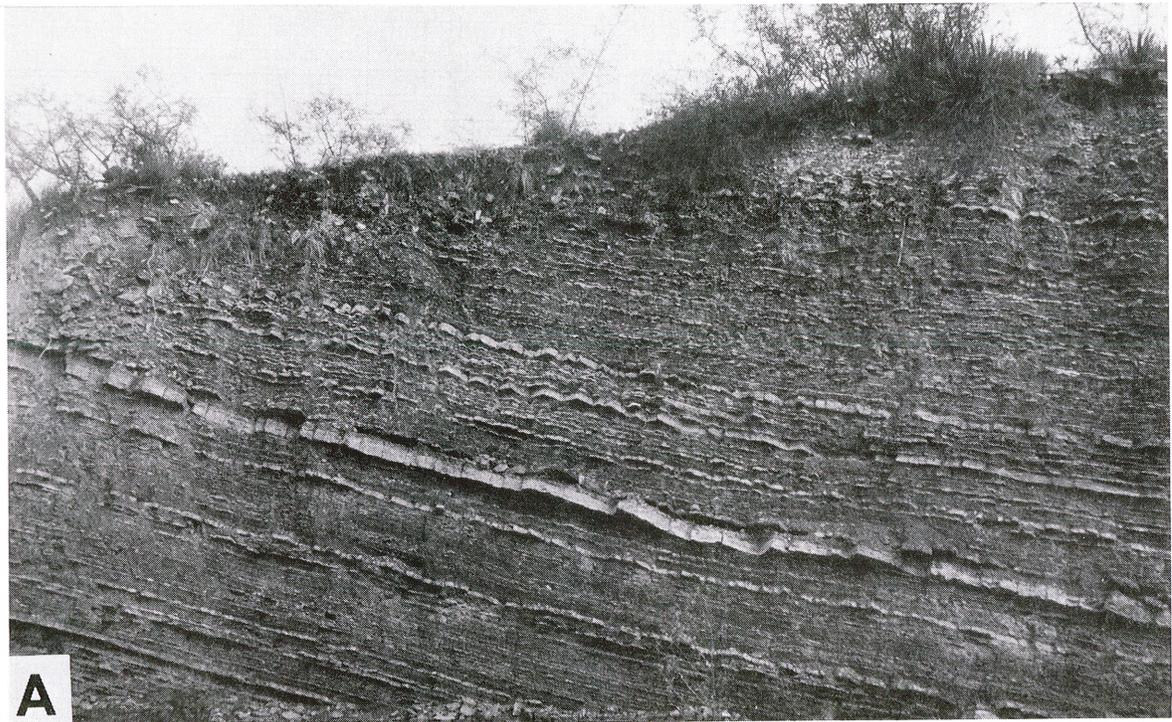


Figure 21, A and B

(See descriptive text on succeeding page below Figure 21, C)



Figure 21, A, B, and C, -- Upper Pennsylvanian rocks in lower plate of Dugout Creek thrust west of Marathon. A-- Thin-bedded flysch in lower part of sequence (approximately of late Haymond age); cut on Dugout Creek south of Dugout Mountain, 8 miles west of Marathon. B-- Resedimented limestone (approximately of early Gaptank age), with crumbly spherical weathering, as in an arkose; south of Dugout Mountain, 10 miles west of Marathon. Dip is toward observer, so that bedding structures are not visible. C-- interbedded layers of conglomeratic limestone, sandstone, and shale (approximately of middle Gaptank age); cut on Southern Pacific Railroad at Milepost 580, 4 miles west of Marathon.

probably much thicker than those in the Gap Tank area, and have been strongly and complexly deformed; it is not surprising that they were identified as Tesnus Formation during the first geological work in the region (Baker and Bowman, 1917, p. 103-105). They are not in sequence with the older rocks of the Marathon Region, and lie beneath the great frontal Dugout Creek thrust. Moreover, deep drilling in recent decades has demonstrated that they do not lie on the Marathon geosynclinal sequence, but on thin lower Pennsylvanian, and this on lower Paleozoic formations of the cratonic sequence (left-hand column of Fig. 12). More than 6,500 feet of upper Pennsylvanian and Wolfcampian rocks overlie the lower Pennsylvanian in the Slick-Urschel Oil Co., No. 1 Mary Decie-Sinclair well at the foot of the Glass Mountains escarpment northwest of Marathon (Flawn and others, 1961, p. 237-238), but their structure is so complex that this gives little idea of their original thickness.

Most or all of these upper Pennsylvanian and lower Wolfcampian rocks are of flysch facies, like the generally older flysch of the Marathon geosynclinal sequence to the southeast. The lowest beds exposed, approximately of upper Haymond age, are demonstrably flysch (Fig. 21 A). Younger rocks include several units of resedimented limestone with prominent graded bedding (Fig. 21B), and

alternations of shale, sandstone, sandy limestone, and and gravelly chert conglomerate (Fig. 21 C).

The area of upper Pennsylvanian-lower Wolfcampian rocks in the northwestern Marathon Basin is an exposed part of the Val Verde basin (Hall, 1956, p. 2252-2254), otherwise known only from drilling, which extends along the front of the Marathon salient of the Ouachita orogenic belt for at least 150 miles, from Brewster County eastward into Val Verde County (Fig. 22). The basin is notable for its inordinate thicknesses (more than 13,000 feet) of fine-grained clastic rocks (Flawn and others, 1961, p. 137-138); commonly these are ascribed to the Wolfcampian on the basis of fusulinids, but I suspect that they also include upper Pennsylvanian. The deposits are dominantly shaly, and include dark gray shale, sandy shale, and thin-bedded gray sandstone, shown by electrical logs to be interbedded in thin layers; very likely they are flysch. The underlying lower Pennsylvanian and lower Paleozoic, and the overlying Permian are of normal thickness, indicating that this flysch marks a unique sedimentologic and tectonic event -- the creation of a foredeep along the front of the orogenic belt that was filled by the erosion products of a late Virgilian-early Wolfcampian orogeny in the Ouachita belt.

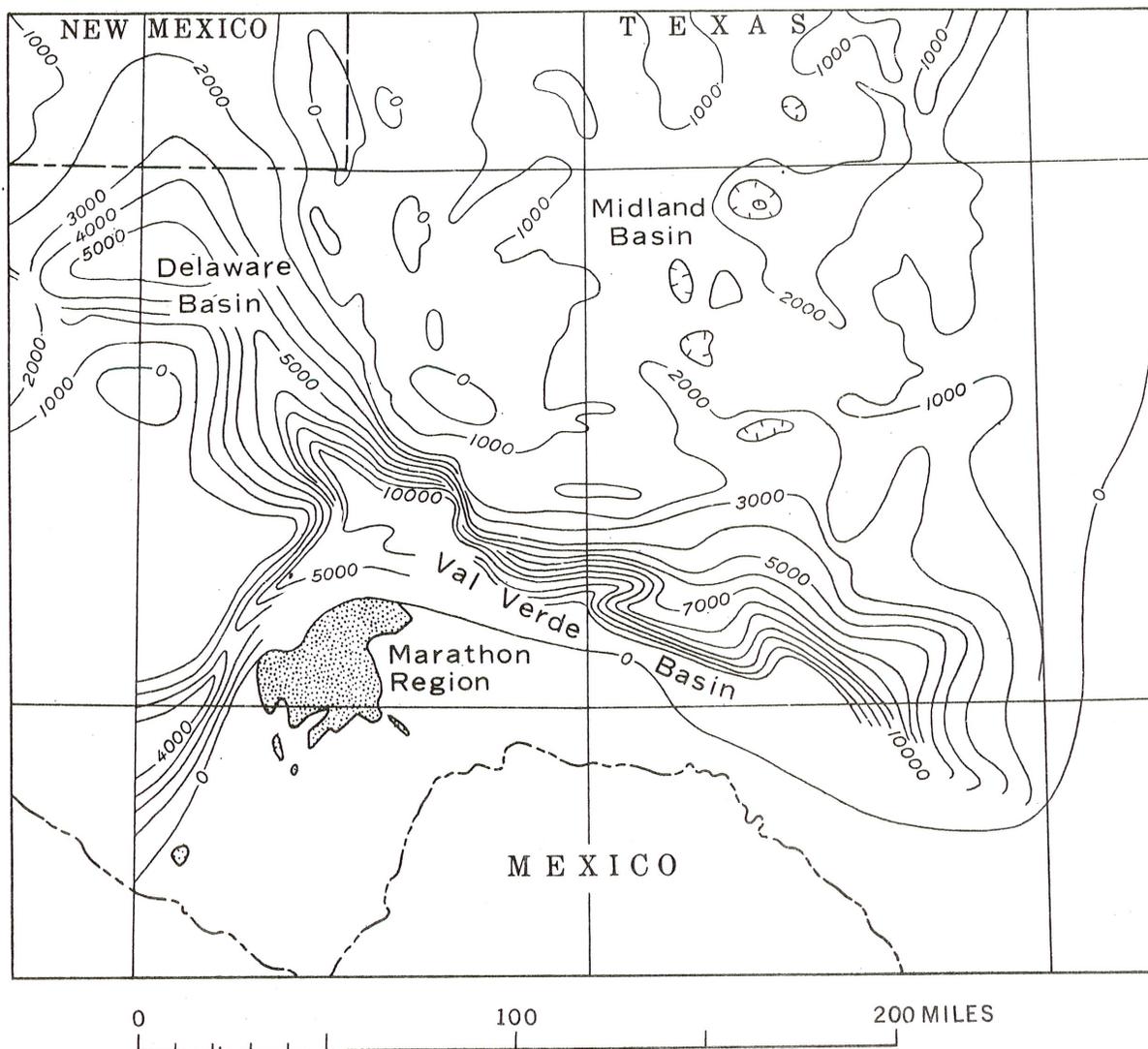


Figure 22. -- Map of part of west Texas, showing thickness of "Interval A," ascribed to the Wolfcampian Series but probably including uppermost Pennsylvanian. The Val Verde basin along the front of the orogenic belt stands out prominently, because of its great thickness of rocks of the interval (as much as 13,000 feet). Isopach interval is 1,000 feet. After Oriol and others (1967, Fig. 12, p. 35).

Dugout Creek thrust. -- Within the Marathon region this orogeny culminated in the emplacement of the Dugout Creek thrust, along which the Marathon geosynclinal sequence, of Cambrian to mid-Pennsylvanian age, was carried northwestward over the upper Pennsylvanian-lower Wolfcampian flysch of the foredeep. Where exposed, the thrust surface is nearly flat, and extends 6 miles across the strike; drill holes indicate that the thrust extends beneath the upper plate for at least 4 miles farther, and dips beneath it at a low angle (upper section, Fig. 23). In the Gulf Oil Corp., No. 1 D. S. C. Combs a mile south of Marathon it was penetrated at a depth of about 6,000 feet (Flawn and others, 1961, p. 234-235).¹¹ Mapping along its exposed

trace indicates that the deformed rocks of the upper plate are truncated at their bases by the fault.

The Dugout Creek thrust differs significantly from the frontal thrusts of the Southern Appalachian Mountains and Ouachita Mountains. These are décollement thrusts that originated by differential slip along weak layers in stratified sequences that had not been materially deformed hitherto, whose leading edges were deflected upward along diagonal shears (lower section, Fig. 23). By contrast, the Dugout Creek thrust broke across rocks already deformed by the Des Moinesian-Missourian orogeny, sheared off the bases of the folds, and carried the whole geosynclinal mass over the edge of its foredeep.

The Dugout Creek thrust is exposed along the strike for about 15 miles, passing at either end beneath Permian and Cretaceous cover rocks. How much farther it extends is conjectural, but thrusts like it may exist elsewhere along the concealed edge of the Val Verde basin. Its projected trace would extend it northeastward beneath the Permian rocks of the Glass Mountains north of Gap Tank. If so,

¹¹The two wells cited here (Gulf, Combs and Slick-Urschel, Decie-Sinclair) are considered sufficient to illustrate the stratigraphy and structure of the northwestern Marathon Basin (see Fig. 23). In addition, half a dozen or so others have been drilled in the area. Records of the earlier ones are given by Flawn and others (1961, p. 233-238) and in various geological society guidebooks. Complete records of the later ones are not yet available, although general information on some of them is known from hearsay. The records of the other wells amplify the picture presented and do not materially modify it.

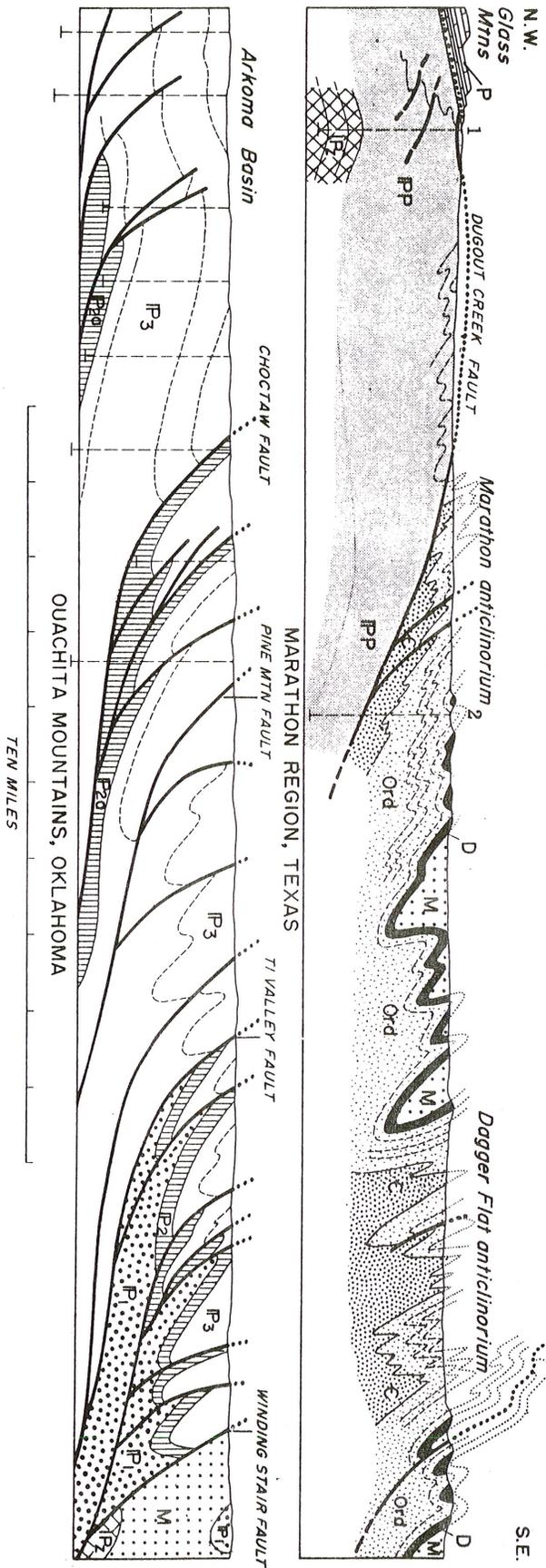


Figure 23. -- Structure sections comparing the frontal thrusts of the Marathon and Ouachita Mountains segments of the Ouachita orogenic belt. Above-- Dugout Creek thrust of Marathon Region (based on King, 1937, section E-E', pl. 21). Below-- Frontal thrusts between the Ouachita Mountains and Arkoma Basin, eastern Oklahoma (based on Berry and Trumbly, 1968, section A-A', p. 102). Well control in both sections is indicated by vertical dashed lines; wells in upper section are: 1 - Slick-Urschel Oil Co., No. 1 Mary Decie-Sinclair, T. D. 9, 741 ft.; 2 - Gulf Oil Corp., No. 1 D. S. C. Combs, T. D. 9,500 ft. Letter symbols as follows: Above--C, Upper Cambrian; Ord, Ordovician; D, Caballos Novaculite; IPz, lower Paleozoic of cratonic sequence; M, Tesnus Formation; PP, upper Pennsylvanian and lower Wolfcampian foredeep deposits; P, upper Wolfcampian (Lenox Hills Formation). Below-- IPz, lower Paleozoic of interior Ouachitas; M, Stanley Shale; IP1, Jackfork Sandstone; IP2, Johns Valley Shale; IP2a, equivalent strata of frontal Ouachitas, including Wapanucka Limestone; IP3, Atoka Formation.

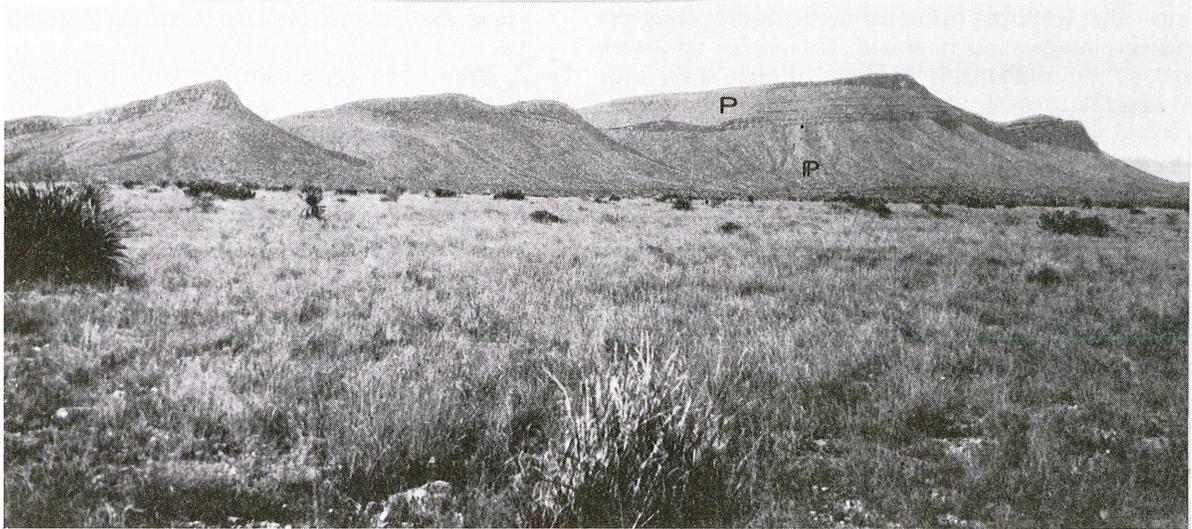


Figure 24. – The south-facing scarp of Dugout Mountain, looking northeastward from about 11 miles west of Marathon. The mountain is capped by upper part of Wolfcampian Series (P), tilted northwest but otherwise undeformed. Lowest ledge is coarse conglomerate of Lenox Hills Formation, which lies with right-angled unconformity on upper Pennsylvanian (IP) that was deformed during the Virgilian-early Wolfcampian orogeny.

the upper Pennsylvanian shelf deposits of that area (Gaptank Formation) which are postorogenic to the Des Moinesian-Missourian orogeny, are nevertheless allochthonous as a result of the Virgilian-early Wolfcampian orogeny, their time of movement being expressed by the seemingly modest unconformity between the Gaptank Formation and the Hess Limestone.

Along the foot of the Glass Mountains in the northwestern part of the Marathon Region the upper Wolfcampian Lenox Hills Formation rests with profound structural unconformity on the Dugout Creek thrust and the deformed rocks associated with it, with hundreds of feet of conglomerate at the base derived from erosion of the complex. Northwest of Marathon, at the site of the Slick-Urschel, Decie-Sinclair well, is a klippe of the thrust sheet 1 1/2 miles long, composed of Caballos, Tesnus, and Dimple Formations, whose northwestern edge is partly overlapped by the conglomerate. Farther southwest, on the face of Dugout Mountain, the Lenox Hills conglomerate lies with right-angled unconformity on vertical to overturned upper Pennsylvanian strata (Fig. 24). The Lenox Hills and higher formations of the Permian are tilted northwestward at a low angle, but are otherwise undeformed. The climax of the terminal orogeny in the northwestern part of the Marathon Basin can therefore be dated closely as occurring within the Wolfcampian epoch as currently defined. This was the last major orogenic event in the Marathon Region during Paleozoic time, and is the end of our tectonic story.

DISCUSSION AND SYNTHESIS

From the observations so far presented, I infer the following sedimentary and tectonic events in the Marathon segment of the Ouachita orogenic belt:

(1) A leptogeosynclinal phase from the Ordovician through the Devonian into the early Mississippian, during

which a small thickness of sediments accumulated slowly in deep water -- graptolite shale and shaly limestones at first, later cherts and other siliceous sediments. Lengthy pauses in deposition occurred, without uplift or erosion, notably during the Silurian, and probably from time to time during the Ordovician.

(2) A flysch phase from late Mississippian to mid-Pennsylvanian time, during which clastic sediments were deposited to great thickness in a deep, rapidly subsiding trough. The phase was initiated by tectonic activity in the backlands of the geosyncline, which were eroded to provide the flood of clastic debris that was shed into the trough; such activity continued during much of the phase. Late in the phase, exaggerated subsidence in the trough produced wildflysch, or boulder beds.

(3) An orogenic phase (Des Moinesian to Missourian) closed up the flysch trough, which had already been nearly filled by sediments. Roundstone conglomerates were shed from structures in the geosynclinal area that were in process of deformation, and postorogenic shallow water deposits were laid down along the northern margin of the orogenic belt.

(4) At about the same time, and continuing into the early Wolfcampian, a new flysch trough formed along the edge of the craton in front of the orogenic belt (Val Verde basin), which received another thick body of clastic sediments.

(5) A second orogenic phase during Virgilian and early Wolfcampian deformed the new flysch trough, which was overridden from the southeast by the previously deformed Marathon geosynclinal rocks along the frontal, or Dugout Creek thrust.

(6) A final postorogenic phase, from late Wolfcampian through the remaining Permian, at the beginning of which

the rocks that had been deformed by the earlier orogenies were deeply eroded and truncated, after which they were covered by the overlapping Permian, at least along their outer edge. At a later time, these postorogenic deposits were tilted away from the orogenic belt, but were not otherwise notably deformed.

It is of interest that the boundaries between at least some of these phases lie within rather than between conventional stratigraphic epochs -- especially the boundary between the leptogeosynclinal and flysch phases within the Mississippian, and the one between the final orogenic and postorogenic phases within the Wolfcampian.

This is the record of the rocks exposed within the Marathon Basin, but its incompleteness must be emphasized. The Marathon Basin is only a small exposed sample of a much larger orogenic belt, and of the marginal part at that. Much more extensive parts are concealed by cover rocks. Consequently, many fundamental questions remain unresolved, although for some of them a few clues can be inferred.

What was the nature of the edges of the geosynclinal trough? The external (northwestern) edge is now concealed beneath the frontal thrusts of the orogenic belt. However, deposits laid down in the marginal parts of the geosyncline are exposed in the northwestern and northern parts of the Marathon Basin -- the dominant chert facies of the Caballos Novaculite, the thinned wedge of the Tesnus Formation, the shelf facies of the Dimple Limestone, and the delta-front delta-plain facies of the upper part of the Haymond Formation. At times, at least, the geosynclinal trough was separated from the higher standing craton by an abrupt shelf break, from which shelf deposits slumped or slid as exotic bouldery debris onto the floor of the geosyncline -- at intervals during the Ordovician and during the middle Haymond.

Much less can be ascertained regarding the internal (southeastern) border of the geosyncline, as it is everywhere concealed by cover rocks and well control is sparse. The original width of the geosyncline, the nature of its boundary with the backlands, and the nature of the backlands themselves are virtually unknown. During the flysch phase, at least, tectonic activity raised the backlands, and perhaps internal parts of the geosyncline as well, to provide the flood of clastic sediments that were deposited in the trough.

How much of the growth of the geosyncline was taphrogenic (that is, tensional), and how much was orogenic (that is, compressional)? I infer that taphrogenic forces acted for a long period along the external border; the slumping of bouldery debris from the shelf into the geosyncline during the Ordovician and mid-Pennsylvanian may have been triggered by tensional faulting. Whether there was taphrogenic activity along the internal border is unknown; possibly taphrogeny prevailed at first and gave way to orogeny later, leading eventually to destruction of the geosyn-

cline itself during the Des Moinesian-Missourian orogeny.

Did the orogenic activity occur in discrete pulsations, separated by times of quiescence (Hall, 1956, p. 2254), or was it a continuing process (King, 1937, p. 134)? The answer inferred depends to a large extent on the predilections of the individual geologist. In this account I have chosen to distinguish a Des Moinesian-Missourian orogeny and a Virgilian-Wolfcampian orogeny, yet the pause between them may have been indistinct.

Up to this point, my discussion has been in terms of traditional geosynclinal tectonics, but the time has arrived when the Ouachita orogenic belt, including its Marathon segment, will be reinterpreted in terms of plate tectonics (Keller and Cebull, 1973), with results that cannot be foreseen. So far, attempts to construct a plate tectonics history of the Ouachita belt have resulted in models that create more problems than they solve. Greatest modifications of traditional tectonic interpretations in terms of plate tectonics will be in the poorly known backlands of the Ouachita orogenic belt. Probably much less reinterpretation of the tectonic history will be needed of the external part of the orogenic belt, such as the small segment that has been described in the Marathon region.

ACKNOWLEDGEMENTS

All of us who have worked upon the different parts of the Ouachita tectonic belt owe a deep debt of gratitude to Hugh D. Miser, whose lifelong devotion to the subject was a source of inspiration and encouragement. Through the years, I discussed with him its problems, and the twists and turns of its interpretation, as well as many related problems in the West Texas Permian basin and the Southern Appalachians. I owe another debt to W. A. J. M. van Waterschoot van der Gracht, Dutch nobleman turned Shell geologist, who first brought to us a European tectonic interpretation of the Ouachita orogenic belt, and whose larger insights are still valid today. And more specifically in regard to the Marathon region, I am indebted to my predecessor Charles Laurence Baker, one of the great reconnaissance geologists of North America, whose initial exploration of the region laid the groundwork for subsequent investigation. Like the rest of us, he could not let the region go, and during my own field work he was a frequent visitor and advisor (as well as a useful gadfly).

For my many successors, I myself have become a gadfly, but I value my associations with them all, and their guidance in the region to see the new data which they are accumulating. I cannot list them here, but their assistance is indicated by the references at appropriate places in the text. The figures accompanying the article have been beautifully executed by Gertrude J. Edmonston. All the photographs are by the present author.

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CORRELATION OF THE CARBONIFEROUS ROCKS OF THE OUACHITA TROUGH WITH THOSE OF THE ADJACENT FORELAND

By Mackenzie Gordon, Jr.¹ and Charles G. Stone²

ABSTRACT

Field studies, literature review, and examination of fossil collections from Ouachita-, Arbuckle-, and Ozark-facies rocks in western Arkansas and eastern Oklahoma lead to the following conclusions about contemporaneous deposition of the trough, unstable slope, shallow basin, and shelf rocks and the structural setting of these depositional sites.

Early Mississippian time was marked by deposition of the upper half of the Arkansas Novaculite in the generally stable, probably moderately deep Ouachita trough; of the Boone Formation and minor pre-Boone rocks on the Ozark shelf; and a thin interval in the upper Woodford Chert in the east-central Oklahoma basin. After deposition of carbonate and siliceous sediments ended in early Meramecian time, brief elevation of the shelf rocks was accompanied by a period of instability and local submarine erosion in the trough.

Beginning in late Meramecian time and continuing almost to the close of the Mississippian (Chesterian), the Moorefield Formation, including its Spring Creek and Ruddell Shale Members, the Batesville Sandstone, Fayetteville Shale, Pitkin Limestone, and the Mississippian part of the Cane Hill Formation of USGS usage were deposited on the Ozark shelf, the Caney Shale in the east-central Oklahoma basin, and the Stanley Shale, including the Hot Springs Sandstone Member at the base, in the then actively downwarping Ouachita trough. Local uplift and erosion of the shelf rocks at or near the end of this period contributed Pitkin and Fayetteville boulders to the Chickasaw Creek Member of the Stanley.

Contemporaneous early Morrowan deposition of most of the Jackfork Sandstone in the Ouachita trough, of the Hale Formation on the Ozark shelf, and the major part of the "Springer" Formation in the east-central Oklahoma basin, is indicated by ammonoid faunas. A florule from the lowermost part of the Jackfork, near Talihina, Oklahoma, suggests a possible Mississippian age for the containing beds.

Middle and late Morrowan time witnessed the deposition of (1) the Bloyd Shale on the Ozark shelf, (2) the upper part of the "Springer" Formation and all of the overlying Wapanucka Limestone in the east-central Oklahoma basin, and (3) the Johns Valley Shale in the Ouachita trough. Basinward movement of olistoliths produced a wildflysch facies at places in the western upslope parts of the Johns Valley throughout the late Morrowan; the emplacement of immense slump blocks of Caney Shale in the lower part of the Johns Valley was completed by the end of Brentwood time. This concentration of exotic blocks in the Johns Valley probably records continued movements along the postulated Bengal submarine ridge and scarp system during tectonic events that included tensional faulting and gentle crustal warpings. Along the western portions this system probably served as a partial barrier between the Ouachita trough and the adjacent foreland.

In the early Atokan, the northern part of the Ouachita trough deepened and slope and flysch deposits, derived in part from the southeast, east, and northeast extended farther northward in Arkansas than heretofore. Flysch deposition continued into middle Atokan time, but gradually much of the trough was filled with deltaic deposits brought in primarily from the north and northeast.

INTRODUCTION

Studies leading to the revision of the geologic map of Arkansas, a joint project of the Arkansas Geological Commission and the U. S. Geological Survey, have yielded considerable evidence as to the age and correlation of the Paleozoic strata. Paleontologic and biostratigraphic support for the Carboniferous part of this work was provided by Gordon. New fossil localities were found, many of them by Stone, who also was very active in the field mapping. From field and paleontologic studies and from restudy of earlier U. S. Geological Survey fossil collections, a reasonable picture has begun to emerge of the sequence of Carbon-

iferous geologic events that took place contemporaneously within the Ouachita trough and on the adjacent slope, shelf, and shallow basin that flanked it to the north and northwest (Stone, 1966; Gordon and Stone, 1969; 1973). The threefold purpose of the present report is to discuss the geologic setting of the Ouachita trough and its environs, to reassess old and present new paleontologic evidence, and to offer interpretations of the depositional history.

OUACHITA TROUGH

Rocks deposited in the Ouachita trough are now exposed from Little Rock, Arkansas, westward to Atoka, Oklahoma. The Ouachita trough was a part of the geosyncline that extended northeastward from northern Mexico across Texas and perhaps as far east as northwestern Alabama. The map of the conterminous United States in Figure 1 shows in simplified form the main structural features that controlled the distribution of the seas on and adjacent to the North American craton in middle Missis-

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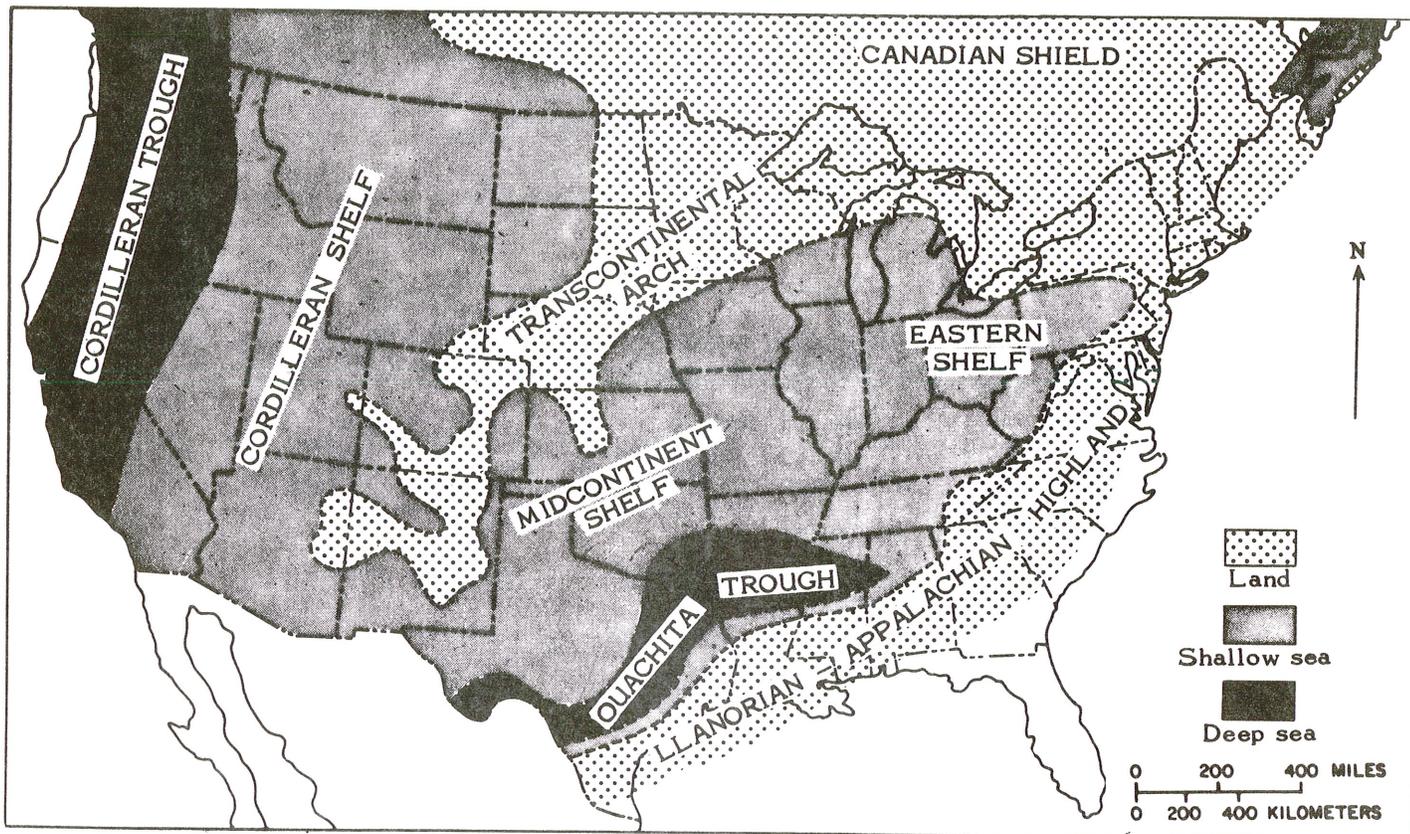


Figure 1. -- Paleogeographic sketch map of the conterminous United States in middle-Mississippian time. (Modified from Gordon, 1974.)

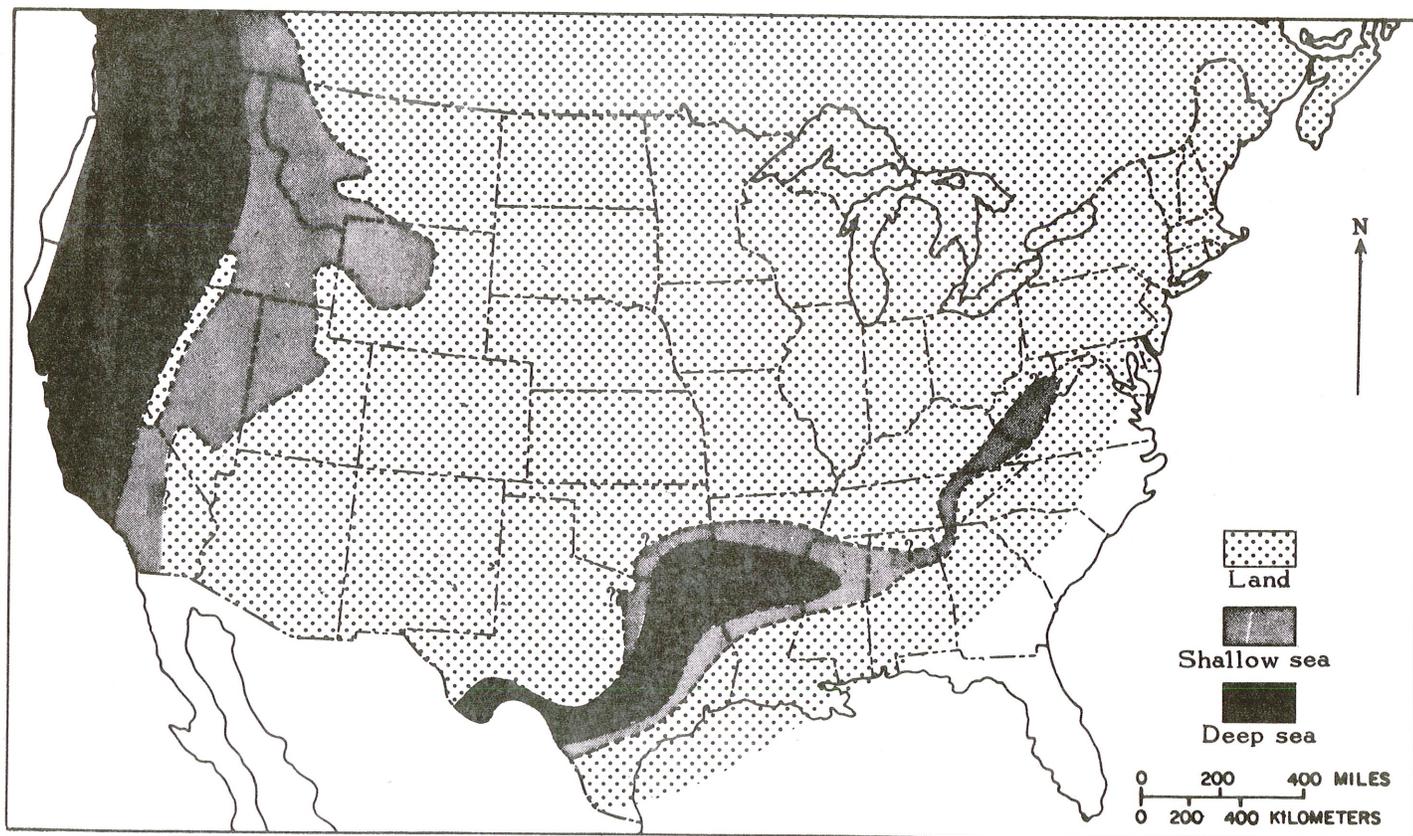


Figure 2. -- Paleogeographic sketch map of the conterminous United States at the close of Mississippian time. (Modified from Gordon, 1974.)

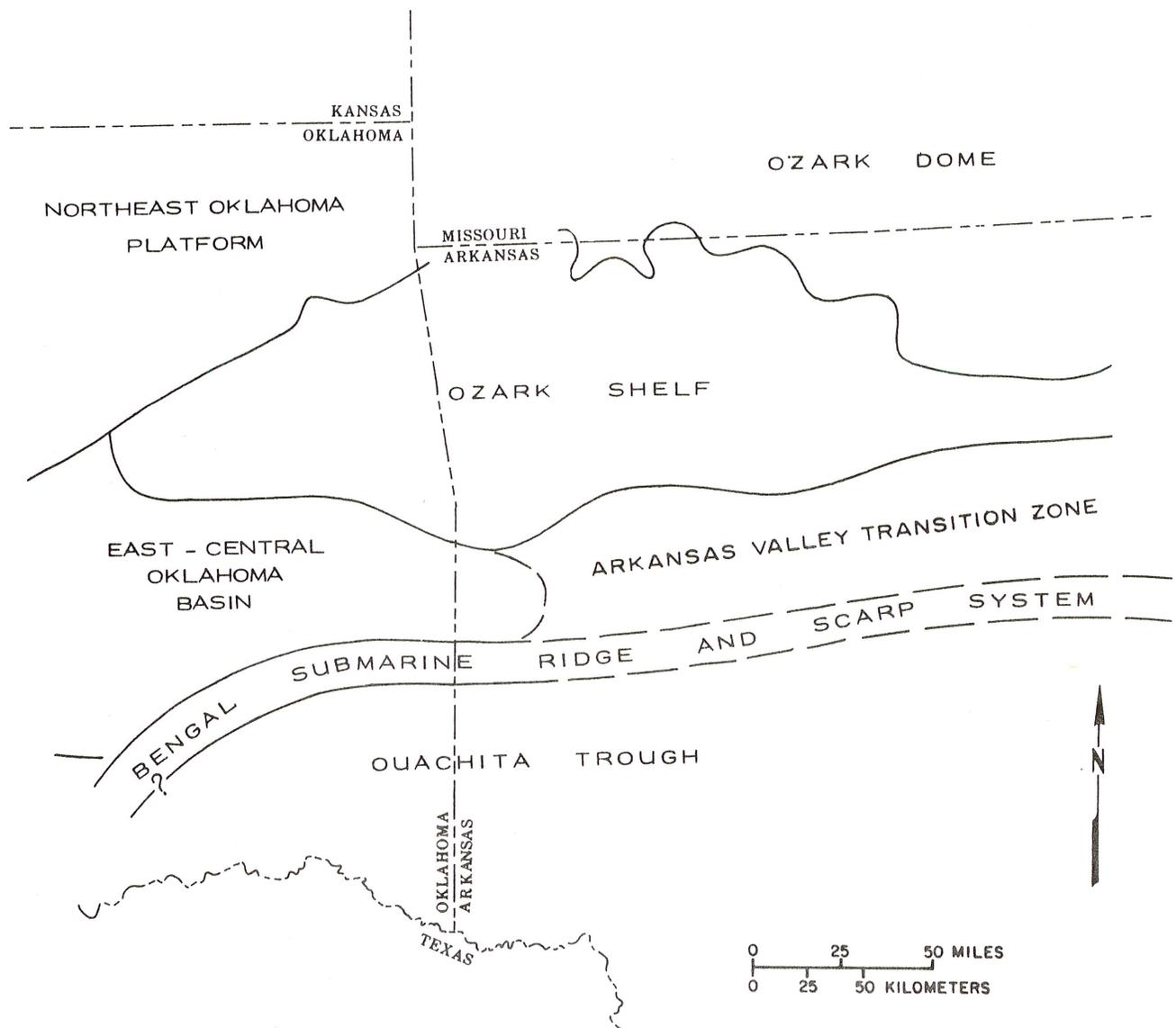


Figure 3. -- Principal structural features controlling Carboniferous deposition in the region that includes the Ouachita trough.

sippian time. The shallow epeiric seas are shown at their maximum extent during that period. The same area is shown in Figure 2 at the close of the Mississippian time, when encroachment of the seas on the craton was at a minimum.

More than 12,500 m of clastic sediments were deposited in the actively downwarping Ouachita trough in late Mississippian and early to early middle Pennsylvanian time. Most of the rocks in the deeper parts of the trough have been recognized as flysch sediments (Waterschoot van der Gracht, 1931, p. 998; Cline, 1959; 1960, p. 87-92; 1966; Cline and Shelburne, 1959, p. 177). After deposition of the lower Atokan sequence, the trough progressively filled with sediments and was eventually uplifted by folding and by thrust faulting from the southeast; this orogeny apparently lasted into Permian time (Waterschoot van der Gracht, 1931, p. 1014, 1028; Flawn, 1959, p. 26).

Both of these diagrams show an Appalachian-Llanorian landmass that provided significant quantities of sediment from the southeast and south to the Ouachita trough in late

Mississippian to early middle Pennsylvanian time. We will not conjecture here as to what tectonic plate or plates provided such a landmass. Some geologists have suggested that instead of a Llanorian landmass, an island arc lay south of the Ouachita geosyncline (King, 1950; Weeks, 1952; Morgan, 1952), but it is rather difficult to conceive of such a limited source providing materially to the quantity of sediments that rapidly filled the Ouachita trough. Whatever explanation is correct, the Ouachita trough was nevertheless an important conduit during a considerable part of the Carboniferous Period for the waters of the shallow epeiric seas that covered the midcontinent and eastern shelves.

STRUCTURAL CONTROL OF CARBONIFEROUS DEPOSITION

At least three major facies of rocks occupy different areas within the region under consideration: the Ouachita facies within the Ouachita trough, the Arbuckle facies north and west of the trough in east-central Oklahoma, and the Ozark facies on the north side of the trough in Arkansas extending westward into northeastern Oklahoma (Fig. 3).

The typical Ozark stratigraphic sequence is restricted to the shelf along the southern and western margins of the Ozark dome (Maher and Lantz, 1953; Caplan, 1957; and Frezon and Glick, 1959). In the Arkansas Valley transition zone the shelf sequence gradually grades southward into more distal shelf and, in part, upper slope deposits. Highly lenticular clastic deposits representing unstable submarine conditions probably were deposited farther downslope along the proposed Bengal submarine ridge and scarp system. Generally, a progressive deepening to the south is indicated across this region by deposits that grade from deltaic to slope to flysch conditions of deposition.

Another shelf, similar to the one on the north side, may have been present on the south side of the Ouachita trough, but no remnants of it are present at the surface. However, some evidence exists for shallower water sediment sources from this direction. Miser (1934, p. 980) thought that a landmass (Llanoria) existed south or southeast of the Ouachita region and made special reference to the northward thinning and decrease in fragment and grain size in the tuffs near the base of the Stanley Shale, the general northward thinning of sandstone beds in both the Stanley Shale and Jackfork Sandstone, and the northward disappearance of grit in the Jackfork. Paleocurrent and petrographic studies of the rocks of the Stanley Shale and portions of the Jackfork Sandstone supports a southern provenance for some sediments, which moved northwestward and northward into the trough and then westward along its axis (Briggs and Cline, 1967; Hill, 1966; Johnson, 1966; Klein, 1966; Walthall and Bowsher, 1966). Sedimentary structures that suggest northward transport were also noted in the Johns Valley Shale and Atoka Formation at places in the Athens Plateau (Stone and others, 1973, p. 101). Moreover, Walthall and Bowsher (1966, p. 131-132) and, in part, other authors describe a microfauna and additional fossils in nodules within clastic olistostromal masses in the lower part of the Johns Valley and a mold fauna in some fairly thick but minor upper Johns Valley and lower Atoka sandstones at places in the Athens Plateau in southern Pike County (locs. 7-9 in Fig. 5). Walthall and Bowsher (1966) further suggest that some of these fossiliferous rocks were transported into the trough from a shallower water environment derived from a tectonically active welt to the south.

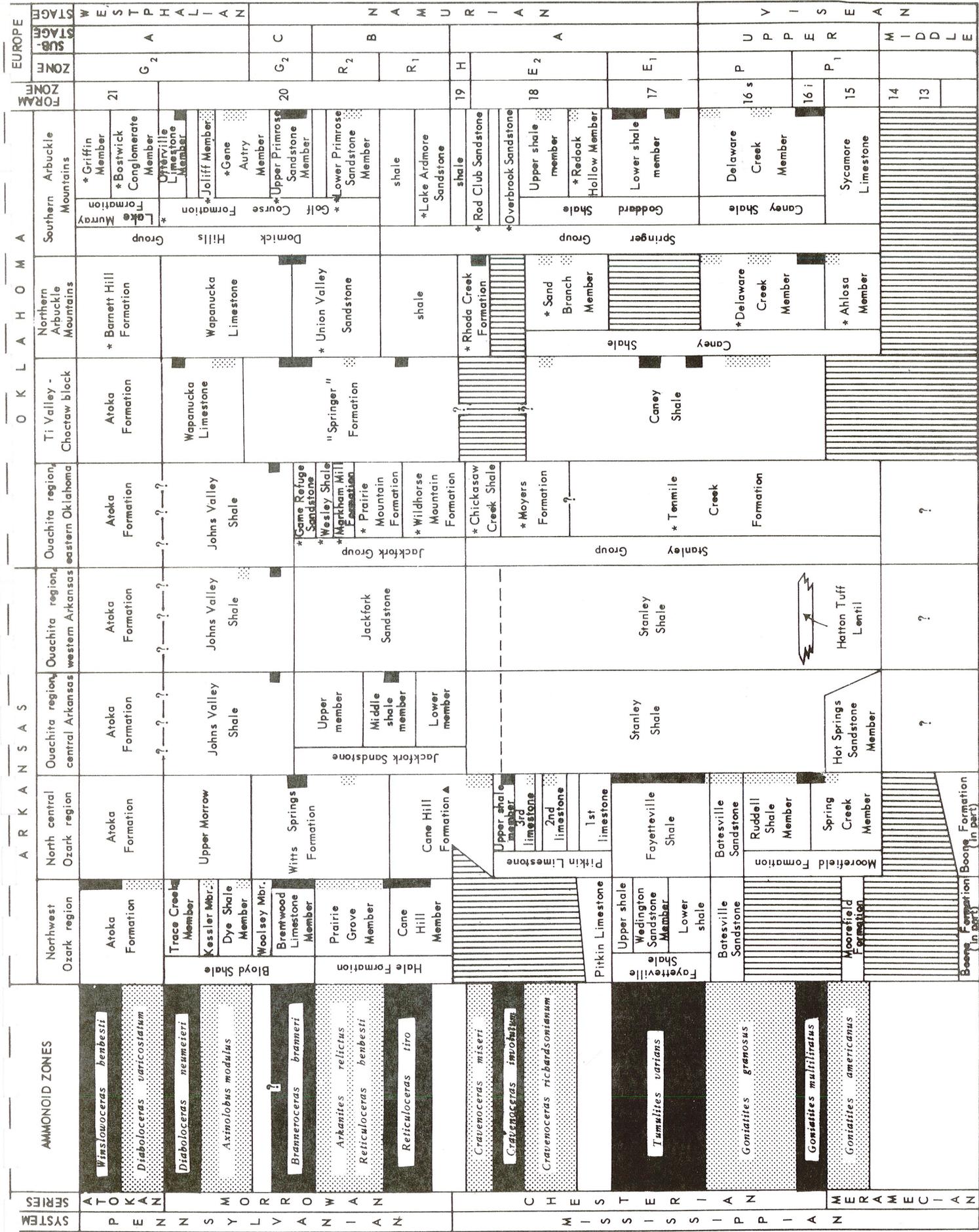
Rocks of the Arbuckle facies were deposited in the east-central Oklahoma basin, southwest of the Ozark shelf. This basin includes the beds underlying Middle Pennsylvanian rocks in parts of the McAlester and Arkoma basins, as well as the Arbuckle facies rocks now within the frontal belt of the Ouachita Mountains. It was limited at the northeast by the shallow-water environment of the Ozark shelf and at the northwest by the northeast Oklahoma platform; southwestward, it joined with the similar but slightly shallower water sediments flanking the Arbuckle uplift. Probably no sharp boundary existed to the east, where the Arbuckle facies merged with shelf, distal shelf and, in part, unstable upper slope sediments in the Arkansas Valley transition zone.

Bounding the east-central Oklahoma basin on the south side and separating it from the Ouachita trough was a long north-northeast to east-trending flexure and zone of normal faulting that was expressed topographically and is herein referred to as the Bengal submarine ridge and scarp system. Because we do not believe that this feature surfaced as a chain of islands, we do not use the terms, "a small rocky uplift" (Powers, 1928, p. 1042-1049), "Ti Valley-Bengal uplift" (Miser, 1934, p. 1003), "Bengalia" (Kramer, 1933, p. 612), "Ancestral Ouachita Mountains" (Moore, 1934, p. 447-452), or "mobile highland" (Harlton, 1938, p. 861). We agree with Fellows (1964, p. 75-79), who reviewed much of the previous literature, that this ridge and scarp system was a submarine flexure at the foreland rim of the Ouachita trough. It was also shown this way diagrammatically by Goldstein and Hendricks (1962, p. 410). This system is probably comprised of a series of submarine ridges, scarps, valleys, and canyons in or near the Ti Valley-Choctaw thrust fault block, but because of shortening by subsequent overthrusting (Hendricks, 1959, p. 48-51), it originally may have been several miles to the southeast. Evidence as to a possible southwestward continuation of this system is buried beneath the cover of Cretaceous and younger rocks near Atoka, Oklahoma. Eastward, it probably extended to about the town of Boles, Arkansas, where a huge block of Caney Shale, now in the Johns Valley, must have been transported from a point upslope. Presumably the ridge was less pronounced farther east but likely continued as a scarp and slope system along the north side of the Ouachita trough.

During the early Carboniferous, the Bengal submarine ridge and scarp system was an extremely unstable site of deposition, with (1) the often meager shelf derived sediments being mostly transported across the area into the trough through dissecting submarine canyons or channels, and (2) numerous spasmodic slide and slurry detachments into the trough from various rock units exposed along the edges of the submarine ridges and scarps. However, the Ouachita trough in southeastern Oklahoma and western Arkansas received the bulk of its detritus from source areas to the northeast, east, and southeast with sediment being transported by turbidity currents westward and southwestward along its axis, as shown by paleocurrent studies in Stanley and Jackfork rocks (Klein, 1966; Johnson, 1966; Briggs and Cline, 1967; Morris, 1971).

USE OF AMMONOID ZONATION IN CORRELATING BASIN AND SHELF SEQUENCES

Study by Gordon of fossil collections from this region, many of which contain ammonoids, has permitted us to assemble the correlation chart shown as Figure 4. Most of these collections belong to the U. S. Geological Survey and are deposited at the National Museum of Natural History in Washington, D. C. The chart diagrammatically summarizes the results of our biostratigraphic studies of the Ouachita trough and of the Ozark shelf and east-central Oklahoma basin to the north. Its framework is the sequence of ammonoid zones resulting from Gordon's Arkansas cephalopod studies (Gordon, 1964, [1965]); these zones



▲ Includes uppermost Mississippian Age "Peyton Creek beds" of McCaleb et al (1964).
 Figure 4. -- Correlation chart of some Carboniferous formations within the Ouachita trough and those of the shelf and basin sequences north of it. Vertically ruled spaces indicate hiatuses. An asterisk in front of the formation name indicates its formal adoption by the U. S. Geological Survey. Ammonoid zones are identified by alternating patterns.

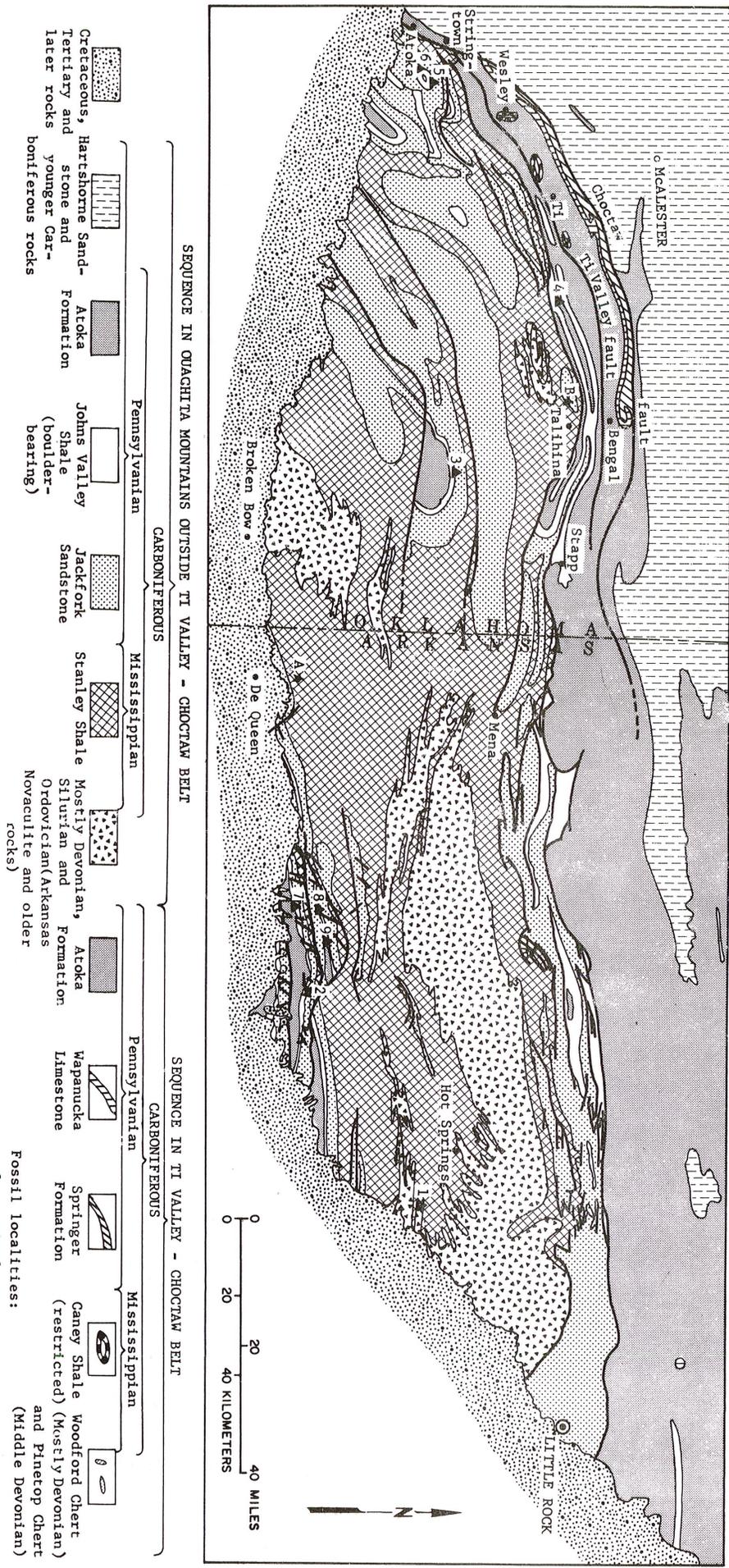


Figure 5. -- Surface geology of the Ouachita trough, showing location of several fossil localities mentioned in text. (Modified from Miser, 1934, Fig. 1)

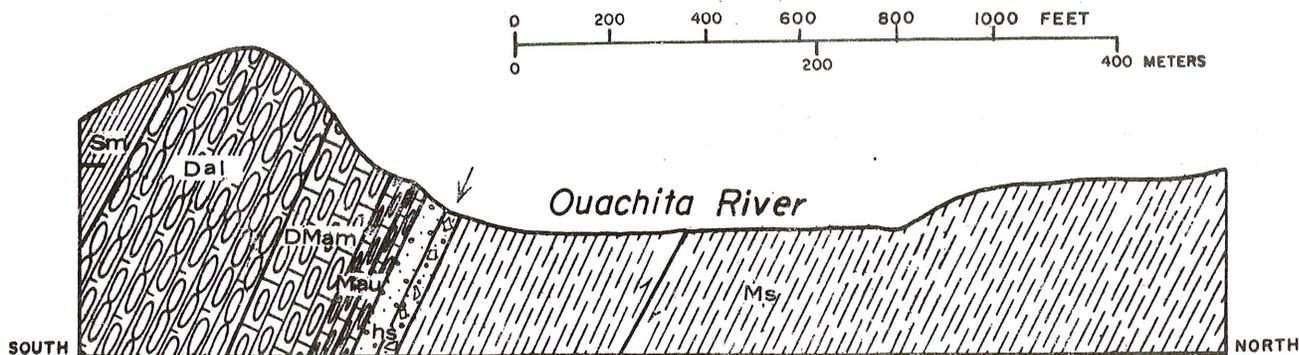


Figure 6. -- Cross section of rocks exposed at Rempel Dam on the Ouachita River, Hot Spring County, Arkansas. Arrow shows microfauna collecting locality. Sm = Missouri Mountain Shale, Dal = Arkansas Novaculite lower division, DMam = Arkansas Novaculite middle division, Mau = Arkansas Novaculite upper division, hs = Hot Springs Sandstone Member of Stanley Shale, Ms = main part of Stanley Shale.

were proposed at the Sixth International Congress of Carboniferous Stratigraphy and Geology in 1967 (Gordon, 1970). The zones are named in the broad left-hand column, where they are also identified by two alternating patterns. Small patches of these patterns in the columns to the right indicate the approximate stratigraphic position of the ammonoids in each section.

FOSSILS WITHIN THE OUACHITA FACIES

Fossils are rare in the flysch sediments; remains indigenous to the trough are restricted mainly to the conodonts in some beds and the *Nereites* assemblage of trace fossils studied by Chamberlain (1971). All the megafossils found in the Stanley Shale and Jackfork Sandstone are believed to have been transported and redeposited in olistoliths, in large slump blocks, or in turbidite intervals that moved from adjacent shallow-water environments into the trough. Most of the blocks and certainly all of them in the frontal belt were derived from the north side of the trough.

Because the fossiliferous materials formed and were deposited in shallower water, and later transported and redeposited in the flysch sequence, their presence at any one level in the Stanley-Jackfork sequence indicates only that this level is no older than the fossils. The flysch unit containing a given slump block may be slightly or appreciably younger than the date indicated by the fossils within the slump block. The sequence of transported fossils in the Stanley and the Jackfork is nearly the same as one finds in place on the shelf, therefore, the lag time for most of these blocks is inferred to be relatively short, and the fossil assemblages are thought to be essentially penecontemporaneous with the enclosing strata. Were it otherwise, a reversal of the order of the transported faunas might be expected. The Johns Valley Shale, on the other hand, contains boulders derived from rocks as old as Cambrian, graded, fossiliferous turbidite intervals, as well as material regarded as indigenous to that formation, and these often occur mixed together in a wildflysch sequence.

The distribution of fossil collections in the Ouachita trough outside of the frontal belt in Arkansas, is shown in Figure 5. The Arkansas part of Miser's sketch map has been changed to conform with recent mapping in the frontal

belt and along the southern edge of the basin; the Oklahoma part has been left essentially as Miser (1934, Fig. 1; 1954) showed it. The collecting localities are discussed later in the text.

DATING THE LOWER PART OF THE STANLEY SHALE BY MICROFOSSILS

The uppermost beds of the Hot Springs Sandstone Member at the base of the Stanley Shale at Rempel Dam on the Ouachita River, Hot Spring County, Arkansas, (fossil loc. 1, Fig. 5) contain contorted blocks of limestone, forming a deposit 1 to 4 m thick. The Hot Springs Sandstone Member of the Stanley Shale pinches out about 140 m eastward from the base of the dam; beyond the pinchout, the limestone blocks are incorporated in the basal 60 cm of the typical Stanley Shale. The rocks exposed at Rempel Dam are shown in cross section in Figure 6.

While it is not completely clear, the twisted shapes of many of the blocks suggest that they occur within transported slurry masses from an unknown formation (or formations) along the unstable edges of the Ouachita trough to the north. Supporting this supposition, is evidence that the clastics in the Hot Springs Sandstone Member were derived from the north and likely represent submarine channel-fill deposits.

Thin sections from separate blocks of this limestone, collected at two localities about 90 m apart, were studied independently by A. K. Armstrong of the U. S. Geological Survey and B. L. Mamet of the University of Montreal. The rocks contain essentially the same foraminiferal fauna, indicating that they are all from the same source. The fauna is listed in Table 1. Both investigators recognized this cosmopolitan fauna as being late Meramecian in age and as representing foraminiferal Zones 13 through 15 (see Mamet and Skipp, 1970 [1971]). Mamet suggested that a sorting process during deposition eliminated grains larger than 300 microns in diameter and, because of this, part of the fauna that might normally be present. For this reason, the assemblage cannot be narrowed down to a single foraminiferal zone. The three zones mentioned occur in the Mississippi Valley section in the St. Louis and Ste. Genevieve Limestones. Limestones of this age were hitherto largely unrepresented in Arkansas.

TABLE 1. -- FORAMINIFERA IN TWO COLLECTIONS OF LIMESTONE BLOCKS FROM THE LOWER PART OF THE STANLEY SHALE AT REMMEL DAM. [Identifications by B. L. Mamet, Univ. of Montreal.]

FORAMINIFERA	Locality	
	a	b
<i>Archaediscus</i> sp.	X	X
<i>A. krestnikovi</i> Rauzer-Chernousova		X
<i>A. ex gr. krestnikovi</i> Rauzer-Chernousova	X	X
<i>Brunsia</i> sp.	X	X
<i>B. lenensis</i> Bogush and Yuferev	X	
<i>B. aff. B. lenensis</i> Bogush and Yuferev		X
<i>Planoarchaediscus</i> sp.	X	X
<i>P. aff. P. eospirillinooides</i> Brashnikova	X	X
<i>Priscella prisca</i> (Rauzer-Chernousova and Reitlinger)	X	X
<i>Pseudocornuspira</i> sp.	X	X
<i>Tetrataxis</i> sp.	X	X
INCERTAE SEDIS		
<i>Calcisphaera laevis</i> Williamson	X	X
<i>C. pachysphaerica</i> (Pronina)	X	

The same rocks at Remmel Dam have yielded a varied conodont assemblage. In addition to the two collections that figured in the foraminiferal study, three collections made by Danilchik and others in 1955 were studied by W. H. Hass. The conodonts in all five collections have been studied by J. W. Huddle, and the fauna is listed in Table 2.

Both conodont specialists called attention to evidence of considerable reworking of elements of this assemblage. No appreciable difference in the fauna was noted between weathered surficial parts and fresh interior parts of the blocks. Of nearly 50 parataxa identified, roughly half are Devonian forms, 30 percent are Mississippian, and the remaining 20 percent cannot be assigned definitely to one or the other system. They range in age from Late Devonian or possibly late Middle Devonian to early Late Mississippian. Although all the conodonts could be older than the foraminifers, such forms as *Cavusgnathus?* sp., *Gnathodus texanus*, and *Taphrognathus varians* are compatible with the late Meramecian age indicated by the foraminifers. One species, *Gnathodus texanus*, makes up 35 percent of the total specimens in the samples.

Table 3 lists the conodonts in the lower part of the Stanley Shale in west-central Arkansas. Two assemblages were recognized by Hass (1950, p. 1578; 1953, p. 72), the lower one consisting solely of the aforementioned *Gnathodus texanus*, abundant in the basal 15 feet of the formation. A younger fauna was recorded 75 to 145 feet above the base of the Stanley in rocks that contain the Hatton Tuff Lentil and some commercial barite deposits. Although regarded by Hass as Meramecian in age, the

TABLE 2. -- CONODONTS FROM LIMESTONE BLOCKS IN THE LOWER PART OF THE STANLEY SHALE AT REMMEL DAM. [Identifications by J. W. Huddle, U. S. Geological Survey.]

DEVONIAN FORMS
<i>Ancyrodella curvata</i> (Branson and Mehl); <i>A. nodosa</i> Ulrich and Bassler; <i>A. sp.</i>
<i>Ancyrognathus</i> sp.
<i>Avisgnathus</i> sp.
<i>Icriodus</i> sp.
<i>Palmatolepis distorta</i> Branson and Mehl; <i>P. glabra</i> Ulrich and Bassler; <i>P. glabra lepta</i> Ziegler and Huddle; <i>P. glabra prima</i> Ziegler and Huddle; <i>P. marginifera</i> Ziegler; <i>P. minuta</i> Branson and Mehl; <i>P. pectinata</i> Ziegler; <i>P. punctata</i> (Hinde); <i>P. quadrantinodosa</i> Branson and Mehl; <i>P. quadrantinodosalobata</i> Sannemann; <i>P. cf. P. regularis</i> Cooper; <i>P. rugosa</i> Branson and Mehl; <i>P. superlobata</i> Branson and Mehl; <i>P. sp.</i>
<i>Polygnathus</i> cf. <i>P. decorosus</i> Stauffer; <i>P. linguiformis</i> Hinde; <i>P. varcus</i> Stauffer s. 1.
<i>Polylophodonta</i> sp.

MISSISSIPPIAN FORMS (Kinderhookian)

Elictognathus sp.
Gnathodus aff. *G. delicatus* Branson and Mehl
Polygnathus communis Branson and Mehl; *P. inornatus* Branson; *P. purus* Voges
Pseudopolygnathus aff. *P. triangularis* Voges
Siphonodella isosticha (Cooper); *S. obsoleta* Hass; *S. sp.*

(Osagean to Meramecian)

Bactrognathus? sp.
Cavusgnathus? sp.
Gnathodus texanus Roundy; *G. aff. G. semiglaber* Bischoff
Taphrognathus varians Branson and Mehl; *T. sp.*

AGE UNCERTAIN

<i>Bryantodus</i> sp.	<i>Nothognathella</i> sp.
<i>Hindeodella</i> sp.	<i>Ozarkodina</i> sp.
<i>Ligonodina</i> sp.	<i>Polygnathus</i> sp.
<i>Lonchodina</i> sp.	<i>Prioniodina</i> sp.
<i>Neoprioniodus</i> sp.	<i>Spathognathodus</i> sp.

upper assemblage is now known to be early Chesterian (J. W. Huddle, written commun., 1967). This assemblage is found also in the Barnett Formation of Texas, the Caney Shale of Oklahoma, and in the "Chester Group" of Illinois. It occurs commonly with the ammonoid faunas present in the Ruddell Shale Member of the Moorefield, Batesville Sandstone, and the Fayetteville Shale of the Ozark region in northwestern and north-central Arkansas.

TABLE 3. -- CONODONTS IN THE LOWER PART OF THE STANLEY SHALE IN GARLAND, MONTGOMERY, AND POLK COUNTIES, ARKANSAS. [Identifications by W. H. Hass, 1950, p. 1578; 1953, p. 72.]

Upper fauna (Chesterian), 75 to 145 feet above the base:
Cavusgnathus cristata Branson and Mehl
Geniculatus claviger (Roundy)
Gnathodus bilineatus (Roundy); *G. inornatus* Hass
Hindeodella ensis Hass; *H. undata* Branson and Mehl
Ligonodina roundyi Hass
Metalonchodina sp. A
Prioniodus inclinatus Hass
Roundya barnettana Hass
Subbryantodus roundyi Hass
Lower fauna (Meramecian), 0-15 feet above base:
Gnathodus texanus Roundy

PLANT FOSSILS IN THE STANLEY SHALE

The next higher fossils stratigraphically are somewhat macerated carbonaceous plants in the upper middle part of the Stanley Shale. According to S. H. Mamay (oral commun., 1967), paleobotanist of the U. S. Geological Survey, only one Stanley Shale locality has yielded a flora both sufficiently varied and well preserved to be of use in correlation. This floral assemblage was found by H. D. Miser (Miser and Purdue, 1929, p. 66-67) near Gillham, Arkansas, in the Athens Plateau of the southern Ouachita Mountains (fossil loc. A in Fig. 5). Miser (oral commun., 1967) estimated that the plants occur roughly 760 m below the top of the formation, which probably would place them in the upper part of the Tenmile Creek Member (or Formation) of Harlton (1938). The following genera are represented in the flora: *Mesocalamites*, *Sphenophyllum?*, *Adiantites*, *Alliopteris?*, *Aphlebia*, *Heterangium?*, *Neuropteris?*, *Palmatopteris*, *Rhodea*, *Rhynchogonium*, and *Trigonocarpium*. The plants were studied by David White (in Miser and Purdue, 1929, p. 66-68; and White, 1934; 1937); the identification of the *Mesocalamites* was by Mamay (in Miser and Hendricks, 1960, p. 1831).

White recognized the close relationship of the plants from the Stanley to well-documented European floras from the Waldenburg and Ostrau Series of the then Austro-Silesian coal region. In the upper Silesian basin, now part of Poland, the Waldenburg plants occur in a section about 3,000 m thick, interspersed with ammonoid-bearing marine tongues and beds (Bojkowski, 1960, Namurian A map). The plants are concentrated in the middle of the Eumorphoceras ammonoid zone, which in Arkansas ranges through most of the Fayetteville Shale, Pitkin Limestone, and Mississippian part of the Cane Hill Formation of USGS usage. The Polish beds are clearly Chesterian equivalents.

White (1934) first suggested that the Stanley plants are similar to those of the Wedington flora of the Ozark region. Later he stated that they are younger than the Wedington flora and, in fact, are Pennsylvanian in age (White, 1937, p. 45-52). In part, this was because (1) he believed that the Mississippian-Pennsylvanian boundary in the

United States was equivalent to the Viséan-Namurian (Lower Carboniferous-Upper Carboniferous) boundary of Europe, and (2) he was aware that the Waldenburg beds had been transferred from upper Viséan to lower Namurian by the Heerlen Congress of 1927. We now know that the Viséan-Namurian boundary more nearly approximates that between the lower and middle divisions of the Chesterian Series in the United States.

Mamay (in Miser and Hendricks, 1960, p. 1831) believes that the Stanley plants are closer to those of the Wedington than any other flora. Read and Mamay (1964, p. K6) assigned the Stanley plants to their flora zone 3, which is Chesterian in age.

MARINE FOSSILS IN THE STANLEY SHALE IN THE FRONTAL BELT

Most of the new discoveries of fossils in the Carboniferous rocks of the Ouachita Mountains have been made in the frontal belt; the geology of this belt is shown in Figure 7. Unlike its extension in Oklahoma, where both Arbuckle and Ouachita-facies rocks are present, the frontal belt in Arkansas is limited to rocks of the Ouachita facies. The squares on the map are townships. On this map we have plotted fossil localities, S numbers for those from the Stanley Shale, J numbers for Jackfork Sandstone localities, and JV numbers for those from the Johns Valley Shale.

Marine invertebrates of Chesterian age occur in a claystone and fine-grained sandstone lens at the spillway of Lake Winona Dam in Saline County, 30 miles west of Little Rock, Arkansas, (fossil loc. S-1 in Fig. 7). The fossiliferous beds are estimated to be roughly 350 to 460 m below the top of the Stanley. Faulting and folding in the vicinity have made it impossible to determine the stratigraphic position precisely, but the beds are believed to be equivalent to the uppermost part of the Tenmile Creek Member (or Formation) or possibly to the lowermost part of the Moyers Member (or Formation) of Harlton (1938). Like other megafossils in the Stanley Shale, these probably are contained in an allochthonous block. Some of the more diagnostic forms are shown in Table 4. The fossils occur

TABLE 4. -- CHESTERIAN INVERTEBRATE FOSSILS FROM THE LOWER PART OF THE UPPER STANLEY SHALE (UPPER TENMILE CREEK OR POSSIBLY LOWER MOYERS EQUIVALENT) AT LAKE WINONA DAM SPILLWAY, SALINE COUNTY, ARKANSAS. [Identifications by Mackenzie Gordon, Jr.]

Bryozoan: *Archimedes* sp.
Blastoid: *Pentremites* cf. *P. laminatus* Easton
Brachiopods: *Rugosochonetes* cf. *R. oklahomensis* (Snider)
Anthracospirifer aff. *A. leidyi* (Norwood and Pratten)
Reticulariina spinosa (Meek and Worthen)
Mollusks: *Phestia* sp. indet.
Leptodesma sp.
Retispira sp.
Worthenia tenuilineata Girty
Leptoptygma sp.
Trilobite: *Paladin girtyianus* G. and R. Hahn

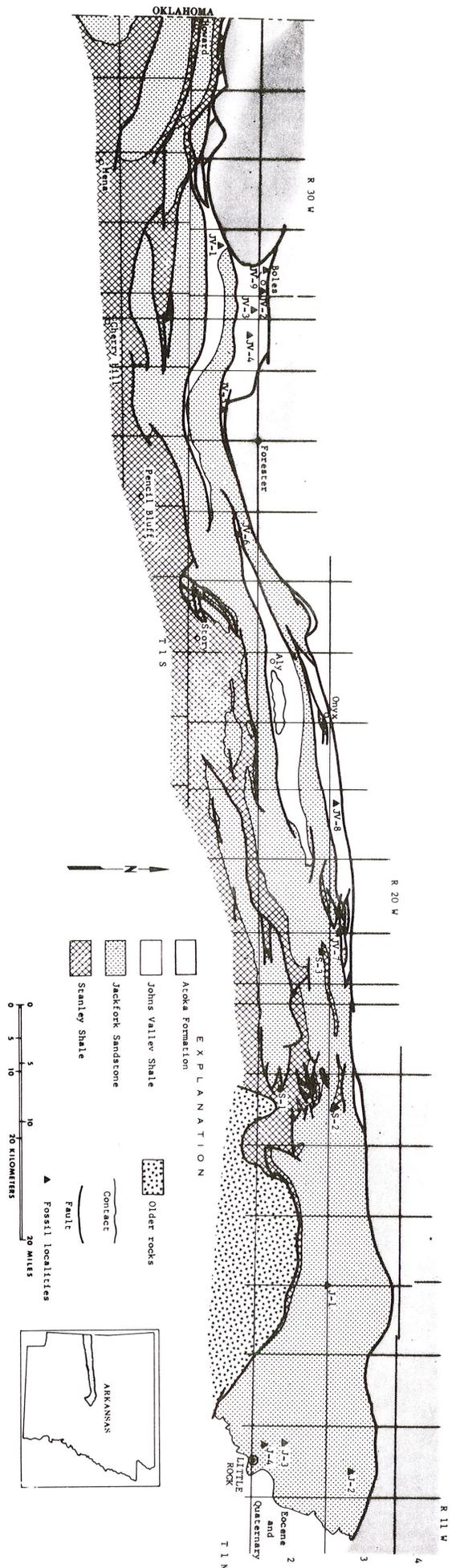


Figure 7. -- Geology of the Arkansas part of the frontal belt of the Ouachita trough, showing fossil localities mentioned in text. (Geology by C. G. Stone and B. R. Haley.)

as molds and include some 25 species. Besides Chesterian types of *Archimedes* and *Reticulariina*, the fauna includes a pentremite that may be the Pitkin species *Pentremites laminatus* Easton, and a recognizable pleurotomareacean gastropod *Wortheria tenuilineata* Girty. The type specimen of this gastropod came from the upper member of the Fayetteville Shale in the Eureka Springs quadrangle, Arkansas, but the same species occurs in Utah associated with ammonoids that suggest an equivalence to the Pitkin Limestone. These fossils date the beds at Lake Winona as post-Wedington Mississippian in age.

The Chickasaw Creek Member, the uppermost member of the Stanley, has been recognized in Arkansas as far east as the Lake Sylvia-Paron area, about 50 km west of Little Rock. Between this point and Little Rock, the member is cut out by faulting. Conodonts are sparsely distributed in the typical siliceous shales of this member, but diagnostic forms have not been collected.

At two localities, one near Lake Sylvia and the other near Forked Mountain (S-2 and S-3 on Fig. 7) boulders derived from the Pitkin Limestone on the shelf sequence are present in the Chickasaw Creek Member (Gordon and Stone, 1969; Morris, 1971, p. 388). The more westerly locality, near Forked Mountain, where the largest block of Pitkin age has a diameter of 6 m, also has erratic blocks of Fayetteville Shale provenance. More than 50 species of

Pitkin fossils have been found in these boulders, 20 percent of them common to the two localities. Some of the more diagnostic Pitkin fossils are listed in Table 5. Several of the species are more common in the upper part of the Pitkin than in the lower part. The same species of *Leptagonia*, for example, has been recognized in the Boston Mountains at two localities, both in the shale member at the top of the formation. This would suggest that deposition of the Pitkin Limestone may have been completed before uplift and that subsequent submarine erosion exhumed the boulders now resting in the uppermost part of the Stanley Shale. It indicates an equivalence of the Chickasaw Creek Member with higher beds, perhaps with the Mississippian part of the Cane Hill Formation of USGS usage, which overlies the Pitkin Limestone in some areas of Newton and Searcy Counties, Arkansas. The Mississippian part has been called the "Peyton Creek beds or shale" and included in the Pitkin Formation by some geologists (McCaleb, Quinn, and Furnish, 1964; Furnish, Quinn, and McCaleb, 1964). The "Peyton Creek beds", together with equivalents of the Cane Hill Shale Member of the Hale Formation were called the Imo Formation by Gordon (1964, [1965]).

A characteristic Fayetteville Shale fossil, the large orthoconic nautiloid, *Rayonnoceras solidiforme* Cronis (Fig. 8), was collected by Stone at the Forked Mountain locality; it was found in place as a cobble lying at an angle to the bedding. This species is fairly common in the lower shale

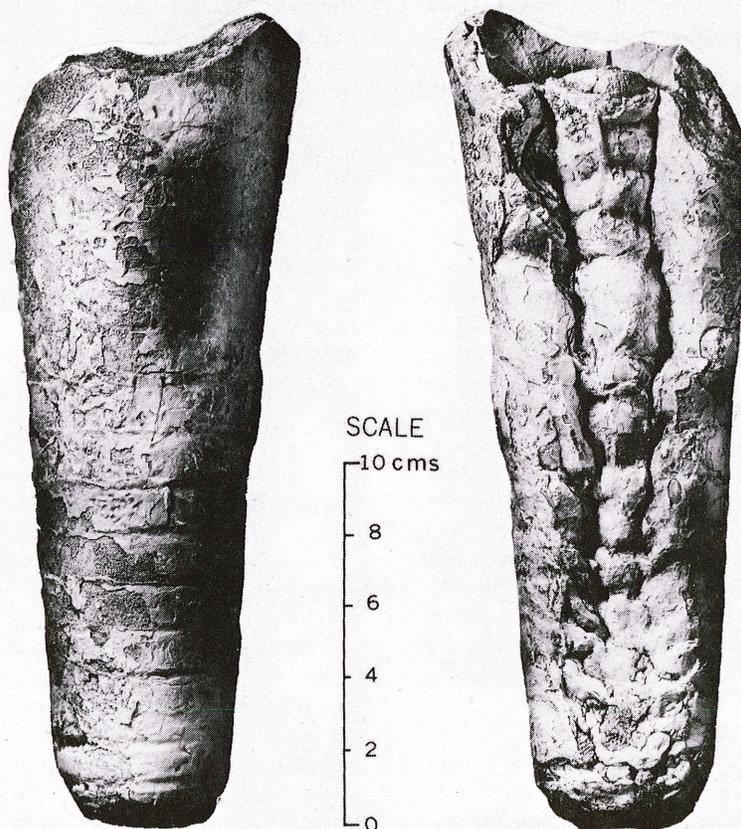


Figure 8. -- The nautiloid *Rayonnoceras solidiforme* Cronis, an exotic cobble in the Chickasaw Creek Member of the Stanley Shale.

TABLE 5. - CHARACTERISTIC FOSSILS FROM PITKIN LIMESTONE BOULDERS IN THE CHICKASAW CREEK MEMBER OF THE STANLEY SHALE, PERRY AND SALINE COUNTIES, ARKANSAS. [Coral identified by W. J. Sando, U. S. Geological Survey; other fossils, by Mackenzie Gordon, Jr.]

Coral:	<i>Lonsdaleia major</i> Easton?
Bryozoan:	<i>Archimedes</i> cf. <i>A. lunata</i> Condra and Elias
Blastoid:	<i>Pentremites laminatus</i> Easton
Brachiopods:	<i>Leptagonia</i> n. sp. <i>Streptorhynchus suspectum</i> Girty <i>Rotaia neogenes</i> (Girty)? <i>Athyris pitkinensis</i> Snider <i>Hustedia multicostata</i> Girty
Pelecypods:	<i>Schizodus chesterensis</i> Meek and Worthen <i>Edmondia pitkinensis</i> Snider <i>Sphenotus quadruplicatus</i> Snider

member of the Fayetteville Shale of the shelf sequence and is believed to have been derived from that source. The nearest surface exposures where Morrowan rocks overlap the Pitkin Limestone and rest directly on the Fayetteville Shale are about 95 km north-northwest of this locality.

Conodonts are common in certain layers in a 2 m thick carbonaceous shale exposed in the stream bank about 90 m west of the nautiloid locality. These include *Gnathodus bilineatus* (Roundy) and *G. commutatus* (Branson and Mehl), common Chesterian forms which are not, however, known in upper Chesterian rocks in the American Midcontinent. This outcrop is also interpreted as an exotic block of Fayetteville Shale.

DATING THE JACKFORK SANDSTONE BY MEGAFOSSILS

The Jackfork Sandstone rests with apparent conformity on the Stanley Shale. The lowermost beds of the Jackfork have yielded a flora of Late Mississippian (Chesterian) age in the Buffalo Mountain syncline, 6.5 km west of Talihina, Oklahoma (fossil loc. B on Fig. 5). Plants from this locality were described by David White (1937) and include the genera *Mesocalamites*, *Lepidodendron*, *Lepidostrobus*, *Rhabdocarpus*, *Rhynchogonium*, and *Trigonocarpium*. As in the Stanley Shale floral collection, *Mesocalamites* was recognized by Mamay who also has stated (S. H. Mamay, oral commun., 1969) that this is the only datable flora known from the Jackfork. It is similar to but less varied than the Stanley Shale flora discussed earlier. The presence of these plants suggests, but since they could be re-sedimented remains, they do not necessarily prove a Mississippian age for the lower part of the Jackfork.

Collections of marine fossils have been obtained in the Jackfork at five localities in Arkansas. The age of the fossils at all of these localities is Early Pennsylvanian (early Morrowan). One of these localities is in the Athens Plateau, 10 km southwest of Amity in Pike County (fossil loc. 2 in

Fig. 5). A slightly distorted sandstone cast of an ammonoid without sutures was found in place on top of a ridge of Jackfork Sandstone (Reed and Wells, 1938, p. 27-28). This ridge was later mapped as part of the Wildhorse Mountain Formation of Harlton (1938), by Walthall and Bowsher (1966, map, p. 133), but Haley and Stone (written commun., 1972) believe that because of structural complications, it is difficult to assign it with certainty to any particular part of the Jackfork.

This poorly preserved specimen is shown as *Cymoceras* sp. in Figure 9-i. It preserves faintly the surface sculpture characteristic of this genus, which ranges from the Cane Hill Member of the Hale Formation to the Brentwood Limestone Member of the Bloyd Shale in the Ozark Region.

Morrowan fossils have been collected in small to large, decalcified, sandstone slump blocks in the middle part of the Jackfork Sandstone at three localities in the vicinity of Little Rock, Arkansas. In this region, the middle part of the Jackfork is composed predominantly of shale and most of it was mapped as upper Stanley Shale on the previous State geologic map (Branner, 1929) and preliminary investigations by Stone (1966, p. 197-199). The sandstone masses are incorporated in this shale member, at horizons roughly 150 to 360 m below the base of the upper sandstone member. The westernmost of the three localities (J-1 in Fig. 7) is at Maumelle Pinnacles, 21 km west-northwest of Little Rock. The predominantly molluscan fauna at this locality aggregates some 40 species, including ammonoids of the *Reticuloceras tiro* Zone (Gordon, 1968, 1970), shown in Figures 9a, d-h, j, k and listed in Table 6. *Retites*, together with *Spirifer gorei* Mather? (Fig. 9b, c) occurs only at the locality on Miles Creek (J-2 on Fig. 7), 6.5 km north of the city limits of North Little Rock. The third locality, at Burns Park on the north side of the Arkansas River (J-3 in Fig. 7), includes the coral *Lophophyllidium* and brachiopods, but lacks the cephalopods.

TABLE 6. - AMMONOIDS OF THE *RETICULOCERAS TIRO* ZONE IN THE MIDDLE SHALE MEMBER OF THE JACKFORK SANDSTONE, PULASKI COUNTY, ARKANSAS. [Identifications by Mackenzie Gordon, Jr.]

<i>Syngastrioceras globosum</i> (Easton)
<i>Bisatoceras</i> (<i>Schartymites</i>) <i>paynei</i> (Gordon)
<i>Reticuloceras tiro</i> Gordon
<i>Retites semiretia</i> McCaleb
<i>Cymoceras adonis</i> Gordon
<i>Stenopronorites quinni</i> Gordon

The highest collection stratigraphically was found by Miser (1934, p. 989-991) in a "grit" interval near the base of the massive, channeled and graded lower sandstone beds of the upper part of the Jackfork in what is now an apartment building complex on the south bluff of the Arkansas River in Little Rock (J-4 on Fig. 7). The fauna at this locality was studied by Girty (in Miser, 1934), who concluded that it was of Pottsville age; it has been reexamined by Gordon. More than 35 species of invertebrates are present, including

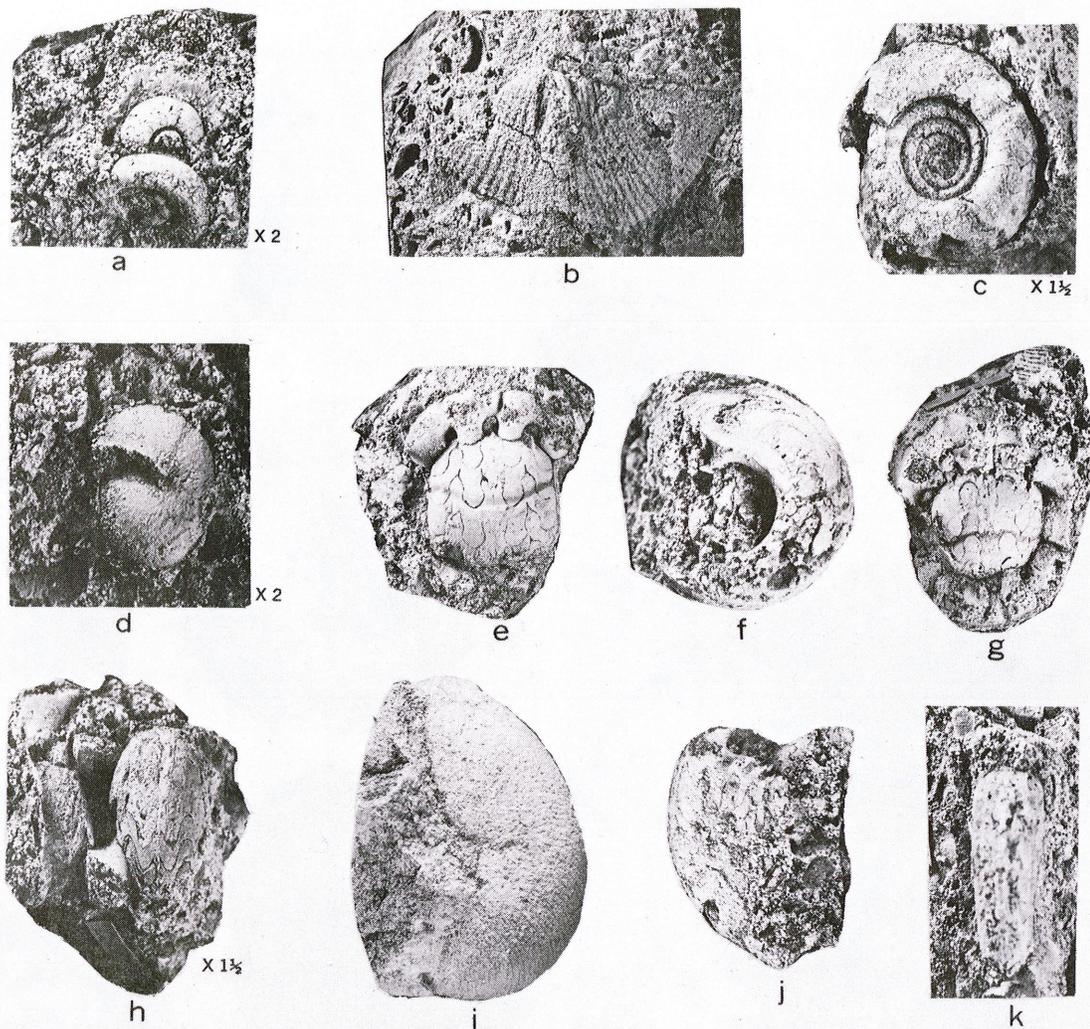


Figure 9. -- Morrowan fossils from slump blocks in the Jackfork Sandstone in Pulaski and Pike Counties, Arkansas. a, *Reticuloceras tiro* Gordon, side view; b, *Spirifer gorei* Mather? brachial valve; c, *Retites semiretia* McCaleb, side view; d, *Bisatoceras (Schartymites) paynei* Gordon, oblique side view; e-g, *Syngastrioceras globosum* (Easton), front, side, and ventral views of three specimens; h, *Cymoceras adonis* Gordon, ventral view; i, *Cymoceras* sp., side view; j, k, *Stenopronorites quinni* Gordon, side and ventral views. Views a and d-h, and j, k are of specimens from locality J-1 (on Fig. 7); b and c are from locality J-2; i is from locality 2 (on Fig. 5). All views natural size unless otherwise indicated in Figure.

the Morrowan brachiopods "*Chonetes*" *arkansanus* Mather and *Hustedia miseri* Mather and the pelecypod *Girtyana honessi* Elias. Whether or not the beds at this locality are Hale or Brentwood equivalents is not presently known, but the stratigraphic position, about 850 m below the top of the Jackfork Sandstone, would suggest Hale as the more likely.

DATING THE JOHNS VALLEY SHALE BY MEGAFOSSILS

The recent mapping program has extended the Johns Valley Shale for a considerable distance farther east along the frontal belt (Fig. 7) than previously known. The Johns Valley is a wildflysch unit in most of the frontal belt of Oklahoma and western Arkansas, containing olistostromes of boulders from formations ranging in age from Cambrian to Early Pennsylvanian; present also in this formation are huge slump blocks, mainly from the middle part of the

Caney Shale. All the slump blocks and most of the olistostromes were derived from the north, but Walthall and Bowsher (1966, p. 131-132) indicate that some clastic olistostromes in the Athens Plateau likely came from the south. Significant erratic-boulder deposits occur eastward in the frontal belt as far as the environs of the abandoned village of Forester, in extreme southeastern Scott County, Arkansas.

An indigenous fauna, predominantly of cephalopods, is found in the lower part of the Johns Valley Shale in clay ironstone concretions; these have been collected in place at 10 localities along the frontal belt, seven of them in Arkansas (JV-1 through JV-7 in Fig. 7). At Sessions Creek near Boles in Scott County (loc. JV-3), these concretions occur in crumpled shales within a few tens of feet stratigraphically above a huge Caney Shale block (Gordon, 1964, [1965], p. 52). About 20 species of fossils, nearly all of them mollusks, make up this fauna, which belongs in the

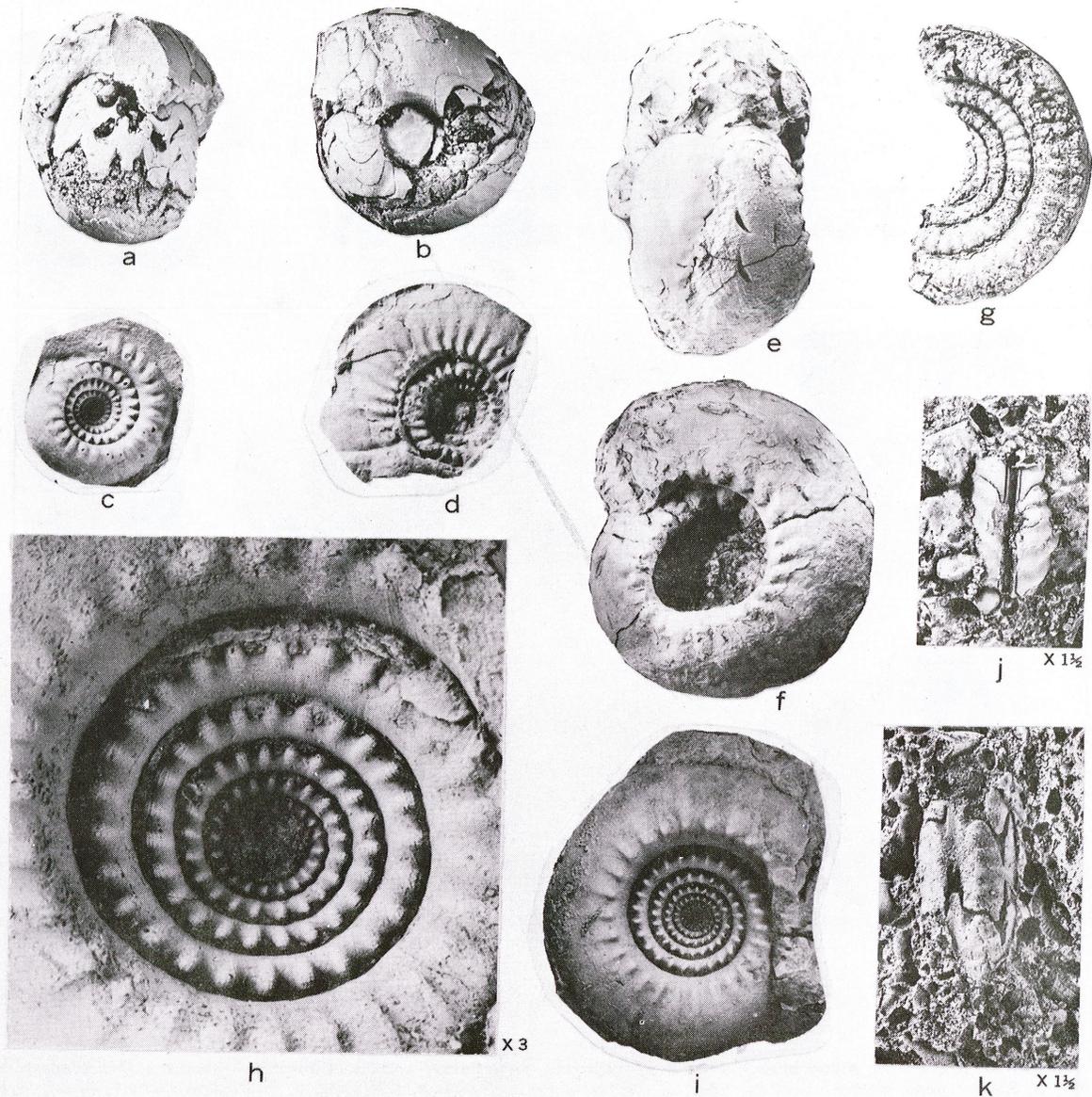


Figure 10. -- Late Morrowan fossils from the lower and middle parts of the Johns Valley Shale. a, b, *Syngastroceras oblatum* (Miller and Moore), front and side views, locality JV-3; c, h, i, *Gastroceras fittsi* Miller and Owen, side views: latex cast of small specimen (c), enlarged view of latex cast made from external mold of larger specimen (h), and an internal mold (i) of same specimen as h, locality JV-4; d, e, f, *G. adaense* Miller and Owen, side view (d), latex cast, locality JV-4, and front (e) and side (f) views of a larger specimen from locality 6 (on Fig. 5); g, *Branneroceras* aff. *B. branneri* (Smith), side view, locality JV-3; j, k. *Axinolobus quinni* McCaleb and Furnish, fragments of internal mold of venter, localities JV-9 and 3 (on Fig. 5). All Figures natural size unless otherwise indicated in Figure.

Branneroceras branneri zone and indicates Brentwood age for the lower part of the Johns Valley. Ammonoids from this zone are listed in Table 7, and also are shown in Figure 10 a-i, most of them from Scott County, Arkansas. The specimen of *Gastroceras* shown in Figure 10e, f, was collected by Harlton near Prairie Mountain, Atoka County, Oklahoma (fossil loc. 6 on Fig. 5), and is the same one figured by Harlton (1938, p. 900, Fig. 20). This ammonoid zone is known also in northern Arkansas in the Brentwood Limestone Member of the Bloyd Shale and in the Witts Springs Formation. In southeastern Oklahoma, ammonoids of this zone have been found in the top of the "Springer Shale" of the Ti Valley-Choctaw fault block and locally in the lower part of the Wapanucka Limestone, as well as in

TABLE 7. -- AMMONOIDS OF THE *BRANNEROCERAS BRANNERI* ZONE IN THE LOWER PART OF THE JOHNS VALLEY SHALE OF THE FRONTAL BELT. [Revised from Gordon, 1964 [1965].]

Syngastroceras oblatum (Miller and Moore)
Verneuillites sp.
Cymoceras miseri McCaleb
Gastroceras fittsi Miller and Owen;
G. adaense Miller and Owen
Branneroceras aff. *B. branneri* (Smith)
Stenopronorites arkansiensis (Smith)

the upper part of the Union Valley Sandstone.

At four other localities, two in Arkansas (JV-8, JV-9 in Fig. 7) and two in Oklahoma (fossil locs. 3 and 5 in Fig. 5), a stratigraphically higher fauna was found, generally in a friable, limonitic sandstone, near the middle part of the formation. Rea (1947, p. 47) has suggested that some of these beds are continuous, whereas others are lenticular and of limited extent. One of these localities is the often-cited Morrowan locality of Honess (1924, p. 14-18; Miser and Honess, 1927, p. 18-20) in the Boktukola syncline in northwestern McCurtain County, Oklahoma (fossil loc. 3 in Fig. 5). The fauna includes more than 80 species of invertebrates and contains ammonoids of the *Axinolobus modulus* Zone, of which the species known from the Ouachita trough are listed in Table 8. Two specimens of *Axinolobus quinni* McCaleb and Furnish are shown in Figures 10j, k. The larger one is from Honess's locality, and the smaller one is from near Boles (fossil loc. JV-9 in Fig. 7). The same ammonoid fauna occurs in the shelf sequence in the Dye Shale Member of the Bloyd Shale and also in the middle part of the Wapanucka Limestone in Oklahoma. Distinctive fossils are not yet known from the uppermost part of the Johns Valley Shale.

TABLE 8. -- AMMONOIDS OF THE *AXINOLOBUS MODULUS* ZONE IN THE MIDDLE PART OF THE JOHNS VALLEY SHALE. [Identifications by Mackenzie Gordon, Jr.]

Syngastrioceras oblatum (Miller and Moore)?
Phanerocheras kesslerense (Mather)
Gastrioceras sp.
Axinolobus quinni McCaleb and Furnish
Stenopronorites arkansiensis (Smith)

Fossils from limonitic clay nodules in the Johns Valley Shale in the Athens Plateau, chiefly microfossils including ostracodes and foraminifers, were regarded by Walthall and Bowsher (1966, p. 130-131) as identical with the Morrowan fauna of the Gene Autry Member of the Golf Course Formation of the southern Arbuckle region. These are associated with boulders and olistostromes in the Johns Valley. A slightly higher fauna was indicated by Walthall and Bowsher (1966, p. 132) and other authors from the upper sandy part of the Johns Valley Shale and from near the base of the Atoka Formation. This interval contains similar ostracodes and a mold fauna which was correlated with the "Morrow fauna" of Honess (1924, p. 14-18). The locations of these fossil collections are shown in Figure 5 (locs. 7-9).

The correlations that have been suggested by the ammonoid distribution are shown diagrammatically in Figure 4. Three out of five Morrowan ammonoid zones present in the type Morrowan region of northwest Arkansas have been recognized in the Ouachita trough: the *Reticuloceras tiro* Zone, common to the Cane Hill Member of the Hale Formation and to large sandstone slump blocks

in the middle shale member of the Jackfork Sandstone near Little Rock, and the *Branneroceras branneri* and *Axinolobus modulus* Zones common to the Bloyd and Johns Valley Shales. Ammonoids of the *Reticuloceras henbesti*-*Arkanites relictus* and *Diaboloceras neumeieri* Zones have not yet been identified in the basin.

ORIGIN OF LARGE EXOTIC BLOCKS IN THE JOHNS VALLEY SHALE

The immense masses of Caney Shale within the Johns Valley Shale, found locally within the region from Boles, Arkansas, southwest to the vicinity of Eubanks, Oklahoma, have presented a difficult problem for many years to geologists working in the Ouachita Mountains. Observable dimensions of these masses range from several meters to nearly a kilometer and a half in length. The largest ones are generally in a stratigraphic position in the lower middle part of the Johns Valley. One of them, in Johns Valley, is the type locality for the Caney Shale. One, near Boles, mapped by Reinemund and Danilchik (1957), exceeds 1,220 m in length. Still larger ones may occur in Jerusalem Hollow, southeast of the Kiamichi River near Stanley, Pushmataha County, Oklahoma, where they are found along a 32 km belt and have been regarded by Cline (1956, 1960) and others as constituting a tongue of Caney Shale in place. In that area, however, the surface cover exceeds 99 percent, and the Caney Shale sections are exposed only in gullies spaced about 2 to the kilometer. It is not possible, therefore, to walk the Caney Shale outcrop and determine whether it is continuous or whether it is made up of a succession of large blocks along strike. Nearly all the Caney sections in Jerusalem Hollow are highly calcareous and contain some limestone. All of them yield the same ammonoid fauna, that of the *Tumulites varians* Zone (Gordon, 1970), demonstrating that they represent the middle part of the Caney Shale. Smaller blocks, down to cobble size, scattered through the Johns Valley Shale but generally more concentrated in the lower part, usually contain the older *Goniatites granosus* Zone fauna.

At two localities in this belt of Caney Shale, clusters of exotic boulders as large as 5 by 15 m with individual boulders as much as 60 cm in diameter are embedded in typical Caney Shale. Miser and Hendricks (1960) concluded that the matrix in which these clusters are embedded is part of the Johns Valley Shale rather than an exotic mass of Caney Shale which is itself embedded in Johns Valley Shale. Both have expressed their general agreement with the present interpretation based upon biostratigraphic evidence (T. A. Hendricks, written commun., 1975).

Completely negating the concept of the Caney Shale being in place at the base of the Johns Valley Shale is the fact that the Pitkin boulders we have found locally in the Chickasaw Creek Member of the Stanley Shale contain fossils younger than those in the large middle Caney blocks, yet they are in a stratigraphic position two formations lower. Furthermore, the conodont fauna of the lower part of the Caney Shale (Branson and Mehl, 1940) is the same one that Hass (1950) collected in the lower part of

the Stanley Shale. Moreover, the Jackfork Sandstone has yielded only Morrowan marine fossils. Obviously at least 2,100 m of younger beds lies stratigraphically below those middle Caney blocks in the Johns Valley Shale.

Our investigations have brought us back to the same conclusion stated by Ulrich (1927), Moore (1934), and Miser (1934)--that all the occurrences of Caney Shale associated with the Johns Valley Shale are exotic. The hypothesis that the outcrops of the Caney Shale in the lower part of the Johns Valley Shale are in place (Miser and Honess, 1927; Cline, 1956, 1960; Miser and Hendricks, 1960) is not borne out by our biostratigraphic study of the underlying formations.

TIME OF DEPOSITION OF OLISTOSTROMES IN THE JOHNS VALLEY SHALE

The olistoliths in the lower part of the Johns Valley Shale range in age from Late Cambrian to Early Pennsylvanian, according to Shideler (1970), who showed in a series of sketch maps the distribution pattern of boulders of different geologic ages. The youngest boulders known to us in the lower part of the Johns Valley are middle Morrowan (Brentwood) in age. A Pennsylvanian boulder from the vicinity of Boles, Arkansas, cited by Miser (1934, p. 998), contains a molluscan faunule now known to include *Syngastrioceras oblatum* (Miller and Moore) and is regarded by Gordon as Brentwood in age. This species is common in the *Branneroceras branneri* Zone, ammonoids of which occur in place in the lower part of the Johns Valley Shale and also in the type Brentwood. Most of the boulders are from rocks of the Arbuckle facies, but some in the eastern part of the boulder area come from the Ozark facies, according to Reinemund and Danilchik (1957) and Shideler (1970).

The emplacement of the immense Caney Shale masses in the Johns Valley Shale must have begun and been completed in Brentwood time. Similar stratigraphic position of all the huge Caney blocks in the Johns Valley supports this concept of a brief time span for the period of emplacement. Smaller boulders, however, either individually or in olistostromes, continued to be deposited by submarine slumping and turbidity currents in the upper part of the Johns Valley until the end of Morrowan time, as they had been in earlier Carboniferous time in some of the more shaly parts of the Stanley Shale and Jackfork Sandstone. These later boulder deposits included blocks of Wapanucka Limestone, particularly in the vicinity of the Windingstair Range.

SUBMARINE TRANSPORTATION OF LARGE ROCK MASSES

One of the most difficult obstacles to the acceptance of Pennsylvanian age for the entire Johns Valley Shale has been the sheer size and local profusion of the Caney Shale blocks incorporated within it. A few years ago it was not easy to accept an exotic origin for such a quantity of rock. Recent investigations, however, have made the concept of slump-block origin for this material considerably more

palatable. A growing body of literature supports a slump-block origin for large rock masses ranging from Cambrian to Tertiary in age in stratigraphic sequences in many parts of the world. Investigations of present-day oceanic slopes by such organizations as JOIDES (Joint Oceanographic Institutions for Deep Earth Sampling) has shown that tremendous submarine slump masses, miles in extent, are relatively common on these slopes.

The massive Caney blocks in the western part of the Ouachita trough are for the most part, concentrated in the frontal belt or upslope parts of the central belt and are now generally less than 25 kilometers from their presumed points of origin within or adjacent to the Ti Valley-Choctaw fault block. The original distance traveled may have been as much as 50 kilometers or more for some blocks but it was shortened by subsequent thrust faulting from the southeast.

The basinward movement of these rock masses seems to have occurred because of the peculiarly favorable nature of the slick substrate, and it was triggered by a series of tectonic events. Vertical to steep uplift on seismically known tensional faults forming a series of submarine scarps, and local steepening of the slope north of the slide area was probably the basic mechanism. The locus of these faults was along what we have termed the Bengal submarine ridge and scarp system. In concurrence with this hypothesis is the widespread early Pennsylvanian orogeny indicated by Powers (1928, p. 1049-1062), Tomlinson (1929, p. 20-27), and others in southern Oklahoma, including the Criner Hills and the Arbuckle Mountains, that Miser (1934, p. 1008) noted was coincident or nearly coincident with boulder deposition in the Johns Valley Shale.

DEPTH OF DEPOSITION IN JOHNS VALLEY TIME

In the frontal belt of the Ouachitas indigenous ammonoids have been found in concretions and nodules in the lower parts of the Johns Valley Shale. These fauna are not considered to be absolutely diagnostic of depth of sedimentation, because under optimum conditions they may survive in deep as well as shallow water. A diverse molluscan fauna is found distributed through small lenticular boulders in the middle of Johns Valley Shale in the frontal belt. These boulders are interpreted as submarine slump masses derived from nearly concurrent shallower deposits on the adjacent platform or from the Bengal submarine ridge and scarp system, and therefore the contained mollusks are not indicative of depth of deposition of the Johns Valley. The authors believe that there may have been some moderation in the depth of the Ouachita trough but agree with Chamberlain (1971, p. 78) who on the basis of trace fossils assigned a lower bathyal-abyssal depth of deposition to the Johns Valley in the frontal Ouachitas.

In the central Ouachita Mountains of Oklahoma (fossil locality 3, Fig. 5) the middle and upper Johns Valley is comprised of friable, graded sandstone containing some fragmental to well preserved fauna (Honess, 1924, p. 14-

16 and Shelburne, 1960, p. 39-40). It is believed that these deposits are deep water turbidites that constitute a series of submarine fan-channel sequences.

In the Athens Plateau similar fossiliferous sandstones with some minor olistostromes were described in the Johns Valley and lower Atoka Formations by Walthall and Bowsher, (1966, p. 130-132) and, in part, other authors and a deep water submarine fan-channel regime is also proposed for these sediments.

MAGNITUDE OF CARBONIFEROUS DEPOSITION

A comparison of the maximum thickness of rocks of Carboniferous age in the three distinct depositional areas is given in Table 9. The correlations indicated in this table are based upon the faunal studies discussed in this report. The sources for the thickness figures are listed. All the maximum thickness for rocks of the Ouachita trough are from western Arkansas. In separating the middle and lower Morrowan in Table 9, we have for convenience used the Hale-Bloyd contact as our dividing line.

Owing to the complexity of the structure in the Ouachitas, few, if any, complete sections exist, so some of the thickness figures in Table 9 are of necessity only approximations. Nevertheless, they give a reasonable idea of the relative rates of deposition and of changes in those rates. A notable change occurred, for example, in middle and late Morrowan time, when the depositional rate slowed more rapidly in the Ouachita trough than in the other two areas. Maximum thickness of lower Morrowan strata in the Ouachita trough is nearly five times that of rocks of the same age in the east-central Oklahoma basin and 25 times that on the Ozark shelf. In middle and late Morrowan time, the deposition was less in all three regions, but the greatest decline proportionally was in the Ouachita trough--to about one-fifth of early Morrowan deposition. The maximum thickness of strata of middle and late Morrowan age in the Ouachita trough is slightly less than twice that of strata of the same age in the east-central Oklahoma basin and about seven times that on the Ozark shelf. This was immediately followed by a marked increase in the rate of deposition in early Atokan time.

DEPOSITIONAL HISTORY IN MISSISSIPPIAN TIME

Let us review the sequence of events that accompanied the deposition of the sediments in the three areas: the Ouachita trough, east-central Oklahoma basin, and Ozark shelf, from earliest Carboniferous time to middle-Middle Pennsylvanian time. In early Carboniferous time, the Ouachita trough was relatively stable and probably moderately deep; deposition was primarily chemical, but included some minor influxes of coarse material resulting from turbidity currents and sedimentary slumping. The middle member of the Arkansas Novaculite was deposited in this basin, beginning in the Late Devonian and continuing through the Early Mississippian (Kinderhookian). Within the same time span, the Woodford Chert was deposited in the east-central Oklahoma basin, and the

Chattanooga Shale and several local and minor formations of Kinderhookian age were laid down on the Ozark shelf, both of which areas were at least locally elevated and rather unstable.

Osagean time was marked by submergence of the Ozark shelf and the deposition upon it of the carbonates and cherts of the Boone Formation. Presumably at this time, the upper chert member of the Arkansas Novaculite was deposited in the Ouachita trough, but fossil evidence is lacking to date this member. Quartzitic sandstones occur locally within this unit and suggest some detachments from the north into the trough and further deposition by bottom and turbidity currents. Carbonate present in the novaculite suggests deposition above carbonate compensation depth and presumably above lysocline. The absence of beds of known Osagean age within the east-central Oklahoma basin suggests that this region was uplifted and was not receiving sediments during the Osagean, or that a thin shale unit might represent this time interval.

These conditions in each of the three areas prevailed into early Meramecian (Warsaw) time. Then the Ozark shelf was again elevated with respect to sea level; a brief period of erosion ensued, preceding the resubmergence of the shelf in late Meramecian time.

Very little is known of middle Meramecian events within the Ouachita trough, but present evidence indicates that continuous deposition prevailed generally. Niem (1972) reported a gradation, 3 to 30 m thick, between beds of the Arkansas Novaculite and the overlying Stanley Shale over most of the southern and central Ouachita Mountains, possibly recording a gradual change from predominantly biological/chemical precipitation to clastic sedimentation. He also recognized a local high or highs indicated by novaculite conglomerate lenses, angular unconformity, and thinning of tuffs and the strata between them in the lower Stanley. Stone and Haley, in a paper they are preparing for a later Ouachita Symposium volume, suggest alternatively that this sequence probably records a change in regimes from less vigorous, perhaps largely chemical deposition accompanied by some spasmodic influxes of northerly-derived olistostromal material and submarine channel debris to generally much more vigorous deposition of clastic turbidite, and in part volcanic material from a predominantly southern source.

The limestone blocks found at Remmel Dam are believed to have been derived from lenses and pods, some of them within sandstone beds, that were originally deposited somewhere upslope, perhaps along the Bengal submarine ridge and scarp system, in early late Meramecian time. These presumably were transported in later Meramecian time to form a deposit in the Hot Springs Sandstone Member at the base of the Stanley Shale.

Evidence from ammonoids and conodonts suggests that the deposition of the Stanley Shale in the Ouachita trough, of the Caney Shale in the east-central Oklahoma basin, and of the Moorefield Formation on the Ozark shelf all began

TABLE 9. -- APPROXIMATE MAXIMUM THICKNESSES OF ROCKS OF EARLY MISSISSIPPIAN THROUGH EARLY MIDDLE PENNSYLVANIAN AGE RECORDED IN EACH OF THE THREE DEPOSITIONAL AREAS.

SERIES OR SUBSERIES	OUACHITA TROUGH	EAST-CENTRAL OKLAHOMA BASIN	OZARK SHELF
LOWER ATOKAN	Atoka Formation, lower member 5,500 m ¹	Atoka Formation, lower member 760 m ⁹	Atoka Formation, lower member 685 m ¹²
MIDDLE AND UPPER MORROWAN	Johns Valley Shale 615 m ²	Wapanucka Limestone 220 m ¹⁰ "Springer" Formation, upper part 120 m ¹¹	Boyd Shale 90 m ¹³
LOWER MORROWAN	Jackfork Sandstone (in part) 3,050 m ³	"Springer" Formation, lower part 640 m ¹¹	Hale Formation 120 m ¹³
CHESTERIAN	Stanley Shale, main part 3,660 m ⁴	Caney Shale 160 m ¹⁰	Cane Hill Formation, lower part; Pitkin Limestone; Fayetteville Shale; Batesville Sandstone; Moorefield Formation (Ruddell Shale and uppermost Spring Creek Members) 320 m ¹⁴
MERAMECIAN	Lowermost Stanley Shale, including Hot Springs Sandstone Member 60 m ⁵ Intergrading Stanley Shale and Arkansas Novaculite 30 m ⁶	---?-----?--- ---?-----?---	Moorefield Formation, Spring Creek Member 84 m ¹⁴ Boone Formation, upper part 59 m ¹⁵
OSAGEAN	Arkansas Novaculite, most of upper member 55 m ⁷	---?-----?---	Boone Formation, lower and main part 90 m ¹²
KINDERHOOKIAN	Arkansas Novaculite, upper part of middle member 9 m ⁸	Woodford Chert, upper shaly part 0.5 m ¹⁶	Walls Ferry Limestone; uppermost Chattanooga Shale, etc. 3 m ¹⁴

Sources of thickness information:

¹Reinmund and Danilchik (1957), as interpreted by Stone and Haley

²Gordon and Stone, this report

³Walthall and Bowsher (1966, p. 127)

⁴Johnson (1966, p. 141)

⁵Goldstein (1959, p. 103)

⁶Niem (1972) as interpreted in this report

⁷Purdue and Miser (1923, p. 5)

⁸Hass (1951, p. 2527, 2540) as here interpreted

⁹Buchanan and Johnson (1968, Fig. 7)

¹⁰Hendricks and others (1947)

¹¹Hendricks and others (1947) redistributed on faunal basis by Gordon, this report

¹²Croneis (1930, p. 89)

¹³Frezon and Glick (1959, Pls. 29-31)

¹⁴Gordon (1964 [1965]) and this report

¹⁵Gordon and Kinney (1944)

¹⁶Ham (1959, p. 75)

at approximately the same time--during the late Meramecian. The Ouachita trough, at the advent of dominant clastic deposition, began to sink rapidly and received sediments from forelands on both sides, but most sediments were derived from the south or southeast, or from the east along the axis of the trough. The east-central Oklahoma basin also deepened.

During Chesterian time, deposition of the Stanley Shale in the Ouachita trough and of the Caney Shale in the east-central Oklahoma basin continued as the uppermost Spring Creek Member and Ruddell Shale Member of the Moorefield Formation and the overlying Batesville Sandstone, Fayetteville Shale, Pitkin Limestone, and the lower part of the Cane Hill Formation of USGS usage were deposited in the Ozark shelf. The Caney Shale was laid down in somewhat deeper water than was present over adjacent areas of the shelf. Another emergence of the Ozark shelf began in northwest Arkansas in Pitkin time and in north-central Arkansas in latest Mississippian time. Deposition of Pitkin boulders in the Chickasaw Creek Member of the Stanley Shale probably was concurrent with deposition of the Mississippian part of the Cane Hill Formation of USGS usage along the southern margin of the eastern part of the shelf. A predominantly south-southeasterly source of sediments in the Ouachita trough, accompanied by turbidite pulsations and minor detachments of erratic blocks from the slope and shelf to the north, is suggested for the Chesterian parts of the Stanley Shale.

MISSISSIPPIAN--PENNSYLVANIAN BOUNDARY

Over the Ozark shelf, the Mississippian-Pennsylvanian boundary occupies the depositional hiatus that began in latest Mississippian time. The amount of geologic time represented by the unconformity differs from place to place, the greatest gaps occurring in parts of northwest Arkansas and northeast Oklahoma. Deposition seems to have been uninterrupted in the east-central Oklahoma basin and Ouachita trough. Faunal evidence, as to the precise location of the boundary in these two basins, is lacking, however. The gap with question marks shown between the Caney Shale and "Springer" Formation in Figure 4 in the Ti Valley-Choctaw fault-block column represents an absence of faunal evidence but not a known physical break in sedimentation. As suggested earlier in this paper, the Mississippian-Pennsylvanian boundary in the Ouachita trough is believed to be in the basal part of the Jackfork Sandstone. Presumably it also is near the contact between the Caney Shale and the "Springer" Formation in the east-central Oklahoma basin.

EARLY MORROWAN DEPOSITIONAL HISTORY

During early Morrowan time, deposition of the major part of the Jackfork Sandstone (or Group) in the Ouachita trough went on simultaneously with deposition of the Hale Formation on the Ozark shelf. The Cane Hill Member of the Hale Formation was laid down more or less contemporaneously with the middle and probably the lower member of the Jackfork in the vicinity of Little Rock.

The sediments of the Prairie Grove Member of the Hale Formation were likewise laid down contemporaneously with the bulk of the upper member of the Jackfork. Evidence for the presence of a river system that emptied southward or southwestward into the Ozark shelf in north-central and northeastern Arkansas in Morrowan time is contained in the sedimentary pattern of the Witts Spring Formation (Glick, 1976), which the deposition of the lower part was at the same time as that of the Prairie Grove. Paralic conditions are indicated by occasional beds containing marine fossils in this formation.

In the east-central Oklahoma basin, early Morrowan time witnessed the deposition of the "Springer" Formation, which contains Prairie Grove ammonoids in its middle part. Deposition of this shale continued until approximately middle-Brentwood time; two species of *Gastrioceras* typical of the *Branneroceras branneri* Zone, restricted on the Ozark shelf to the Brentwood Limestone Member of the Bloyd Shale, have been found in its upper part. The same ammonoid fauna occurs in the top 2 m of the Union Valley Sandstone, deposited under shallower water conditions on the northeast side of the Arbuckle dome.

TECTONIC EVENTS AND DEPOSITION IN MIDDLE AND LATE MORROWAN TIME

In the earlier part of the Carboniferous, the tectonic events that controlled the depositional pattern in this region were restricted to the sinking of the Ouachita trough and some fluctuation of the Ozark shelf. In Brentwood time, this pattern changed. The northern part of the Ouachita trough was probably only moderately deep during this period, as likely suggested by indigenous ammonoids in concretions in the lower Johns Valley Shale and by the proximal-appearing turbidite deposits of the Game Refuge Sandstone the uppermost Jackfork formation of Harlton (1959).

The large assemblages of exotic blocks in the Johns Valley Shale derived from the Arbuckle and to a lesser extent the Ozark facies to the north records the breaching and at least partial destruction of the Bengal submarine ridge and scarp system that separated the foreland from the Ouachita trough. This was a complex process and seems to have been initiated by extensive movements along normal faults trending east-northeast and generally downthrown on the south side (Harlton, 1938, p. 859-865, and others). These movements likely were associated and occurred simultaneously with widespread early Pennsylvanian orogeny indicated by Powers (1928, p. 1042-1049) and others in southern Oklahoma. These postulated faults were along the Bengal system and in adjacent parts of the east-central Oklahoma basin and the Arkansas Valley transition zone to the east and were the predecessors and in part related to the growth faults that greatly furthered the development of the Arkoma basin in early and middle Atokan time. The fault scarps exposed the older rocks of the Arbuckle facies and provided a source for the aprons of olistostromes, which were deposited along the north slope of the Ouachita trough, in some places as much as 55 km or more down-

slope from their source. Faulting probably was accompanied by the development of southward-trending submarine canyons which locally breached the submarine ridge and scarp system, this process controlled much of the wildflysch deposition in the northern part of the Ouachita trough. Many of the recently deposited and barely consolidated sediments of the "Springer" Formation were also swept southward into the Ouachita trough, providing detritus to the Johns Valley Shale. Along the top of the submarine ridge, the sediments of the Arbuckle facies were removed in some places down to the first resistant layers, the calcareous middle part of the Caney Shale. In two areas, one large area upslope from Johns Valley and Jerusalem Hollow, Oklahoma, and the other in the vicinity of Boles, Arkansas, great masses of middle Caney Shale, tilted gently basinward because of further gentle warping, broke loose from the subjacent 'incompetent' shales or muds of the lower Caney and moved downslope on the slick substrate beneath (essentially on the upper surface of the Jackfork Sandstone lubricated by a relatively thin layer of Johns Valley Shale). The calcareous middle Caney beds behaved competently and held together in huge blocks, whereas the soft shaly lower Caney beds, which in part they moved upon, tended to break up into smaller fragments. Emplacement of these great middle Caney blocks seems to have taken place within a brief interval and to have been completed in Brentwood time.

Deposition of the main part of the Wapanucka Limestone, 220 m thick, (Hendricks and others, 1947), began in middle Brentwood time in the east-central Oklahoma basin, which was now more openly connected with the Ouachita trough, the Chickachoc Chert facies was laid down in somewhat deeper water in this basin. The western part of the Ozark shelf, which had been somewhat unstable in early Morrowan time, was uplifted in northwestern Arkansas at the end of Brentwood time. A series of coastal swamps formed along the new shoreline as the sediments of the Woolsey Member of the Bloyd, including the Baldwin coal, were laid down. This brief emergence was suddenly terminated in middle Bloyd time by the rapid incursion of the late Morrowan sea, introducing the fauna of the *Axinolobus modulus* Zone, beginning with a widespread conglomeratic bed (that on some of the 1929 Arkansas State geologic map was mistaken for the base of the Atoka Formation). Within the Ouachita trough, the upslope depositional area of the Johns Valley Shale was receiving spasmodic submarine slump masses or channel fills containing a large and varied shallow-water molluscan faunal assemblage derived from nearly concurrent deposits on the adjacent platform or from elevated portions of the Bengal submarine ridge and scarp system. During this time fossiliferous sands from the platform facies were also being carried by turbidity currents and deposited in a postulated southward prograding submarine fan-channel sequence in the central Ouachita Mountains of Oklahoma. A similar deep-water deposit is suggested for comparable rocks in the Athens Plateau area in Arkansas, except that, they were likely derived from a southern platform facies. The middle part of the Wapanucka Limestone was deposited in the east-central Oklahoma basin at this time.

In Late Morrowan time deposition of sandy sediments of fluvial and deltaic origin continued in northeast-central Arkansas. In northwest Arkansas, finer clastic sediments of the Trace Creek Shale Member of the Bloyd Shale, containing ammonoids of the *Diaboloceras neumeieri* Zone, were deposited, as was the upper part of the Wapanucka Limestone in east-central Oklahoma. A collection of ammonoids of the *Diaboloceras neumeieri* Zone made by P. K. Sutherland (written commun., 1973) from the upper part of the Wapanucka in the Ti Valley-Choctaw fault block, 22.4 km north of Daisy, Oklahoma, has been examined by Gordon. Although deposition of the Wapanucka Limestone was contemporaneous with that of the Johns Valley Shale, transported blocks derived from it are found rarely in the Johns Valley within 8 km of the edge of the east-central Oklahoma basin. Some intercalated lenses in the upper part of the Johns Valley in the area between the Ti Valley and Windingstair faults have been regarded as representing the Wapanucka Limestone (Fellows, 1964, p. 50-53, 98).

ATOKAN DEPOSITION

The Ouachita trough migrated northward and probably deepened in early Atokan time. Turbidite deposition took place farther north than previously reaching a northern limit in Arkansas near a line that extends through Conway westward roughly through Danville and Ola, and a little north of Waldron (Stone, 1968; Haley, 1966; Haley and Stone, Unpub. data). In some places this limit is more than 10 miles north of the Ti Valley Fault. As much as 5,500 m of lower Atoka sediments is present in this belt. Deltaic deposits reached great proportions in northeast-central Arkansas, as vast quantities of medium-to-coarse clastic sand as well as finer material were brought in from the northeast, perhaps through the Illinois basin. Considerable sediment also came from the east-southeast. The deposition formed a clastic wedge that in the region of the present Arkansas Valley thickened rapidly basinward; it kept more or less equal to the rate of subsidence, and the shoreline oscillated greatly (Merewether, 1961).

In the Athens Plateau of the southern Ouachita Mountains, an incomplete lower Atoka section of alternating sandstone and shale, more than 1,525 m thick, in faulted synclinal structures, was recorded by Walthall and Bowsher (1966, p. 131) and by the recent work of Haley and Stone on the Arkansas State geologic map. Stone and others (1973), however, assigned the lower part of this sequence to the middle and upper parts of the Johns Valley Shale, including some of the minor fossiliferous mold-fauna-bearing units that were correlated by Walthall and Bowsher (1966, p. 132) with the "Honest mold fauna" of late Morrowan age in the Boktukola syncline in Oklahoma.

In northwest Arkansas the lower part of the Atoka Formation contains several beds of marine fossils deposited apparently in a paralic environment. These include ammonoids of the *Diaboloceras varicostatum* and *Winslowoceras henbesti* Zones. While these rocks were being laid down, a shelf facies of very dark gray shale and intercalated limestone was deposited in the northeast Arbuckle area. This

was called the Barnett Hill Formation by Harlton (1938) and includes beds containing *Profusulinella*. Although the early Atokan ("Winslow") ammonoids are not known in the Barnett Hill Formation, most geologists working in this region consider it to be part of the Atokan Series. This contention has now been confirmed by the discovery mentioned above of latest Morrowan ammonoids in the Wapanucka Limestone, which underlies the Barnett Hill Formation.

In the Oklahoma Ouachita Mountains at least 1,675 m of Atoka are preserved in major synclines (Stark, 1966, p. 164). North of the Choctaw fault the formation ranges in thickness from a featheredge to slightly more than 4,575 m in the vicinity of Wilburton, Oklahoma (Buchanan and Johnson, 1968, p. 77). Sudden increases in thickness were noted by Buchanan and Johnson (1968) on the south side of growth faults, which seem to have controlled the formation of the Arkoma basin during early and middle Atoka time. The Atoka Formation in this area is composed of sediments that are largely deltaic.

The youngest flysch deposits in the Ouachita trough are middle Atoka in age; their remnants are restricted to parts of the frontal belt in central western Arkansas.

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**CONODONTS FROM GRAPTOLITE FACIES
IN THE OUACHITA MOUNTAINS, ARKANSAS AND OKLAHOMA**

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ABSTRACT

Conodonts have been recovered from almost one-half of the samples collected from the lower Paleozoic rocks in the core of the Ouachita Mountains. The samples came from thin limestone units in a succession that consists predominantly of shales and sandstones. Preservation of the conodonts ranges from poor to good; most specimens can be identified at least to the generic level and many can be identified with species. Abundance of conodonts in most samples is very low. Elements from the lowest unit, the Collier Shale, represent a fauna that is widespread in North America. Age of the Collier is demonstrated to be Early Ordovician rather than Cambrian as has been assumed. Higher units in the succession have yielded conodonts typical of the North Atlantic Province. Previously established ages of these units, based on graptolite occurrences, generally are confirmed.

Strata in McCurtain County, Oklahoma, believed by Pitt (1955) to underlie the Collier were identified by him as the Lukfata Sandstone. Limestones from the type section of the Lukfata contain conodonts that indicate late Early or early Middle Ordovician age. The overlying strata are correlative with the Collier of Arkansas on the basis of conodonts. An inverted stratigraphic sequence is indicated.

INTRODUCTION

The stratigraphy of the lower Paleozoic formations of the core of the Ouachita Mountains has presented taxing problems to the geologists who have worked in that region. The succession consists primarily of shales, interrupted at several levels by major sandstone units, with minor interbeds of limestone. The structure is exceedingly complex, and the region is heavily forested so that exposures are relatively small and discontinuous. As a consequence, no thick continuous sequences are available and the succession must be reconstructed through regional mapping and interpretation.

Purdue (1909) and Miser (1917) subdivided the stratigraphic sequence and established the formation names now in use (Fig. 1). They recognized the several thick shale sequences as separate formations with intervening sandstone formations. Although variations in character among the shales of the several formations have been described, these differences are subtle so that the stratigraphic position of a small and isolated exposure commonly is difficult to determine. In addition, stratigraphic studies are hindered by the paucity of fossils in these formations. Graptolites have been collected from several levels within the succession in Arkansas, but their occurrences are quite localized and sporadic. Nevertheless graptolites have served to establish generalized ages for all but the two lowest units in the

section, the Collier Shale and the Crystal Mountain Sandstone. The traditional ages assigned to these two formations have been based solely on interpretation of their stratigraphic position. Hendricks et al (1947) reported conodonts, inarticulate brachiopods, trilobites, radiolaria and spicules from the Bigfork Chert and the Womble Shale at Black Knob Ridge in Oklahoma, the westernmost exposures of the lower Paleozoics of the Ouachitas. Harlton (1953) reported these fossils as well as ostracodes from these units in the same area. However none of the fossils were identified or described, and their stratigraphic occurrences were not documented precisely.

The presence at several levels within the Ouachita sequence of carbonate units suggested to us the possibility of using conodonts to confirm and hopefully to refine the ages previously established for these lower Paleozoic units. Carbonates can be reduced by treatment with organic acids to free the conodonts, whereas the well consolidated shales do not respond to disaggregating procedures. Conodonts have been studied intensively over the past two decades, and a detailed zonation is becoming established for lower Paleozoic rocks. Accordingly we collected limestone samples from localities (Fig. 2) in Arkansas where carbonate units representing the several formations are exposed.

We are indebted to the staff of the Arkansas Geological Commission, and particularly to Charles Stone and O. A. Wise, for helping us to locate suitable outcrops and for collecting additional samples. Colleagues at the University of Missouri, particularly David Kennedy and Timothy McHargue, assisted with field collecting, and George Viele offered counsel regarding problems of structural geology. The laboratory work was supported in part by grant GP-4786 from the National Science Foundation.

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TRADITIONAL AGE ASSIGNMENTS	LOWER PALEOZOIC SEQUENCE IN ARKANSAS	THIS REPORT	LOWER PALEOZOIC SEQUENCE BROKEN BOW ANTICLINE
M. ORDOVICIAN	BIGFORK CHERT	M. ORD.	LUKFATA FM.
	WOMBLE SHALE		
L. ORDOVICIAN	BLAKELY SANDSTONE	L. ORD.	
	MAZARN SHALE		
	CRYSTAL MOUNTAIN SDST.		
CAMBRIAN	COLLIER SHALE		COLLIER SHALE

Figure 1. -- Correlations of the lower Paleozoic units in the core of the Ouachita Mountains.

Conodonts were recovered from all of the units from which samples were collected, although about one-half of the samples were barren. To date our collections from all units include approximately 4000 conodont specimens. Preservation is poor in the samples from the lower units (Collier and Mazarn) but generally is good in the material from the Womble. Our investigations, although representing only a reconnaissance, demonstrate that conodonts can be useful to geologists working in the Ouachita region. Because of the structural complexity of the area, systematic collecting of many exposures will be required if a complete conodont sequence is to be established. Nevertheless our preliminary results show that most of the ages and correlations previously based on graptolites are essentially correct, although the conodonts may offer greater precision. However, the conodonts demonstrate that earlier correlations and age determinations that were effected in the absence of fossils and on the basis of interpretation of stratigraphic position have been erroneous. In the following sections, we report the conodonts recovered from each formation and our evaluation of their biostratigraphic significance. All specimens illustrated on Plates 1 and 2 are deposited in the conodont collection of the University of Missouri-Columbia (UMC 1040-9 through UMC 1042-18); the remaining specimens are deposited in the bulk conodont collections of the University.

COLLIER SHALE

The age of the Collier Shale has remained an enigma since the formation was reported for the first time by Purdue (1909). Almost all of the geologists who have published on the rocks of the core area of the Ouachita Mountains have commented on the absence of fossils from the Collier. Age assignments have been based on assump-

tions about its stratigraphic position in the Ouachita sequence coupled with interpretations of the significance of its lithology in the depositional regime of the southern midcontinent.

Most workers have consigned the Collier to the Cambrian System. This interpretation was expressed first by Ulrich (1911, p. 676-677) who stated that: "Although fossils have not been found in these two basal formations [Collier and Crystal Mountain], I am strongly inclined to refer them to the lower Cambrian." Miser and Purdue (1929, p. 24) expanded upon this conclusion and established the traditional age assignment of the Collier. They reported a conglomerate to be present in the basal part of the Crystal Mountain Sandstone in the De Queen and Caddo Gap Quadrangles in Arkansas. Clasts in this conglomerate were described as derived from limestones and cherts like those of the underlying Collier. The Crystal Mountain was stated to be transitional upward into the overlying Mazarn Shale which already was known to be of Early Ordovician age on the basis of graptolite occurrences. Miser and Purdue interpreted the conglomerate at the base of the Crystal Mountain to indicate a significant unconformity at the top of the Collier and assigned the latter formation to the Cambrian on the basis of stratigraphic position and relationships. Most subsequent workers accepted this interpretation.

A few dissenters have preferred to include the Collier in the Ordovician System, although no strong evidence has been advanced in support of this conclusion. Ham (1959, p. 82) sought to equate the rocks of the shaly Ouachita facies with the carbonate sequence of the Arbuckle Mountains. He was able to effect correlations of the higher part of the Ouachita section (Polk Creek Shale through Mazarn Shale)

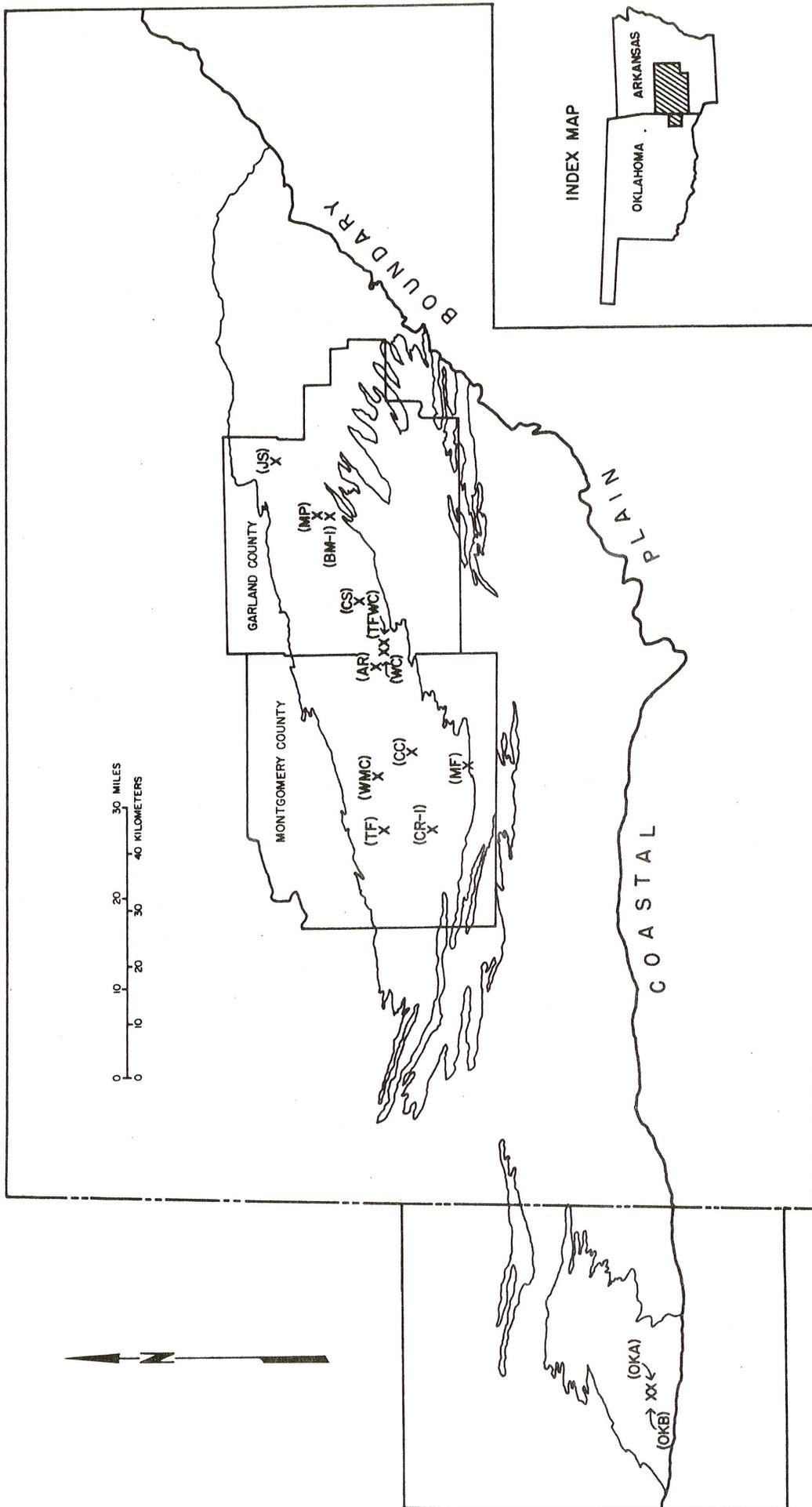


Figure 2. -- Localities in the Ouachita Mountains from which conodonts were recovered. Sinusoidal lines enclose areas of pre-Pennsylvanian outcrops.

with the Sylvan Shale through Kindblade Formation of the Arbuckle area on the basis of common fossils. He equated the Collier with the lower part of the Cool Creek Formation (Early Ordovician) of southern Oklahoma, although no justification was provided for this correlation. Apparently Ham's conclusion about correlation of the Collier was based wholly on comparative positions in stratigraphic sequences.

Berry (1960, p. 35) "arbitrarily" correlated the Collier with the McKenzie Hill Formation of southern Oklahoma and equated both formations with the lowest of the Ordovician graptolite zones he recognized in North America. This correlation of the Collier was influenced by the presence in it of indeterminate fragments that Berry believed might be fragmented graptolite specimens. Pitt, et al. (1961, p. 75) reported the inability to find a basal conglomerate in the Crystal Mountain Sandstone in its type area along Collier Creek and could find no other evidence for an unconformity at the top of the Collier. Therefore they tentatively rejected the Cambrian age assignment of the Collier and placed it in the Ordovician System.

The conodonts reported here (Table 1) are the first identifiable fossils that have been recovered from the Collier Shale. Sixteen samples representing six localities were processed. Because of the limited exposures at the localities, no sections were measured. Fifteen of the samples came from the core area of the Ouachita Mountains in Arkansas near the type area of the Collier as outlined below; the remaining sample is from the core of the Broken Bow Anticline in Oklahoma. Localities in Arkansas from which productive samples of Collier were obtained are as follows; the sample from Oklahoma is considered under the discussion of the Lukfata Sandstone.

AR-1 - From ledges exposed along the north side of

U. S. Highway 270 just east of the west boundary of sec. 26, T. 2 S., R. 23 W., Montgomery County, Arkansas (McGraw Mountain 7.5-minute quadrangle).

WC-1 and 2 - Ledges just downstream from where U. S. Highway 270 crosses Murphy Creek, SW¼, NE¼, sec. 31, T. 2 S., R. 22 W., Garland County, Arkansas (McGraw Mountain 7.5-minute quadrangle).

WMC-1 and 3 - Ledges exposed along and in the unsurfaced road that leads southeast from Arkansas Highway 27, SE¼, SW¼, sec. 26, T. 2 S., R. 25 W., Montgomery County, Arkansas (Mount Ida 15-minute quadrangle).

CC-1 and 2 - Exposures in and along Collier Creek, SW¼, sec. 17, T. 3 S., R. 24 W., Montgomery County, Arkansas (Glenwood 15-minute quadrangle).

TF-1 - Ledges along unsurfaced road near center sec. 35, T. 2 S., R. 26 W., Montgomery County, Arkansas (Mount Ida 15-minute quadrangle).

Only nine of the Collier samples were productive of conodonts. Although poorly preserved in most cases and very sparing in their abundance, the conodonts include elements of a fauna that is widespread in its geographic distribution. These conodonts establish unequivocally that the Collier belongs in the Ordovician and that it represents a level somewhat above the base of the system.

The presence of *Cordylodus angulatus* Pander serves to establish the Early Ordovician age of this unit. This cosmopolitan conodont element has been found in almost all of the older Ordovician conodont faunas reported in North America. This species also occurs in earliest Ordovician (Tremadocian) strata of Australia (Druce and Jones,

TABLE 1. -- DISTRIBUTION OF CONODONT ELEMENTS AMONG SAMPLES FROM THE COLLIER SHALE OF OKLAHOMA AND ARKANSAS, ALL ELEMENTS ARE LISTED *SENSU FORMO*.

	WC-1	WC-2	WMC-1	WMC-3	TF-1	AR-1	CC-1	CC-2	OKB	Total
<i>Acanthodus lineatus</i> (Furnish)	1	—	—	—	—	—	1	—	—	2
<i>Acodus oneotensis</i> Furnish	1	—	—	—	—	—	—	—	—	1
<i>Acontiodus propinquus</i> Furnish	—	—	—	—	—	—	—	3	1	4
<i>Chosonodina herfurthi</i> Müller	—	—	—	—	—	—	—	1	—	1
<i>Cordylodus angulatus</i> Pander	—	—	—	—	1	—	—	—	1	2
" <i>Drepanodus</i> " <i>subarcuatus</i> Furnish	—	—	—	—	—	—	1	—	—	1
Drepanodiform elements	—	—	1	1	—	—	2	—	9	13
<i>Loxodus bransonii</i> Furnish	—	—	1	1	—	1	—	—	—	3
<i>Oistodus</i> aff. <i>O. lanceolatus</i> Pander	—	—	—	—	—	—	3	—	1	4
" <i>Oistodus</i> " <i>triangularis</i> Furnish	—	—	—	—	—	—	1	—	—	1
<i>Paltodus bassleri</i> Furnish	2	1	—	—	—	—	2	—	3	8
aff. <i>Scolopodus warensis</i> Druce and Jones	—	—	—	—	—	—	4	—	5	9
aff. <i>Scolopodus sexplicatus</i> Jones	—	—	—	—	—	—	—	—	5	5
New Genus	2	1	1	—	1	—	—	6	—	11
Total	6	2	3	2	2	1	14	10	25	65

1971) and of the Baltic region of northern Europe (Lindström, 1960). The Ordovician age of *C. angulatus* is clearly established in its other occurrences by association with a diverse fauna, particularly including trilobites whose Ordovician age cannot be contested.

Cordylodus angulatus has a moderately long range in the lower Ordovician of North America. In the lower part of its range, this species occurs with few associated elements. By contrast, in the higher part of its range, *C. angulatus* is joined by a diversity of distinctive elements including both "simple cones" and multidenticulate forms. These include *Paltodus bassleri* Furnish, *Loxodus bransoni* Furnish, *Acanthodus lineatus* (Furnish), "*Oistodus*" *triangularis* Furnish, and *Chosonodina herfurthi* Müller, all of which are represented in at least some of the Collier samples. This association was designated "Fauna C" by Ethington and Clark (1971, p. 69-73) who reviewed the known occurrences of this fauna in North America. The presence of this fauna in the Collier and in the McKenzie Hill Formation of the Arbuckle Mountains (Mound, 1968) corroborates the correlation of these formations as was inferred by Berry. No conodonts have been reported from the earliest Ordovician of the Ozark region. However, the conodont faunal assemblage of the Collier is present in the Oneota Dolomite of the Upper Mississippi Valley (Furnish, 1938).

LUKFATA FORMATION

The sequence in the core of the Broken Bow Anticline in McCurtain County, Oklahoma, was investigated by Honess (1923). He recognized the Collier as the oldest unit that crops out in the area but also reported (p. 36) a sequence of "shales and related materials" exposed along Lukfata Creek. He could establish no clearly defined stratigraphic relationships for the latter succession but concluded that they were "probably Collier." Pitt (1955) reconsidered the sequence along Lukfata Creek and recognized an upper unit of shale and limestone that he identified with the Collier. Underlying the Collier and dipping in the same direction is a series of sandstones and shales which Pitt segregated as a new formation, the Lukfata Sandstone.

Pitt did not indicate any age or correlation for the Lukfata other than its position physically beneath the Collier for which the traditional Cambrian age was assumed. Decker (1959, p. 92, and Fig. 1) equated the Lukfata-Collier sequence with the Reagan Sandstone and Honey Creek Formation, both of Cambrian age, of the Arbuckle Mountains. Goldstein (1961) also correlated the Lukfata with the Reagan and with the Dagger Flat Formation in the Marathon region of Texas as well.

Charles Stone of the Arkansas Geological Commission provided us with two large samples collected from the core of the Broken Bow Anticline. One of these, collected in the NW¼, SE¼, sec. 17, T. 5 S., R. 24, E., is from a "flaggy, dense, gray limestone interbedded with gray shale and some thin quartzitic sandstone." This sample (OKA) is from the type locality for the lower member of the Lukfata Sandstone as defined by Pitt (1955, p. 13). The other sample

(OKB) is from "flaggy, dense to finely crystalline gray limestone with interbedded shale" and was collected in the NW¼, NE¼, SW¼, sec. 18, T. 5 S., R. 24 E. This is an area mapped as Collier by Pitt and by Honess (1923).

Two kilograms of rock from the sample of Collier yielded 25 identifiable although poorly preserved conodonts. *Cordylodus angulatus* Pander, *Acontiodus propinquus* Furnish and *Paltodus bassleri* Furnish confirm the correlation of these rocks with the Collier of Arkansas and reaffirm the Ordovician age of that formation. Conodonts are much less abundant in the carbonate from the Lukfata Formation. Eleven kilograms were processed and yielded 59 specimens, most of which can be identified with or compared to previously described conodont elements. In the list that follows, the numbers refer to the number of specimens.

3	Acodiform elements
2	aff. <i>Acodus deltatus</i> Lindström s. f.
7	<i>Distacodus expansus</i> (Graves and Ellison) s. f.
1	aff. <i>Distacodus stollus</i> Lindström s. f.
5	Drepanodiform elements
1	<i>Drepanodus</i> aff. <i>D. arcuatus</i> Pander s. f.
1	<i>Drepanodus</i> aff. <i>D. longibasis</i> Lindström s. f.
10	Oistodiform elements
5	?aff. <i>Oistodus elongatus</i> Lindström s. f.
1	<i>Oistodus</i> aff. <i>O. inaequalis</i> Pander s. f.
1	<i>Oistodus parallelus</i> Pander s. f.
	<i>Paracordylodus gracilis</i> Lindström
7	Cordylodiform elements
10	Oistodiform elements
1	<i>Scolopodus gracilis</i> Ethington and Clark s. f.
1	<i>Scolopodus quadraplicatus</i> Branson and Mehl s. f.
1	<i>Scolopodus</i> aff. <i>S. rex</i> Lindström s. f.
1	<i>Scolopodus triangularis</i> Ethington and Clark s. f.
1	New genus?

This assemblage clearly demonstrates the Ordovician age of the Lukfata strata, and that they are younger rather than older than the Collier. *Acodus deltatus* s. f., and *Distacodus stollus* s. f. occur in undescribed collections from the El Paso Group (Bed 15, subunit B₁, of Cloud and Barnes, 1948) in west Texas, the Monument Spring Member of the Marathon Formation of Texas, and in the Kindblade Formation of southern Oklahoma. These elements are present above the middle of the Fillmore Formation in western Utah (Hintze, et al., 1972), and together with *Paracordylodus gracilis* and *Scolopodus rex*, are characteristic components of the conodont fauna of the Ninemile Formation of central Nevada (Ethington, 1972). A fauna from eastern Pennsylvania with many elements in common with the Lukfata assemblage has been reported recently by Bergström, Epstein and Epstein (1972). Most of these North American occurrences are in strata whose age has been established on grounds other than conodonts as being in the higher part of the Canadian Series.

Bergström, et al., (1972) reported difficulty in placing the conodont-bearing rocks from the Hamburg klippe of

eastern Pennsylvania in the standard Ordovician sequence for North America. They were able to correlate them with confidence on the basis of conodonts with strata from Sweden that also are characterized by *Didymograptus balticus* and *Phyllograptus densus*. Although the North American equivalents of some European graptolite zones have been the subject of disagreement (Skevington, 1968; Berry, 1968), it is clear that the Swedish zones just mentioned are equivalent at least in part to strata in North America that contain *Tetragraptus fruticosus*. Berry (1960, p. 35) established that the Mazarn Shale contains graptolites that indicate that the formation includes the zone of *T. fruticosus*.

Thus the Lukfata can be correlated by conodonts via strata from eastern Pennsylvania with rocks in Sweden containing graptolites. The strata from Sweden can be correlated in turn on the basis of the graptolites with the Mazarn Shale. As outlined below, we have examined some samples from the Mazarn for conodonts and recovered a modest fauna. Although not conclusive, they are compatible with equivalence of the Mazarn and Lukfata and support the indirect correlation of these units already discussed.

If the sequence along Lukfata Creek is continuous as Pitt (1955) believed it to be, it is inverted with the Collier lying structurally above the Lukfata. According to this interpretation, the sandstone member of the Lukfata, situated at the base of the formation instead of at the top, corresponds in stratigraphic position to the Crystal Mountain Sandstone. The younger limestone and shale part of the Lukfata would have the same position in the stratigraphic sequence as the Mazarn. Thus the succession in the Broken Bow Anticline generally corresponds to that in Arkansas, but is overturned along Lukfata Creek.

MAZARN SHALE

Three limestone samples provided by the Arkansas Geological Commission were processed to obtain conodonts from the Mazarn. None of the samples produced an abundance of conodonts, but each one yielded at least some specimens that permit a tentative age assignment. Of the three collections, the only one that can be stated with confidence to be from the Mazarn is that one collected near Crystal Springs and south of Lake Ouachita (C S). The other two samples (J S and M P) may have been obtained from horizons in the lower part of the Womble Shale.³

Center sec. 34, T. 1 N., R. 19 W., about 3 miles east of Jessieville, Arkansas (J S).

- 1 Oistodiform element

³The samples of limestone obtained at sites J S and M P are presently considered from the lower portions of the Womble Shale by Boyd Haley, U. S. Geological Survey and Charles Stone, Arkansas Geological Commission, (personal communication, 1975).

- 4 *Oneotodus* sp.
2 *Scolopodus triplicatus* Ethington and Clark s. f.

Center north section line, sec. 33, T. 1 S., R. 20 W., about 2.5 miles northeast of Mountain Pine, Arkansas (M P).

- 3 *Gothodus communis* Ethington and Clark s. f.
4 *Oepikodus quadratus* (Graves and Ellison) s. f.
2 Oistodiform elements
1 *Scolopodus quadraplicatus* Branson and Mehl s. f.
2 New genus (?)

SE¼, NE¼, sec. 18, T. 2 S., R. 21 W., approximately 4 miles northeast of Crystal Springs, Arkansas (C S).

- 4 Acodiform elements
1 "*Drepanodus*" *subarcuatus* Furnish s. f.
3 Drepanodiform elements
1 *Oistodus* aff. *O. lanceolatus* Pander s. f.
5 *Oistodus* aff. *O. linguatus* Lindström s. f.
2 *Oistodus* aff. *O. parallelus* Pander s. f.
1 *Oneotodus* sp.
1 *Scolopodus quadraplicatus* Branson and Mehl s. f.
1 *Scolopodus* aff. *S. rex* Lindström s. f.

Scolopodus quadraplicatus s. f., which was found in the samples from near Mountain Pine and from near Crystal Springs respectively, is a widespread element in middle and upper Canadian faunas in North America. It commonly is associated with "*Drepanodus*" *subarcuatus* s. f. and *Scolopodus triplicatus* s. f. in these occurrences. *Gothodus communis* s. f. and *Oepikodus quadratus* s. f. are more restricted in their stratigraphic distribution and occur in the upper part of the range of *S. quadraplicatus*. Ethington and Clark (1971) designated this assemblage "Fauna E." The fauna occurs in the El Paso Group (Ethington and Clark, 1964) and in the Marathon Limestone (Graves and Ellison, 1941) of Texas. It is present in undescribed collections from the upper Fillmore Formation of western Utah and in the Ninemile Formation of central Nevada (Ethington, 1972). Berry (1960) earlier correlated the graptolites from the Mazarn identified by Ulrich (*in* Miser and Purdue, 1929) with his graptolite zone 4 of the Marathon Limestone. The conodont occurrences reported here support this earlier correlation of the Mazarn.

WOMBLE SHALE

The most productive samples in our collections from the Ouachitas are those from limestones in the Womble Shale. All of the samples produced at least a few conodonts, and most of them yielded conodonts in moderate abundance. More than one-half of our total collection came from one sample (CR-1-4) which was dissolved entirely (about 3 kg) and yielded over 2000 specimens. Distribution and abundance of the conodonts recovered from Womble samples is shown in Table 2.

Although the conodonts on which this study is based

were assembled exclusively from limestones, these fossils are not restricted in the Womble to rocks of that lithology. O. A. Wise of the Arkansas Geological Commission provided for us a collection of graptolitic shales from exposures near the Crystal Springs Landing road south of Lake Ouachita (SE corner, sec. 21, T. 2 S., R. 22 W., Garland County, Arkansas). Microscopic examination of the surfaces of the shale slabs revealed scattered depressions whose outlines demonstrated that they are molds of conodont elements. Generally the impressions are not defined well enough to permit specific identification, and only a few form types have been observed. Nevertheless the impressions are sufficiently abundant to make them potentially useful for the identification of Womble shales in the absence of other criteria and to establish the general stratigraphic position of the shales. Elements identified in the shales include *Periodon* sp., *Pygodus anserinus*, prioniodiniform elements, and "simple cones." These types of elements are present in the more diverse population recovered from limestones collected in the Womble in the area east of Lake Ouachita.

The conodont collections from the Womble limestones are dominated by the several elements of *Periodon aculeatus* Hadding, *Pygodus anserinus* Lamont and Lindström, and coarsely costate scolopodiform elements. These species are typical of the faunas that Sweet and Bergström (1972) have described as characterizing a North Atlantic Faunal Province. This province encompasses northwestern Europe and regions in the eastern Appalachians from Newfoundland to Alabama. Elements representing the North Atlantic Province also have been found in central Nevada (Ethington and Schumacher, 1969). Minor components of the Womble conodont fauna, including *Plectodina* sp., *Phragmodus* sp., and various hyaline form-genera, are representative of the North American Mid-continent Province of Sweet and Bergström. As the name suggests, these conodonts have been reported widely from the Middle and Upper Ordovician strata of central North America. The preponderance of elements representative of the North Atlantic Province indicates direct communication between the Ouachita depositional area and the eastern belt of the Appalachians and thereby to northern Europe during the time of Womble deposition.

Bergström (1971) proposed a zonation of the Middle Ordovician strata of eastern North America and Europe based on his extensive investigations of conodont occurrences in the two areas. Two distinct conodont faunas, each representing one of Bergström's zones, can be recognized within our collections from the Womble.

The older of the two faunas was recovered from a series of samples (BM-1-1--BM-1-13) collected in a large quarry in the west half of sec. 34, T. 1 S., R. 20 W., (Mountain Pine quadrangle). These strata have been interpreted by some workers as occurring in the lower part of the Womble Shale (Charles Stone, pers. commun., 1976). The lithology at this locality is bluish gray micritic limestone with "floating" quartz grains distributed through the rock. Some ledges show indications of graded bedding, and cross-lamination is developed at some levels. Both criteria indi-

cate that the limestone body is inverted. The beds dip steeply into the quarry face at about 60°; exposures of the limestone are confined to the area of the quarry. All of the samples from the quarry contained conodonts; specimens from samples collected in the highest part of the quarry are quite poorly preserved.

Periodon aculeatus Hadding is the most abundant component in the fauna from the quarry. It is associated with *Pygodus anserinus* Lamont and Lindström, *Scolopodus varicostatus* Sweet and Bergström, *Scandodus unistriatus* Sweet and Bergström and a variety of acontiodiform elements. *Polyplacognathus sweeti* Bergström, a distinctive platform species, together with *Belodina monitorenensis* Ethington and Schumacher, and ? *Paltodus* sp. of Ethington and Schumacher (1969, p. 468) are represented by occasional specimens. Although not abundant in any sample, these latter species are distributed throughout the entire succession in the quarry. All of these species occur together in European and North American strata that Bergström (1971) has assigned to the *Pygodus anserinus* Zone. Approximate correlations in North America include parts of the Valcour Formation (upper part of type Chazyan) in New York (Bergström, 1971, p. 121) and the Arline Formation in the type Porterfield Stage in Virginia (Bergström, 1971, p. 119). This characteristic fauna also is present in the lower part of the Copenhagen Formation of central Nevada (Ethington and Schumacher, 1969). Sweet and Bergström (1973, p. 355) have reported *P. sweeti* to characterize the lower part of the "Lower Bromide" in southern Oklahoma.

The fauna of the North American Mid-continent Province is represented in the collections from the quarry by a variety of hyaline (= fibrous of authors) elements including *Erismodus* sp., *Cardiodella* sp. and *Truchero-gnathus* sp., as well as occasional specimens of *Curtognathus* sp. and *Polycaulodus* sp. Somewhat more abundant are elements that represent a species of *Plectodina*. The distribution, both geographically and stratigraphically, in North America of the hyaline forms is not as well documented as is the case for the non-hyaline components of the fauna of the Mid-continent Province. The association of hyaline and non-hyaline forms in the samples of Womble from the quarry is suggestive of the fauna of the Joachim Formation in Missouri (Branson and Mehl, 1933; Andrews, 1967). However, the hyaline forms persist upward in the section of the Ozark region into the Platin and the Kimmswick Formations, although diminishing considerably in abundance above the Joachim.

The younger of the faunas that we have obtained from the upper portions of the Womble Shale was collected in outcrops along a northward flowing tributary to the Caddo River (NW¼, sec. 35, T. 3 S., R. 26 W., Athens quadrangle). These collections (CR-1-1--CR-1-5) were made about 4.75 miles west of Norman, Arkansas, and about 30 miles southwest of the quarry near Mountain Pine. The most productive lithology is a thin-bedded, laminated limestone with abundant fine-grained pyrite (Sample CR-1-4). The same fauna is present in a sample collected southwest of Caddo Gap (SW¼, sec. 13, T. 4 S., R. 25 W;

Sample MF-1).

Although many of the elements from these latter localities also are present in the collections from the quarry, key zonal indices are different in the two faunas. Particularly significant in the collections from near Norman and from near Caddo Gap are *Eoplacognathus elongatus* (Bergström) *Prioniodus gerdæ* Bergström and *Scolopodus* n. sp. cf. *S. insculptus*. Bergström (1971, p. 100) reported the mutual occurrence of these forms in the lower part of the *Prioniodus gerdæ* Subzone (the middle subzone of the *Amorphognathus tvaerensis* Zone) in his zonal scheme for the Middle Ordovician of Scandinavia and eastern North America. The *A. tvaerensis* Zone is the next zone above that of *P. anserinus*, thus indicating that these collections are younger than those from the quarry. Bergström (p. 101) reported the *A. tvaerensis* Zone to occur in North America in the lower part of the Rich Valley Shale and in the Lincolnshire, Botetourt and Edinburg Formations, all in the Appalachian Mountains of Virginia. Sweet and Bergström (1973, p. 355) indicate that *P. gerdæ* is present in the upper half of the "Lower Bromide" of Oklahoma.

The limestones sampled so far in the Womble Shale of Arkansas thus correlate on the basis of conodonts with the lower part of the Bromide Formation of southern Oklahoma. In a general way these limestones are equivalent to the middle member of the Copenhagen Formation of central Nevada which makes their age somewhat younger than White-rockian. The conodonts indicate that the Womble limestones can be correlated with at least the lower part of the Porterfield Stage at its type locality in Virginia (Bergström, 1971, p. 127).

The lower part of the Womble has been declared to be of Early Ordovician (Beekmantown) age based on graptolites collected near Hot Springs and identified by E. O. Ulrich (in Miser and Purdue, 1929, p. 33). Collections from near Jessieville and from near Mountain Pine, listed under Mazarn above but perhaps from Womble, may confirm this lower age limit. Upper Womble has been reported to include the Zone of *Nemagraptus gracilis* (Berry, 1960, p. 37-38). The conodonts from the Womble limestones include elements that occur in the eastern Appalachians in strata that also include the graptolites of the *N. gracilis* Zone. The traditional correlation of this part of the

Ouachita section thus is supported by the conodont evidence. Still younger graptolite zones in the Womble reported by Ulrich (in Miser and Purdue, 1929, p. 35) and by Berry (1960, p. 37-38) place the upper limit of the formation in the Trentonian Series. If the limestones of the formation are concentrated in the upper 100 feet or so of the Womble as has been assumed (Miser and Purdue, 1929, p. 32), the upper part of the formation is not as young as suggested by these correlations. The only sample in our collections that unquestionably came from the general level of the boundary between Womble and Bigfork Chert (Sample TFWC-1; SW¼, sec. 32, T. 2 S., R. 22 W.,) did not yield a diagnostic conodont fauna. Additional collecting in highest Womble and in the Bigfork should clarify the relationships in question.

CONCLUSIONS

1. The Collier Shale of Arkansas and Oklahoma is of Early Ordovician (Tremadocian) age rather than Cambrian as previously supposed. It is equivalent to widespread lower Canadian carbonates of the continental interior.

2. The Lukfata Sandstone of Pitt (1955) in the Broken Bow Anticline of Oklahoma is younger than the Collier instead of older. The carbonate member of the Lukfata probably is generally equivalent to some part of the Mazarn Shale of Arkansas. The sandstone member of the Lukfata corresponds in stratigraphic position to the Crystal Mountain Sandstone. The section along Lukfata Creek in the Broken Bow Anticline is inverted; it displays the same general succession of lithologic units that is found in the core of the Ouachita Mountains in Arkansas.

3. The few conodonts recovered from the Mazarn agree with the middle-upper Canadian age already assigned on the basis of graptolites.

4. Limestones in the upper Womble contain conodonts having European and eastern Appalachian affinities. These conodonts demonstrate that this part of the Womble is equivalent to the lower part of the type Porterfield Stage in Virginia and reaffirm earlier correlations based on graptolites.

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PLATE 1. CONODONTS FROM THE COLLIER AND WOMBLE SHALES, ARKANSAS AND OKLAHOMA

Figure

1. *Paltodus bassleri* Furnish; Collier, sample WC-2, UMC 1041-9, X 52.
2. *Loxodus bransonii* Furnish; Collier, sample WMC-3, UMC 1041-6, X 52.
3. *Cordylodus angulatus* Pander; Collier, sample OKB, UMC 1040-19, X 49.
4. *Chosonodina herfurthi* Müller; Collier, sample CC-2, UMC 1040-18, X 52.
5. *Acontiodus robustus* Hadding *sensu* Sweet and Bergström, 1962; Womble, sample CR-1-4, UMC 1040-13, X 21.
6. New genus; Collier, sample CC-2, UMC 1042-18, X 52.
7. *Acanthodus lineatus* (Furnish); Collier, sample CC-1, UMC 1040-9, X 52.
8. *Acontiodus propinquus* Furnish; Collier, sample OKB, UMC 1040-12, X 52.
9. *Walliserodus ethingtoni* (Fähræus); Womble, sample MF-1, UMC 1040-17, X 55.
10. *Acontiodus cooperi* Sweet and Bergström; Womble, sample CR-1-4, UMC 1040-11, X 46.
- 11, 19. *Scolopodus giganteus* Sweet and Bergström; Womble, samples BM-1-7 and CR-1-5 respectively, UMC 1042-10, UMC 1042-11, X 20 and X 17.
- 12, 13. *Scolopodus* n. sp. cf. *S. insculptus* (Branson and Mehl) *sensu* Bergström, 1972; Womble, sample CR-1-4, UMC 1042-15, UMC 1042-14, X 28 and X 33.
14. *Leptochoirognathus* sp.; Womble, sample CR-1-4, UMC 1041-5, X 47.
15. *Acodus similis* Rhodes *sensu* Hamar, 1966; Womble, sample CR-1-4, UMC 1040-10, X 46.
- 16 - 18, 20, 22, 24. *Periodon aculeatus* Hadding. 16, Roundyaform element; Womble, sample CR-1-4, UMC 1041-14, X 52. 17, Cladognathiform element; Womble, sample CR-1-4, UMC 1041-10, X 54. 18, Tortiliform element; Womble, sample BM-1-9, UMC 1041-15, X 61. 20, Falodiform element; Womble, sample CR-1-4, UMC 1041-11, X 41. 22, Prioniodiform element; Womble sample CR-1-4, UMC 1041-13, X 38. 24, Ligonodiniform element; Womble, sample CR-1-4, UMC 1041-12, X 40.
21. *Curtognathus* sp.; Womble, sample BM-1-6, UMC 1041-1, X 55.
23. *Erismodus* sp.; Womble, sample CR-1-4, UMC 1041-4, X 38.
25. *Cardiodella* sp.; Womble, sample CR-1-4, UMC 1040-17, X 53.

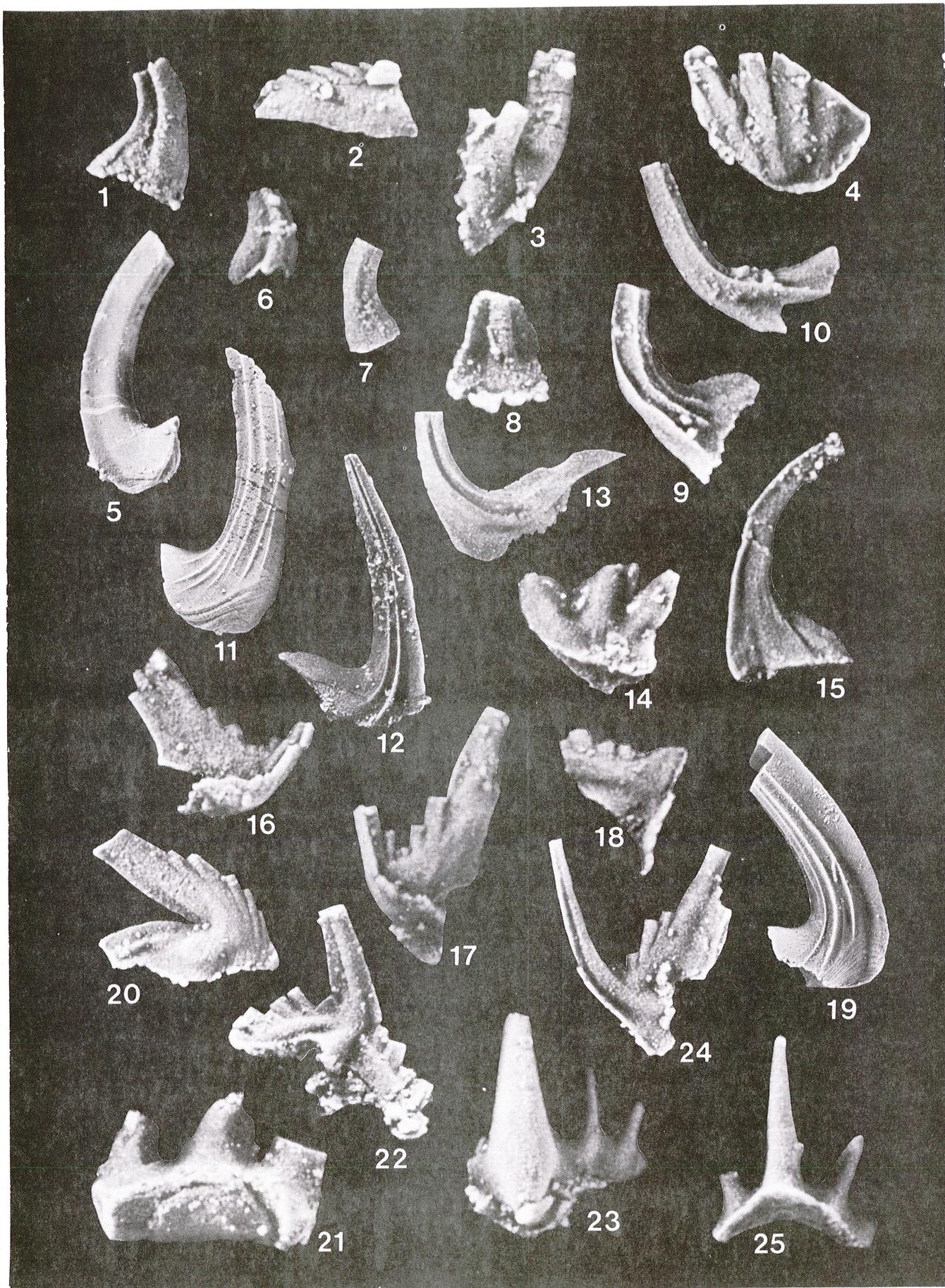


PLATE 1. CONODONTS FROM THE COLLIER AND WOMBLE SHALES, ARKANSAS AND OKLAHOMA

PLATE 2. REPRESENTATIVE CONODONT ELEMENTS FROM THE MAZARN AND WOMBLE SHALES

Figure

- 1, 4, 5, 7. *Prioniodus gerdae* Bergström; cordylodiform (UMC 1042-3), oistodiform (UMC 1042-4), prioniodiform (UMC 1042-5) and amorphognathiform (UMC 1042-2) elements respectively; Womble, sample CR-1-4, X 48, X 44, X 31, and X 36.
- 2, 8. *Eoplacognathus elongatus* Bergström; polyplacognathiform (UMC 1041-3) and ambalodiform elements respectively; Womble, sample CR-1-4, X 35 and X 44.
3. *Polyplacognathus sweeti* Bergström; ambalodiform element; Womble, sample BM-1-3, UMC 1042-1, X 35.
6. "*Tvaerenognathus*" sp.; Womble, sample BM-1-5, UMC 1042-16, X 50.
- 9, 10. *Pygodus anserinus* Lamont and Lindström; pygodiform (UMC 1042-7) and haddingodiform (UMC 1042-6) elements; Womble, samples BM-1-1 and BM-1-7 respectively, X 29 and X 42.
11. *Roundya pyramidalis* Sweet and Bergström; Womble, sample BM-1-8, UMC 1042-8, X 45.
- 12, 18. *Phragmodus* sp.; oistodiform (UMC 1041-17) and dichognathiform (UMC 1041-16) elements; Womble, sample CR-1-4, X 39 and X 43.
13. *Appalachignathus* sp.; fragment of eoligonodiniform element, Womble, sample BM-1-8, UMC 1040-14, X 65.
- 14, 16, 17. *Plectodina* sp.; zygognathiform (UMC 1041-20), trichonodelliform (UMC 1041-19) and cordylodiform (UMC 1041-18) elements respectively; Womble, sample CR-1-4, X 44, X 42, and X 41.
15. *Scolopodus quadruplicatus* Branson and Mehl; Mazarn, sample CS, UMC 1042-12, X 46.
19. *Oistodus excelsus* Stauffer *sensu* Hamar; Womble, sample CR-1-4, UMC 1041-18, X 50.
20. "*Cordylodus*" *spinatus* Hadding; Womble, sample CR-1-4, UMC 1040-20, X 23.
21. "*Belodella*" sp.; Womble, sample BM-1-7, UMC 1040-15, X 55.
22. "*Oepikodus*" *copenhagenensis* Ethington and Schumacher; Womble, sample BM-1-5, UMC 1041-7, X 55.
23. *Belodina monitorenensis* Ethington and Schumacher; belodiniform element, Womble, sample MF-1, UMC 1040-16, X 42.

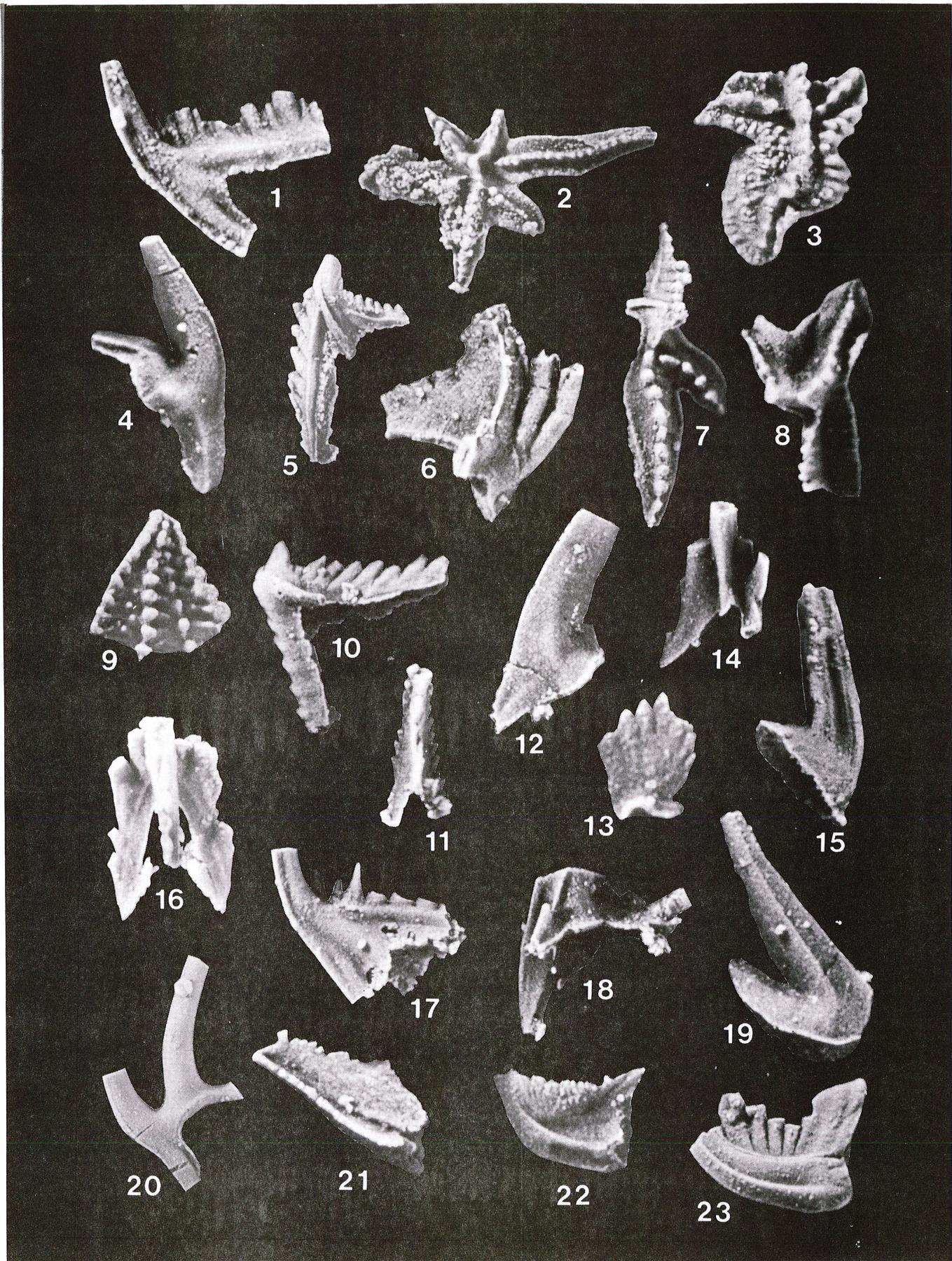


PLATE 2. REPRESENTATIVE CONODONT ELEMENTS FROM THE MAZARN AND WOMBLE SHALES

THE OCCURRENCE AND ORIGIN OF THE GRANITE --

META-ARKOSE ERRATICS IN THE ORDOVICIAN BLAKELY SANDSTONE, ARKANSAS

By Charles G. Stone¹ and Boyd R. Haley²

INTRODUCTION

The Blakely Sandstone of probable late Early Ordovician age was formally named by H. D. Miser (1917, p. 67) from outcrops on Blakely Mountain in Garland County, Arkansas (sec. 32, T. 1 S., R. 20 W.) as a result of the definitive work by A. H. Purdue and H. D. Miser (1923) on the Hot Springs District. They reported that the formation is about 500 feet thick and is composed of interbedded sandstone and shale. They also described several thick beds of conglomeratic sandstone in the Blakely, along the northern border of the Hot Springs quadrangle in eastern Garland County. Sandstone, limestone and chert clasts were found in the sandstone but no granitic fragments were recognized. Miser and Purdue (1929, p. 29) mapped the Blakely from the Hot Springs quadrangle west to the vicinity of Norman (Womble), in Montgomery County, Arkansas, but did not describe more conglomeratic sandstone. Recent mapping for the state geologic map has shown that the Blakely crops out from Norman west to near the western boundary of Montgomery County (Fig. 1).

DESCRIPTION OF BOULDER DEPOSITS

In 1943, H. D. Miser (written and verbal commun., 1962) and A. E. J. Engel found a small cobble of "anorthosite" in the Blakely Sandstone, in a newly excavated trench at the quartz mine on Miller Mountain (S½, sec. 2, T. 1 S., R. 21 W.) in northwestern Garland County (Fig. 1). Engel (1952, p. 189-192) describes many conglomerates and breccias in the Blakely of northern Garland County but does not refer to granitic rock or to the "anorthosite" cobble.

D. F. Holbrook (1955) investigated the Uebergang uranium prospect and adjoining properties along a small east-west trending ridge in secs. 3 and 4, T. 1 N., R. 15 W., in northern Saline County (Fig. 1). In this area, rounded boulders, as much as 2 feet in diameter, of vuggy, coarse-grained, quartz-feldspar rock are scattered on the surface. Samples of the radioactive rock were chemically analyzed by the U. S. Geological Survey and the radioactivity was ascribed primarily to an unidentified thorium mineral (maximum of 1.5 percent Th²³² and 0.019 percent U^{3O8}).

Prospect trenches on the Uebergang property in the SE¼ of sec. 3, were examined by P. J. Sterling, C. G. Stone, and D. F. Holbrook in 1959-60 while mapping the Benton

quadrangle. They found quartz-feldspar and feldspathic quartzite boulders, as much as 5 feet in diameter in the uppermost part of the Blakely Sandstone; in a zone of red clay residuum that was interbedded with altered "talcose" shale, siltstone and dense glossy gray-black chert. The quartz-feldspar rock was tentatively classified as igneous. Later work disclosed that the granite and feldspathic quartzite boulders occur in several beds in the Blakely Sandstone, along a narrow outcrop belt which trends from about 1/2 mile south of Paron eastward a short distance into Pulaski County. In the SW¼, SE¼, sec. 4, T. 1 N., R. 16 W., southeast of Paron (Fig. 1), in the middle of the Blakely Sandstone, an isolated mass of quartzitic sandstone about 75 feet long contains many feldspar-rich layers. At roadcuts along Arkansas Highway 9 north of Crows (sec. 1, T. 1 S., R. 17 W.) in central Saline County (Fig. 1), small partially decomposed granitic boulders, small gray-black chert masses, and some red clay residuum were observed, in 1960-61, in beds of decalcified brown siltstone and maroon shale at the top of the Blakely. Small, decomposed granite pebbles were found in the upper and middle parts of the Blakely at several localities between the Crows area and an area near Mountain Valley, in eastern Garland County. Samples from these sites were only slightly radioactive.

In 1960, a sample of the coarse-grained granitic rock from the Uebergang prospect was submitted to K. C. Jackson, University of Arkansas, for petrographic study. Jackson (written commun., 1962) reported that the rock contained microcline, microcline micropertite and minor quantities of oligoclase feldspar, fibrous amphibole and replaced mica. He concluded that the rock was a meta-arkose and that the original detritus was derived from a simple granitic terrane (see Jackson, this Volume). Stone and Sterling (1962, p. 388) briefly summarized the descriptions of the meta-arkose boulders in the Blakely Sandstone of the eastern Ouachita Mountains, Arkansas. Sterling and others (1966, p. 181) observed that the meta-arkose boulders in the Blakely were larger in northern Saline County, and suggested that they were derived from outcrops of Precambrian rocks during Blakely time. The Precambrian rocks probably formed submarine highs which were northeast of the main area of Blakely deposition.

In the fall of 1967, O. A. Wise, H. D. Miser and N. A. Sommers (Sommers, 1971, p. 23) investigated the boulder-bearing strata in the middle and upper part of the Blakely Sandstone at Point 50 on the eastern end of Lake Ouachita (NW¼, sec. 36, T. 1 S., R. 21 W.) in Garland County (Fig. 1). At this locality, meta-arkose and granite cobbles and boulders, as much as 5 feet in diameter, are embedded in an intricately folded limy conglomeratic sandstone which is 20-30 feet thick. Other clasts in the sandstone consist of limy siltstone, silty limestone, gray-black chert,

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EXPLANATION



Generalized outcrop of Blakely Sandstone

X

Location of granite-meta-arkose erratics

→Crows

Location described in text

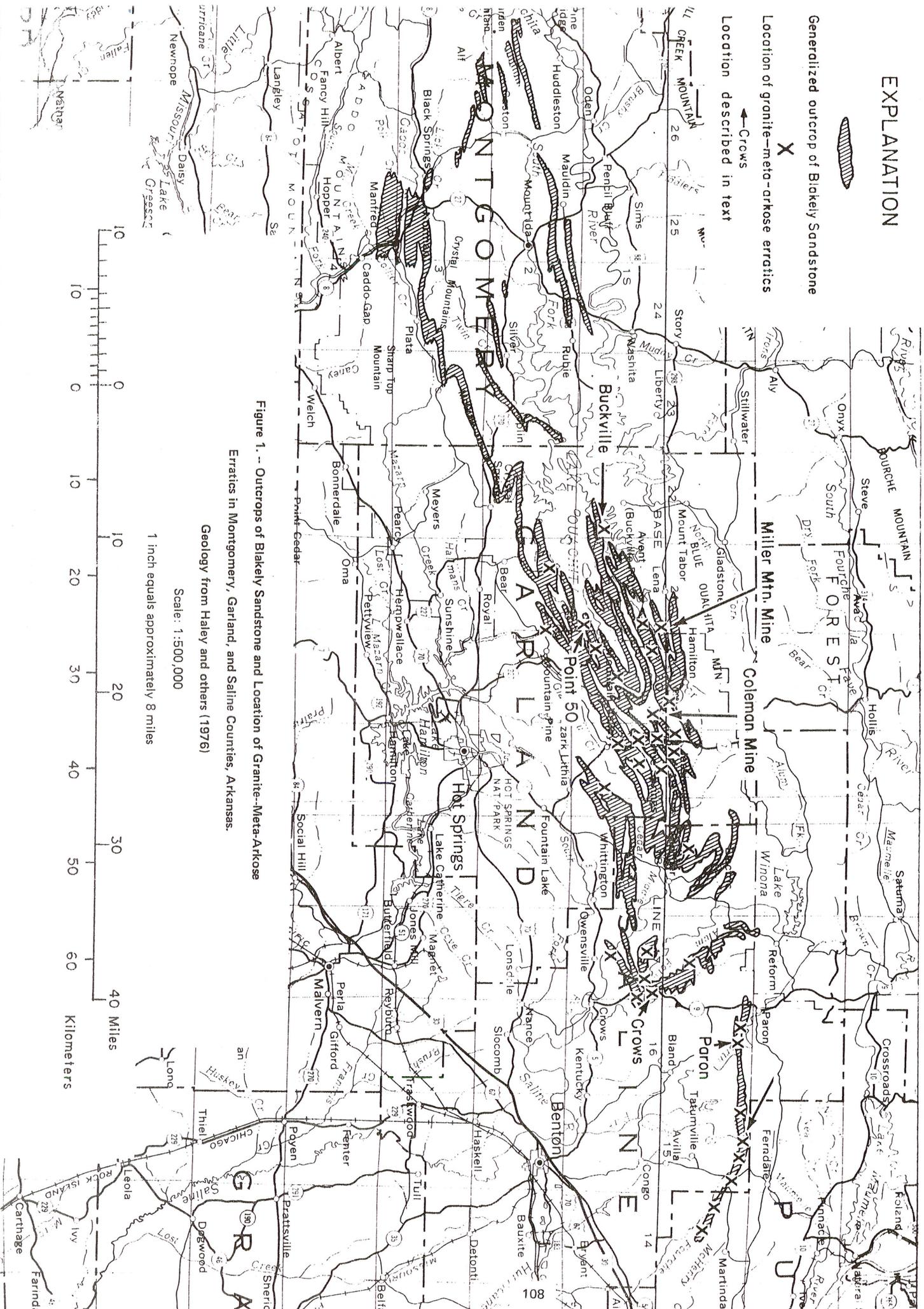


Figure 1. -- Outcrops of Blakely Sandstone and Location of Granite--Meta-Arkose Erratics in Montgomery, Garland, and Saline Counties, Arkansas.

Geology from Haley and others (1976)

Scale: 1:500,000

1 inch equals approximately 8 miles



Figure 2. -- Small, round to angular, light-colored, altered, granite, meta-arkose and arkosic siltstone erratics in the mélangé part of decalcified Blakely Sandstone, with dissecting milky quartz veins; along north wall of Coleman quartz mine near Blue Springs, in northern Garland County.

dense dolomitic limestone, limy sandstone, sandstone, quartzite, and one small cobble of very light colored, dense, fine-grained limestone which resembles updip (platform) Lower Ordovician strata.

In the spring of 1968, B. R. Haley, C. G. Stone, M. B. Woodward, and others examined thick mélangé zones mostly in the middle part of the Blakely Sandstone at the Coleman quartz mines (SW¼, NW¼, and the NW¼, SW¼, sec. 12, T. 1 S., R. 20 W.,) and at other quartz mines near the community of Blue Springs, in Garland County. The mélangé contains many round to subangular pebbles, cobbles and boulders of kaolinized, meta-arkose, arkose and granite, in a soft, weathered, sandstone along the walls of open pits (Fig. 2). The boulders at the Coleman quartz mine attain a maximum diameter of 45 feet, and their long axes generally parallel the bedding of the sandstone. These granite erratics are very similar to those in northern Saline County, except that a few contain abundant muscovite, some contain a distinctive smoky-gray quartz, and most seem more decomposed. Several rounded kaolinized granite cobbles, enclosed in meta-arkose and arkose boulders, were found at the Coleman mine.

In the fall of 1968, several geologists of the Arkansas Geological Commission and H. D. Miser revisited the Miller Mountain quartz mines, in northwestern Garland County, where Engel and Miser had earlier found the "anorthosite" cobble. A thorough examination of the small pits resulted in the discovery of many pebbles, cobbles and small boulders of kaolinized granite, meta-arkose and sedimentary rocks, but no anorthosite, in poorly exposed beds of decalcified conglomeratic sandstone. Miser then offered the opinion that the "anorthosite" cobble found in the earlier visit was probably a fragment of feldspar-rich granite.

While the authors and others were mapping the Ouachita Mountains for the new Arkansas State Geologic Map, granite and meta-arkose pebbles, cobbles, and boulders were discovered in several parts of the Blakely Sandstone in the region between Ferndale and the Montgomery-Garland County line (Fig. 1). Only a few kaolinitic pebbles and cobbles which may be weathered meta-arkose or granite were found in the Blakely in Montgomery County. Brieve (1963, p. 50) and Williamson (1973, p. 30) describe minor quantities of detrital feldspar in the Blakely west of Garland County, but they apparently did not find erratics. Brieve (1963, p. 68) also states that the non-tronite in shale of the Mazarn and Blakely was probably derived from the alteration of volcanic ash and glass.

Samples of granite from the Blakely Sandstone in northern Saline County were submitted to R. E. Denison of Mobil Research in 1970 for petrographic study and age determination. Denison (written commun., 1971) stated that these rocks are comparable to epizonal granites found in the south-central United States. He also reported that the age determinations were erratic, ranging from 283 to 489 + or - million years, and probably did not indicate true age (Denison and others, this Volume).

Studies by Williamson (1973), Williamson and Davies (1973) and Davies and Williamson (1973) (see Davies and Williamson, this Volume) describe two general types of sandstone in the Blakely. Williamson (1973, pp. 54 and 60) states that "the medium-grained sandstone was derived from a stable platform north of the Blakely outcrops and that the fine-grained sandstone and kaolinized granite and meta-arkose erratics were derived from a continental (sialic) source south of the outcrops. Perhaps Llanoria or Africa was the southern source of sediment."

CLASTS OF IGNEOUS AND METAMORPHIC ROCKS IN OTHER FORMATIONS IN THE OUACHITA MOUNTAINS

Small pebbles of granitic rock in limestone of the Early Ordovician Collier Shale were first reported by H. D. Miser and O. A. Wise (personal commun., 1961) in the SE¼, NW¼, sec. 35, T. 2 S., R. 26 W., in Montgomery County. We have found small fragments of granitic rock in limestone and conglomeratic limestone in the Collier Shale and the Early Ordovician Crystal Mountain Sandstone, at several localities in Montgomery County and western Garland County. Small cobbles of red phyllite were also found in a 10-15 foot thick bed of conglomeratic limestone in the upper Collier, on the south shore of Lake Ouachita near the Garland and Montgomery County line. Flores (1962, p. 21) and others have described the oolites and pellets associated with some of the limestones and the conglomeratic limestones.

The Collier and Crystal Mountain Formations do not crop out in the eastern Ouachita Mountains of Arkansas and, consequently, have been examined only in a small area. Near Broken Bow, Oklahoma, Honess (1923, pp. 45 and 49) described boulder and conglomeratic zones in the Collier, probable Collier, and Crystal Mountain Formations and stated that granitic gravel, arkosic fragments, and oolitic structures occur in the Collier limestone.

Some granitic erratics have also been tentatively identified in the black chert and shale breccias in the upper part of the Early Ordovician Mazarn Shale near Jessieville in northern Garland County.

Thin silty sandstone and conglomeratic (commonly phosphatic) sandstone or limestone in the Lower and Middle Ordovician Womble Shale of the Ouachita Mountains, are described by Honess (1923, p. 62), Miser and Purdue (1929, p. 32), Stone and Sterling (1962, p. 389-390), Lozano (1963, p. 9), Sterling and others (1966, p. 181), and Stone and others (1973, p. 52). Only scant igneous and metamorphic clasts have been reported in the Womble. The sill-like soapstone-serpentine masses in the Womble and Bigfork Formations of Middle and Upper Ordovician age in northern Saline County, Arkansas, are altered peridotite that was probably intruded in Middle Ordovician to Late Pennsylvanian time or slightly later (Sterling and others, 1966, p. 181).

Near Broken Bow, Oklahoma, Honess (1923, p. 210-212) describes a thin diorite sill intruded into the Womble sandstones. He further notes (p. 261) the highly fractured nature of this rock and indicates that it antedates the deformation in the area. The diorite sill thus was probably intruded in Middle Ordovician to Late Pennsylvanian time or slightly later.

J. Perrin Smith (Williams, 1891, p. 409) found decomposed igneous boulders (as much as 4 feet in diameter) in limestone and shale along small streams in secs. 23, 24, and 28, T. 1 S., R. 22 W., near Buckville (Avant), Arkansas, in

western Garland County. He concluded that these boulders represented five separate syenite dikes. Most of the boulders are now beneath Lake Ouachita, but we are assuming that they are granitic erratics in a conglomerate of the Blakely Sandstone or possibly of the lowermost Womble Shale.

A few biotite-rich granite pebbles were discovered by the authors in thick beds of conglomerate in the Missouri Mountain Shale of Silurian age, the Arkansas Novaculite of Devonian and Mississippian age, and the Stanley Shale of Mississippian age, near Goosepond Mountain in western Saline County and at other localities in northern Garland County. Because of the spatial distribution of the conglomerates, we propose that the granite pebbles were derived from submarine scarps or ridges that were north of Garland County along the continental slope of a craton.

Honess (1923, pp. 108, 126, 135, and 136) reported granite, porphyritic basalt, and volcanic ash fragments in conglomerates of the Missouri Mountain Shale and Arkansas Novaculite, in the Broken Bow area of Oklahoma. He concluded that the volcanic ash was deposited contemporaneously with some beds in these formations.

Acidic volcanic tuffs in the Mississippian Stanley Shale are described by Miser and Purdue (1929, p. 62-64), and others. The Hatton tuff occurs near the base of the Stanley and is thickest in the southwestern part of the Ouachita Mountains. Regional changes in the thickness and grain-size of the tuff are evidence that the volcanic source was south of the Ouachita Mountains of Arkansas. Thin tuff beds containing andesine plagioclase feldspar are also indicated in the Chickasaw Creek Member of the upper Stanley in the frontal Ouachita Mountains by Seely (1963, pp. 67 and 132), and others. It has been tentatively proposed by some workers that this volcanic material was derived from the south or southeast, but this has not yet been confirmed by studies in the southern Ouachita Mountains.

CONCLUSIONS

The spatial distribution of the clastics and erratic boulders in the Blakely Sandstone are not easily determined because: (1) the complex structure includes refolded and recumbent strata, thrust faults with displacements of at least twelve miles, and windows and klippen; (2) the boulder-bearing beds are generally calcareous, weather rapidly, and are therefore poorly exposed; and (3) most of the granitic clasts weather to kaolinitic clay, particularly where they are above the ground-water table.

The authors herein postulate that during deposition of the Blakely, clay, silt, fine-to coarse-grained sand, and siliceous and calcareous muds were carried southward across the continental shelf. Some of the sediment was deposited on the shelf and some was transported across the shelf to accumulate on the continental slope. Sediment was also transported through submarine canyons to spread southward and westward from the canyons by turbidity

and bottom currents. Sediment was possibly also carried from the east or southeast by these currents. The granitic clasts were eroded from outcrops along the edge of the shelf, from the sides of submarine canyons, or from submarine ridges or highs near the site of deposition. The boulder-bearing strata of the Blakely have resulted from the mixing action that occurred during submarine slumps, slides, and in turbidity currents which moved down and along the continental slope.

The authors believe that these granitic rocks are both igneous and metamorphic in origin and are probably of Precambrian age, but possibly of Cambrian age, and that the source area of these rocks is now buried beneath the northward-thrusted Ouachita Mountains. In all probability the granite--meta-arkose fragments in the other Ordovician formations were emplaced by the same sedimentary processes and are from the same general source area. The possible volcanic detritus in the Blakely and other formations cannot be ignored but the eruptive centers, if present, were not in the area of deposition, because the soapstone-serpentine in Saline County, Arkansas, and the diorite sill near Broken Bow, Oklahoma,

are the only known intrusives of probable Paleozoic age in the Ouachita Mountain region.

The igneous and volcanic rock fragments described by Honess (1923, pp. 108, 126, and 135) in the Missouri Mountain Shale and Arkansas Novaculite of Silurian-Devonian-lowest Mississippian age, in the Broken Bow area of Oklahoma, are evidence of either a new source area or of rejuvenated igneous activity in portions of the older Blakely source area. Since no similar fragments have been reported in these formations to the north or northwest in the Ouachita Mountains, in the Black Knob Ridge or Potato Hills of Oklahoma (Hendricks and others, 1937, p. 11-12), it seems that another younger source area existed to the south. The minor fragments of granite found in conglomerates in the north central Ouachita Mountains of Arkansas, are related to submarine detachments derived from a northern source, from either the older metamorphic-igneous complex or possibly from concurrent igneous activity. The areal distribution of the Hatton Tuff, in the lower Stanley Shale of Mississippian age, indicates that a volcanic area was southwest of the depositional site during the Mississippian Period.

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META-ARKOSE BOULDER FROM BLAKELY SANDSTONE (ORDOVICIAN),

BENTON QUADRANGLE, ARKANSAS

By Kern C. Jackson¹

INTRODUCTION

During the Ouachita Symposium Sessions at the 7th Annual Meeting of the South-Central Section of GSA held at Little Rock, Arkansas, April 5-7, 1973, the subject of certain "granite" or "meta-arkose" boulders from the eastern Ouachita region came up repeatedly. Inasmuch as the "meta-arkose" interpretation was based on a petrographic description by me, the chairman of the symposium asked me to submit a description of that material for this publication. The petrographic study was based on a single thin section and two hand specimens collected by the Arkansas Geological Commission from boulders at the Uebergang uranium prospect in the Benton Quadrangle, northern Saline County, Arkansas. The following location and field description of the occurrence is by C. G. Stone.

LOCATION AND FIELD DESCRIPTION

The Uebergang uranium prospect consists of very small trenches and pits made in 1955 on a radioactive anomaly primarily along the south flank of a small hill in the SE¼, sec. 3, T. 1 N., R. 15 W., about 1000 feet north of the dirt road (old Gray Hill School area) in northern Saline County, Arkansas.

Numerous vuggy quartz-feldspar and some feldspathic quartzite boulders, up to 5 feet in diameter were exhumed at the site from a decalcified (?), red clay residuum in a sequence of black, glossy chert, siltstone, and gray "talcose" shale apparently at the top of the Blakely Sandstone of probable Lower Ordovician age. A thorium mineral in the boulders and clay was indicated by the U. S. Geological Survey as the primary cause of radioactivity (maximum of 1.5 percent Th^{232} and 0.019 percent U_3O_8).

MEGASCOPIC DESCRIPTION

The rock is moderately weathered, friable, very pale orange but locally iron stained, coarse grained granitic textured material. Quartz and a non-striated feldspar are the only recognizable minerals. The quartz occurs as colorless, apparently anhedral, single grains or clusters of grains up to 1 cm in maximum dimensions. Feldspar appears as subhedral blocky crystals 2x5 to 5x10 mm. It is turbid and shows some alteration but some cleavages are still lustrous. Many cleavages, however, are partially covered with a film of alteration products. No ferromagnesian

minerals are identifiable megascopically but there are some small dark grains. The limited amount of iron staining would indicate either a very low initial ferromagnesian content or thorough removal. About five percent of the rock consists of partially filled interstitial voids ranging from 1 to 4 mm. Most of these voids are straight sided with shapes controlled by adjacent feldspar and quartz. The voids are partially filled by a boxwork aggregate of fine, white, granular material which disaggregates between the teeth to a pasty aggregate of clay sized material with abundant silt. On a fresh break, however, the material has considerable cohesion and maintains its porous boxwork texture.

MICROSCOPIC CHARACTERISTICS

The thin section shows a wider range of grain size than the hand specimen would indicate. The slide is dominated by large subhedral to anhedral blocky feldspars and large rounded to irregular quartz grains, but in addition there are numerous patches of silt sized materials in two occurrences. First there are irregular patches of coarse silt to fine sand sized material, dominantly quartz but with some recognizable microcline and zircon grains. These grains are imbedded in abundant opaque material and are associated with extensive iron staining. The second occurrence is the megascopically recognizable fine grained interstitial material. Most of this material has been lost in grinding the thin section but thin rims of very finely granular rhombohedral carbonate were left in some of the cavities.

Quartz appears as clear anhedral with abundant inclusions. The crystals are quite angular but tend to be equidimensional. Grain boundaries vary from smooth curved surfaces to highly sutured boundaries. Liquid filled cavities, many with bubbles, are abundant as trains, most as the result of healed fractures. There are at least two generations of fracturing in that one set is iron stained and contains ameboid shaped cavities whereas others lack staining and contain smaller more regularly shaped cavities, some approaching negative crystal shape. Locally quartz has been replaced or penetrated by a fibrous mineral probably in the cummingtonite-grunerite series.

The most important feature of the quartz is the presence of well developed Boehm lines which show up as polysynthetic twin-like patterns when the grain is near extinction. Most grains show only one set of Boehm lines, however some show two sets. The orientation of twenty five sets of lines were measured relative to the long dimension of the thin section and these show a definite preferred orientation as indicated in the accompanying pie diagram (Fig. 1). Undulatory extinction is present in the quartz but is not universal nor markedly developed.

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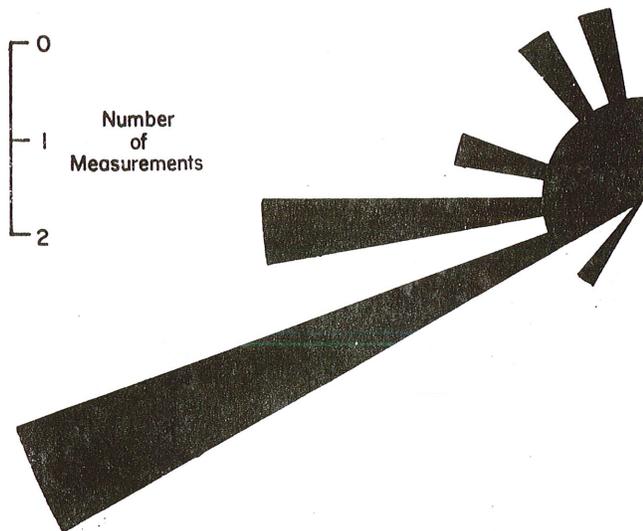


Figure 1. -- Pie diagram showing orientation of Boehm lines in quartz relative to long dimension of unoriented thin-section.

Quartz in the silt patches consists of fine sand to coarse silt sized polygonal single grains. The grains fit together with smooth curved contacts and no suturing of boundaries. Many single grains are outlined by films or trains of granules of opaque minerals and locally opaque materials are more abundant than quartz grains. The large feldspar crystals are anhedral against these silt patches and some of the feldspars contain abundant quartz grains identical to those of the silt patches. Chert is also present in the slide as an apparent replacement of original mica flakes. The largest such flake, approximately 1 mm long in the section, is represented by three patches of chert in which the original cleavage of the mica is recognizable in the chert under plane polarized light.

Microcline, microcline microperthite, and minor oligoclase (An ca. 12 percent) are the feldspars present. Microcline appears as large blocky to irregular rectangular crystals with well developed tartan twinning. Albite twinning is relatively coarse with well defined twin boundaries. Extinction to the albite twinning on 001 cleavage is 15° . Pericline twinning is broader and more diffuse with extinction angles to these twins on 010 cleavages of 5° . Microcline microperthite is identical in its twinning characteristics to microcline but includes irregular patches of plagioclase elongated in the 100 plane. The plagioclase is untwinned in the perthite, but birefringence, relief and extinction angles relative to the surrounding microcline would indicate a sodic oligoclase composition. Plagioclase as a separate phase occurs only rimming nonperthitic microcline. It is in parallel orientation to the adjacent microcline as indicated by the continuity of albite law twins across grain boundaries. Extinction angles indicate a composition near An_{12} . Inclusions in the potassium feldspars are abundant as gas filled cavities, shreddy alteration products, iron staining, and quartz grains. Silt sized quartz grains with random orientation are abundant in the margins of some microperthite crystals particularly in those portions of the feldspars which are close to the silt patches.

The original ore minerals seem to have been pyrite with minor amounts of titaniferous magnetite or ilmenite but now are completely altered to limonite and two grains of leucoxene. The ore minerals are the only primary ferromagnesian mineral recognizable. No pyrite remains in the slide but well developed cubic outlines or aggregates of tiny cubes of limonite are readily recognized. The pyrite was apparently euhedral and authigenic and was most abundant in the residual siltstone patches. Some limonite has crystallized to recognizable goethite crystals showing crossed axial plane dispersion.

A fibrous mineral, tentatively identified as a member of the cummingtonite group, occurs as fibers and tufts. The mineral is colorless but iron stained at the margins of tufts, anisotropic with interference colors to the first order yellow, maximum extinction angle of 17° , and consistently length slow. Most of this material occurs in and along the margins of the silt patches but the fibers penetrate into adjacent quartz and potassium feldspar grains.

Zircon occurs as subhedral to rounded grains and is particularly abundant in the silt patches. One such patch contains thirteen zircons in the section. A rhombohedral carbonate marginal to holes in the slide represents the white granular boxwork in the hand specimen. It is very fine grained and granular. Both indices of refraction are greater than balsam and it does not effervesce with cold HCl, therefore it is tentatively identified as siderite or ankerite.

PETROGENESIS

Evidence for a metamorphic origin of the rock includes: (1) Boehm lines in quartz; (2) character of the alkali feldspars; and (3) presence of the fibrous amphibole. Evidence for the sedimentary character of the rock prior to metamorphism include: (1) fine sand and coarse silt patches; and (2) relict rounded quartz sand grains with secondary enlargement in optical continuity. Also suggestive of a sedimentary origin is the paucity of ferromagnesian silicates and the abundance of limonite after pyrite.

Boehm lines in quartz are the result of strain and are most common in metamorphic rocks. They are rare in igneous rocks unless the rock has been metamorphosed. They may be inherited from a parent metamorphic rock by quartz clasts in sediments, but would show random orientation. The definite preferred orientation of the Boehm lines in the slide is clear evidence of metamorphism.

At a few places in the slide quartz grains are rimmed by a train of opaque grains. Such quartz grains are somewhat smaller than average for the rock and are rounded rather than angular. In one large quartz grain such a rounded train of inclusions is visible within the grain and indicates a sand sized quartz clast with secondary overgrowth probably developed during metamorphism rather than a sedimentary enlargement.

The feldspars appear to be porphyroblasts developed from clastic grains. Microcline partially rimmed by

oligoclase and microcline microperthite probably represent two different initial feldspars on the basis of the apparent ratio of volumes of microcline to plagioclase. No statistical measurements were made but the plagioclase content of microperthite appears to be significantly higher than that associated with nonperthitic microcline. The microcline-oligoclase association may have been a single phase of solid solution which underwent exsolution and migration of the plagioclase component during metamorphism. The plagioclase partial rims on microcline are clearly not from an earlier (pre-clastic) origin as they show no abrasion. The fact that the feldspars have grown as porphyroblasts is clearly indicated by the inclusion of silt sized quartz grains at their margins. Chertification of mica flakes was probably the result of the porphyroblastic growth of feldspars requiring additional potassium to upgrade clay minerals.

The silt patches are clearly clastic and cannot be a granulation product. There is some granulation along contacts between the larger crystals and this results in material of

a very different appearance from the silt patches. The silt is dominantly quartz with minor feldspar and a high concentration of zircon, cemented initially by a highly ferruginous cement. The fibrous amphibole has developed by reaction between the matrix of the silt patches and quartz. The high iron content of the matrix suggests a member of the cummingtonite-grunerite series and thus suggests siderite or ankerite in the original matrix. Optically the amphibole fits into the tremolite-actinolite series which would suggest initial dolomite. In either case the amphibole is of metamorphic origin.

The rock, thus, is metasedimentary and developed from an arkose. The arkose was derived from a simple granitic terrane, was moderately to poorly sorted, and accumulated in a reducing environment. These conclusions are based on the simple accessory mineral assemblage of zircon and possibly cassiterite, the coarse silt to medium sand size of recognizable clasts, and the abundance of pyrite. Metamorphism was probably into the lower amphibolite facies.

PALEOENVIRONMENTS AND PALEOBATHYMETRY OF
LOWER PALEOZOIC CRYSTAL MOUNTAIN AND BLAKELY FORMATIONS,
OUACHITA MOUNTAIN CORE

By David K. Davies¹ and Eddie A. Williamson²

ABSTRACT

Sedimentary structures and textures of the Ordovician Crystal Mountain and Blakely Formations, Arkansas, indicate that the widely accepted deep-water origin of these two formations is sedimentologically untenable. Sandstones of these two formations are well-sorted, medium- to fine-grained orthoquartzites which occur either as amalgamated beds with little or no finer sediment between (Crystal Mountain) or as individual beds separated by clay and silt (Blakely). Bouma sequences are absent, grading occurs in only 15 percent of the beds, and the majority of upper surfaces of sandstones show evidence of penecontemporaneous erosion. Parallel lamination is the most common internal sedimentary structure, although some beds may show very long, low-angle (5°) foresets or short, high angle (15°) foresets. These sandstones are interpreted as sublittoral sheet sandstones which were deposited in a shallow marine shelf association. The sandy Crystal Mountain Sandstone represents deposition in near-shore inner-shelf environment; the flysch-like Blakely Sandstone represents deposition in delta-front environments. Conglomeratic deposits which occur at various levels within the Blakely and locally in the Crystal Mountain are the result of debris flows from fault scarps produced during rifting associated with basin subsidence.

The Crystal Mountain and Blakely Formations were deposited in a shallow marine basin which was subsiding at a rate in approximate balance with sedimentation rate. The provenance for sandstones in these formations probably was not the North American continent. Rather, these sediments may have been derived from a southerly or southwesterly source, and deposited at shelf depths on the southern flanks of the lower Paleozoic Ouachita basin.

INTRODUCTION

An understanding of the detailed sedimentology of Ouachita rocks is critical to any paleogeographic reconstruction of the lower Paleozoic southern margins of proto-North America. Various plate tectonic models have been erected to explain the evolution of the southern margins of this continent. (Keller and Cebull, 1973; Morris, 1974). Central to modern plate tectonic models is the inference that "in the general area of the Ouachita Trough . . . a shelf, slope, and abyssal plain extended southward from the North American craton during the early and middle Paleozoic" (Morris, 1974, p. 133). Lower Paleozoic sediments are presumed to have accumulated in slope and abyssal environments as a result of deposition from turbidity currents and through differential pelagic settling. Modern plate tectonic models demand deep-marine paleoenvironments for Ouachita lower Paleozoics. However, a large volume of literature exists which repudiates, either in general or detail, this deep-marine origin for lower Paleozoic rocks of the Ouachita trough (Barton, 1945; Brieve, 1963; Eardley, 1951; Flores, 1962; Goldstein, 1961; Honess, 1923; Lozano, 1963; Miser and Purdue, 1929; Severson, 1963. Indeed, various authors have concluded that lower Paleozoic rocks of the Ouachita trough were

deposited in neritic and related shallow marine environments (Brieve, 1963; Flores, 1962; Folk, 1970; Goldstein, 1961; Miser and Purdue, 1929).

Much of the earlier paleoenvironmental work in the Ouachitas suffered because of a lack of information concerning mechanisms responsible for transportation and deposition of coarse detritus in deep ocean basins. Recent advances in the general area of deep marine sedimentation (ably summarized by Middleton and Bouma, 1973) have given the sedimentologist criteria necessary to interpret the varied deposits of the dominant mechanisms of deep ocean sedimentation.

This paper focuses attention on the sedimentary structures, textures, and composition of sand-size and coarser detritus in two Ordovician formations of the Ouachita Mountains--the Crystal Mountain and Blakely Formations (Table 1). Of the lower Paleozoic formations in the Ouachitas, these two contain the largest volume of sandstone and conglomerate. These lithologic varieties are the most likely to yield data on sediment mechanisms, paleoenvironments, and paleobathymetry in this sparsely fossiliferous lower Paleozoic sequence.

Lower Paleozoic rocks of the Ouachita Mountains form a part of a belt (Fig. 1) which stretches from far West Texas through Oklahoma and Arkansas, through the Appalachian Mountains to Great Britain and Scandinavia.

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TABLE 1. -- LOWER PALEOZOIC STRATIGRAPHIC COLUMN, OUACHITA MOUNTAIN CORE (Thicknesses courtesy of Charles Stone and Boyd Haley).

SYSTEM	FORMATION	AVERAGE THICKNESS (m)	MAXIMUM THICKNESS (m)
SILURIAN	Missouri Mountain	30	75
	Blaylock	9	450
ORDOVICIAN	Polk Creek	22	62
	Bigfork	195	255
	Womble	480	1,200
	Blakely	120	<180
	Mazarn	540	1,050
	Crystal Mountain	210	270
	Collier	<300	>300

The lower Paleozoic sediments of West Texas, the Appalachians and Great Britain have been the subject of considerable study. Relatively little work has been carried out on lower Paleozoic strata of the Ouachita Mountains in Arkansas. Previous investigations have yielded a generalized stratigraphic column for the area (Table 1), but the petrology of the various units and their structural interrelationships are as yet known only in general, for the area is extensively faulted and folded, outcrops are small and scattered, and well data are nonexistent. Due to scanty paleontological evidence, faunal or time lines can be inferred on only the most general of scales. The sedimentary rocks of the Ouachita Mountains have been subjected to diagenesis and low-grade metamorphism.

The lower Paleozoic rock sequence in the Ouachita Mountain core is dominantly detrital, and limestones, sandstones, and shales commonly display a flysch-like aspect. Thin to thick beds of gray to light brown sandstone occur interbedded with dark gray to black shales through much of the rock sequence. It is this overall flysch-like aspect of the rocks that has lent weight to hypotheses of their deep-marine origins (Morris, 1974, p. 124). The shallow water origin of the same rocks has been hypothesized largely because of the quartzose nature of the sandstones, and the frosted, well-rounded and well sorted character of the quartz grains. Unfortunately, flysch-like rocks do not always originate in deep marine environments, and frosted, mature quartz arenites do not, of necessity, characterize shallow marine environments. We feel that previous investigators have relied too heavily on characteristics which are non-diagnostic of both environment and bathymetry. Flysch-like rock sequences can occur in deltaic, shelf, slope, and abyssal environments. Quartz arenites owe their composition and grain textures largely to their source and transport history, and not to their ultimate site of deposition.

Considering the degree of confusion surrounding the paleobathymetric status of Ouachita lower Paleozoic rocks, it is critical at this juncture to examine the detailed

sedimentologic evidence preserved and to attempt paleogeographic reconstruction consistent with plate tectonic theory.

METHODS

Seven Blakely Sandstone and three Crystal Mountain Sandstone exposures were chosen for sampling (Fig. 1 and Table 2). Each exposure was identified as belonging to either the Crystal Mountain or Blakely on the basis of extensive field mapping by Boyd Haley and Charles Stone (personal communication). These specific localities were selected because each is large and well-exposed both vertically and laterally (when compared to other Ouachita outcrops), and each shows little apparent internal strain owing to folding. Haley and Stone (personal communication) consider each locality to be typical of the formation it represents. Because of complex regional deformation, and because paleontologic evidence is too scanty to allow detailed faunal or time-line correlation, intraformational correlation between outcrops is impossible.

Each bed in five localities (A,E,K,P, and S, Fig. 1, Table 2) was analyzed sedimentologically in detail, beginning with the stratigraphically lowermost exposed sandstone or shale bed and continuing through the uppermost exposed bed. Features or information for which specific search was made in each of 283 beds included:

1. Lithology (sandstone, shale)
2. Thickness
3. Grain size (Wentworth scale)
4. Sorting
5. Graded bedding (present, absent, normal, reverse, multiple)
6. Nature of upper and lower contacts (gradational, sharp, erosive)
7. Intrastratal sedimentary structures (cross-bedding, cross-lamination, parallel laminations)
8. Interstratal sedimentary structures (sole marks,

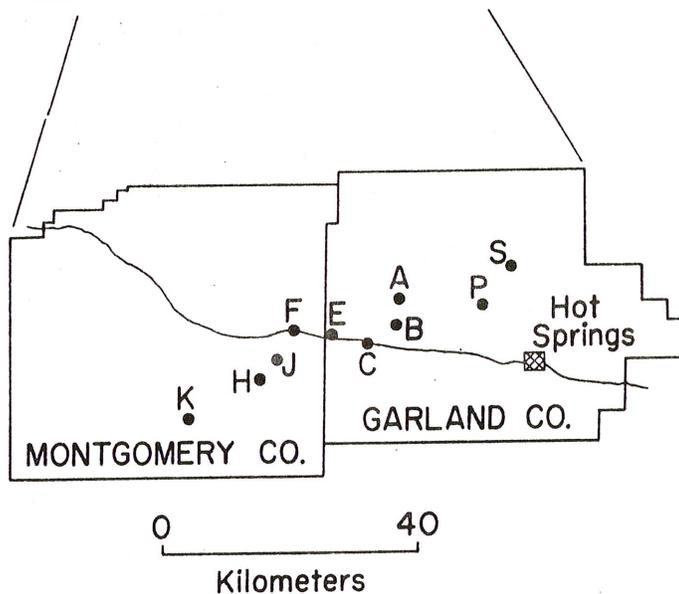
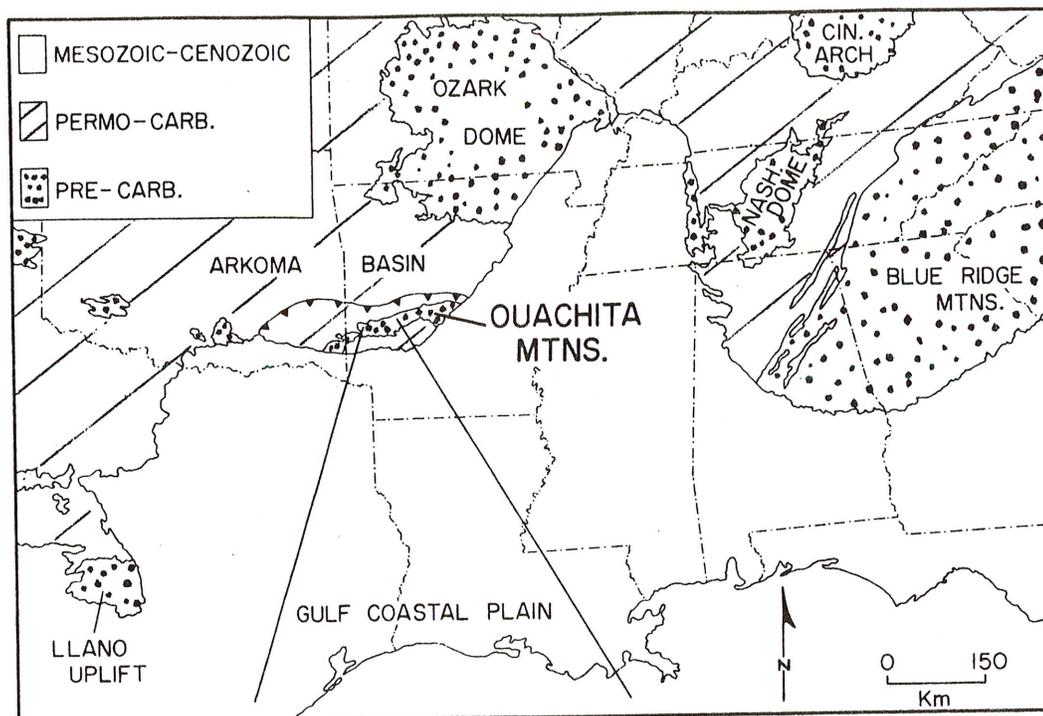


Figure 1. -- Localities studied (see Table 2 for coordinates).

TABLE 2. -- LOCATIONS OF SAMPLE LOCALITIES IN FIGURE 1

OUTCROP. FM .	LOCATION
A	. CM . NW¼, SW¼, Sec. 16, T. 2 S., R. 21 W.
B	. BI . S½, NW¼, Sec. 20, T. 2 S., R. 21 W.
C	. BI . SW¼, SE¼, Sec. 32, T. 2 S., R. 22 W.
E	. CM . SE¼, NW¼, Sec. 31, T. 2 S., R. 22 W.
F	. BI . Center NE¼, Sec. 27, T. 2 S., R. 23 W.
H	. CM . SE¼, NW¼, Sec. 9, T. 3 S., R. 24 W.
J	. BI . NW¼, SE¼, Sec. 35, T. 2 S., R. 24 W.
K	. BI . NE¼, SE¼, Sec. 35, T. 3 S., R. 25 W.
P	. BI . SE¼, Sec. 12, T. 2 S., R. 25 W.
S	. BI . NW¼, NW¼, Sec. 33, T. 1 S., R. 20 W.

CM=Crystal Mountain Sandstone BI=Blakely Sandstone

- ripples, dunes, lebenspürren)
9. Bioturbation
 10. Vertical change in sedimentary structures either within or between beds.
 11. Lateral variation (thickness, sedimentary structures)
 12. Texture of individual sand grains (rounding, overgrowths, frosting, etching, polishing).

Samples for thin sections were taken from ten percent of the sandstone beds, plus a smaller number of the shale beds at all 10 localities. In addition, beds which exhibited unusual features were purposively sampled. In conglomeratic zones of the Blakely Sandstone, samples were taken of both "erratics" and the sandstone matrix.

A standard thin section was made from each sample. In each thin section the long axes of fifty randomly selected quartz grains were measured. The arithmetic mean of these measurements is used as an expression of grain size for each sample. A further discussion of this technique is given by Griffiths (1968). The composition of each sample was determined by point-counting 200 points per thin section.

FORMATION DESCRIPTIONS

Crystal Mountain Sandstone

The Crystal Mountain Sandstone has a variable field aspect. Commonly and characteristically this formation consists of thin-to thick-bedded, fine- to medium-grained orthoquartzites. Shale interbeds in some outcrops may be present, and in others may comprise up to 50 percent of an exposure (Fig. 2). The base of the Crystal Mountain Sandstone is marked by a conglomerate which forms a massive bed ranging from about 3 - 4.3m thick. It consists of pebbles and granules of quartz, chert, limestone, shale and argillite in a matrix of sandy limestone or calcareous sandstone. In the De Queen and Caddo Gap quadrangles of western Arkansas, Miser and Purdue (1929, p. 25) noted that the "limestone and chert pebbles are identical in character with the limestone and chert in the Collier Shale and were probably derived from that formation." Recent studies (George Viele, personal communication, 1974) have revealed that such conglomerates are not restricted to the base of the Crystal Mountain Sandstone. Indeed there appear to be several additional conglomeratic units at ill-defined levels within this formation.

Individual sandstone beds in the Crystal Mountain Sandstone extend laterally at least 30m, and can range widely in thickness (Table 3). Bed termination is often a result of erosion preceding the deposition of succeeding beds. The bases of sandstone beds are sharp, usually flat to undulose, with occasional evidences of erosion. The most common internal stratification is flat, or at very low angle (~5°) (Figs. 2B and 3). Bouma sequences are absent in all beds at localities A and E. The upper portions of sandstone beds are not always completely preserved, having been removed by penecontemporaneous erosion. Individual sandstone beds may be graded, the greatest grain size change occurring near the top of individual beds (Table 4).

TABLE 3. -- BEDDING DATA FOR LOCALITIES A AND E OF THE CRYSTAL MOUNTAIN SANDSTONE (For locations, see Figure 1 and Table 2).

OUTCROP FEATURES	OUTCROP	
	A	E
Outcrop Thickness	. 341 cm	. 1,103 cm
Number of Beds	. 27	. 16
Percentage Sandstone Beds	. 97%	. 84%
Sandstone Bed Thickness	range . 3-84 cm	. 27-204 cm
	average . 23 cm	. 100 cm
Shale Bed Thickness	range . 1-10 cm	. 10-69 cm
	average . 4 cm	. 28 cm

TABLE 4. -- INTERSTRATAL AND INTRASTRATAL SEDIMENTARY CHARACTERISTICS OF SANDSTONE BEDS IN LOCALITIES A AND E OF THE CRYSTAL MOUNTAIN SANDSTONE

SEDIMENTARY CHARACTERISTICS	PERCENTAGE OF SANDSTONE BEDS DISPLAYING EACH CHARACTERISTIC	
	OUTCROP A	OUTCROP E
Graded Bedding (normal, well sorted)	. 44%	. 0%
Parallel Lamination	. 44%	. 67%
Cross Lamination	. 11%	. 67%
Bouma Sequence (C-D-E)	. 0%	. 0%
Erosive Base	. 33%	. 33%
Gradational Base	. 0%	. 0%
Sharp, Non-Erosive Base	. 67%	. 67%
Load Casts, Flute Casts, Grooves	. 0%	. 0%
Gradational Top	. 11%	. 17%
Sharp Top	. 89%	. 83%



A



B



C



D

Figure 2. -- Variable field aspect of the Crystal Mountain Sandstone. A. Thin-bedded sandstones; no shale interbeds (Outcrop H). B. Medium-bedded sandstones, with low angle ($\leq 5^\circ$) cross bedding, and thin shale interbeds (Outcrop A). C. Thin- to medium-bedded sandstones, with shale interbeds of varying thickness (Outcrop A). D. Medium- to thick-bedded sandstones interbedded with medium- to thick-beds of shale (Outcrop E).



Figure 3. -- Parallel laminated, very thin beds of alternating sandstone and shale, characteristic of shale interbeds at Outcrop E, Fig. 2D.

Fossils are generally absent. Sandstone beds are amalgamated where interbedded shale is rare to absent.

The two localities which yielded data for Table 4 are end members of a lithologic series within the Crystal Mountain Sandstone. This series is characterized by variations in the relative abundance of sandstone and shale beds. Locality A is a typical shale-poor, sandstone-rich example; locality E is a more shale-rich, sandstone-poor example. Through sampling both localities we hope to have maximized significant sedimentological variance which may exist within the Crystal Mountain Sandstone.

Three schools of thought have placed the Crystal Mountain Sandstone in three different paleoenvironments: (1) eolian (Hones, 1923); (2) shallow marine--epicontinental

neritic (Flores, 1963), littoral (Goldstein, 1961), and beach (Miser and Purdue, 1929); and (3) deep marine (Haley and Stone, 1973; Morris, 1974)--"proximal turbidites and/or watery slides derived from the northern shelf via submarine canyons" (Morris, 1974, p. 128).

Blakely Sandstone

The Blakely Sandstone is, at first glance, a typical flysch sequence consisting dominantly of alternating beds of sandstone and shale (Fig. 4A). Sandstones are very thin- to thick-bedded, fine- to medium-grained, medium to well sorted orthoquartzites, separated by thin to very thick beds of gray to black shale (Table 5). Fossils are sparse, with graptolites and probably conodonts being the most common. Locally, the Blakely Sandstone can consist of up to 50 percent sandstone beds (outcrop P, Table 5, Fig. 4B).

TABLE 5. -- BEDDING DATA FOR THREE LOCALITIES OF THE BLAKELY SANDSTONE.

OUTCROP FEATURES	OUTCROP						
	K		P		S		
Outcrop Thickness	5881	cm	4466	cm	2296	cm	
Number of Beds	69		100		71		
Percentage Sandstone Beds	6%		50%		28%		
Percentage Shale Beds	94%		50%		72%		
Sandstone Bed Thickness	range	3-60	cm	2-427	cm	3-48	cm
	average	10	cm	43	cm	19	cm
Shale Bed Thickness	range	15-480	cm	2-282	cm	3-457	cm
	average	168	cm	40	cm	45	cm



A



B



C



D

Figure 4. -- Blakely Sandstone. A. Very thin- to thin-bedded sandstones and interbedded shale (Outcrop S). B. Thin-bedded sandstones and interbedded shale (Outcrop K). C. Medium- to thick-bedded sandstones with varying amounts of interbedded shale (Outcrop P). D. Medium- to very thick-bedded sandstones (Outcrop P).



A



B



C

Figure 5. – Thickness variations in sandstone beds of the Blakely Sandstone. A. Bed termination due to lateral thinning and basal scour (Outcrop S). B. Close-up of downward thickening of sandstone bed at X in Figure 4A. C. Thickness variation due to upper surface dunes (Outcrop K).

It is extremely difficult for the field geologist to distinguish between the more sandy Blakely and the more shale-rich Crystal Mountain.

Sandstone beds of the Blakely Sandstone can be variable in lateral and vertical extent and thickness, thinning and thickening in response to (1) basal scour, (2) upper surface ripples and dunes, and (3) bed termination due to lateral thinning or erosion prior to deposition of overlying beds (Fig. 5). Sandstone bases are characteristically sharp, with many showing evidences of local scour (Outcrop S, Table 6). Locally the basal scours may be loadcasted. Parallel lamination is the most common interstratal sedimentary structure, and is generally the only structure in individual sandstone beds. Cross lamination is common

locally with foresets being short and steep (generally $<15^\circ$), formed from the migration of ripples (Fig. 6). Bouma sequences are rare, and where they do occur they consist of the upward sequence cross lamination (C) → parallel lamination (D) → structureless (E). Tops of sandstone beds may be gradational or sharp, and succeeded by shale. Graded bedding is not common (Table 6).

Localities K and P are end members of a lithologic series in non-conglomerate units of the Blakely Sandstone. As in the case of the Crystal Mountain lithologic series, the Blakely series is characterized by variations in the relative abundance of sandstone and shale beds. Locality S is intermediate in this series.

TABLE 6. -- INTERSTRATAL AND INTRASTRATAL SEDIMENTARY CHARACTERISTICS OF SANDSTONE BEDS IN THREE LOCALITIES OF THE BLAKELY SANDSTONE

SEDIMENTARY CHARACTERISTICS OF SANDSTONE BEDS	PERCENTAGE OF SANDSTONE BEDS DISPLAYING EACH CHARACTERISTIC		
	OUTCROP K	OUTCROP P	OUTCROP S
Graded Bedding (normal, well sorted)	9%	9%	17%
Parallel Lamination	68%	57%	33%
Cross Lamination	24%	9%	39%
Bouma Sequence (C-D-E)	4%	6%	8%
Erosive Base	26%	0%	72%
Gradational Base	6%	9%	3%
Sharp, Non-Erosive Base	68%	91%	25%
Load Casts	21%	0%	11%
Flute Casts, Grooves	0%	0%	0%
Gradational Top	40%	38%	49%
Sharp Top	60%	62%	51%



Figure 6. -- Cross-lamination in very thin-bedded sandstones (Outcrop S).



Figure 7. -- Conglomerate layer within the Blakely Sandstone. Light colored boulders generally are kaolinitic sandstones, granites, and kaolinitic shales. Unit 1B is a large portion of a bed of well-sorted, medium-grained sandstone, petrographically identical to sandstone beds in the Blakely Sandstone. As such it represents an intrabasinal fragment within the conglomerate.

Sandstones and shales of the Blakely Sandstone have been attributed to deposition in either (1) the neritic environment of an epicontinental sea (Brieva, 1963), or (2) the deep marine environment (Haley and Stone, 1973; Keller and Cebull, 1973; Morris, 1974), the sandstones being "structureless quartz arenites, pulled apart in places. . . apparently deposited rapidly as proximal turbidites or watery slides upon an unstable slope" (Morris, 1974, p. 128).

Conglomeratic units are interspersed at various, and at present, ill-defined stratigraphic levels within the Blakely Sandstone (Fig. 7). Individual units are each some tens of meters thick. These conglomerates contain pebbles, cobbles, and boulders of varied composition, including orthoquartzite, shale, limestone, chert, kaolinitic sandstone, and altered granite. Fragments of kaolinitic sandstone, and granite are unlike other rocks in the Ouachita area. Most of the boulders have been subjected to considerable diagenesis or weathering (Fig. 8). Individual fragments in the conglomerates range widely in size, from a few centimeters to more than ten meters in longest dimension (Fig. 7). The largest boulder known to date had a maximum diameter of 15m and a volume of $\sim 8.0 \times 10^8 \text{ cm}^3$ (Charles Stone, personal communication, 1973). The long axes of the fragments are often parallel or sub-parallel to one another and to the base of the conglomerate units. No sedimentary structures were observed. The matrix of the conglomerates is bimodal, limonite stained, orthoquartzitic sandstone (Fig. 8B).

SANDSTONE PETROGRAPHY

Sandstones of both the Crystal Mountain and Blakely Formations are silica-cemented orthoquartzites, which locally may contain secondary dolomite rhombs, calcite and limonite cement (Table 7). Diagenesis and low grade metamorphism have complicated the original depositional composition and texture. Quartz overgrowths are common particularly in sandstones with little or no original matrix

(Fig. 9). Quartz overgrowths are at times oriented radially with respect to detrital quartz grains, and "interlayered" with chlorite, producing a foliation. In sandstones with matrix, the matrix has been recrystallized to microcrystalline quartz and muscovite, or more rarely, microcrystalline quartz and chlorite (Fig. 10).

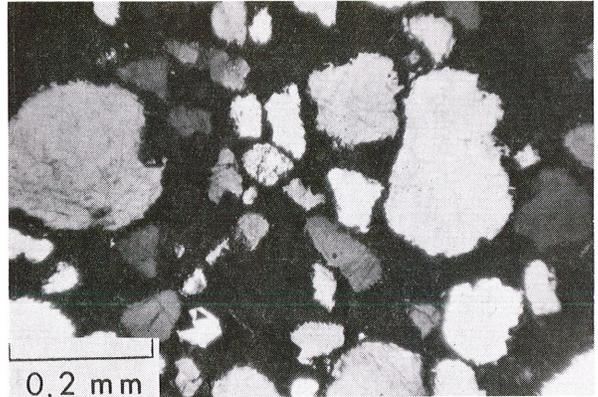
On the average, sandstones of the Blakely Sandstone contain more recrystallized matrix and fewer quartz overgrowths than sandstones of the Crystal Mountain Sandstone (Table 7). Apart from this distinction the petrography of both formations is remarkably similar (see Williamson, 1973, for detailed petrographic descriptions).

TABLE 7. -- PETROGRAPHIC ANALYSES OF SANDSTONES OF THE CRYSTAL MOUNTAIN AND BLAKELY FORMATIONS, OUACHITA MOUNTAIN CORE

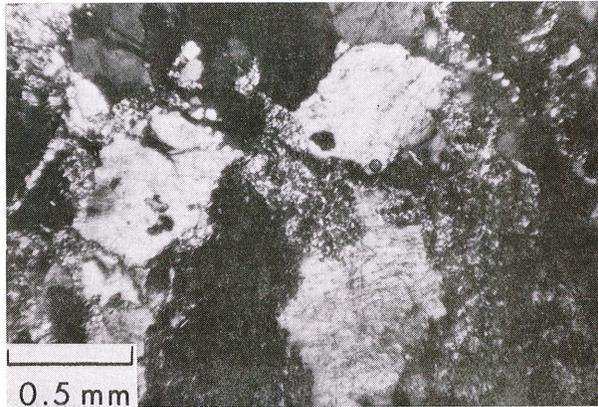
MINERAL	CRYSTAL MOUNTAIN	BLAKELY
Quartz	53%	51%
Mica	Trace	1%
Feldspar	1%	1%
Secondary Quartz (overgrowth)	27%	11%
Microcrystalline Quartz	5%	21%
Carbonate	Trace	2%
Limonite	4%	5%
Pore Space	10%	8%



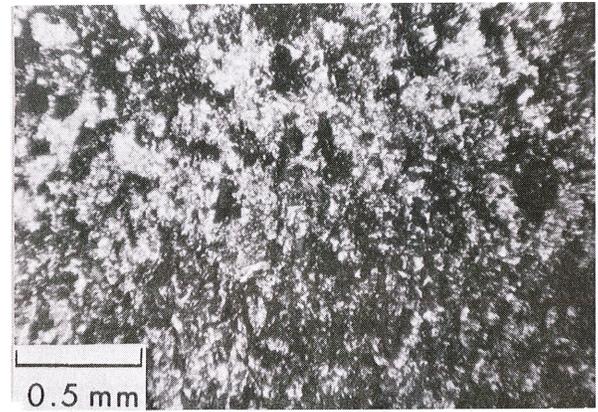
A



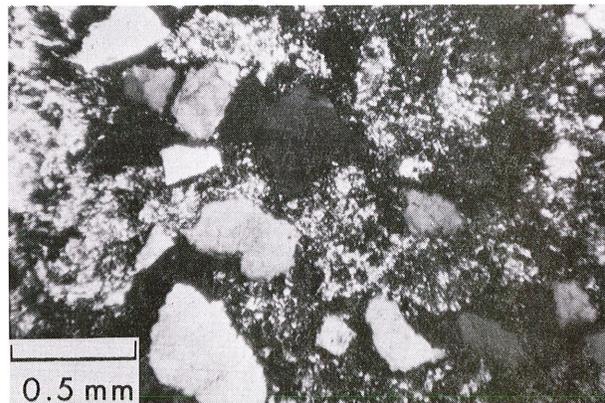
B



C



D

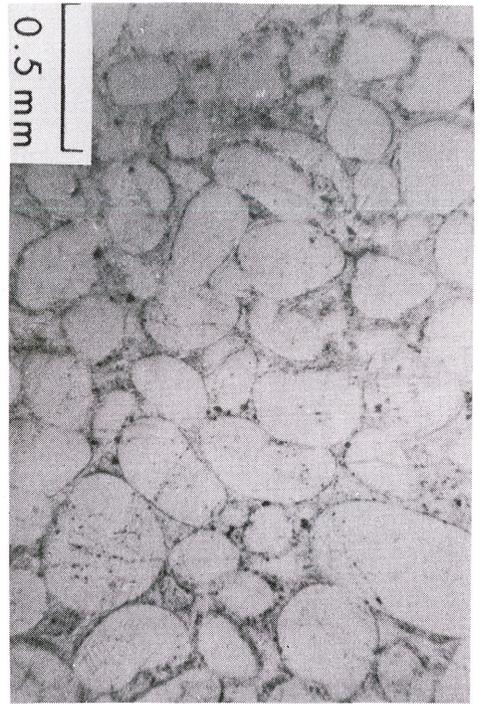


E

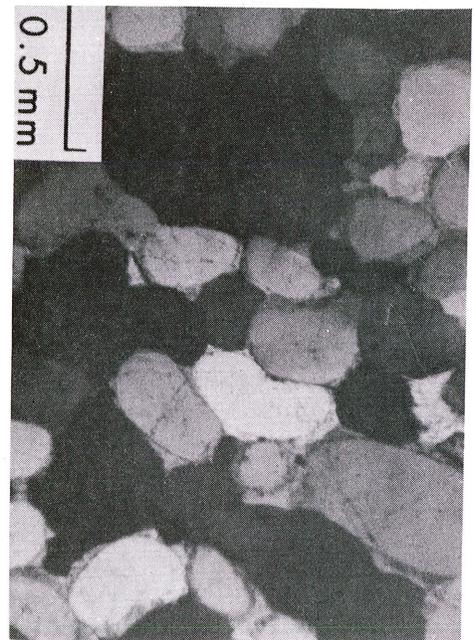


F

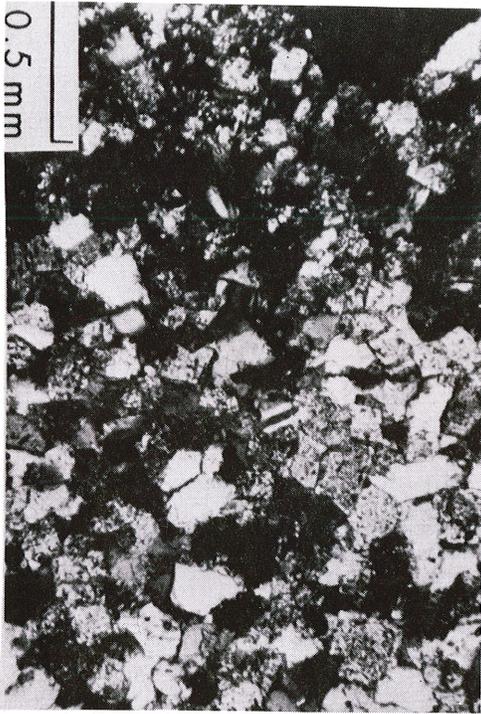
Figure 8. -- Blakely conglomerates. A. Weathered, extrabasinal fragments in conglomeratic unit. B. Bimodal, limonite stained matrix which surrounds the boulders in the conglomerate. C. Granitic extra-basinal fragment with kaolinitic matrix. D. Kaolinite-rich extra-basinal fragment. Dark patches are limonite stained areas of kaolinite. E. Granitic extra-basinal fragment. Megacrystalline grains are quartz, microcrystalline grains are kaolinite, muscovite, and limonite stained kaolinite. F. Muscovite grain surrounded by kaolinite and microcrystalline muscovite - characteristic of granitic extra-basinal fragments.



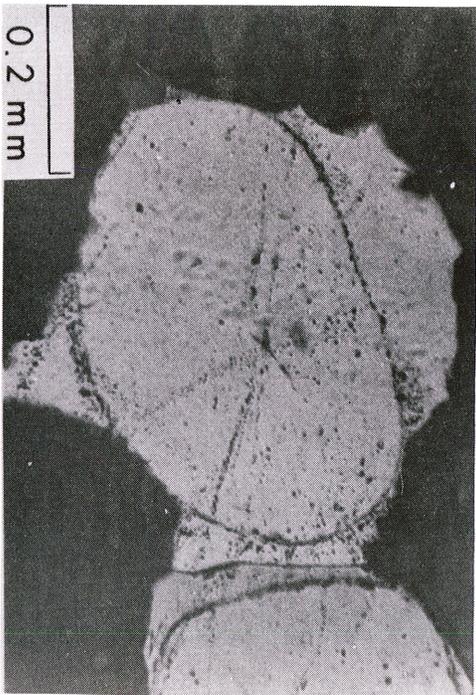
A



B

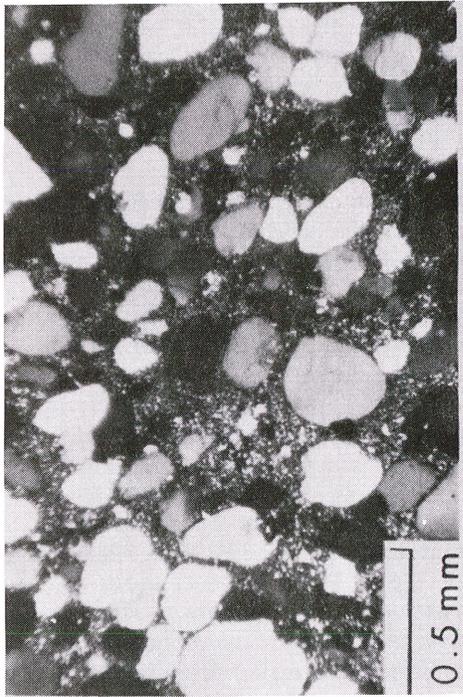


C

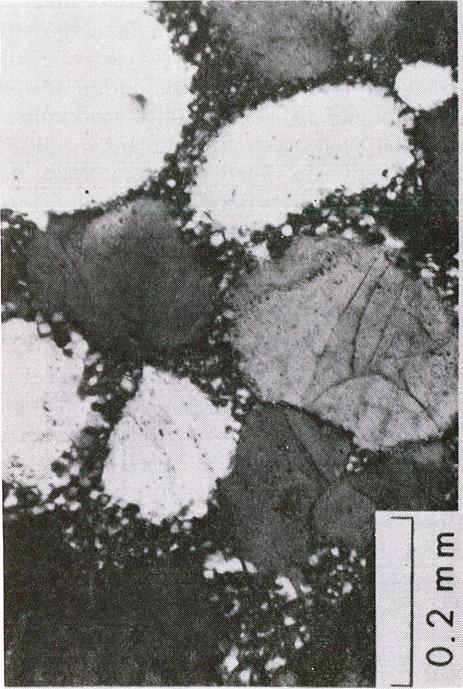


D

Figure 9. -- Sandstone photomicrographs, Crystal Mountain and Blakely Formations. A. Well-sorted, well-rounded quartz grains (plane polarized light). B. Well-sorted, well-rounded quartz grains, showing silica overgrowths (crossed nicols view of 9A). C. Fine-grained sandstone (crossed nicols). Note feldspar grains and dolomite rhombs. D. Close-up of overgrown quartz grain in a medium-grained sandstone (crossed nicols).



A



B



C



D

Figure 10. -- Sandstone photomicrographs, Crystal Mountain and Blakely Formations. A. Medium-grained sandstone with recrystallized matrix of microcrystalline quartz (crossed nicols). B. Close-up of 10A. C. Radial intergrowth of microcrystalline quartz and muscovite (plane polarized light). D. Same as 10C (crossed nicols).

DISCUSSION OF PALEOBATHYMETRY AND PALEOENVIRONMENTS

Although the fauna of the Crystal Mountain and Blakely Formations is sparse, there is no doubt that both formations are of marine origin. The sparseness of body and trace fossils and the lack of colonization evidence indicates that marine conditions were far from optimal, resulting in dominantly lethal (non-life) facies. Unfortunately, these sparse fauna offer no clues to water depth. In the absence of such faunal evidence, sedimentary mechanisms may be used as relative depth indicators since the balance among major marine sedimentary mechanisms varies with water depth. In deep marine environments (slope, rise, and abyssal plain) coarse sediment is transported and deposited by turbidity currents, fluidized sediment flows, grain flows, debris flows and ocean bottom currents. The sedimentary characteristics of beds deposited by each mechanism are well known and are tabulated herein (Table 8). In shallow marine (shelf) environments a large number of sedimentary mechanisms interact (Swift, Duane and Pilkey, 1972), but the characteristic sedimentary features of sequences deposited in this environment result dominantly (but not exclusively) from non-turbid, bottom current action.

The sedimentary structures and textures of sandstone beds in the Crystal Mountain and Blakely Formations indicate that processes operating during their deposition must have been quite variable. This variability is reflected in the grain size of the sediment being deposited and in the amount of penecontemporaneous erosion. At times the scoured sediment-water interface was covered with fine or medium-grained sand; at other times silty clays were deposited. The sharp tops characteristic of sandstone beds in sandstone-shale sequences, indicate either; (1) removal of arenaceous sediment prior to the deposition of succeeding sediments, or (2) sudden changes in the velocity of transport mechanisms. The undulose nature of soles in outcrops characterized by amalgamated sandstone beds indicates penecontemporaneous erosion. Changes in sediment size and bottom current intensity resulted in the transportation of sand either as dunes, ripples, or in flat-bed configuration. Effective traction currents operated continuously throughout the deposition of sandstone beds up to 4m thick.

It is readily apparent that the sandstone beds possess few of the sedimentary characteristics diagnostic of deposition from turbidity currents, fluidized sediment flows, and grain flows (compare Tables 4, 6, and 8). Particularly important is the low percentage occurrence of graded bedding and Bouma sequences, and the abundance of sharp upper surfaces on individual beds. However, the sedimentary characteristics of the sandstones do approximate those diagnostic of deposition from deep-ocean bottom currents, with one notable exception--bed thickness (compare Tables 3, 5, and 8). According to Bouma (1972), deep-ocean bottom current deposits (contourites) are thinner than 5cm. Sand supply and winnowing ability in deep ocean basins are limited, and thus contourites are very thin. Traction reworking of previously deposited turbidites does occur in deep ocean areas, but is generally

restricted to the upper few centimeters of the turbidite bed (Kelling, 1964). It could be suggested that thick (>5 cm), non-graded sandstone beds of these formations result from deep ocean bottom current activity which operated at a higher level of intensity during the early Paleozoic than in the Cenozoic. Thus thick sand sequences transported into the deep ocean environment by turbidity currents, fluidized sediment flows, or grain flows, could have been completely reworked by unusually intense bottom currents. Such an argument is tenuous and unprovable.

We feel that the sedimentary characteristics of these sandstones do indicate highly variable and effective bottom current activity. Such activity is characteristic not of deep marine but of shelf environments. Indeed, the sedimentary characteristics of Crystal Mountain and Blakely sandstones place them in a group of rocks which have been referred to as "sublittoral sheet sandstones" (Goldring and Bridges, 1973). Sublittoral sheet sandstones "occur either as individual beds separated by interbeds of clay and silt, or as virtually amalgamated beds with little or no finer sediment between. Individual beds are typically 5-70cm thick but may exceptionally reach 2.0m. Units of amalgamated beds are known up to 5.0m. Typically the main part of the bed is fine- to very fine-grained, well sorted, parallel laminated sandstone. Very low angle cross-stratification dipping 5° or less is also common. . . While some sandstone tops are gradational, the majority show penecontemporaneous erosion and it is this feature which most clearly distinguishes this facies from typical turbidite sequences" (Goldring and Bridges, 1973, p. 737). This description of sublittoral sheet sandstone fits well the rock sequences in the Crystal Mountain and Blakely Formations--indeed it might have well been written explicitly for these formations.

Within the context of sublittoral sheet sandstones, it would appear that the Crystal Mountain and Blakely Formations represent a shallow marine shelf association. Amalgamated sandstone units (outcrop A, Fig. 1, Table 2 and 3) represent deposition at inner shelf depths, close to the input source of detrital material. The well-sorted and well-sorted texture of such units would suggest an interdeltic environment of deposition, probably as a series of subtidal sand bars or barriers. Medium- to thick-bedded sandstones set in shale sequences (outcrop P, Fig. 1, Tables 2 and 5) we interpret as representing deltaic deposition in distal bar and distributary mouth bar environments. Very thin-bedded sandstones set in thick shale sequences (outcrops K and S, Fig. 1; Tables 2 and 5) are, at least superficially, the most difficult to interpret. A series of investigations of Holocene continental shelves have demonstrated that in areas removed from major supply points, most mud is deposited not on the outer shelf, but along the coast line (McCave, 1972; Meade, 1972; Schubel and Okubo, 1972). Outcrops K and S (Fig. 1) may either represent deposition in prodeltaic environments close to major supply points, or in shallow-marine interdeltic sediment

TABLE 8. - OCCURRENCE OF DIAGNOSTIC INTERSTRATAL AND INTRASTRAL SEDIMENTARY CHARACTERISTICS OF SANDSTONE AND SILT-
STONE BEDS DEPOSITED BY SEDIMENTARY MECHANISMS RESPONSIBLE FOR SEDIMENT TRANSPORT IN DEEP MARINE ENVIRONMENTS (Modified
from Bouma 1972, and Middleton and Bouma, 1972)

SEDIMENTARY CHARACTERISTIC	TURBIDITY CURRENT DEPOSITS (TURBIDITES)	FLUIDIZED SEDIMENT FLOW DEPOSITS	GRAIN FLOW DEPOSITS	DEBRIS FLOW DEPOSITS	BOTTOM CURRENT DEPOSITS (CONTOURITES)
Graded Bedding	Common	Poor to Absent	Absent	Absent	Generally Absent
Parallel Lamination	Common in B and D intervals	Absent	Present (but faint)	Absent	Common
Cross Lamination	Common in C interval	Absent	Absent	Absent	Common
Convolute Lamination	Common toward top of bed	May be present at top of bed	Absent	Absent	Absent
Bouma Sequence (A-E or C-E)	Common	Absent	Absent	Absent	Absent
Erosive Base	Common	Common	Common	Common	Common
Gradational Base	Generally Absent	Absent	Absent	Absent	Absent
Sharp (Non-Erosive) Base	Common	Common	Common	Common	Common
Load Casts	Common	Common	Common	Absent	Absent
Flute Casts, Grooves	Common	May be present	Absent?	May be grooves	Absent
Gradational Top	Common	Absent	Absent	Absent	Absent
Sharp Top	Generally Absent	Common	Common	Common	Common
Dish Structure	Absent	May be present	Common in center of bed	Absent	Absent
Fluid Escape Pipes	Generally Absent	Present	Absent?	Absent	Absent
Bed Thickness	5-400 cm	Not known	"Thick" (Stauffer, 1967)	Highly variable	<5 cm

sinks. The lack of vertical control, due to extensive folding, and the resulting inability to measure large vertical sections with any degree of confidence, makes more precise environmental interpretation difficult at this time.

If the stratigraphic sequence in Figure 1 is a true temporal sequence then sedimentation on the inner shelf kept pace with subsidence and both the Crystal Mountain and Blakely Formations were deposited at inner shelf depths. This subsidence was accompanied by local submarine rifting particularly during deposition of the Blakely Sandstone. This rifting produced fault scarps which affected intrabasinal Paleozoic sediments as well as crystalline continental basement of dominantly granitic composition. Sedimentation from these scarps took place as a series of debris flows, each of limited lateral extent, producing the "erratic" zones well displayed at various horizons in the Blakely Sandstone. (For a comparison of Blakely erratics and debris flow characteristics see Figures 7 and 8, and Table 8).

PALEOGEOGRAPHIC RECONSTRUCTION

Traditionally, Ordovician sediments of the Ouachita Mountains are considered to have been deposited on the southern continental margin of North America, at least 500 km basinward of the paleo-shoreline. Acceptance of such an hypothesis demands outer shelf or deep marine environments of deposition for these sediments. The suggestion that these same sediments were deposited in deltaic and interdeltic environments poses a problem with respect to this traditional paleogeographic hypothesis, and introduces considerable difficulty in interpreting interrelationships between Ordovician rocks of the Ouachita Mountains and their North American shelf equivalents. The conclusions of this paper and those of Barton (1945) Brieva (1963), Flores (1962), Goldstein (1961), and Miser and Purdue (1929) are that the present area of the Ouachita Mountains in Arkansas was extremely shallow and close to shore during the Ordovician.

We postulate that the Crystal Mountain and Blakely Formations were deposited close to the shoreline of a continental area which lay to the south or southwest of the present Ouachita Mountains. The possibility of a southern or southwestern land mass during the lower Paleozoic is not a new idea, having been discussed more than fifty years ago by Miser (1921). Miser (1921) named this southern

continent "Llanoria". If indeed sandstones of the Crystal Mountain and Blakely Formations were deposited at inner shelf depths, as our evidence suggests, Miser's Llanoria gains credence.

In our model, the Ordovician Ouachita trough was a shallow basin underlain by granitic crust. This basin received sediments from both southern and northern continental areas. Sediments derived from the southern continental area now are exposed in the Ouachita Mountains. Sediments derived from North America now are exposed largely to the north of the Ouachitas. The quartzose nature of Ouachita Ordovician sandstones, their extreme diagenesis and low rank metamorphism make detailed provenance determinations exceedingly complex for sand-size detritus. Perhaps trace element analysis of quartz grains, as described by Dennen (1967) may ultimately be useful in determining the precise source of Ouachita Ordovician detritus.

The Crystal Mountain and Blakely Formations were deposited at approximately the same water depths, indicating that basin subsidence and sedimentation rate were in an approximate balance during their deposition. Subsidence was accompanied or accomplished by rifting which may signify the start of the foundering of this Ouachita basin which eventually produced the deep-marine trough of Carboniferous time.

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THE ARKANSAS NOVACULITE:
SOME ASPECTS OF ITS PHYSICAL SEDIMENTATION

By Donald R. Lowe¹

ABSTRACT

The Arkansas Novaculite (Devonian-Lower Mississippian) can be spatially subdivided into two distinct facies: (1) a northern facies characterized in western exposures by interbedded black chert and organic-rich shale and in eastern exposures by chert- and shale-clast breccia, black chert, organic-rich shale, and quartz sandstone and siltstone; and (2) a southern facies showing classic upper and lower white novaculite members and a middle member of black chert and shale similar to that in western exposures of the northern facies. The white novaculites of the southern facies consist largely of thick to massive bedded homogeneous microcrystalline quartz containing sparse floating grains of fine terrigenous quartz sand and silt. This fine quartz is here interpreted as being of aeolian origin. Locally thin (generally < 1 cm) alternating quartz-rich and quartz-poor laminations are interpreted as varves. A second population of well rounded medium- to coarse-grained quartz sand occurs in and just below the lower part of the lower novaculite. This sand was transported as the bed load of currents.

The middle novaculite shows a variety of exotic grains including terrigenous quartz, glauconite, detrital carbonate, and conodonts as well as intraformational chert and shale clasts. These grains occur mainly in the lower, coarser parts of thin, continuous, graded chert beds which are interbedded with shale and commonly show Bouma structural intervals. These chert beds, and many others which do not contain coarser detritus, are probably turbidites.

It is critical to correct interpretation that full consideration be given to physical processes in studies of novaculite sedimentology.

INTRODUCTION

The term "novaculite" is applied to widespread deposits of white, microcrystalline silica or chert in the Ouachita structural belt of the southern United States. The chemical purity and massiveness of the Arkansas Novaculite in classic exposures such as those in and about Hot Springs, Arkansas, are often taken as proof *ipso facto* of the non-detrital, chemical or biogenic origin for these sediments. Yet within even the purest and most massive novaculite are identifiable detrital grains, and interbedded and correlative units show evidence of deposition by currents. It is the objective of this paper to outline some of the physical attributes of the novaculite and to discuss the role of physical processes in its formation.

The results presented here are only the most preliminary observations in a continuing study of the siliceous deposits of the Ouachita system. The author gratefully acknowledges partial support provided for this study by the Louisiana State University Graduate Research Council.

NOVACULITE FACIES

Purdue and Miser (1923) recognized that the Arkansas Novaculite changes in lithology when traced northward from Hot Springs. In fact, as brought to the author's attention by Charles G. Stone of the Arkansas Geological Commission, an entirely different facies characterizes the

northern exposures of the Arkansas Novaculite. This facies, dominated by black chert, shale, sandstone, siltstone, and chert-and shale-pebble conglomerate and breccia, is here termed the northern facies. The better known sections to the south including thick units of white novaculite are here termed the southern facies. The distribution of these facies is shown in Figure 1.

The northern facies can be further subdivided into two lithologic sequences. Western exposures consist predominantly of thinly bedded, black, argillaceous chert interbedded with black, gray, and green claystone and shale. Coarser grained detrital sediments are limited largely to thin beds of intraformational breccia near the top of the formation. These beds, often including a wide variety of otherwise exotic grains including quartz, glauconite, and conodonts, consist largely of angular to subrounded clasts of organic-rich claystone, siliceous claystone, and chert. The eastern exposures of the northern facies, north and northeast of Hot Springs, show a strikingly different aspect. Structure here is complex and the stratigraphic section is difficult to resolve, but the interval represented elsewhere by the Arkansas Novaculite consists of shale, massive chert- and shale-chip breccias up to many ten's of feet thick, black chert, and beds of terrigenous quartz siltstone and sandstone. The conglomerate and breccias appear to represent mass flow deposits. Many of the beds of siltstone and sandstone show erosive bases, cross-bedding, and grading and are strongly suggestive of turbidites. Current directions where measured as part of this study suggest southerly currents. Little is really known of these northern rocks and their significance. They will not be discussed in detail in the present report.

The southern novaculite facies includes the rocks more

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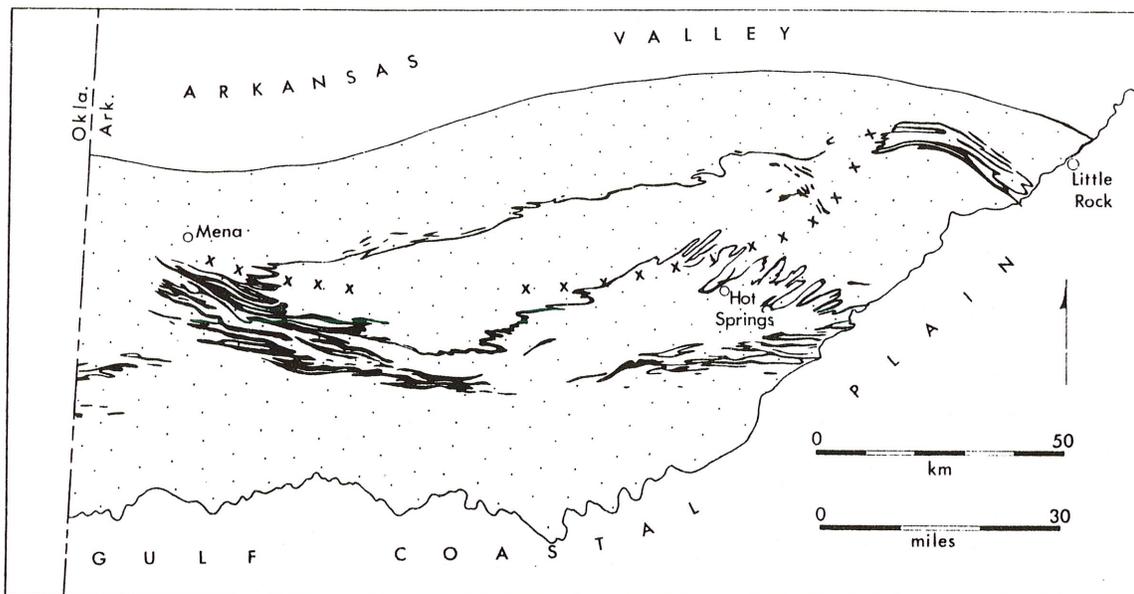


Figure 1. -- Outcrop map of the Arkansas Novaculite (black) in the Ouachita Mountains (stippled) of Arkansas. The approximate location of the northern-southern facies transition is indicated by X's.

commonly identified with the Arkansas Novaculite. The formation is divisible vertically into three members: the massive white novaculite of the lower member; a middle member of shale, black chert, and detrital grains; and an upper member of white novaculite. These members will be discussed in detail in subsequent sections.

GRAIN POPULATIONS

Most of the massive white "true" novaculite consists of an interwoven mass of individually irresolvable microcrystalline silica domains and cannot be visually separated into individual grains. When the members of the southern novaculite are carefully examined, however, all include identifiable detrital grains set within this massive microquartz background. For the purposes of the present discussion, four principal grain populations will be recognized: (1) detrital quartz, (2) detrital carbonate, (3) intraformational clasts, and (4) organic particles. Small amounts of reworked glauconite occur locally in all facies and members of the novaculite.

Detrital Quartz

Detrital quartz occurs in two distinct size and shape populations within the Arkansas Novaculite (Fig. 2): (1) fine, angular quartz sand and silt (Fig. 2 a, c), and (2) medium- to coarse-grained, well rounded sand (Fig. 2 b, c). The first population occurs throughout the massive white novaculite as angular grains of very fine quartz sand and coarse quartz silt. Finer-grained detrital quartz, if present, would be visually inseparable from the surrounding microcrystalline silica. This quartz forms less than 1% of the novaculite in all sections examined. In the massive nova-

culite which makes up the bulk of the upper and lower members, this quartz occurs as isolated, randomly distributed grains, but in most sections of the lower novaculite, there are zones in the lower half of the member which show fine, cyclic laminations separated by intervals of massive novaculite 1 to 10 mm thick (Fig. 3a). In thin section, these laminations show a high concentration of fine detrital quartz, commonly accompanied by a high content of brownish organic grains and other fine-grained exotic particles. Even in these laminations, however, the framework is generally not grain supported. The intervening chert shows relatively minor amounts of detrital quartz, but locally contains siliceous organic particles such as spherical, radiolarian-like forms and spicules (Fig. 3b). In the middle novaculite member, fine detrital quartz commonly forms laminations only a few grain-diameters thick at the base of thin argillaceous chert beds and occurs mixed with other clasts in chert- and shale-chip breccias. These cyclic laminations have been interpreted as varves by Lowe (1976).

The second population of detrital quartz is made up of well-rounded, highly spherical medium- to coarse-grained sandstone in thin beds within the lowest 70 feet of the lower novaculite and uppermost Missouri Mountain Shale between Caddo Gap and Little Rock (Fig. 2b). The beds are generally less than 1 foot thick, grain supported, silica-cemented, and commonly cross-laminated. Excellent exposures are found at the base of the lower member in Gulpha Gorge near Hot Springs. This quartz is remarkably similar to that making up the Blakely Sandstone (Lower Ordovician) and quartz in limestones in the Womble Shale (Middle Ordovician). It is texturally and compositionally identical to that in the St. Peter Sandstone on the shelf to the north and may indicate a shelf contribution to geosynclinal sedimentation.

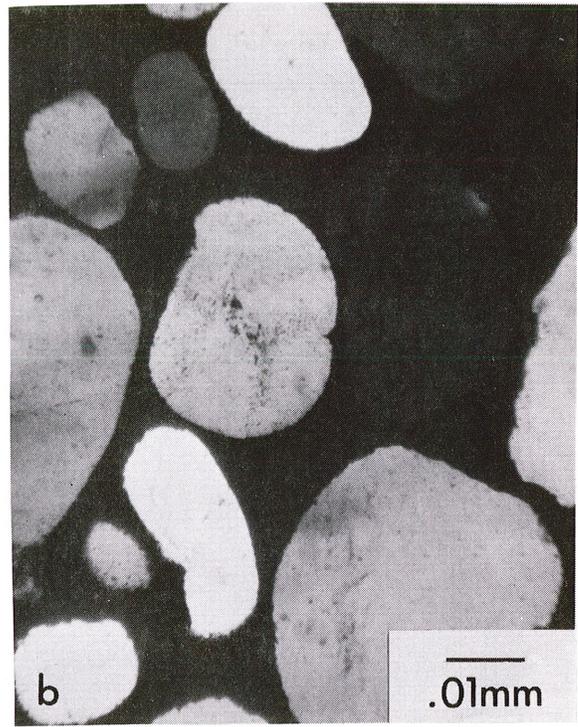
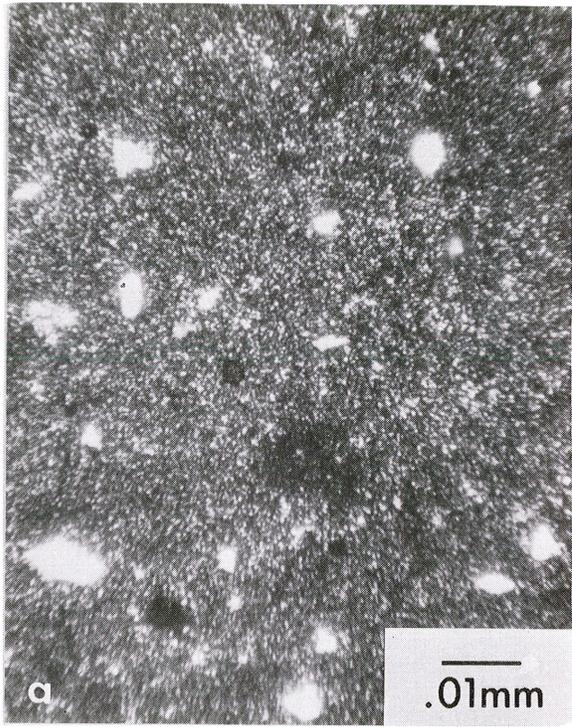


Figure 2. -- Quartz populations. a. Fine quartz sand and silt disseminated throughout massive novaculite of lower member. b. Coarse-grained quartz sand cemented by hematite. Collected from thin quartzite bed 66 feet above base of lower member at Caddo Gap highway section. c. Mixed fine- and coarse-grained quartz populations in bioturbated lower member. West Mountain overlook, Hot Springs, Arkansas.

The importance of these quartz populations has long been overlooked in studies of the Arkansas Novaculite. Both are sufficiently coarse that they could not have been carried significant distances in suspension by normal ocean currents. The finer population, distributed throughout the otherwise massive, pure novaculite, strongly resembles aeolian silts found throughout present ocean basins and on oceanic islands. The fine cycles present

in the lower novaculite would seem then to represent aeolian cycles possibly related to wind and storm periodicity rather than evaporite cycles as hypothesized by Folk (1973). The uniform distribution of this fine quartz within the massive novaculite is anomalous in terms of particles that must have been current (wind or water) deposited. The sorting effect of currents on the quartz and the primary siliceous particles, whether they were colloidal

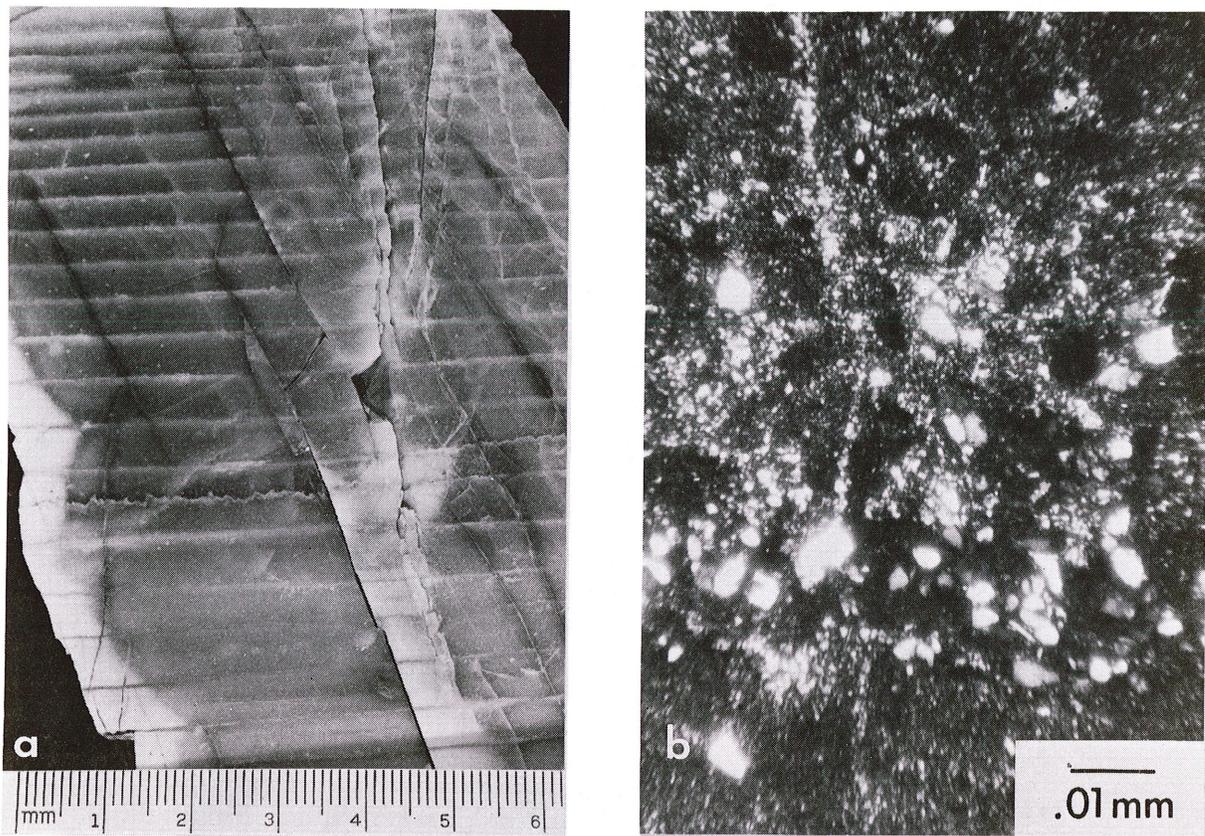


Figure 3. -- Cyclic laminations in lower novaculite member. a. Megascopic appearance. Thin whitish laminations are richer in quartz silt and other terrigenous detritus than intervening zones of darker, translucent novaculite which contain abundant sponge spicules. Caddo Gap, Arkansas. b. Photomicrograph of quartz-rich lamination and overlying and underlying quartz-poor chert with Radiolaria-like spheres. Potato Hills, Oklahoma.

silica or radiolarian tests, would have produced quartz-rich and quartz-poor laminations rather than a completely homogeneous sediment. The present uniformity suggests post-depositional modification, possibly the result of burrowing organisms.

The coarser population of quartz could only have been deposited as the bed load of currents. The thin, discrete units, interbedded with clay shale at the top of the Missouri Mountain Shale (Silurian ?) and with white novaculite at the base of the lower member, strongly resemble rapid, single-shot sedimentation units such as turbidites.

Carbonate

Very fine sand- and silt-sized carbonate euhedra are common throughout the novaculite, but the present form of these grains appears to be largely secondary, and they will not be considered here. Identifiable wholly detrital carbonate grains have not been observed within the true novaculite, but can be found locally in the coarse bases of graded units in the middle novaculite member and in chert- and shale-chip breccias at the Arkansas Novaculite - Stanley transition. The carbonate grains are largely rock fragments and mineralogy has not been determined. Many have been silicified.

Intraformational Clasts

Like detrital carbonate grains, interformational clasts occur largely within the lower parts of coarse-grained graded sedimentation units in the middle novaculite member and in thin intraformational breccia beds at the Novaculite-Stanley contact. The clasts are typically angular to subrounded and range up to 10 cm across. Most are composed of black, organic shale; siliceous shale; or black or gray chert. Radiolaria and spores are common within the shale and siliceous shale, much of which shows bedding. Clasts of white novaculite are rare as are clasts of coarser-grained terrigenous sediments.

Organic Particles

Identifiable organic particles constitute an especially important grain population because of the probable biogenic origin of the bulk of the novaculite's silica. Within the white novaculite, biogenic grains include a few siliceous spheres and spherical ghosts closely resembling Radiolaria (Fig. 3b), and very abundant siliceous sponge spicules. In the organic mudstones and black cherts of the middle member and also characteristic of the northern facies, Radiolaria-like siliceous spheres, sponge spicules, and non-

siliceous spore capsules are abundant. Conodonts are common in the coarser grained bases of graded beds in the middle member and within intraformational breccias. Many of these particles, notably the Radiolaria, spores, and conodonts, were transported to their sites of deposition.

SEDIMENTARY STRUCTURES

Lower and Upper Novaculite Members

Other than bedding, sedimentary structures are strikingly rare in the massive white novaculites. Three will be discussed as part of the present study: cross-laminations, cyclic laminations, and burrowing.

Cross-laminations

Cross laminations occur sporadically throughout the lower novaculite member, in graded beds in the middle member, and in a single bed of novaculite from the upper member (Fig. 4a). They are particularly common in sandstone and translucent chert beds of the chert-and-shale subdivision and in thick translucent chert beds of the translucent novaculite subdivision of Lowe (1976). They indicate that much of the silica had as a precursor particulate sand- to coarse-silt-sized material.

Varves

Varves (Lowe, 1976) are common throughout the lower member of the novaculite, but have also been observed in the middle and upper units. They include two types (Lowe, 1976). Terrigenous varves, discussed previously, are defined by alternations of thin laminations enriched in terrigenous detritus and nearly pure chert layers. Calcareous varves are characterized by alternating carbonate-rich and carbonate-poor laminations.

Burrowing

Animal burrows or tracks and trails have been unequivocally identified on the bases of and within only a very few beds in the novaculite. Evidence suggests, however, that the present homogeneity of the novaculite is in large part a result of the burrowing activities of animals: (1) the even distribution within the massive novaculite of fine-grained detrital quartz deposited by currents, either aeolian or aqueous, suggests post-depositional mixing of what must have originally been more structured beds; (2) in one instance where discrete laminations of the medium- to coarse-grained quartz were present within the massive novaculite, the quartz laminations were swirled and mixed with novaculite along their margins, effects similar to those produced by burrowers; and, (3) thin beds of medium- to coarse-grained quartzite near the base of the lower member commonly show cross-bedding but are also commonly massive or swirled due to bioturbation.

Middle Novaculite Member

The middle novaculite member differs from the super-

jacent and subjacent massive white novaculite members in two important respects: (1) it includes an important terrigenous component, especially clay, but also coarse silt, sand, and fine-gravel sized material; and, (2) it contains unoxidized organic carbon which lends a black color to the shales and cherts. As exemplified by the section at Caddo Gap (Miser and Purdue, 1929), the middle member consists largely of black siliceous clay shale and black, thinly bedded, argillaceous chert interbedded in subequal proportions. The general aspect is similar to the western exposures of the northern facies. Coarse-grained detritus occurs within resistant siliceous beds distributed throughout the middle member.

Some of the resistant siliceous beds show evidence of having been deposited by currents, and their overall evenness and continuity and the rhythmic alternation of shale and silica units suggests turbidity currents. Structures commonly observed within these beds are grading, cross-laminations, and flat laminations (Fig. 4b). The grading is typically marked by a coarse-grained base consisting largely of chert-cemented intraformational debris or coarse-grained St. Peter type quartz abruptly overlain by a top of black or gray chert (Fig. 4b, c). The coarse-grained base, depending on its thickness, may show Bouma intervals a, b, and c (Fig. 4); the silica top represents the d subdivision. This chert interval contains none of the uniformly disseminated fine-grained detrital quartz typical of the massive novaculite and commonly shows a progressive upward increase in argillaceous material. It is apparent that this silica was transported rapidly into the depositional site along with the coarser detritus. The overlying shale, siliceous shale, or mudstone apparently represents the Bouma hemipelagic e subdivision. Figure 4d illustrates three successive sedimentation units of olive green chert from the northern (?) facies in the Potato Hills, Oklahoma which show fine-grained quartz bases only a single grain thick overlain by crudely cross-bedded chert. It seems probable that many of the thin silica units in the middle member which do not contain coarser-grained debris may also represent deposition by currents.

DISCUSSION

The objective of this paper has been to suggest that many aspects of the Arkansas Novaculite are best explained by reference to processes of physical sedimentology. Some of the observations made in the course of this study are summarized below:

(1) Two distinct detrital quartz grain populations occur within the massive white novaculite. The fine population consists largely of coarse silt and very fine sand and appears to be of aeolian origin. Fine quartz is uniformly distributed throughout most of the novaculite, probably the result of the thorough mixing of originally less homogeneous sediment by burrowers. Locally thin detrital quartz-silica alternations appear to represent the cyclic introduction of aeolian detritus into the basin of deposition. Similar cycles or other laminations may have characterized much of the original novaculite sediments.

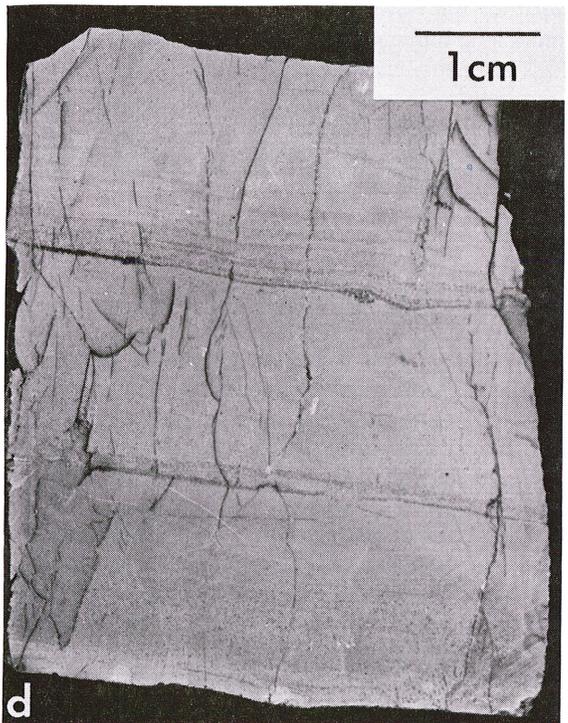
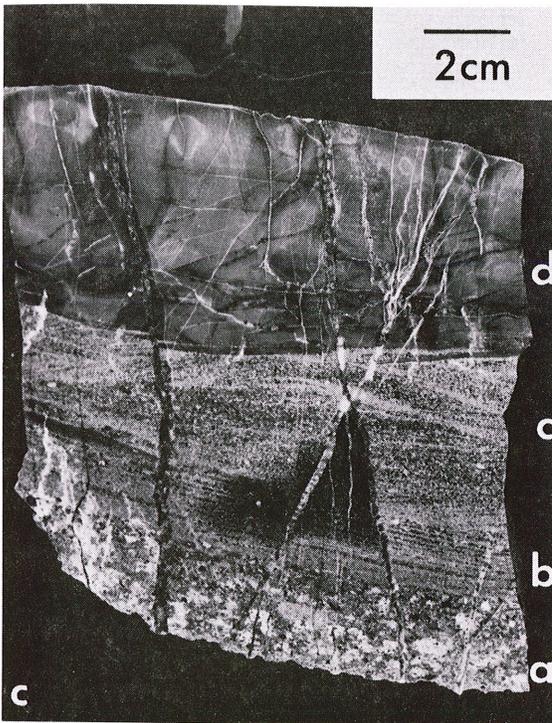
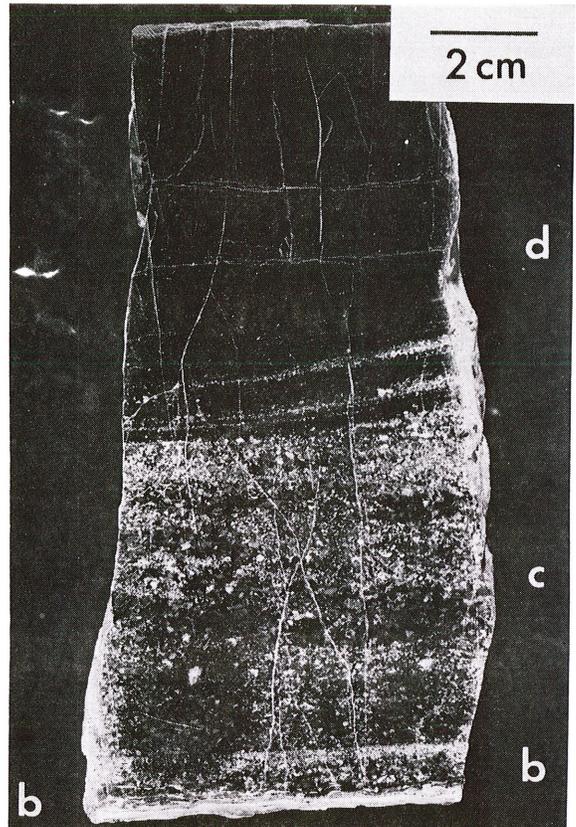


Figure 4. -- Current structures in the Arkansas Novaculite. a. Small ripple-form (arrow) and associated cross-lamination in novaculite. Upper novaculite member, Caddo Gap, Arkansas. b. Graded bed with flat laminations near base (*b* subdivision of Bouma), low angle cross-laminations in middle (*c* subdivision), and massive black argillaceous chert at top (*d* and *e*? subdivisions). Coarser material in lower half of bed is silicified carbonate grains, mudstone, and chert. Middle member, Caddo Gap, Arkansas. c. Graded turbidite bed showing *a*, *b*, *c*, and *d* subdivisions of Bouma. Uppermost *d* subdivision consists entirely of gray translucent chert which lacks disseminated aeolian quartz. Coarse detritus includes silicified carbonate grains, intraformational material, and quartz. Middle member, West of Malvern, Arkansas. d. Three successive thin graded beds. Dark laminations are zones of quartz which fines upward and passes into crudely stratified greenish chert. Middle member, Potato Hills, Oklahoma.

The second, quartz grain population is made up of well-rounded, highly spherical medium- to coarse-grained sandstone in thin beds within the lowest 70 feet of the lower novaculite and uppermost Missouri Mountain Shale. This sand may indicate a shelf contribution from the north by rapid sedimentation processes such as turbidity currents.

(2) The middle novaculite shows an alternation of chert and shale beds. Many chert beds contain coarser grains, mostly of intraformational origin, and where they do, grading and current structures are common. These beds appear to be turbidites, and many of the finer-grained chert beds may also represent turbidity current deposition.

(3) The occurrences of chert as c and d intervals of graded turbidites suggests that it was deposited as discrete, rigid, non-cohesive particles hydrodynamically equivalent to very fine-grained quartz sand or coarse-grained quartz silt. In one bed within the middle member, showing a thin lens of chert interlaminated within cross-bedded intraformational and extraformational sand and granule-sized debris, the chert included abundant large, fine- to medium-grained sized radiolarian-like spheres. Although these latter observations may have little direct bearing on the original character of the upper and lower novaculite members, they do suggest that physical processes of sedimentation could also have played a critical role in their deposition.

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ARKANSAS NOVACULITE STRATIGRAPHY

By Mark A. Sholes¹

ABSTRACT

The Arkansas Novaculite is a maximum of 275 m thick and consists of five members which are in ascending order: lower chert and shale; lower novaculite; middle chert and shale; upper novaculite; and upper chert and shale. The formation is conformable with the underlying Missouri Mountain Shale and the overlying Stanley Shale. Local conglomerate beds are interpreted as submarine slump and slide deposits and the locally overlying Hot Springs Sandstone as a submarine channel fill contemporaneous with deposition of the Arkansas Novaculite.

Novaculite is a siliceous rock composed of 5 to 20 millimicron polyhedral grains of quartz that can generally be distinguished from chert by its gritty rather than smooth fracture surface. The novaculite is spiculitic and pelletal, whereas the chert is predominantly Radiolaria-bearing and not burrowed.

The character of the Arkansas Novaculite changes west and north from Caddo Gap, where the novaculite members thin and become interbedded with shale. Novaculite is absent in Black Knob Ridge and the Potato Hills, but chert in the lower part of the formation is similar to novaculite. It is spiculitic and pelletal.

The roughly isochronous members of the Arkansas Novaculite correlate lithologically with rocks north and west of the Ouachita Mountains by decrease in chert and substitution of shelf carbonates for novaculite. This change is interpreted as a basin to shelf change of depositional environments.

INTRODUCTION

The Arkansas Novaculite crops out in an elongate belt about 340 km long and 50 km wide between Little Rock, Arkansas, and Atoka, Oklahoma (Fig. 1). The stratigraphy of the Arkansas Novaculite has been broadly worked out by Miser (1917), Honess (1923), Miser and Purdue (1929), and Hendricks, Knechtel, and Bridge (1937). Contributions to its stratigraphy and petrography have been made by Goldstein and Hendricks (1953), Goldstein (1959), Sellers (1966), and Park and Croneis (1969). Most previous studies have dealt with only limited areas of the Arkansas Novaculite as part of local areal geology studies. This study, however, deals with the Arkansas Novaculite throughout its outcrop area. Regional variations in the stratigraphic sequence provide clues to the origin of the Arkansas Novaculite and to its relationship with shelf clastics and carbonates to the north and west.

NOVACULITE

A brief discussion of novaculite is necessary to clarify the distinction between chert and novaculite used in this report. Miser and Purdue (1929, p. 49) defined novaculite as "a gritty, fine-grained, homogenous, highly siliceous rock, possessing a conchoidal or subconchoidal fracture and being translucent on thin edges." Tarr (1938, p. 27) defined novaculite as "a very dense, even-textured, light-colored cryptocrystalline siliceous rock; similar to chert but characterized by a dominance of quartz rather than chalcedony." King (1937, p. 54-55) and McBride and Thomson

(1970, p. 46-47) from study of the Caballos Novaculite of Texas made white color a part of the definition. Folk and Weaver (1952), using a transmission electron microscope, recognized a novaculite texture consisting of polyhedral blocks of microcrystalline quartz. The definition preferred by this writer is that novaculite is a siliceous rock composed of 5 to 20 millimicron polyhedral grains of quartz with little or no chalcedony. This microtexture gives novaculite its unique properties as a whetstone and a characteristically gritty rather than smooth fracture surface. In part, the texture of novaculite is due to dynamic metamorphism (Goldstein and Hendricks, 1953, p. 430). Samples from McCurtain County, Oklahoma, and the vicinity of Hot Springs, Arkansas, within Miser's (1943, p. 104-105) metamorphosed zones, contain novaculite with remnant extinction patterns of cavity filling chalcedony. That the novaculite is not entirely a product of metamorphism is shown by rocks at Caddo Gap, Arkansas, where the novaculite members are prominent but metamorphism has not been sufficient to destroy the chalcedony present in the associated chert, or the chalcedony spherulites sparsely present in the novaculite.

The petrographic distinction between chert and novaculite results in recognition of faunal differences. Chert of the middle and upper chert and shale members contains Radiolaria and is finely laminated, indicating absence of a benthonic fauna, whereas novaculite contains sponge spicules and fecal (?) pellets (Fig. 2). Textures are not so clearly distinguishable where metamorphism has recrystallized the chert but ghosts of the original texture scattered sparsely through the rock support the faunal distinction between the middle and upper chert and shale members and the novaculite. Definite bioturbation of the lower novaculite member in western Arkansas is shown by remnant patches of well preserved spicules surrounded by pelletal novaculite. Radiolaria are sometimes present with spines, range up to 0.2 mm in diameter, and are always

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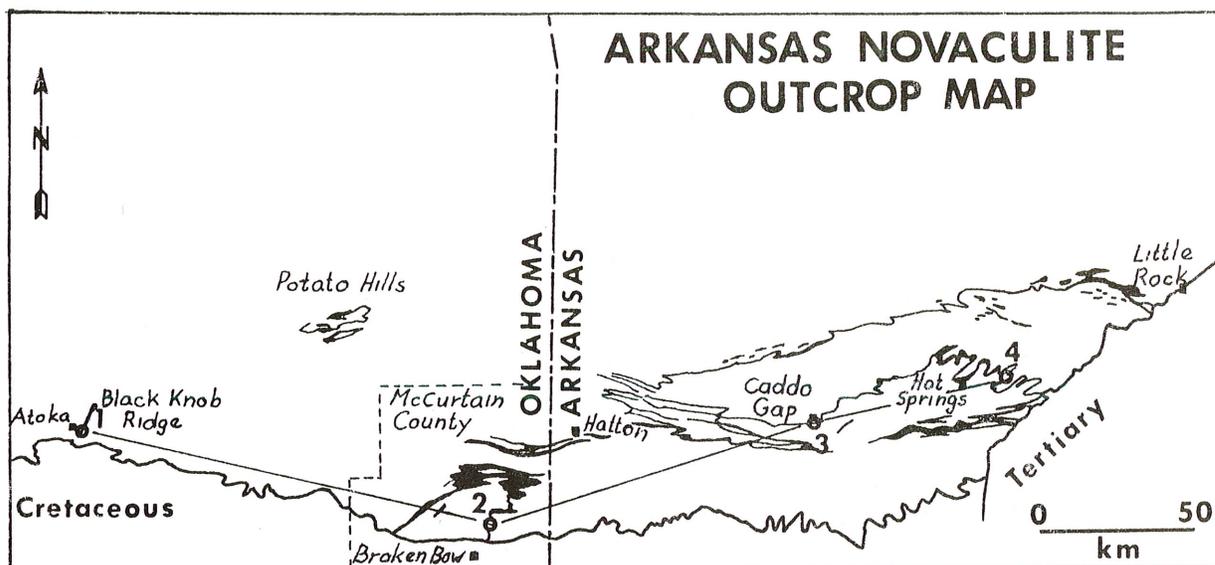


Figure 1. - Generalized outcrop map of Arkansas Novaculite. Numbered points show location of cross section (modified after Park and Cronis, 1969).

coarser grained than the surrounding chert. Fecal pellets do not have spines, are generally less than 0.1 mm in diameter, and are finer grained than the surrounding rock. Pellets containing spicules are present in chert in the lower part of the Arkansas Novaculite at Black Knob Ridge.

STRATIGRAPHY

The sequence of rocks in the Arkansas Novaculite is shown in the diagrammatic section at Caddo Gap, where the formation is 275 m thick (Fig. 3). The recognition of five members in this report instead of three defined by Miser and Purdue (1929) is based on the nearly universal presence of interbedded chert and shale at the base and top of the formation. Lateral variation in the Arkansas Novaculite is illustrated by the diagrammatic east-west cross section (Fig. 4). North-south variation appears similar to the east-west variation but is more difficult to document persuasively. Severe folding and fault displacement of unknown magnitude complicate interpretation of stratigraphy and facies relationships. In the northernmost outcrops of Arkansas Novaculite in Arkansas, the formation is thin, deformed, and difficult to distinguish from the Bigfork Chert. The novaculite members are not present; the entire formation consists of interbedded dark chert and shale. The change is considered a result of depositional facies change rather than erosional removal because of the similarity to the east-to-west rock changes which show more clearly the gradual increase in shale content and thinning of the novaculite members.

The Arkansas Novaculite conformably overlies the Missouri Mountain Shale. The base of the Arkansas Novaculite is considered to be the base of the first chert or chert-cemented chert-clast conglomerate bed above typical Missouri Mountain Shale. There is no detectable change in the shale across the boundary between the two formations.

The lower chert and shale member is 7 m thick at Caddo Gap, but is less than 3 m thick or absent at other locations. The Missouri Mountain Shale grades upward into the lower chert and shale by the appearance of thin, lenticular, chert beds, and locally of chert-clast conglomerate and quartzarenite beds. Near Hot Springs, Arkansas, lenticular beds of medium grained, mature to supermature cherty quartz-

arenite up to 15 cm thick are present. The chert is spiculitic with rare Radiolaria. The lower chert and shale member as defined in this report includes most of the Missouri Mountain Shale as identified by Hendricks, Knechtel, and Bridge (1937) 25 km north of Atoka, Oklahoma, where it contains black chert beds.

In Arkansas, the lower novaculite member is composed almost entirely of slightly silty, spiculitic, pelletal novaculite. Shale and quartzarenite are locally present as thin beds and laminae. Shale increases in abundance to the west, and in McCurtain County, Oklahoma, shale is present in beds up to 4.5 m thick and as lenses and pods in hummocky and contorted novaculite beds. In the Potato Hills a 10 m thick unit of gray, pyritic, banded chert is laterally equivalent to the lower novaculite member in Arkansas. The lower 34 m of the formation exposed at Black Knob Ridge consists of interbedded spiculitic, pelletal, green to brown chert and shale. The lower member ranges from 15 to 135 m thick and thins northward and westward from a point about 16 km south of Hot Springs, Arkansas (Purdue and Miser, 1923).

The middle member consists mainly of Radiolaria-bearing dark gray to black interbedded chert and shale. Most beds are less than 30 cm thick, although shale units up to 9 m thick are locally present. A complete range from pure chert through argillaceous chert to clay shale is present. The position of the middle member 24 km east of Hot Springs is occupied mostly by interbedded novaculite and shale. Graded and nongraded laminae of silt and fine sand 0.1 to 4 cm thick are present in both chert and shale beds throughout the middle member. Finely banded chert is abundant. Chert-clast conglomerates are rare components most commonly present at the base of the middle member. The middle member ranges from a maximum of 105 m to a minimum of 3 m thick (Purdue and Miser, 1923).

The upper novaculite member is composed of white to gray novaculite which locally contains abundant carbonate rhombs or cavities from which carbonate has presumably been removed. Spicules and pellets are locally recognizable. The member is a maximum of 55 m thick in Hot Springs and thins abruptly to the north and south and more gradually to the west. It is not present in the Potato Hills or

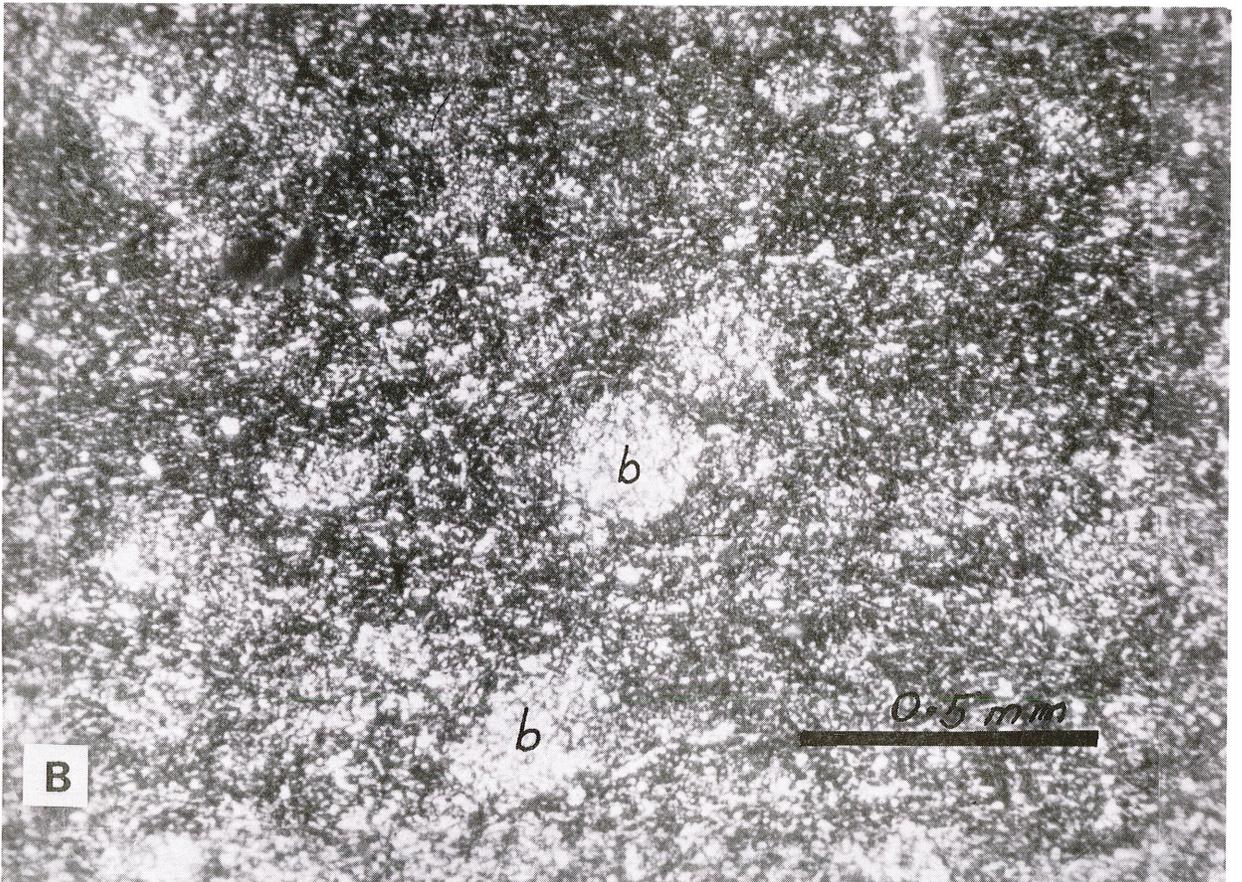
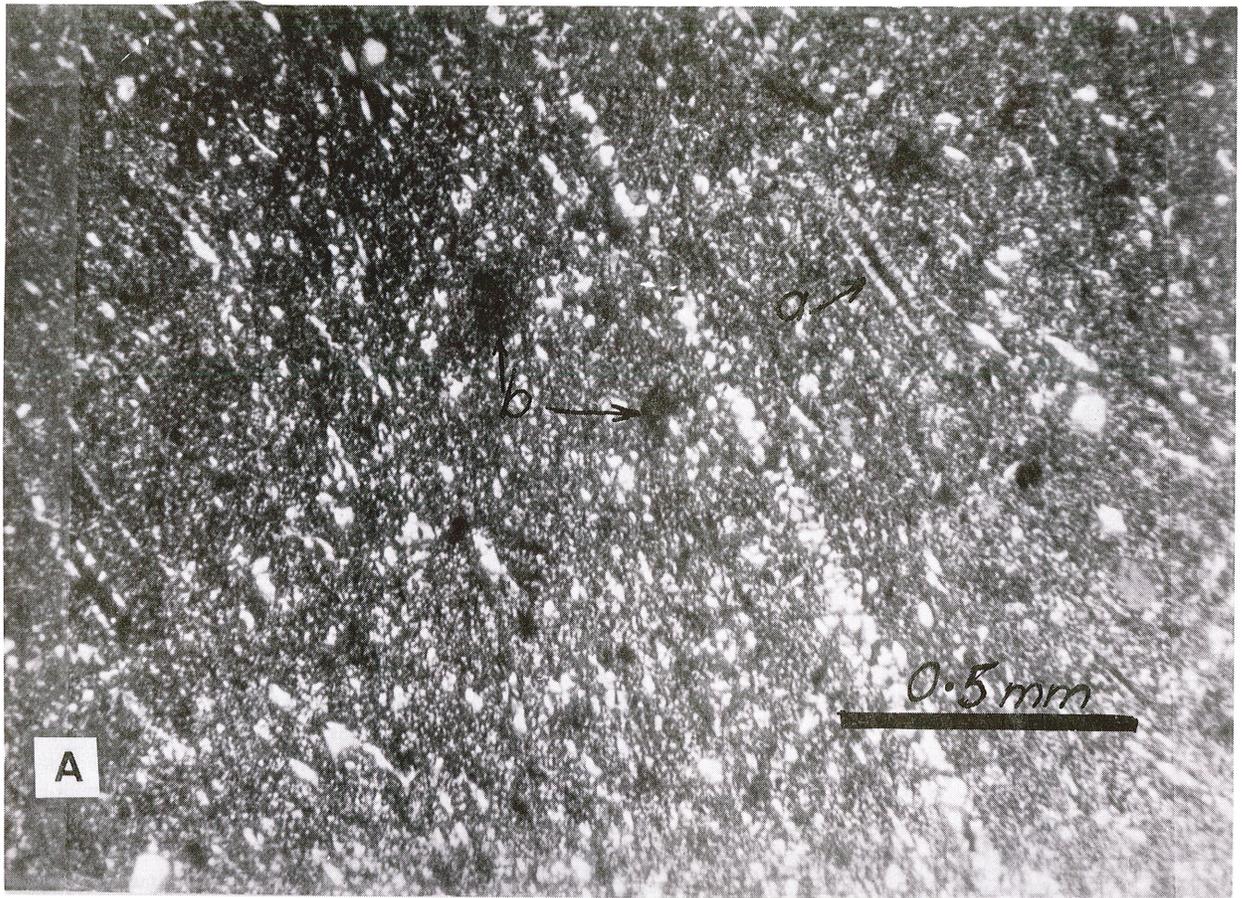


Figure 2. -- A. Novaculite with spicules (a) and faint pellets (b) from lower novaculite member at Caddo Gap, Arkansas. Crossed nicols.
 B. Chert from upper part of Arkansas Novaculite at Black Knob Ridge, Oklahoma. Radiolaria shown at b. Crossed nicols.

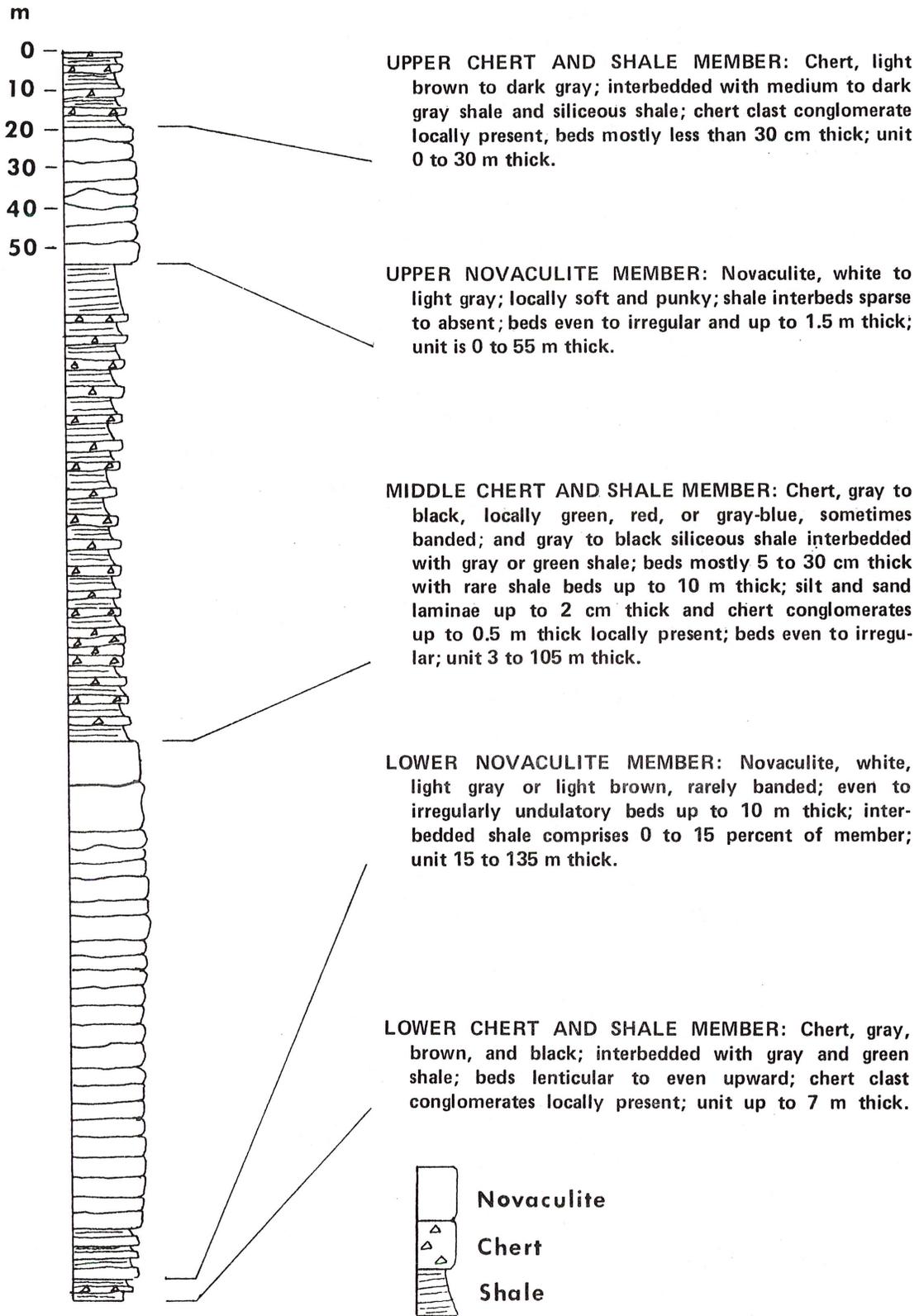


Figure 3. -- Diagrammatic columnar section of Arkansas Novaculite based on the section at Caddo Gap, Arkansas. The total observed range of thickness for each of the members is given.

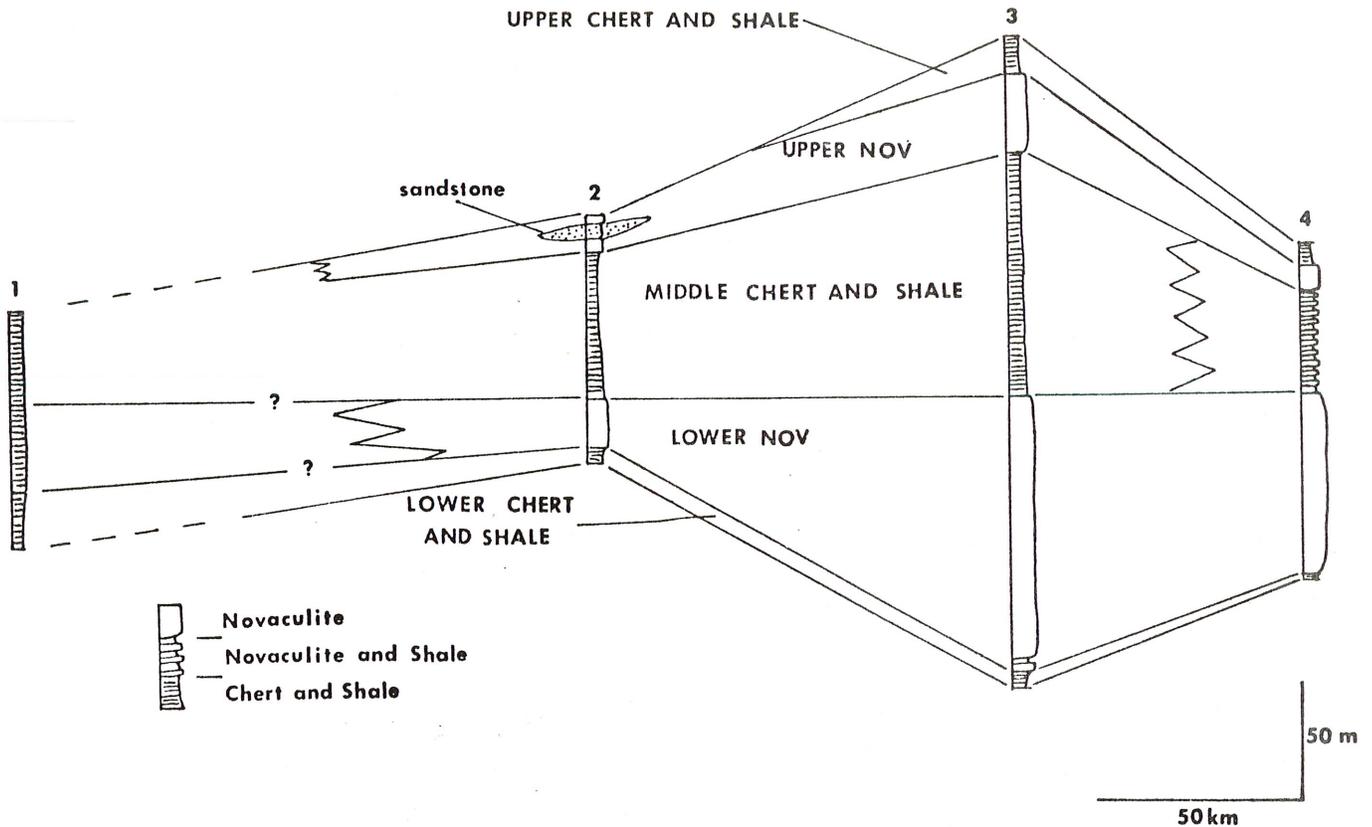


Figure 4. -- Diagrammatic east-west cross section of Arkansas Novaculite. Section 1 is a composite section from Black Knob Ridge, Oklahoma. A fault of unknown displacement is present in the lower novaculite member of section 4. Locations of sections are shown on Figure 1.

Black Knob Ridge. In McCurtain County, Oklahoma, 6 m of fine- to medium-grained, graded, immature, cherty phyllarenites are present locally in the middle of the upper member.

The upper chert and shale member is composed of interbedded gray or brown chert and gray shale in Arkansas and McCurtain County, Oklahoma, and interbedded green chert and shale in Black Knob Ridge and the Potato Hills. Chert-clast conglomerate and phyllarenite beds are present locally. Shale constitutes about 50 percent of the unit at the base and becomes more abundant upward. Radiolaria are the most abundant fossils present. The transition into the Stanley Shale is marked by the replacement of chert with siliceous shale. The upper chert and shale member is a maximum of 15 m thick in Arkansas but is up to 30 m thick in the Potato Hills.

Stanley Shale overlies the Arkansas Novaculite except in the vicinity of Hot Springs, Arkansas, where the Hot Springs Sandstone is present below the Stanley Shale. The upper boundary of the Arkansas Novaculite is gradational except where the Hot Springs Sandstone, with its distinct basal conglomerate, is present. The top of the Arkansas Novaculite is defined as the top of the highest chert bed below typical Stanley Shale. Chert conglomerate beds are present at or near the top of the Arkansas Novaculite at about 50 percent of the outcrops studied. They are not considered important stratigraphic marker beds.

CONGLOMERATE BEDS

The importance of conglomerate beds involves the customary equation of conglomerate with subaerial erosion.

Chert-clast conglomerates are present mainly in the lower chert and shale member, near the base of the middle member, and in the upper chert and shale member. Clasts range from coarse sand-size to about 10 cm, from angular to rounded, and include a variety of rock types. The source of most clasts is clearly Arkansas Novaculite, and clasts of all rocks in the Arkansas Novaculite, including shale, are present locally. The matrix is either spiculitic rhomb-bearing novaculite or clear chalcedonic chert. Some beds are distinctly size graded and grade upward into chert, whereas others show no size grading. In spite of similarity of stratigraphic position, it is not possible to demonstrate stratigraphic continuity of any of the conglomerate beds. Variations in clast types indicate several intrabasinal sources. The conglomerates have been considered indicators of subaerial exposure and erosion (Miser and Purdue, 1929, p. 57). Most conglomerates are not associated with distinct erosional scour and they are not lenticular over distances as small as 10 m. There are no percussion marks on the clasts. Conglomerate beds, locally grading upward into chert, overlie thin shale beds. Some conglomerates contain well-rounded quartz grains, but sedimentary structures associated with modern subaerial channel or beach deposits, such as ripple marks, crossbedding, or desiccation cracks, are absent. Fractured and deformed clasts indicating that the clasts were not completely lithified when deposited are present sparsely. In addition, beds showing semi-plastic deformation and brecciation may have provided an intraformational source for the gravel detritus. These conglomerate beds are, therefore, interpreted as submarine slump and slide deposits which formed on an existing slope because of a decrease in sediment stability. The concentration of conglomerate beds in the lower chert and shale member, near the base of the middle member, and in the

upper chert and shale member roughly coincides with major sediment changes, mainly an increase in the mud content which would decrease the degree of slope allowable for sediment stability. Low slope stability is indicated for the sediment which yielded the interbedded chert and shale members. The geographically restricted Hot Springs Sandstone and sandstones within or above the upper novaculite member are plane bedded or massive, locally graded, and lack sedimentary structures expected in a subaerially formed sand deposit. Destruction of sedimentary structures by bioturbation is possible, but there is no trace of the fauna responsible or of bioturbation structures.

AGE

Fossils that can be dated are rare in the Arkansas Novaculite and reliable age determinations based on conodonts have been made for only part of the middle member (Cooper, 1931, 1933, and 1935; Hass, 1951 and 1956; and Hass and Huddle, 1965). Hass (1951) determined that the upper 8.5 m of the middle member is Early Mississippian (Kinderhook) in age and that the next lower 48 m (no date was possible on the lowest 50 m) of the middle member is late Devonian in age. This part of the Arkansas Novaculite correlates with all but the lowest part of the Woodford Formation and the Chattanooga Shale (Hass, 1956, and Hass and Huddle, 1965). The maximum age of the overlying Stanley Shale is Meramec (Mississippian) (Hass, 1950; Elias, 1959; and Gordon and Stone, 1969). This indicates a possible Osage (Mississippian) age for the upper novaculite member. The lower boundary is less satisfactorily limited because there are no dated fossils in the lower part of the Arkansas Novaculite or in the underlying Missouri Mountain Shale. The Blaylock Sandstone below the Missouri Mountain Shale contains Silurian graptolites (Miser and Purdue, 1929, p. 45). Miser and Purdue (1929, p. 58-59) correlate the lower novaculite member with the Middle Devonian Camden Chert of Tennessee on lithologic similarities. Comparison of the Arkansas Novaculite with the Penters Chert-Chattanooga Shale-Boone Chert sequence north of the Ouachita Mountains (Giles, 1935; and Kinney, 1946) yields striking similarities. The changes in the Arkansas Novaculite required to yield the northern sequence are increases in carbonate content for the lower and upper novaculite members and a decrease in chert content for the entire formation. A similar comparison with the rock sequences present to the west and north in Oklahoma is possible (Harlton, 1953; and Amsden, 1967). Correlation of the middle member with the Woodford Formation is based on fauna and lithology, and lithologic correlation of the Frisco and Sallisaw Formations with the lower member of the Arkansas Novaculite juxtaposes shelf bioclastic, arenaceous limestones and dolomitic limestones, with slightly calcareous novaculites (Amsden, 1960, 1961, and 1967). If these lithologic similarities also indicate similarities in age, then the lower novaculite member is Early to Middle Devonian in age. Scull (1958, p. 11) suggests correlation of the Missouri Mountain Shale with the Haragan Formation of Early Devonian age in the Arbuckle Mountains, thus reducing the maximum age of the Arkansas Novaculite to Middle Devonian. If the correlations suggested are tenable (the middle member of the Arkansas Novaculite-

Woodford-Chattanooga correlation is based on faunal and floral evidence), then the geographical change of novaculite to carbonate and the loss of chert ought to be explicable in terms of related, probably connected, environments of deposition. Additional information on the age of the Arkansas Novaculite will require study of the palynomorphs present throughout the formation at Black Knob Ridge and the Potato Hills.

An important question is whether the vertical succession of rocks in the Arkansas Novaculite indicates similar lateral facies changes or whether the members are isochronous and the vertical succession of rocks is due to large scale environmental changes. Evidence is meagre, but that available indicates the members are generally isochronous, in agreement with McBride's conclusion (in McBride and Thomson, 1970, p. 40-41) that the members of the Caballos Novaculite are isochronous. The middle member of the Arkansas Novaculite correlates with the widespread Woodford and Chattanooga Formations. Parts of the Arkansas Novaculite in Black Knob Ridge and the Potato Hills similar to the middle member elsewhere and in the proper stratigraphic position yield abundant *Tasmanites* which are common in both the Chattanooga and Woodford Formations. The top of the Arkansas Novaculite is very nearly isochronous based on widespread dating of the lower Stanley Shale as Meramec in age (Hass, 1950; Elias, 1959; and Gordon and Stone, 1969). Geologically instantaneous events in the form of slump, slide, and turbidite deposits mark the base of the middle member in many places and are common in the upper chert and shale member. These conglomerate beds cannot be convincingly correlated, although as indicators of changing depositional environment and tectonic regime, they provide some argument for at least roughly synchronous deposition.

DISCUSSION

From Caddo Gap, Arkansas, to Black Knob Ridge, Oklahoma, the Arkansas Novaculite gains shale and loses the novaculite members. The spiculitic pelletal character of the novaculite is retained in the interbedded spiculitic pelletal chert and shale at Black Knob Ridge. This gradual east-west change in the formation is similar to the much more abrupt south to north lithologic changes which probably reflect north-south compression by folding and the lesser distance to the craton to the north. Except for the generally spiculitic lower chert and shale member, the novaculite and chert and shale members reflect environments supporting benthonic and planktonic faunas respectively. The Arkansas Novaculite was formed in a succession of environments in which either chert and shale or novaculite was produced contemporaneously with shale or shelf carbonates respectively to the north and west. The Arkansas Novaculite formed in a depositional environment distinct from but grading into a depositional environment yielding shelf clastics and carbonates. A deep marine basin adjacent to a cratonic shelf provides the simplest model which explains the observed differences. The exclusion of sponges and a benthonic infauna from the middle and upper chert and shale is probably due to increased sediment influx and euxinic (starved) bottom conditions. Exclusion of planktonic faunas from the novaculite is seemingly more difficult, but may be explained in terms

of climatic controls and nutrient supplies (Lisitzin, 1972, p. 41). Recognition of lateral lithologic changes in the Arkansas Novaculite is a necessary condition for understanding the relationships of the Arkansas Novaculite with cratonic shelf deposits to the north and west.

SUMMARY

The Arkansas Novaculite can be divided into five members, which are, in ascending order: a lower chert and shale member; the lower novaculite member; middle chert and shale; upper novaculite; and upper chert and shale. The members are not recognizable everywhere, and the lower and upper novaculite members thin and become interbedded with shale north and west of Hot Springs. Novaculite is not present in Black Knob Ridge or the Potato Hills. The upper and lower boundaries are conformable. Local conglomerates are interpreted as submarine debris slides and turbidites and the Hot Springs

Sandstone as a submarine canyon fill. The basically isochronous members of the Arkansas Novaculite correlate generally with shallow-water carbonates north and west of the Ouachita Mountains, and demonstrate basin-to-shelf lithologic transition.

Novaculite is a siliceous rock composed of 5 to 20 millimicron polyhedral grains of quartz with little chalcedony and is distinguishable from chert in the chert and shale members. The included faunas indicate different depositional environments for the novaculite, and interbedded chert and shale members supporting benthonic and planktonic faunas respectively.

ACKNOWLEDGEMENTS

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**PETROGRAPHY OF STANLEY-JACKFORK SANDSTONES,
OUACHITA MOUNTAINS, ARKANSAS**

By Robert C. Morris¹

ABSTRACT

Quantitative data concerning grain size distributions and mineral compositions is now available from approximately 300 thin sections of Stanley and Jackfork sandstones. They comprise a representative geographic and stratigraphic suit of samples that span the range of flysch subfacies recognized for these rocks. There are considerable differences in textures and compositions between Jackfork and Stanley sandstones. Stanley sandstones are very fine grained (3.7 Mφ) and poorly sorted (1.4 σφ). They average 67.4 percent quartz, 7.5 percent feldspar, 3.1 percent rock fragments, and 18.1 percent matrix. Jackfork sandstones are fine grained (2.7 Mφ) and moderately sorted (0.86 σφ). They average 84.8 percent quartz, 0.9 percent feldspar, 2.2 percent rock fragments, and 11.4 percent matrix.

Within both Stanley and Jackfork sandstones textural variations apparently are related to type of flysch facies. The proximal turbidite facies is characterized by comparatively thick bedding units, coarse mean grain sizes, low matrix contents, low percentages of interbedded shales, and the presence of scour structures. Distal turbidites are characterized by rhythmically-bedded thin sandstones and comparatively larger proportions of interbedded shales.

The consistently feldspathic Stanley turbidites entered the depositional area from the southeast as indicated by paleocurrents. Jackfork sandstones contain varying amounts of feldspar, as well as considerable variation of paleocurrent readings, suggesting that point sources from the northeast as well as the southeast supplied clastics to the deep abyssal plain.

INTRODUCTION

Although there is presently available considerable information on the petrography of sandstones of the Stanley and Jackfork Groups in the Arkansas-Oklahoma area, comparatively little of this information is either available in published form or has been regionally synthesized. The purpose of this paper is to both present some of the unpublished data and synthesize it with published information in consideration of probable source areas and dispersal systems for Stanley-Jackfork rocks. Published reports include those of Bokman (1953), Goldstein (1959), Goldstein and Hendricks (1962), Klein (1966), and Seely (1963). Unpublished reports include those of Burkart (1969), Hamilton (1973), Hill (1967), Howard (1963), Moretti (1958), Morris (1964), Simonis (1967), and Russell (1969), plus the continuing work of the author and his students.

Figure 1 shows the relationships of lithostratigraphic units discussed in this paper. The Stanley nomenclature is from Harlton (1959) and the Jackfork from Morris (1971a). Locations of major sampling sites are shown on Figures 2 and 3. Many of the samples were collected from measured sections. The larger number of samples were collected from the frontal Ouachitas, where thin-section studies have accompanied the field mapping of Burkart (1969), Hamilton (1973), Russell (1969), and Simonis (1967).

TEXTURE

Quantitative textural data was obtained by measuring in thin section the long dimensions of 200 grains per slide, plotting the results on arithmetic probability paper, and calculating phi mean diameter, phi deviation (sorting), and phi skewness (Inman, 1952). In some cases the formulas of Folk (1968) were utilized. Table I contains a summary of the textural data for both Stanley and Jackfork rocks, whereas Tables II and III present the number of times various textural classes were encountered by several investigators.

SYST.	CENTRAL OUACHITAS (OKLA.)	FRONTAL OUACHITAS (ARK.)	OZARK REGION (ARK.)
PENNSYLVANIAN	ATOKA	ATOKA	ATOKA
			BLOYD
	JOHNS VALLEY	JOHNS VALLEY	HALE
	JACKFORK	JACKFORK	
	GAME REFUGE		
	WESLEY	BRUSHY KNOB	
	MARKHAM MILL		
	PRAIRIE MOUNTAIN		
	WILDHORSE MOUNTAIN	IRONS FORK MOUNTAIN	
	CHICKASAW CREEK	CHICKASAW CREEK	PITKIN
MOYERS	MOYERS		
MISSISSIPPIAN	STANLEY	STANLEY	FAYETTEVILLE
	UPPER	TENMILE CREEK	BATESVILLE
	LOWER		RUDELL
	TENMILE CREEK	HOT SPRINGS	MOOREFIELD
	ARKANSAS NOVACULITE (upper part)	ARKANSAS NOVACULITE (upper part)	BOONE

Figure 1. -- Correlation chart showing approximate placement of Stanley and Jackfork rocks.

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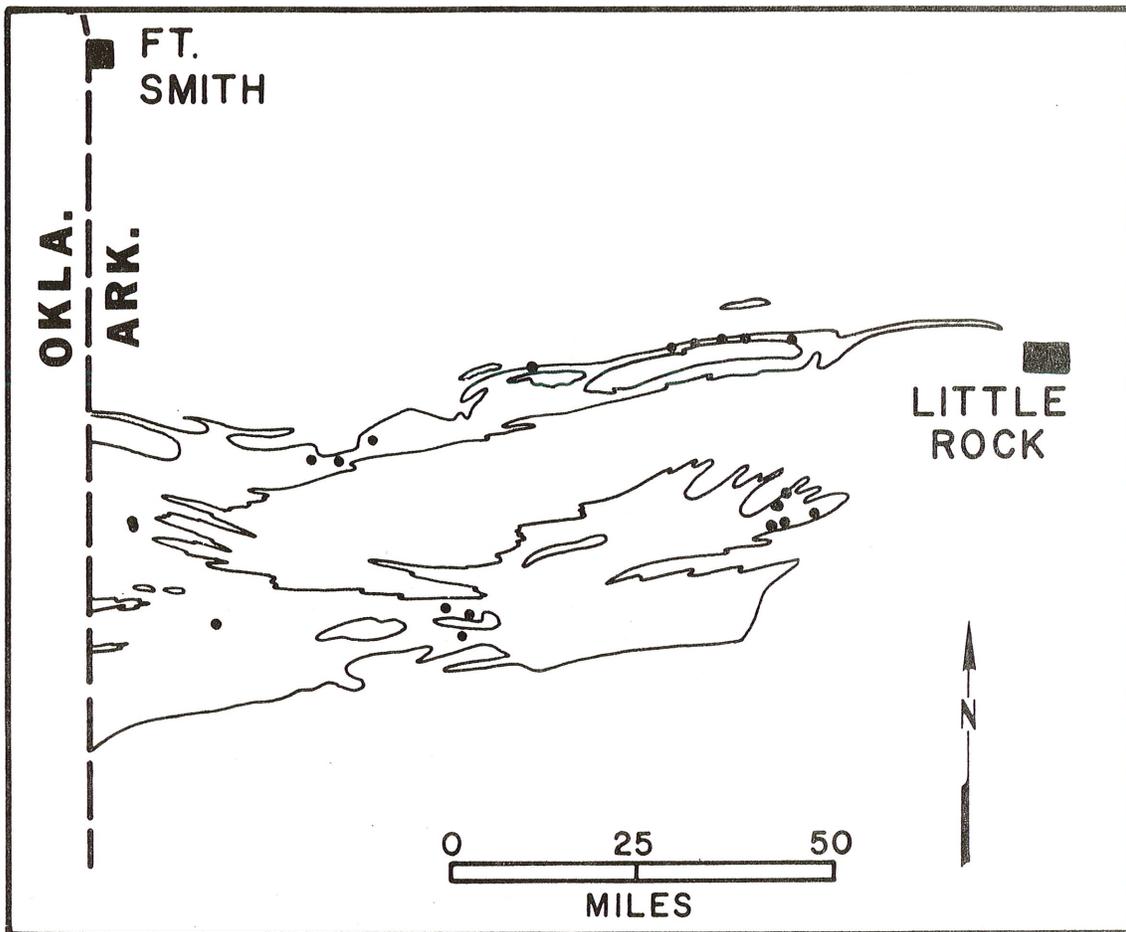


Figure 2. -- Generalized outcrop pattern of Stanley Group showing sample locations. Multiple samples were taken from measured sections.

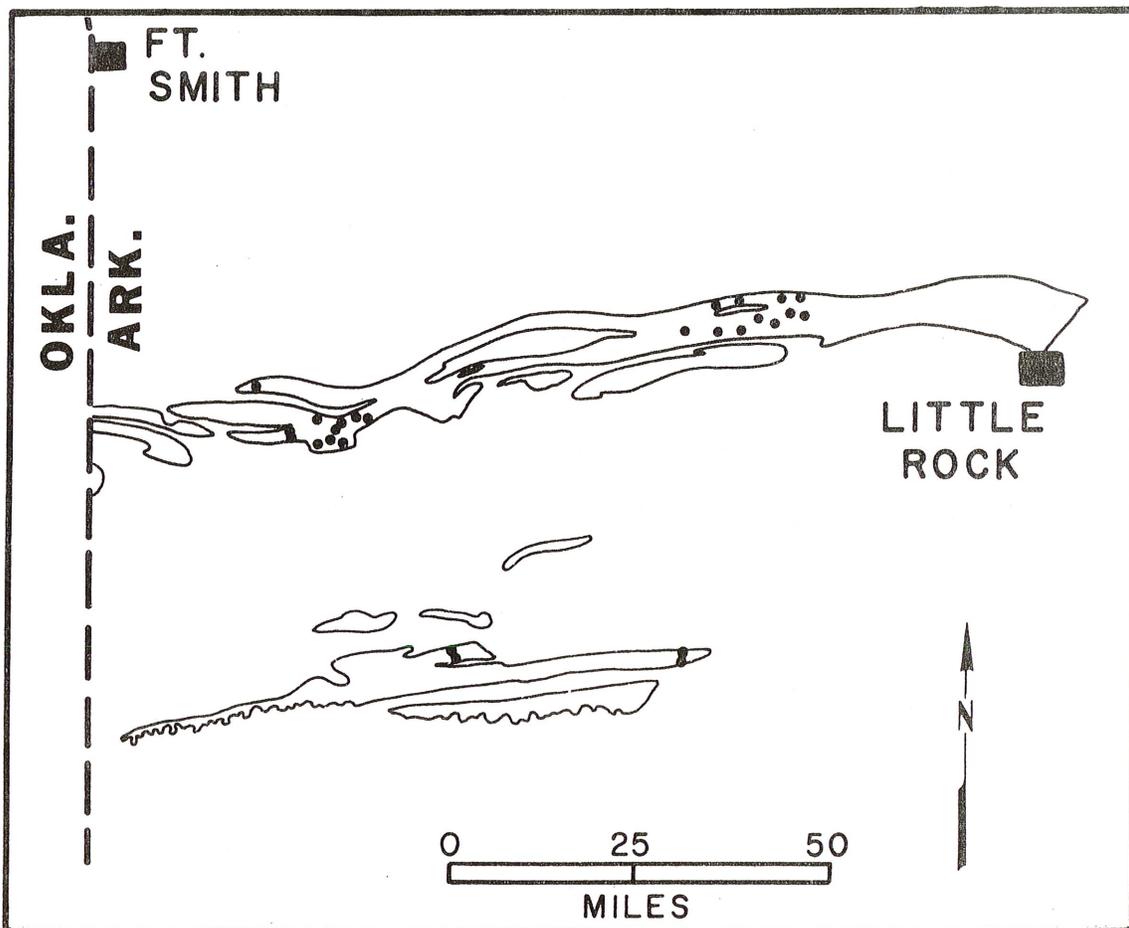


Figure 3. -- Generalized outcrop pattern of Jackfork Group showing sample locations. Multiple samples were taken from measured sections.

TABLE I - SUMMARY OF SANDSTONE PETROGRAPHY

STRATIGRAPHIC UNIT	Mφ	σφ	αφ	QUARTZ %	FELDS-- SPAR %	ROCK FRAGS. %	MATRIX %	OTHER %
Jackfork Group								
Brushy Knob (69)	2.47	0.77	0.18	86	1.1	2.5	9.3	1.1
Irons Fork Mountain (19)	3.02	0.76	0.23	87	0.1	1.4	9.8	1.7
Southern Ouachitas (61)	2.81	1.01	0.21	75	1.0	2.1	14.2	7.7
Undifferentiated ¹ (22)	2.70	0.90	-----	91	0.6	1.3	7.0	0.1
Brushy Knob ² (11)	2.52	0.72	-----	97.9	trace	trace	1.1	---
Irons Fork Mtn. ² (19)	2.89	0.78	-----	95.6	trace	trace	3.4	---
Mean (201)	2.69	0.86	0.20	84.8	0.9	2.2	11.4	0.7
Stanley Group								
Tenmile Creek ³ (25)	3.93	1.67	0.35	66.9	4.7	4.5	21.9	1.9
Moyers (24)	3.69	1.30	0.07	58.5	12.5	4.0	16.6	8.8
Moyers ⁴ (17)	3.48	-----	-----	72.3	9.4	0.5	16.8	1.0
Moyers ² (10)	3.85	1.59	-----	63.5	4.9	7.6	22.8	1.2
Undifferentiated ¹ (21)	3.60	1.30	-----	68.8	9.4	1.4	20.0	0.4
Hot Springs ³ (7)	3.35	0.78	0.25	89.0	trace	----	1.0	10.0
Mean (104)	3.69	1.40	0.22	67.4	7.5	3.1	18.1	3.9

Numbers in parentheses refer to total thin sections studied.

- 1 Data from Simonis (1967)
- 2 Data from Burkhardt (1969)
- 3 Data from Hamilton (1973)
- 4 Data from Russell (1969)

Mean grain sizes of most analyzed sandstones range from the fine sand to coarse silt range (2 to 5 ϕ) for Stanley sandstones and in the medium to very fine sand group (1 to 4 ϕ) for the Jackfork sandstones (Table II). The coarsest grains in each Stanley sample fall in the fine to coarse sand sizes. Jackfork sandstones are similar, although some proximal turbidites in the upper part of the section have grains that range up to pebbly sizes (-2 to -6 ϕ). Undoubtedly, the higher matrix content of most Stanley sandstones is the chief factor contributing to their finer mean grain sizes. Morris (in press) has presented preliminary evidence that suggests some textural correlation between designated flysch facies. There is a progressive decrease in mean grain sizes from proximal turbidites to distal turbidites, a characteristic confirmed by additional, uncompleted work.

Stanley sandstones are poorer sorted than those in the Jackfork. The Tenmile Creek Formation appears to be the poorest sorted Stanley unit (1.67 ϕ deviation). Jackfork sandstones are most poorly sorted across the Southern Ouachitas where they have a higher matrix content. In a few exceptional cases poor sorting is due to pebbles and granules concentrated near bases of proximal turbidites. However, the common trend is that sorting increases with mean grain size, as illustrated by Jackfork sandstones (Fig. 4).

The matrix contents of Stanley sandstones is much higher than those in the Jackfork (Table I) in the frontal Ouachitas, but in the southern Ouachitas matrix contents are approximately equal. The Hot Springs sandstones are all arenites (Williams and others, 1954) but other sandstones studied within the Tenmile Creek have matrix contents which classify them as wackes. The very fine grained sandstones and siltstones in shaly flysch (Morris, 1974a) have the highest matrix contents. The few arenites within the Moyers Formation occur in massive, proximal turbidites with scoured bases (subfacies 3A). Crushing and disintegration of pelitic sedimentary and metamorphic clasts following sedimentation may account for much of the matrix. Of course, within all but the massive sandstones, the matrix content increases upwards within a bed. The most friable sandstones within both the Stanley and Jackfork are those with an appreciable matrix content, a characteristic suggesting that matrix may have inhibited fluid flow, and therefore cementation, during burial. Matrix in Jackfork sandstones has been described as "patchy" by Moretti (1958), but the same characteristic is present in the Stanley as well. In most cases, patchy matrix occurs within sandstones exhibiting slurried bedding (Morris, 1971b), which suggests that they were emplaced as debris flows that never developed into turbidity currents.

We have not made quantitative studies of grain roundness, but observations suggest that subangular to subrounded shapes dominate both Stanley and Jackfork sandstones. Both angular and rounded shapes are also present, however, large sand grains are commonly the better rounded whereas very fine sand or silt is generally the more angular. The common occurrence of etching, intergrowths, and over-

growths normally inhibits determination of original grain shapes in the arenites, but in most wackes the original shapes are better preserved.

Most quartz and feldspar grains are roughly equidimensional in both Stanley and Jackfork sandstones. However, Bokman (1953) measured elongation ratios (long axis/short axis) and concluded that Stanley sandstones are less elongated than those of the Jackfork. Hill (1967) similarly found most elongation ratios of Stanley rocks in Oklahoma to range between 1.0 and 2.0. In some cases Jackfork sandstones in the frontal Ouachitas appear to be elongated due to pressure solution removal of some silica. True elongated grains are present at several stratigraphic intervals within the Jackfork, but do not appear to be common.

Grain sphericity and grain roundness should provide some clues concerning original source and abrasional history. I would agree with Hill (1967) that equant quartz grains are not necessarily indicative of igneous sources as suggested by Bokman (1953). Although the angular quartz grains might have original crystalline sources, most of the grains show considerable abrasional history. It does not seem possible that this much abrasion could have occurred during the emplacement of these turbidite sandstones; therefore, we must conclude that much of the sand-sized sediment originated from pre-existing sandstones and/or weakly metamorphosed quartzites.

Skewness is consistently positive in both Stanley and Jackfork sandstones, with the exception of a few granular or pebbly sandstones (Table III). The "typical" sandstone, therefore, is one with an excess of fine-grained material (matrix). As yet, no one has detected any lateral or vertical trends to the skewness values of these rocks.

GRAIN CONTACTS

Where the framework grains are matrix supported, the grain boundaries are generally sharp and relatively unaffected. Exceptions occur in some rocks with a high clay matrix, where the chemical composition of the matrix and the interstitial water have corroded the grain contacts.

If the framework grains are touching and there has been little solution, tangential contacts can be seen. Generally there has always been a slight amount of solution such that smooth contacts are the rule in the cleaner sandstones.

The most common type of grain contacts, especially in the well-sorted sandstones, are concavo-convex. Generally the heavy minerals, and sometimes calcite grains, embay themselves into the less resistant quartz grains. In the arenites quartz grains will often embay themselves into their neighbors, and subsequent "healing" of the contacts makes it extremely difficult or impossible to distinguish these composite grains from highly strained quartz or quartzite grain.

Where the sands are clean, all gradations exist between

TABLE II – MEAN ($M\phi$ or Mz) DIAMETERS OF STANLEY AND JACKFORK SANDSTONES

DIAMETERS IN $1/2 \phi$ INTERVALS	NUMBER OF SAMPLES PER $1/2 \phi$ INTERVAL							
	JACKFORK				STANLEY			
	(1)	(2)	(3)	(3)	(4)	(5)	(1)	(2)
-0.5 – 0.0	1							
1.0 – 1.5	1							
1.5 – 2.0	3	1	3					
2.0 – 2.5	43	3	2					
2.5 – 3.0	73	19	10	4	2	2	1	
3.0 – 3.5	48	7	5	7	7	3	3	2
3.5 – 4.0	10		2	7	8	12	12	4
4.0 – 4.5				3		7	3	4
4.5 – 5.0						6	3	
5.0 – 5.5						1	2	
5.5 – 6.0						1		

- (1) $M\phi$ – Morris (179 Jackfork, 24 Stanley)
- (2) Mz – Burkart (30 Jackfork, 10 Stanley)
- (3) $M\phi$ – Simonis (22 Jackfork, 21 Stanley)
- (4) Mz – Russell (17 Stanley)
- (5) Mz – Hamilton (32 Stanley)

TABLE III – SORTING AND SKEWNESS VALUES IN STANLEY AND JACKFORK SANDSTONES

TEXTURAL PARAMETER	NUMBER OF TIMES EACH VALUE DETERMINED							
	STANLEY				JACKFORK			
	(1)	(2)	(3)	(4)	(1)	(2)	(3)	
SORTING								
0.35 to 0.50 ϕ	0	0	0	0	16	0	2	
0.51 to 0.70 ϕ	0	1	0	1	39	6	8	
0.71 to 1.00 ϕ	4	3	0	0	62	9	19	
1.01 to 2.50 ϕ	20	17	10	24	62	7	1	
SKEWNESS								
+ 1.00 to +0.31	0	—	—	14	38	—	—	
+0.30 to +0.11	11	—	—	8	100	—	—	
+0.10 to -0.10	11	—	—	0	38	—	—	
-0.11 to -0.30	2	—	—	3	3	—	—	

- (1) Morris (24 Stanley, 179 Jackfork, after Inman, 1952)
- (2) Simonis (21 Stanley, 22 Jackfork, after Inman, 1952)
- (3) Burkart (10 Stanley, 30 Jackfork, after Folk, 1968)
- (4) Hamilton (25 Tenmile Creek, excluding Hot Springs, after Folk, 1968)

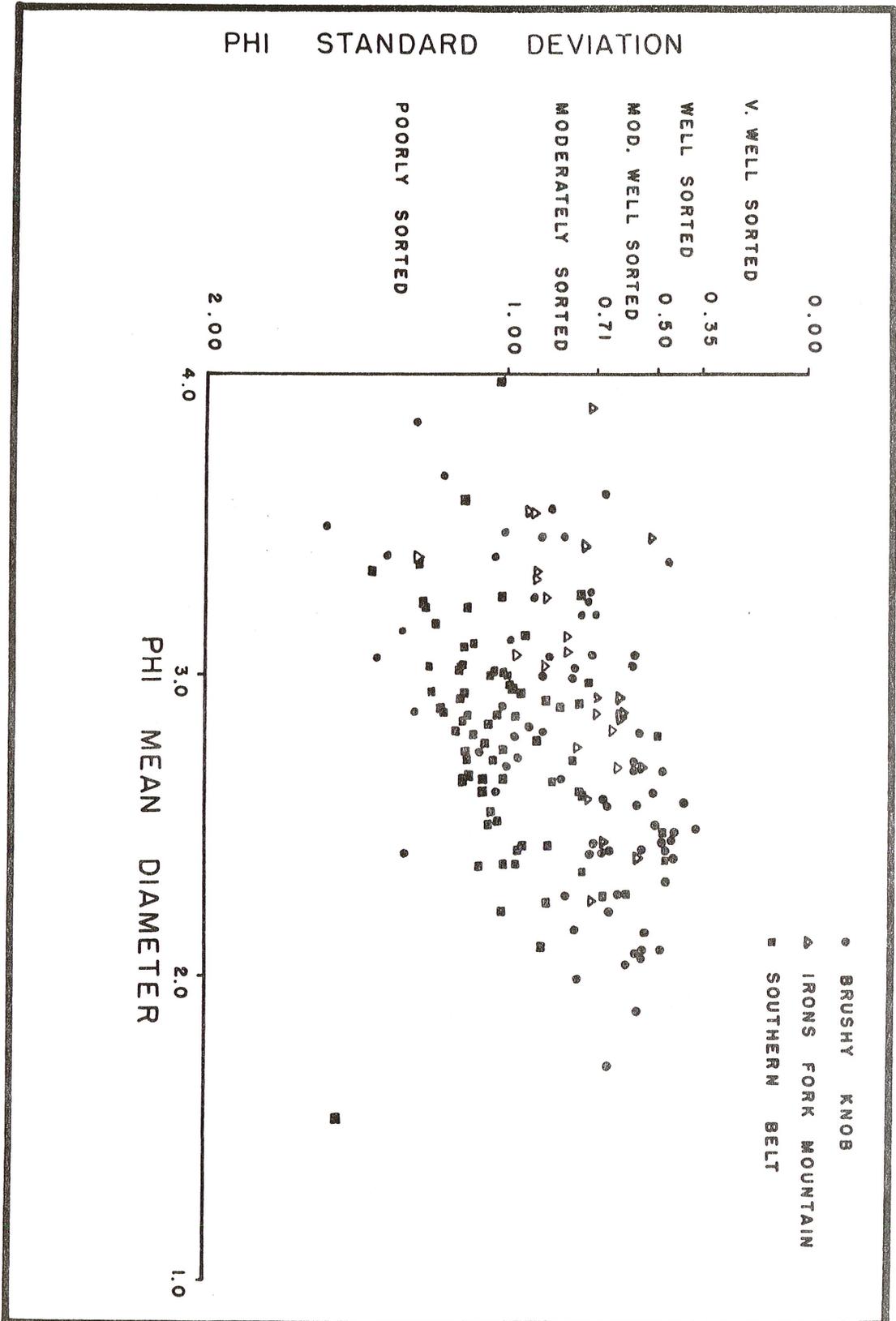


Figure 4. -- Graph of Jackfork samples where sorting ($\sigma\phi$) and mean grain size ($M\phi$) are plotted. There is a general improvement of sorting with increased mean grain size.

concavo-convex and sutured contacts. In sandstones containing a slightly higher amount of matrix, the contacts often exhibit sharp, jagged lines filled with a ferruginous clay. The chert or composite grains do not tend to develop stylolitic contacts against adjoining quartz grains. However, the polycrystalline quartz grains have sometimes been observed to react with the matrix so that it is difficult to pick a grain-matrix contact. The intricate sutured contacts within the polycrystalline quartz grains were developed during the previous diagenetic or metamorphic cycle.

Stylolites were found in both Brushy Knob and Irons Fork Mountain sandstones, with the columns oriented perpendicular to bedding, irrespective of the dip. This suggests that the stylolites developed through sedimentary loading rather than folding. At the outcrop the beds often split along these stylolitic seams, exposing a jagged outline whose trend is parallel to bedding. In thin section the stylolites appear to be composed of clay material rich in organic matter and/or iron oxide. Laminated, carbonaceous sandstones are generally replete with minute stylolitic seams. In one thin section the writer observed a stylolite to change abruptly into a quartz vein, suggesting that some of the quartz veins may be due to endogenetic processes.

The development of those complex contacts and stylolites occurs through a process of pressure solution. Heald (1958) noted the relationship between clay content and reduction in porosity as a result of pressure solution in silica. He believed that clay acts similar to a catalyst, that it accelerates the process if other conditions are suitable. Thompson (1959) believed that silica dissolution would be enhanced at elevated pH values, which in turn could be brought about by the release of potassium from an illitic clay. Dapples (1959) also indicated that temperature may be a more important factor.

In the present study, many of the clearest sutured contacts were in unstrained grains. Lerbekmo and Platt (1962) noted the presence of iron minerals as well as carbonaceous matter along continuous, stylolitic seams. They suggested that ferric oxide in the presence of sulphur-bearing carbonaceous organic matter may be reduced to form iron carbonate and iron sulfide. This reaction releases hydroxyl ions, causing a local increase in pH, and thereby promotes the solution of silica at points of stress. It is significant that many organic-rich horizons in the Jackfork have a considerable amount of small-scale, continuous stylolites parallel to bedding. However, many of the clean sandstones appear to be devoid of carbonaceous matter and still have interlocking grains, generally of the concave-convex variety. As Moretti (1958) indicated, the amount of interlocking bears an inverse relationship to the amount of argillaceous and lithic matrix.

In summary, it seems that two processes are active in altering the grain contacts: (1) true pressure solution brought about entirely by the load of the overlying rock and (2) solution caused by pressure *and* local increases in pH due to the liberation of certain cations which produce

the corrosion effects, the stylolitic contacts, and the true stylolites.

Quartz overgrowths are most common in the matrix-free, well-sorted sandstones with a minimum of grain interpenetration. Where there are tangential contacts of the original grains, it may be that the secondary silica entered from somewhere else, liberated either from pressure solution or perhaps brought in by highly charged siliceous waters forced out of argillaceous material high in adsorbed silica. The overgrowths are generally separated from the detrital grain by small bits of foreign material. Large amounts of adhered argillaceous material apparently inhibits overgrowth development. If given space, the overgrowths will develop euhedral outlines and where 3 grains share a mutual pore area, each will develop its own optically continuous overgrowth out to a common point. Of course, where much pressure solution has occurred, overgrowths are almost impossible to find.

The degree to which Stanley and Jackfork sandstones are cemented is apparently related not only to the extent to which grain contact modifications of the type noted above have occurred, but also to the degree of tectonic disturbance they have undergone. This appears to be witnessed by the fact that Stanley-Jackfork sandstones are generally much less-well cemented in the southern Ouachitas than in the more severely disturbed area of the frontal Ouachitas. However, in both areas cementation is erratic.

MINERAL COMPOSITION

The major compositional categories considered in this study are presented in Table I. These include quartz, feldspar, rock fragments, matrix, and "other" material (accessory minerals). The coarse detrital composition of each sample analyzed, recalculated to omit matrix and accessory minerals, is shown in Figures 5 and 6. The quantitative data was determined by counting from 100 to 200 points per thin section.

Quartz

The Stanley and Jackfork sandstones are composed primarily of quartz, with the Jackfork being the more quartzose. Three major forms of quartz were recognized and counted: unstrained, strained, and polycrystalline grains. The latter include chert, meta-quartzite, and complexly strained quartz.

Common (unstrained) quartz is the dominant quartz type in both the Stanley (Hill, 1967) and the Jackfork (Fig. 7). This type is composed of single quartz crystals with an unique and non-undulatory extinction pattern. Unstrained quartz is probably of little value as a provenance indicator since it is a common type in all quartz-bearing rocks.

Strained quartz is defined as requiring more than one

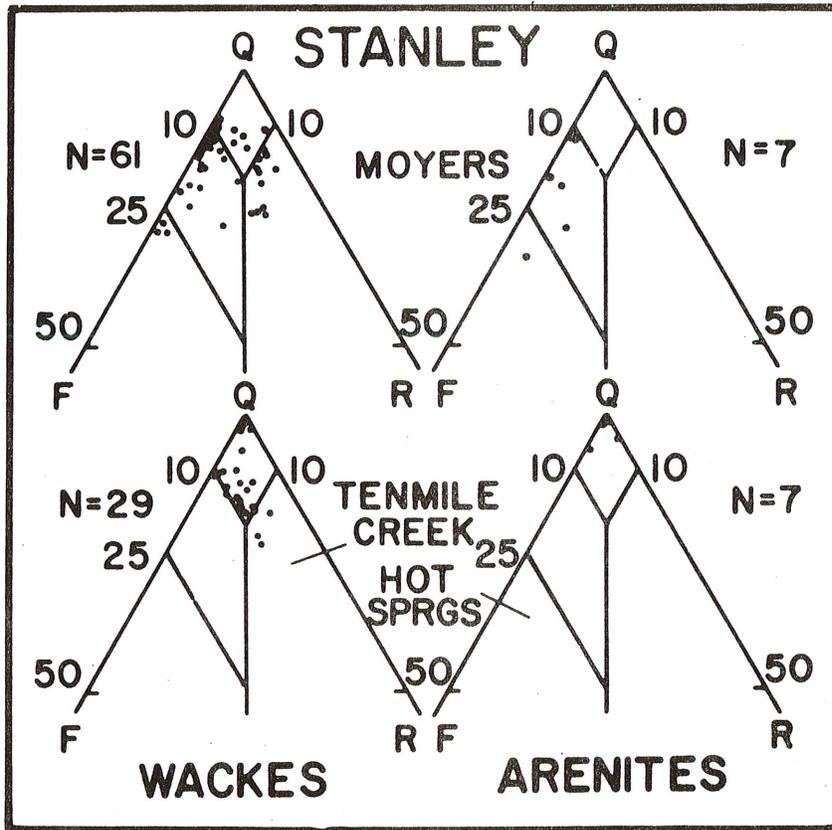


Figure 5. -- Lithology of framework grains, Stanley sandstones, plotted on upper part of diagrams by Williams and others (1954). There were no arenites in the Tenmile Creek or Wackes in the Hot Springs Sandstone.

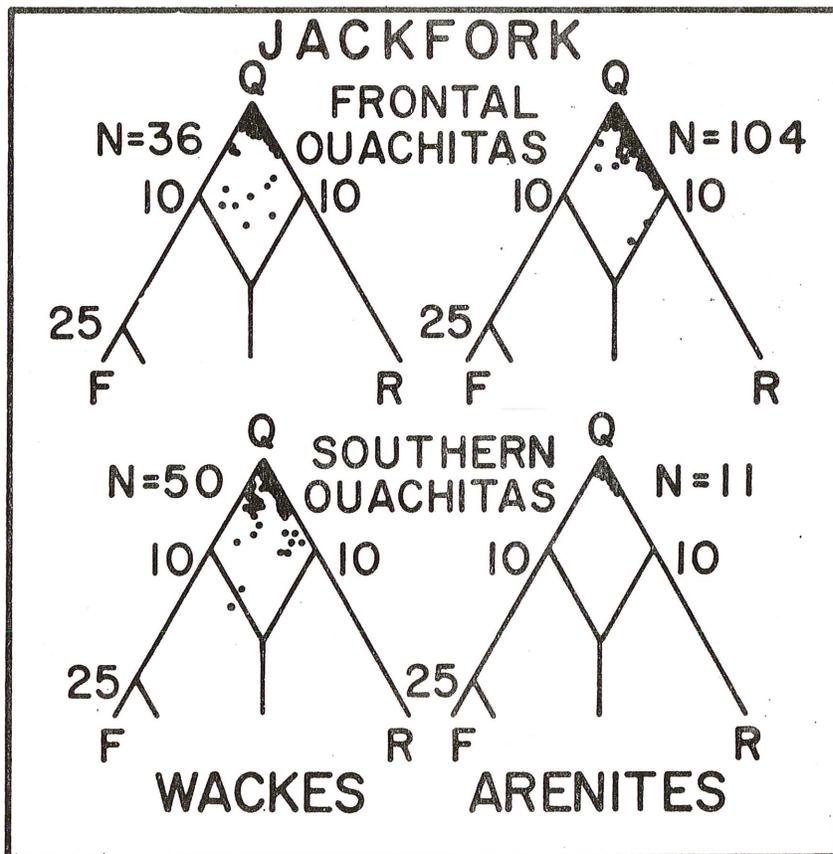


Figure 6. -- Lithology of framework grains, Jackfork sandstones, plotted on the upper part of diagrams by Williams and others (1954).

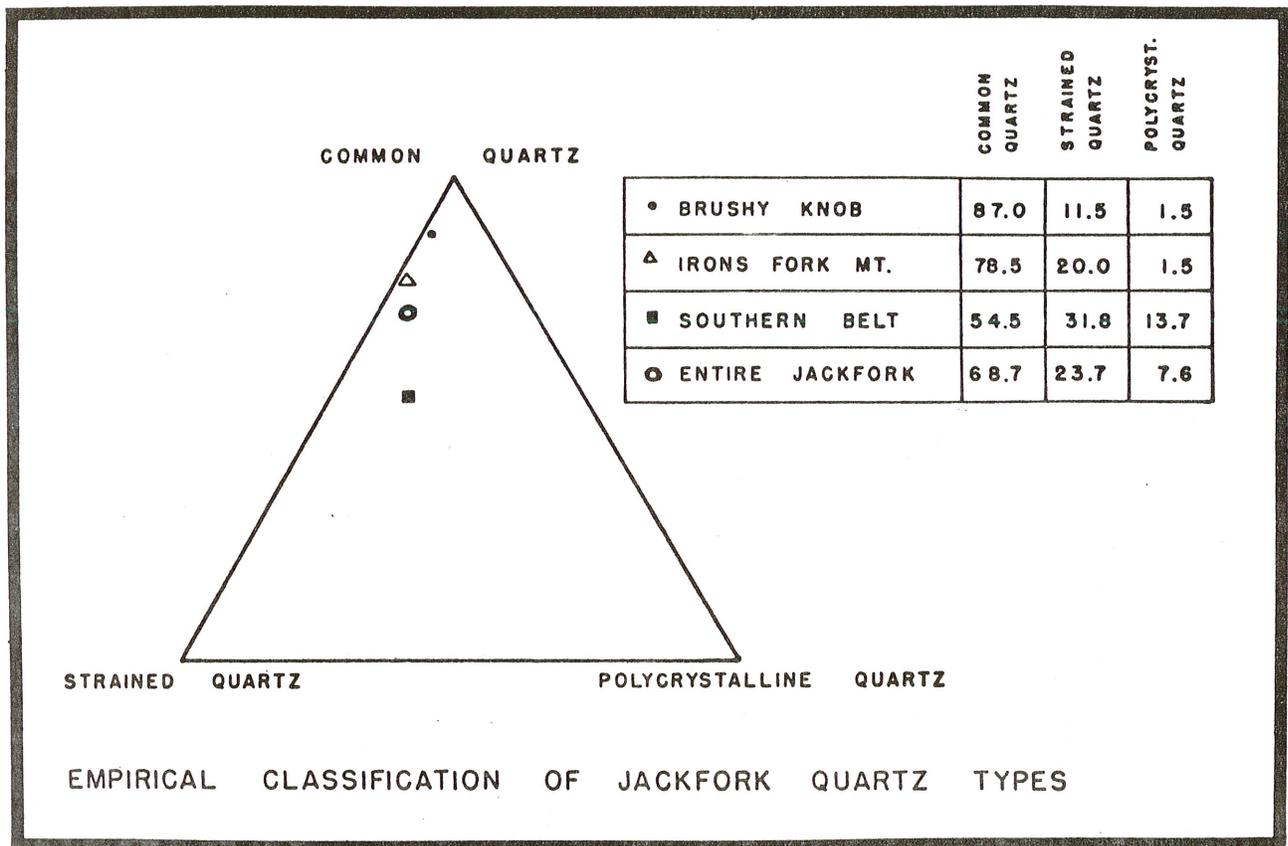


Figure 7. -- Plot of common, strained, and polycrystalline quartz grains of Jackfork sandstones. The greater amount of strained and polycrystalline grains across the southern Ouachitas may reflect contributions from an additional source.

degree of rotation for complete extinction. Probably the greater amount of strained quartz grains inherited their characteristics in the source area. Locally common in more deformed belts are strained grains produced by pinching against a neighboring grain. Strained quartz is also common to all types of quartz-bearing rocks and appears to have little use as an indicator of provenance.

The polycrystalline grains form a varied assortment of highly strained, stretched, mosaic, metaquartzite, and true chert grains. In Oklahoma, Hill (1967) observed that the percent of polycrystalline grains increased with increased mean grain sizes. This suggests that much of the fine-grained, normal and strained quartz may have been derived from fragmentation of polycrystalline quartz into its component grains. There seems to be little doubt that the polycrystalline grains were derived chiefly from meta-sedimentary sources. The greater abundance of polycrystalline grains in the Jackfork of the southern Ouachitas as compared to the frontal Ouachitas suggests the possibility of a different source area for the regions.

The quartz grains commonly contain inclusions, the most common being irregular microlites and vacuoles or bubble inclusions. The recognizable mineral grains within the quartz grains include acicular rutile, calcite, zircon, biotite, chlorite, tourmaline and opaque minerals.

Feldspar

Although Stanley sandstones are much more feldspathic than those of the Jackfork (Figures 5 and 6), the composition and textures of feldspar grains in the two units are otherwise similar. Feldspars that have been recognized include plagioclase (both oligoclase and andesine), orthoclase, microcline, and possibly sanidine. Many of the plagioclase grains are angular. Similar angular plagioclase grains occur in tuffs within the lower Stanley, suggesting that the plagioclase grains may have a volcanic origin. Other feldspars vary from subangular to well rounded, indicating moderate amounts of transport, or perhaps a recycled sedimentary source. Much of the feldspar is altered in different degrees to calcite, sericite, kaolinite, or iron oxide. The replacement may be in irregular, structureless patches or else the entire grain may be replaced.

Rock Fragments

Rock fragments comprise from one to eight percent of the Stanley and around two percent in the Jackfork sandstones (Table I). Hill (1967) similarly reported the Stanley to contain approximately one percent rock fragments in Oklahoma. Because the matrix and argillaceous rock fragments are so similar, some of the apparent discrepancy in volume of rock fragments reported may be due to identification differences. Besides shale clasts,

PROVENANCE AND DISPERSAL SYSTEM

Stanley

other sedimentary rock fragments include limestone (generally dissolved so that only molds remain), siltstone, sandstone, chert and siliceous shale. Metamorphic clasts are more abundant than either volcanic or sedimentary ones, most of which are mica schists or quartz mica schists. The schists are composed of parallel plates and shreds of muscovite, with varying amounts of elongate, silt-sized quartz grains. Chlorite schists and chlorite-garnet schist clasts are common in the Stanley. Both units also contain clasts of slates and phyllites. Only a small quantity of volcanic rock fragments were recognized, most in Stanley rocks. The most common were brown, devitrified glasses with minute feldspar laths. A trace of crystalline rock fragments was reported by Hill (1967) in Stanley sandstones, but they do not appear to be a common constituent. The types of rock fragments do not support the hypothesis of Bokman (1953) that the Stanley clastics originated from a granitic source.

Accessory Minerals

The most abundant light accessory minerals are the micas, biotite and muscovite, and chlorite. The later two occur both as detrital and authigenic minerals. Calcite and chalcedony also occur as authigenic minerals in the Stanley. Jackfork sandstones generally have just muscovite as a light, accessory mineral.

There is considerable disparity in the reported heavy minerals of Stanley and Jackfork sandstones (Bokman, 1953; Goldstein, 1959; Hill, 1967). Most of our work in Arkansas suggests that in the Stanley zircon, tourmaline, and opaque minerals (ilmenite, leucosene, and magnetite) are the most abundant and occur in approximate equal amounts. They are followed in abundance by garnet and rutile, with trace amounts of apatite and questionable staurolite. Jackfork heavy minerals comprise an even more stable suite (Morris, 1964), dominated by zircon, tourmaline, and opaque minerals, with traces of rutile and ilmenite. Most of the heavy minerals in both Stanley and Jackfork sandstones show considerable rounding, indicating a long (and probably multiple) abrasional history. Most commonly the heavy minerals are coarse-silt sized, although scattered ones are larger than the surrounding quartz grains. Hill (1967) suggested that this anomalous grain size in heavy minerals may indicate that they were originally transported with coarser grained, better-rounded quartz that subsequently became fragmented during transportation or deposition. Some of the heavy minerals show solution effects in addition to abrasion.

Heavy minerals occur disseminated throughout the rock in the thicker bedded sandstones. However, in thinly-bedded distal turbidites, high concentrations of heavy minerals commonly occur along stratification planes. This phenomena of heavy mineral placers is ascribed by Bouma (1972) as characteristic of contourites. These deposits are essentially the product of reworking by deep, marine currents.

Stanley paleocurrents of Arkansas and Oklahoma have been summarized by Morris (1974a). These paleocurrents, derived exclusively from sole markings of turbidites, show a northwesterly flow that becomes more westerly towards the Oklahoma Ouachitas. As Stanley time progressed, there was also an overall shift to more westerly flow. Most of the thin-bedded siltstones were emplaced northward whereas the thicker beds were emplaced westward. No significant data are available for proximal turbidites and debris flow sandstones.

During Stanley time the bulk of the sand-sized clastics apparently entered the eastern end of the Ouachita trough as turbidity flows and sandy debris flows derived from the southern Appalachian region (Fig. 8). These built a westward-directed deep-sea fan which merged outward into a flat abyssal plain at the southern margin of the North American craton. Sandy debris flows and scoured proximal turbidites may have been localized within leveed channels and the suprafan depositional lobes. This inner fan area moved laterally as well as forwards or backwards, depending upon rates of sediment influx and basin subsidence. Directly downslope from the depositional lobes was a broadening, outer fan, characterized by rhythmic turbidites and black shales (sandy flysch of Morris, 1974a). Inactive outer fans and broad expanses of the basin plain were characterized by thinner-bedded turbidites and more abundant black shales (shaly flysch).

The origin of the clastic debris is obviously speculative.

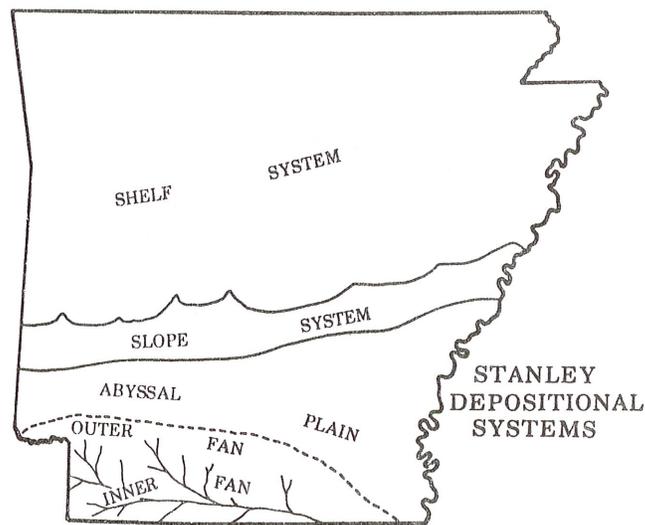


Figure 8. -- Postulated depositional systems during sedimentation of Stanley rocks.

Morris (1974b) has suggested that a NE-SW plate junction east of the present Appalachian Mountains involved collision with the African plate during the late Devonian, resulting in a rising, folded mountain chain of meta-sedimentary and sedimentary rocks. A deltaic system would have carried the eroded material to the shelf, from which it entered the ocean basin to create the ever expanding subsea cone. A small amount of sand entered the trough from the cratonic shelf, fed by deltas that prograded southward across the Illinois basin. Minor zones of contorted bedding containing limestone, chert, and sandstone blocks were created by southward slumping along the unstable northern slope of the trough. Several tuffs at the base and one tuff at the top of the Stanley attest to a volcanic source, probably to the south of the present Ouachitas. This volcanic arc is likely the result of subduction along the African-American plate boundary that curved westward from the Appalachian region to create an arc-trench system at some distance south of the Ouachita deep-sea plain. The small volume of the volcanics suggests two possibilities about the nature of the plate boundary. One possibility is that the plate boundary was characterized by considerable strike-slip movement and hence only minor subduction and volcanism adjacent to the Ouachita area. Another possibility is that the boundary possessed a subduction zone, but that it was south-dipping and an intervening trench acted to inhibit transfer of significant quantities of volcanics into the Ouachita area.

Jackfork

Jackfork paleocurrents show westerly flow with components to the northwest and southwest, but no overall change in current flow between lower and upper Jackfork rocks. Most of these readings were taken from distal turbidites.

At the onset of the Morrowan Epoch lowered sea levels caused the northern shelf to be eroded, resulting in a flood of nonfeldspathic clastics from the northeast that flowed off the shelf into the Ouachita trough. Major slumps and slides moved southward to the edge of the continental slope and rise, probably triggered by earthquakes associated with the plate-boundary system to the south. The southeasterly flood of sparsely feldspathic-lithic sands that had begun in Mississippian time continued into the Pennsylvanian, but ceased to be important before

Jackfork deposition ended. The resulting westward-building, deep sea cone, fed by essentially two point sources, has the most diverse mineral compositions in the extreme northeast and southeast ends of the outcrop belt. Westward, out of the region of the inner fan, the sandstones increase in number, decrease in thickness, whereas shales increase in total volume (Fig. 9). The clastics off the northeast shelf, fed by a substantial delta system, increased in volume as the overlying Atoka flysch succession developed. By Des Moinesian time (Pennsylvanian) the plate boundary was converted to a belt of uplifted tectonic lands that resulted in northward sliding of the sedimentary succession as continentward-directed folds and thrust sheets.

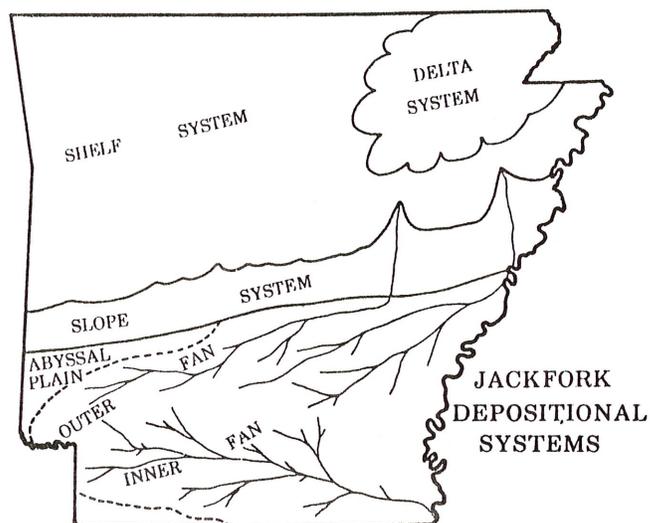


Figure 9. -- Postulated depositional systems during sedimentation of Jackfork rocks.

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**FLYSCH FACIES OF THE OUACHITA TROUGH -- WITH EXAMPLES
FROM THE SPILLWAY AT DE GRAY DAM, ARKANSAS**

By Robert C. Morris¹

INTRODUCTION

The use of the term flysch was chiefly popularized by Cline (1960, 1970) who demonstrated convincingly the similarity of the Late Paleozoic clastic rocks of the Oklahoma Ouachitas to the flysch sediments in the classical areas of the Carpathians, Alps, and the Apennines. Cline (1970) lists seventeen characteristics of flysch in the Alps and Carpathians, and suggests that these characteristics are duplicated in the Stanley-Jackfork-Johns Valley-Atoka sequence in the central Ouachita Mountains of Oklahoma. The occurrence of these diagnostic characteristics in the southern and frontal Ouachitas of Arkansas suggests that the Carboniferous of Arkansas is also a flysch facies.

Recurring rock types or associations can be grouped into four major facies, of which three can be further differentiated into subfacies. Figure 1 is a conceptual model of the facies relationships, designated by number and letter (descriptive) and name (interpretative and genetic). This scheme allows reference to these facies by number without endorsement of the interpreted depositional processes. The facies symbols in Figure 1 are placed alongside the stratigraphic section in this report to designate interpreted subfacies.

The 1000-foot thick succession of upper Jackfork rocks exposed along the spillway at De Gray Dam provides a unique opportunity to study most types of Carboniferous flysch subfacies in the Ouachita Mountains. The following discussion is intended to extend to all Carboniferous rocks throughout the Ouachita Mountains but the measured section (Fig. 2) and all illustrations used here come from the spillway section. Figure 3 is a legend for the stratigraphic section, showing lithologies, bedding contacts, and sedimentary structures.

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FLYSCH FACIES

Facies 1 - Pelagic

The pelagic facies include sediments that accumulated slowly either by settling of suspended material or by chemical precipitation and diagenesis. Such rocks probably formed in the deeper, more stable reaches of the Ouachita trough where incursions of turbidity flows and mass slumps were infrequent. Included in this category are thick, undisturbed shale or mudstone intervals, siliceous shales, and black, impure cherts.

Fissile shales apparently result from the slow settling of dispersed clay minerals in such a way that the flakes orient themselves parallel to the sea bottom. Mudstones are non-fissile, probably due to a random orientation of clay minerals. Such randomness could result from rapid

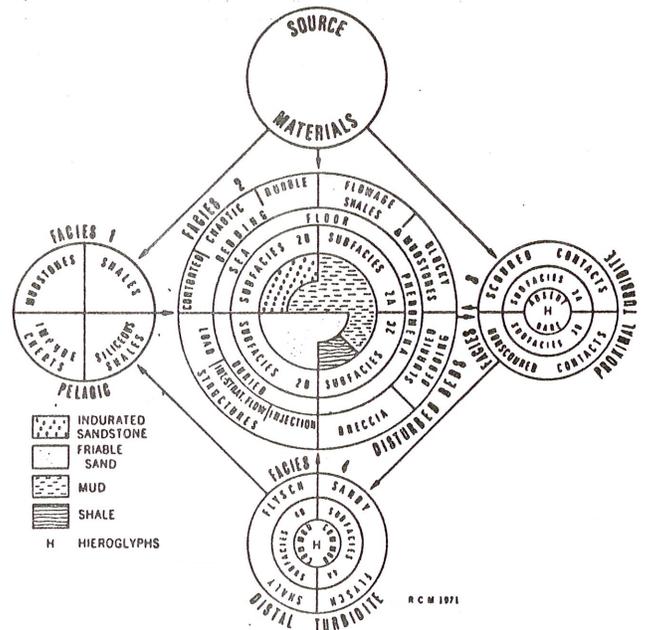


Figure 1. -- Conceptual model for major facies relationships of Carboniferous rocks. Each major facies is represented by number, whereas letters designate subfacies. The facies subdivisions of the measured sections illustrated in this report are designated by these numbers and letters. Sediment entered the basin in three ways: (1) by settling out from suspension to form the pelagic facies, (2) by slumping or sliding to form the disturbed beds, or (3) as turbidity currents. Downslope from their point of origin, proximal turbidites pass into distal turbidites with the very fine material in suspension carried still further to form the pelagic facies. Each facies can be subject to soft-sediment deformation within the basin. Names around the outer edge of the diagram for Facies 2 suggest the types of disturbed bedding to form, dependent upon type of lithology involved (center of diagram).

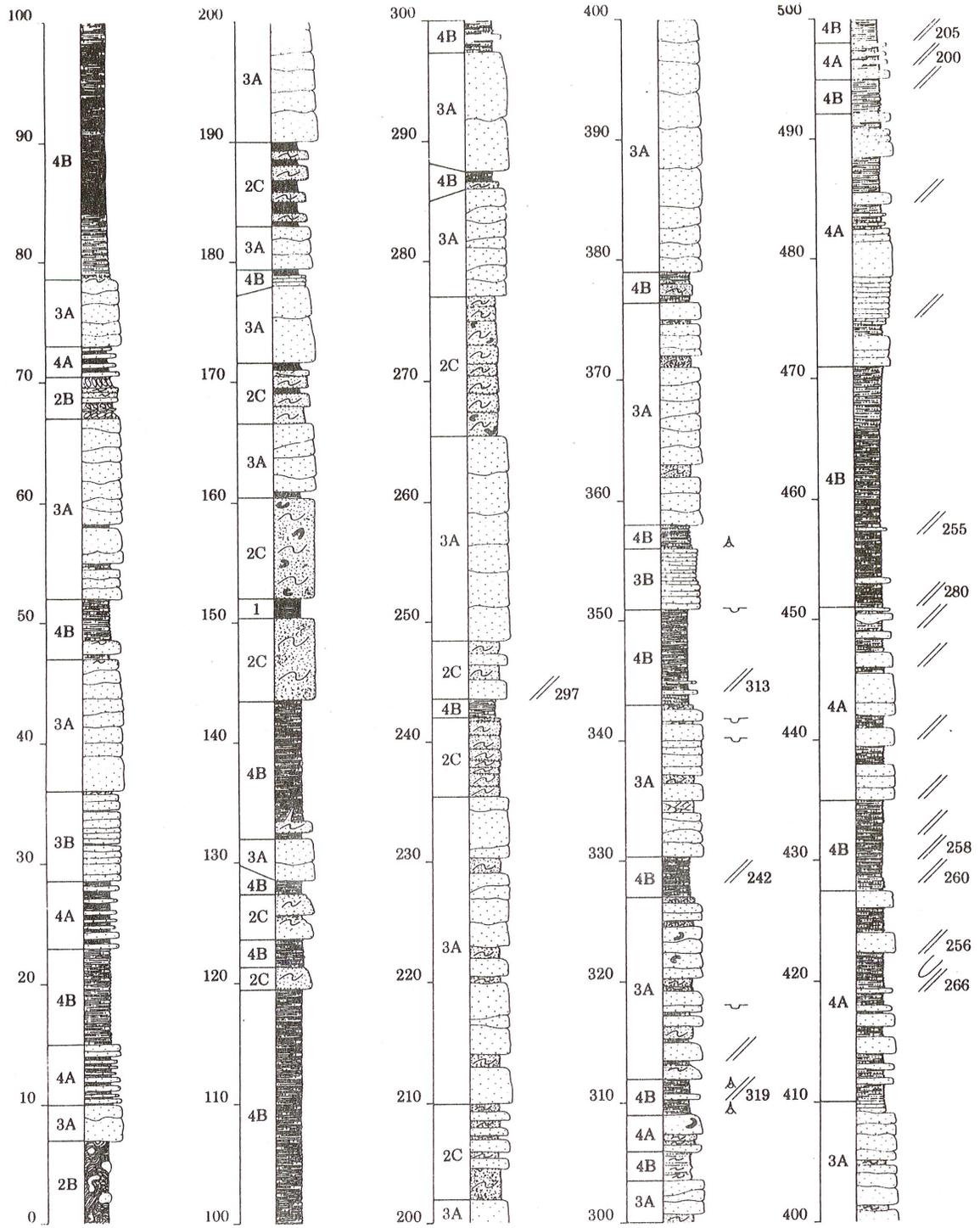


Figure 2. -- Partial stratigraphic section of upper Jackfork rocks exposed along the spillway at DeGray Dam, Sec. 14, T6S, R20W. See Figure 3 for legend. Numbers at left of each column are feet above bed where study began. Numbers and letters within central column are flysch subfacies designated in Figure 1.

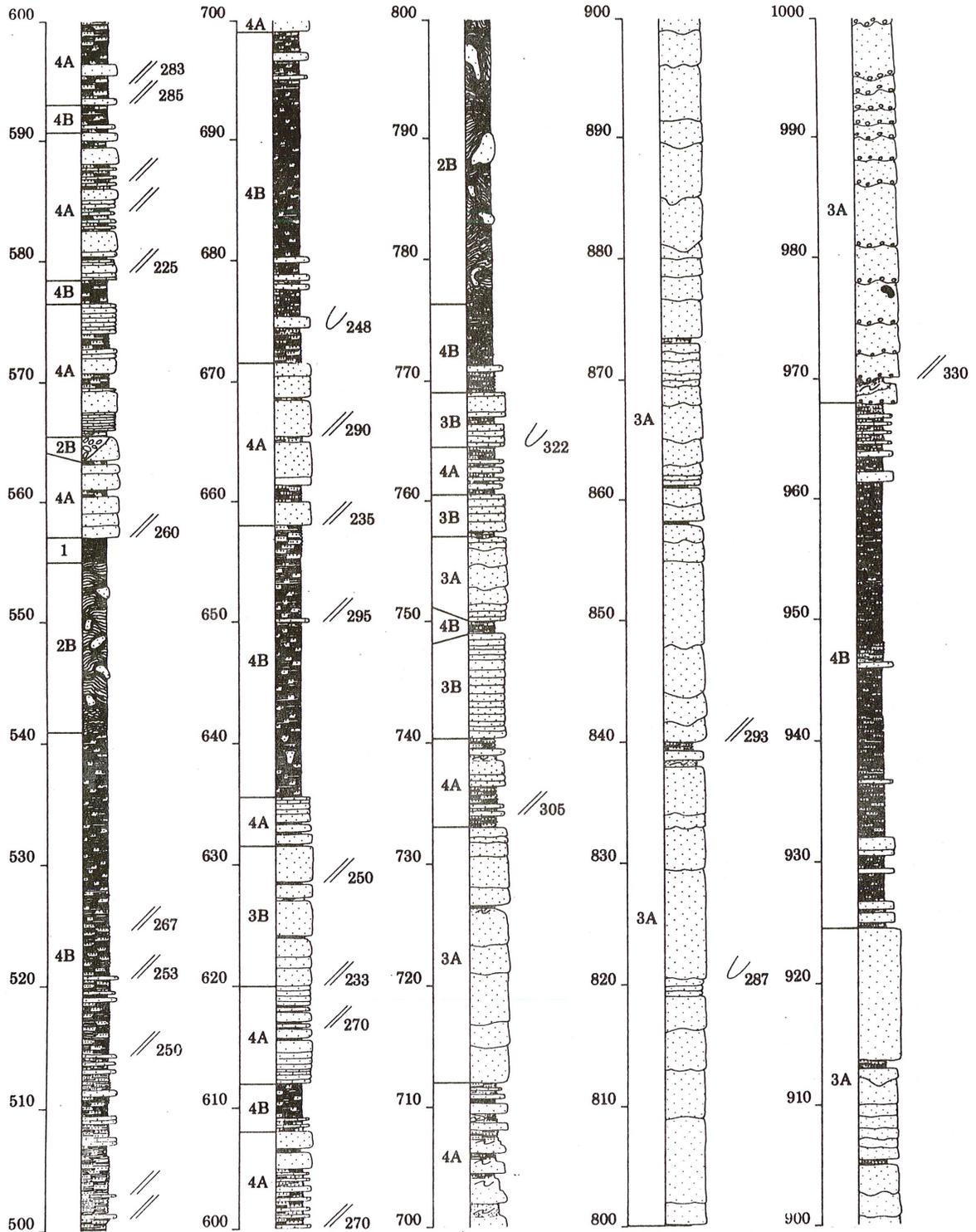


Figure 2. -- (Continued)

LEGEND

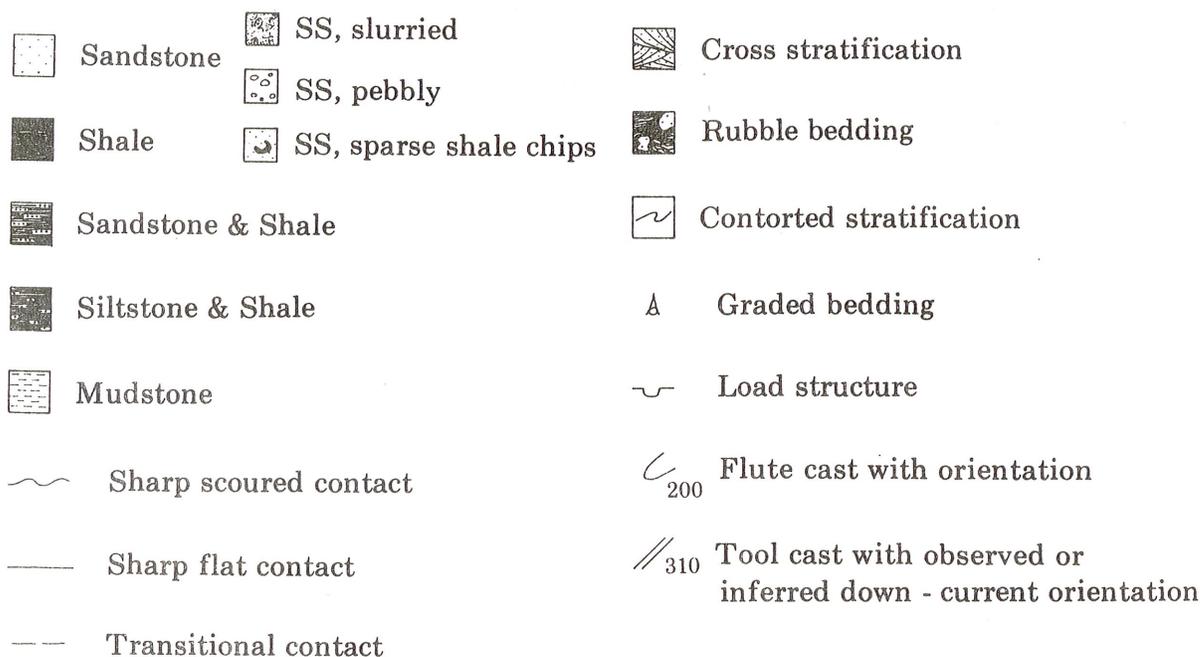


Figure 3. -- Legend for lithologies, bedding contacts, and sedimentary structures in the stratigraphic section of Figure 2.

settling of flocculated clays, slope failure of highly liquified muds, or perhaps intense bioturbation. The shales or mudstones must exceed three feet before they constitute a pelagic facies. Thinner intervals are included with other facies.

Black siliceous shales, commonly associated with impure, locally spiculitic cherts, are uncommon in Stanley-Jackfork rocks of Arkansas. Where present, the siliceous shales are black, brittle, highly siliceous, pyritic, and platy. The impure cherts occur in beds less than six inches thick, and have sharp boundaries, close jointing, discontinuous laminae, and irregular, globular, silica specks. Some silt laminae occur within and between chert layers. These beds probably formed by some settling process in which planktonic siliceous particles accumulated rapidly upon the deep-sea floor for short periods of time. Selective solution of the silica and subsequent precipitation within the pores would have lithified the sediment.

Facies 2 - Disturbed Beds

Disturbed bedding is especially abundant in rocks of the Jackfork Group and has been described and classified by Morris (1971). The basis of this classification is that when soft sediment moves down a depositional slope, the resulting deposit depends upon type of material involved, its state of consolidation, and its distance of transport down the slope. Names were given to distinct, clear cut "end members" although the resulting deposits are gradational into each other. The conceptual model (Fig. 1) includes this classification of disturbed bedding. Four

materials or combination of materials are considered: (1) plastic muds only, (2) a combination of plastic muds and cohesive sandstones, (3) a combination of plastic muds or shales and cohesionless material (chiefly sand-sized), and (4) cohesionless material (chiefly sands) only. Each type of material above should deform into a progression of structures as the state or degree of deformation progresses.

Disturbed beds resulting from disruption of muds only (Subfacies 2A) -- Flowage shales are visibly folded, buckled, squeezed, ruptured, and sheared mudrocks in which bedding planes are still visible. Fold axes are chaotically arranged and cleavage is negligible. Masses of flowage shales occur between bedded turbidites and form part of the sedimentary succession. Blocky mudstones are massive and generally devoid of bedding although traces of contorted surfaces may be present. Fissility is not well-developed, mica flakes are unoriented, and exotics of more fissile shales occur within the mudstone.

The boundary between flowage shales and blocky mudstones is gradational. The two rock types may have resulted from failure of an original muddy deposit upon an unstable slope. Because the original bedding planes in flowage shales were not completely destroyed, the masses must have slumped or slid only a short distance. Flow either over longer distances, or perhaps multiple flows, broke down the fabric of the mud, with the resultant complete destruction of former bedding planes and the production of blocky mudstones.



Figure 4. -- Chaotic bedding, consisting of imbricated slabs of sandstone and shale, interval 70' of stratigraphic section. Two other imbricated zones occur a few feet below this one. Each disturbed zone is interpreted to have been deformed upon the sea floor at successive times, and each was subsequently leveled off and partially removed by turbidity currents that deposited the overlying bed.

Disturbed beds resulting from disruption of cohesive muds and indurated rocks (Subfacies 2B). -- This spectacular facies has three main types: (1) Contorted bedding is a term used for folded and locally faulted units which are sandwiched between undeformed rocks. The competent layers generally have failed by folding with minor fracturing and shearing, whereas the incompetent beds have been deformed both by plastic and shear movement. (2) Chaotic bedding describes a mixture of elongated, angular, cohesive sandstone blocks in which the two long axes are oriented parallel with bedding and each block is separated completely by attenuated shales or mudstones. Because the sheared mudrocks are devoid of sand grains, the sandstone beds must have been reasonably competent when rupture occurred. An example would be the several zones of imbricate bedding as seen in Figure 4. (3) Rubble bedding describes a complex assortment of angular or rounded, non-oriented blocks of relatively competent rocks set in crumpled and compressed flowage shales. The blocks are usually similar to bedded turbidite sandstones above or below. Exotic limestones and chert blocks in rubble beds along the frontal Ouachitas are important indicators of a northward-lying shelf. Several zones of rubble bedding are present in the spillway section (i. e. 775' - 800').

Contorted and chaotic bedding are thought to develop

from only small-scale movement. Long-distance submarine slumping apparently completely disrupts competent beds, so that they become isolated blocks within deformed shales or mudstones. The mudrocks in rubble bedding have no consistent orientation of fold axes, so the slump directions must be inferred as oriented perpendicular to the belts of maximum occurrence. The frontal Ouachitas have the greatest amount of rubble bedding. Along the southern Ouachitas, less numerous and thinner intervals occur. I interpret this to mean that slumping was mainly north to south, as evidenced both by the higher concentration of rubble beds on the rocks of the north and by the presence there of large numbers of shelf-derived, exotic carbonate blocks.

Disturbed beds resulting from disruption of cohesive muds and friable sands (Subfacies 2C). -- This facies consists of two similar kinds of disrupted rocks that differ in whether the mudrocks interbedded with the friable sands were consolidated or unconsolidated at time of failure. The two types are: (1) Slurried bedding is a deposit of widely different slumped masses of friable sand and variable amounts of mud, not always of the same consistency. Brief movement did not completely disaggregate the sand but left angular blocks or incompletely assimilated infolds that blend into more homogenous mixtures (Fig. 5). With further movement, or when the original sand beds

were less coherent, the masses collapsed into a jumbled, hummocky mass of sandstone blocks, friable sand, and scattered streaks, blocks, and pods of shale. This type of bedding can either be confined to tops of turbidite beds, or may comprise the entire bed, as demonstrated repeatedly in the spillway section. (2) Sandstone-shale breccias consist of angular shale and sandstone blocks, set in a well-sorted matrix. The Big Rock Quarry at Little Rock has numerous sandstone-shale breccias associated with channeled proximal turbidites. No good examples of sandstone-shale breccias were observed at the spillway.

The two types above presumably formed by failure on the sea floor. The breccias commonly occur in the proximal turbidite facies. These breccias may be due to bank failure along walls of submarine canyons. Distant transport of sands and muds would tend to produce a slurried bedding with imperfectly disaggregated and blended sand and shale fragments.

Disturbed beds resulting from disruption of friable sands only (Subfacies 2D). -- These include rocks containing load structures, sandstone dikes and sills, and intrastratal flow structures. Load structures are irregular, rounded projections on the base of sandstone beds that resulted from vertical sinking of the sand into underlying unconsolidated mud immediately after deposition. Sand injection structures cut upwards, downwards, or parallel bedding and range in size from less than an inch to a foot in width. Intrastratal flow structures are those irregular laminations occurring either within a bed or including the whole bed. These structures are polygenetic and usually reflect both a depositional as well as compactional (liquefaction) history (Fig. 6).

Facies 3 - Proximal Turbidites

Thick-bedded Stanley and Jackfork sandstones form conspicuous ridges across the Ouachitas. These nonrhythmic, channel-fill beds match Walker's (1967, p. 23) descriptions of proximal turbidites (fluxoturbidites) so well that they must have a similar genesis. Major characteristics of the Carboniferous proximal turbidites are listed in Table I. Close examination of these rocks permits subdivision into two major groups: (1) thick-bedded sandstones with no interbedded mudrocks, so that sandstones rest upon scoured tops of similar sandstones, and (2) regular, thick-bedded, generally structureless sandstones with minor amounts of interbedded mudrocks. The second type is considered to be transitional and somewhat gradational with distal turbidites. Both types are commonly structureless and ungraded -- perhaps because of the considerable bed thickness as well as the high degree of initial sorting.

Subfacies 3A (Scoured Contacts). -- Turbidite sequences with scour-and-fill features are characterized by thickly bedded sandstones with no interbedded mudrocks, so that the A division (Bouma, 1962) of one turbidite unit lies upon the scoured top of a similar A division of the underlying bed. These amalgamated bedding planes can be distinctive or obscure. The beds may be grouped together without any interbedded shales so that "bundles" of such beds approach 500 feet in thickness.

Shallow channels are seldomly observed because of poor exposures, but excellent ones are present in Brushy Knob sandstones at North Little Rock, Arkansas. These channels are positive evidence of one point source for upper

TABLE I - MAJOR BEDDING CHARACTERISTICS OF PROXIMAL TURBIDITES (FLUXOTURBIDITES)

SUBFACIES 3A SCOURED CONTACTS	SUBFACIES 3B NONSCOURED CONTACTS
Beds all thicker than 25 cm and commonly exceed 200 cm.	Beds thicker than 25 cm but seldom exceed 100 cm.
Beds rest upon scoured tops of similar beds (sand-on-sand).	Sharp, even basal contacts generally rest upon dissimilar beds.
Complete absence of interbedded mudrocks.	Sandstones dominate over thin, interbedded mudrocks.
Absence of internally laminated sandstones.	Scarcity of internally laminated sandstones.
Channel-fills locally.	Channels never present.
Amalgamation of sandstones form massive rocks in which individual beds may be difficult to locate.	Rhythmic, even-bedded sequences with sharp, distinct tops and bottoms.
Absence of hieroglyphs.	Scarcity of tool and flute casts, load structures common.
Absence of all types of cross-bedding.	Absence of all types of cross-bedding.
Sandstones consist exclusively of A (graded) intervals although the high initial sorting did not permit conspicuous grading to develop.	Sandstones consist of A (graded) division and overlying E (peletic) division. Some ABE sequences may also be present.

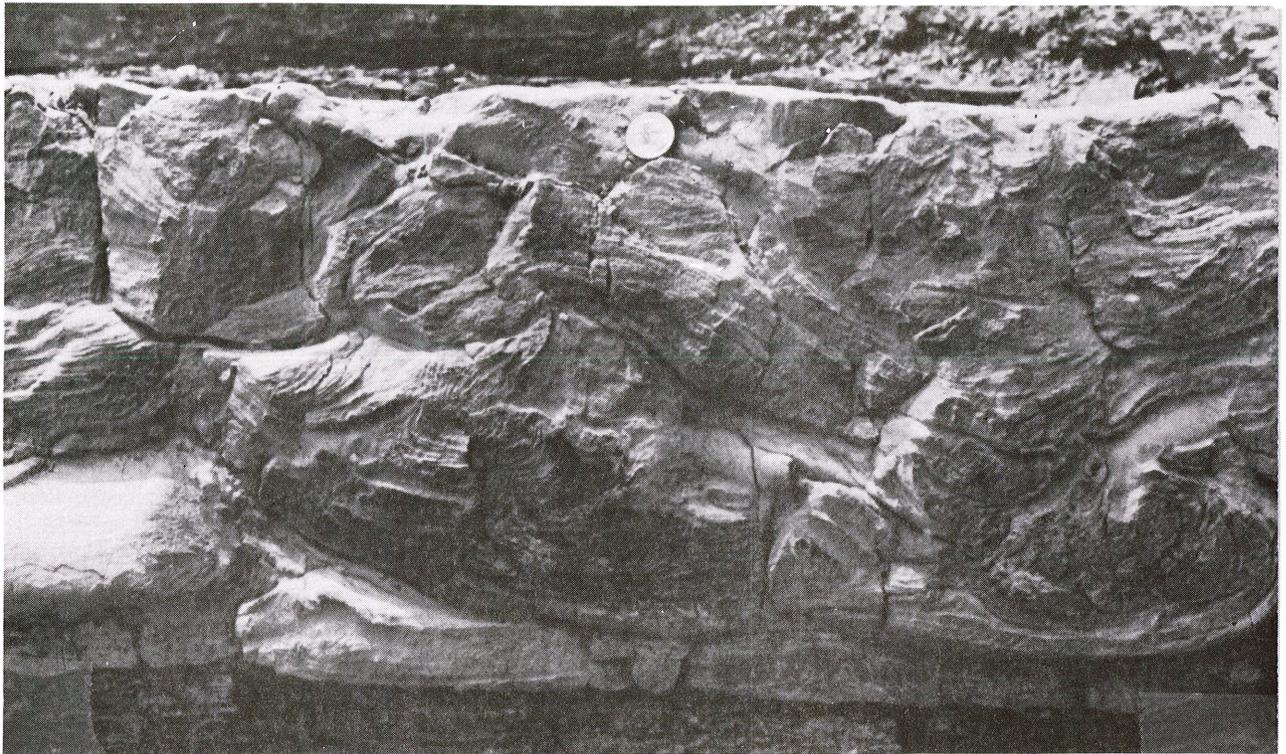


Figure 5. -- Close-up of portion of slurried bed, interval 247' of stratigraphic section. This contorted mass of fine-grained sandstone and shale was incompletely assimilated when some sort of shock phenomena reduced these beds to a structureless mass upon the sea floor. It is impossible to ascertain how much lateral movement is associated with this type of disturbed bedding. The overlying turbidite was deposited later than the development of this bed.



Figure 6. -- Convolute (contorted) laminations due to some sort of intrastratal deformation, interval 133' of stratigraphic section. The laminations, perhaps including a certain amount of undulation of laminations, is a current-related phenomenon, but increased deformation has resulted from liquefaction of the entire bed. The later occurred following deposition of the overlying beds, as evidenced by a small sandstone dike projecting from the top of the bed.

Jackfork sandstones. Irregular or patterned, shallow scours with elongation parallel with the current direction most commonly separate scoured sequences. These scoop-shaped scours can be used to distinguish tops and bottoms of beds. They characteristically have broad, shallow bottoms and narrow, sharp ridges that mark the junctions between troughs.

Thin-section analyses of the better sorted proximal turbidites indicate no significant lateral variations in size of framework grains, but local pebbly zones occur. Pebbly sandstones which locally range to cobble conglomerates occur at several intervals within the upper part of the Jackfork Group. These clasts consist of sandstone, siltstone, limestone, quartz, fossil fragments, and chert. Most commonly, the sandstones above and below the scour have the same texture and result in an obscure contact. However, where the overlying bed has a pebbly base, contrast is striking (Fig. 7).

These deposits may record a high flow regime in which scouring and erosion dominated the initial stage of each turbidity flow. The slowing current allowed deposition of the A division. Because lateral grading had not yet developed, division A is generally poorly graded.

Subfacies 3B (Nonscoured Contacts). -- These deposits are characterized by thick, structureless sandstone beds with sharp, regular bottoms and somewhat distinctive tops. They are rhythmically interstratified with subordinate amounts of shale or mudstone. Amalgamation of beds was rare because a thin peletic E division overlies the poorly graded A division, or, less commonly, the laminated B division. With the exception of load molds, hieroglyphs

are rarely developed. Although the distinction may sometimes be a subtle one, these beds can be differentiated from distal turbidites by their thicker bedding and also by the sparse development of internal laminations. The tops of these beds are nowhere rippled, although there may be a rather sharp textural break between the sandstone and the overlying mudrock.

The absence of scoured contacts between turbidite sequences in this facies would seem to indicate that the turbidity currents were no longer eroding as they passed this point. Therefore, erosion was not sufficient to remove all the muds which covered the previous turbidite. Because of the general lack of hieroglyphs in these rocks, the muds must have been non-cohesive and incapable of standing up under scour or to being etched by objects. Perhaps an undetermined amount of mud was removed by the succeeding turbidite, to explain the extremely thin shale interbeds.

Facies 4 - Distal Turbidites

Distal turbidites are characterized by rhythmically interbedded shales and turbidite sandstones with an abundance of hieroglyphs and internal structures. This facies probably grades laterally into the proximal turbidite sequences with unscoured contacts, but it must be admitted that this relationship is more conceptual than observational. A two-fold subdivision of this facies has been employed, based essentially upon whether turbidites or black shales dominate a succession. Additional characteristics are summarized in Table II. A prolific development of both subfacies is present within the spillway section (Fig. 8).

TABLE II - MAJOR BEDDING CHARACTERISTICS OF DISTAL TURBIDITES

SUBFACIES 4A - SANDY FLYSCH	SUBFACIES 4B - SHALY FLYSCH
Beds predominately 3 to 100 cm thick (average \pm 15 cm). Sharp, even basal contacts always rest upon dissimilar beds.	Beds mostly less than 5 cm thick (average \pm 2 cm). Sharp, even basal contacts always rest upon dissimilar beds.
Subequal amounts of interbedded sandstones and mudrocks.	Mudrocks dominate over interbedded siltstones and/or very fine grained sandstones.
Abundance of internally laminated sandstones.	Abundance of internally laminated sandstones.
Channels never present but broad, shallow scours do cut out entire beds.	Channels never present.
Rhythmic, even-bedded sequences with sharp, distinct bottoms. Tops of sandstones may be rippled, sharp and even, or locally gradational into E (peletic) division.	Rhythmic, even-bedded sequences with sharp, distinct bottoms. Tops may be rippled, sharp and distinct, or gradational.
Abundance of tool and flute casts and load structures.	Tool and flute casts generally present although small. Absence of load structures.
Absence of large-scale cross-bedding. Small-scale cross-stratification (C division) common.	Absence of large-scale cross-bedding. Small-scale cross-bedding (C division) common.
Sandstones commonly demonstrate complete or nearly complete Bouma sequences. ABCDE, ABCE, ABE, BCDE, BCE, CDE, and A \rightarrow E sequences present.	Sandstones commonly demonstrate complete or nearly complete Bouma sequences. ABCDE, ABCE, ABE, BCDE, BCE, CDE, and A \rightarrow E (graded but structureless) sequences present.



Figure 7. -- Bedding characteristics of proximal turbidites showing a scoured contact accentuated by the pebbly base of the infilled bed, interval 995' of stratigraphic section. The actual shape of the scours can be viewed at other places along the spillway where the broad, irregular scours are outlined by sharp, narrow edges.



Figure 8. -- Typical distal turbidite succession consisting of alternating turbidite sandstones and shales, interval 575' - 625' of stratigraphic section. The intervals consisting of thicker bedded sandstones are called sandy flysch whereas the more shaly intervals containing thin turbidites are called shaly flysch.

Subfacies 4A (Sandy Flysch). -- Rocks of the sandy flysch facies are dominated by even-bedded, rhythmically interstratified sandstones with subordinate amounts of undisturbed shales. Sandstone bottoms, which are sharp and nowhere channeled, contain abundant tool-mark casts with subordinate flute casts and load structures. Most paleocurrent readings were taken from these sandy flysch successions. Internally, the sandstones generally show laminations and/or small-scale cross-laminations (Fig. 9). Tops of sandstones may be even, but commonly contain ripples (linguloid, lunate, transverse, interference).

Subfacies 4B (Shaly Flysch). -- Shaly flysch is a succession of interlaminated turbidites and shales, in which shale is the dominant lithology. The turbidites are thin-bedded, fine-grained sandstones or siltstones, commonly graded and generally laminated. Complete or nearly complete Bouma sequences are common, but each division is thin.

Some cross-laminations may have concentrations of heavy minerals. Ferruginous siltstones less than two inches thick occur sparingly within some shaly flysch at the spillway (i. e. 115'). Interbedded shales or mudstones locally pinch out completely within a few feet along strike. Sole markings are not as common in siltstones as in sandstone turbidites. Thin turbidites vary considerably in thickness due to rippled tops; indeed, isolated ripples (flaser structures) with well developed internal laminations have been observed.

PALEOCURRENTS

The Jackfork sandstones display a wonderful array of sole markings that can be utilized to reconstruct ancient paleocurrent directions. Flute casts and prod casts impart a current direction while other types of tool markings furnish

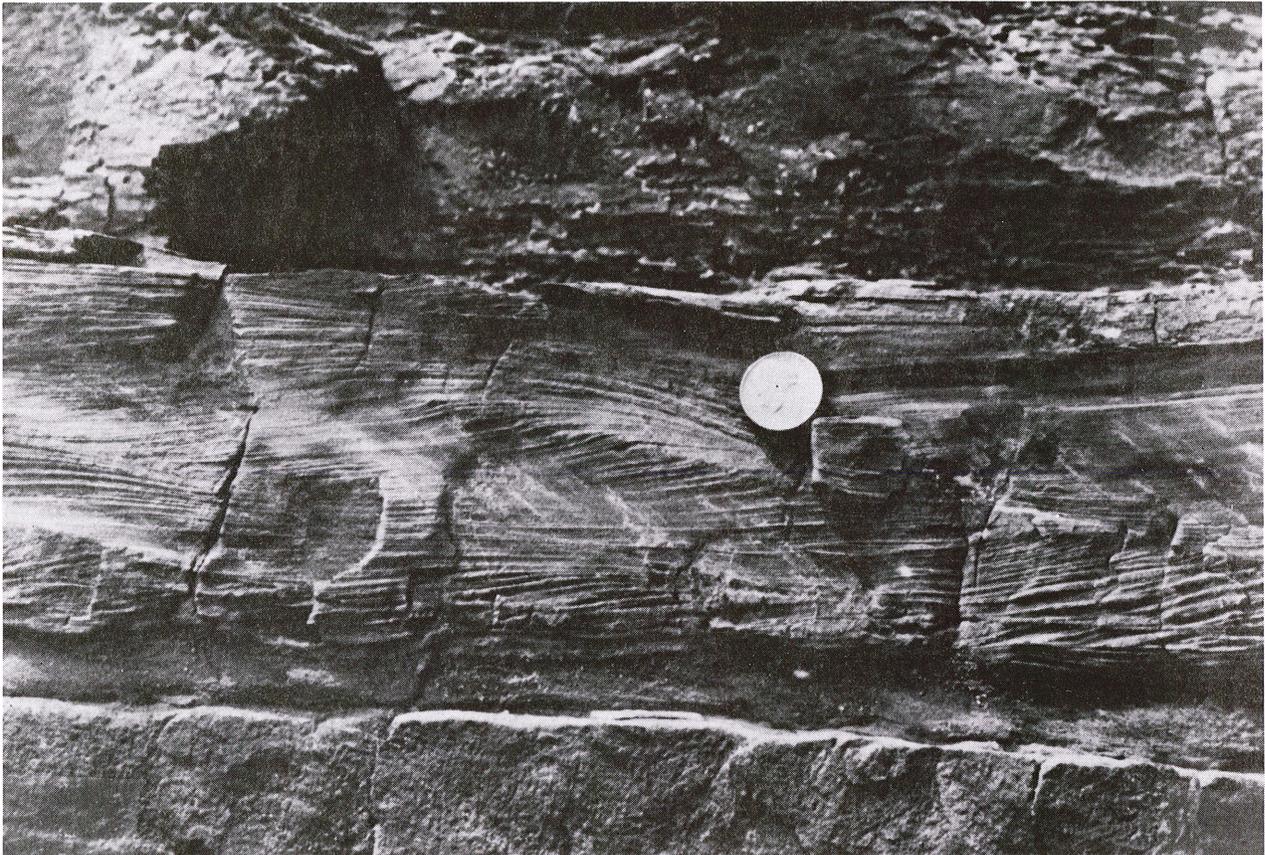
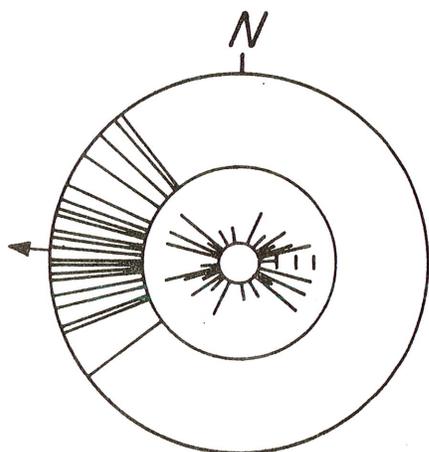


Figure 9. -- Ripple-drift cross-stratification overlying a slurried bed, interval 170' of stratigraphic section. Some cross-stratification in turbidite successions may be due to reworking of turbidites by normal (contour) currents. However, most are associated with the "C" division of the Bouma interval and represent the point where deposition from the turbidity current passed from upper to lower flow regime.



LOC. 21, T 6 S, R 20 W

$N = 44, \bar{X} = 274^\circ$

Figure 10. -- Paleocurrent diagram of sole marks from upper Jackfork sandstones in the spillway area. Inner circle indicates data with current trend only (length of lines correspond with number of readings); outer circle includes current directions from prods or flutes. Arrow indicates mean current direction as determined from all measured or inferred down-current directions. Letter N indicates north.

a current sense only. Figure 10 is a summary diagram of upper Jackfork paleocurrents in the general area of the spillway. The mean direction of 274° is very close to a grand mean of 266° obtained from 141 upper Jackfork readings. Although not readily apparent on this diagram,

there is evidence that the sands entered from two point sources. Non-feldspathic detritus entered from the northeast, whereas sparsely feldspathic detritus entered from the southeast. Down the axis of the trough to the west, the number of sands increased, the thickness of each sand decreased, and a broad, abyssal plain was formed.

SUMMARY

The rocks of the spillway section apparently were deposited upon an ancient abyssal plain not far removed from a continental rise. The dominant pattern of sedimentation was repetitive incursions of turbidity flows whereas minor slumping also contributed to the development of the sedimentary succession. The more powerful turbidity currents, presumably initiated from nearby areas, were capable of carrying tremendous quantities of sand which eroded and scoured significant portions of underlying beds before depositing their load. Natural groupings of such deposits can be called proximal turbidites. Other less frequent, less turbulent currents, perhaps originating from more distant points, developed patterned scours and tool marks over cohesive mud bottoms. The repetitive nature of these turbidites and muds form a natural grouping of distal turbidites. By giving empirical classifications such as proximal or distal turbidite to these abnormally thick turbidite successions, we estimate how "proximal" or how "distal" one or more point sources may have been located. A complicating factor is that different volumes of material were dislodged at different times. In conclusion, thick sequences of proximal turbidites probably did originate nearby and thick sequences of distal turbidites probably had a more distant beginning. Close vertical stacking of these two end members could either be due to dovetailing of turbidites from two or more point sources, or to differences in sedimentary volume resedimented or both.

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INTERNAL SEDIMENTARY STRUCTURES IN SANDSTONES OF THE JACKFORK GROUP, OUACHITA MOUNTAINS, OKLAHOMA

By Robert D. LoPiccolo¹

ABSTRACT

The Jackfork Group (Pennsylvanian) is composed largely of fine-grained quartzose sandstone in beds generally ranging from 3 cm to 1 m in thickness. Primary sedimentary structures include small-scale climbing ripples, flat laminations, and small-scale festoon cross-laminations. Individual sedimentation units typically begin at the base with an interval of flat laminations and grade upwards into climbing ripples. The uppermost layers of some flat laminated beds have been reworked to form small-scale festoons. Coarse grained detritus and large-scale cross-bedding or cut-and-fill have not been observed.

Diagenetic structures include convolute laminations and dish and pillar structure. Convolute laminations develop within beds penecontemporaneously with deposition by the deformation of climbing ripples. Dish and pillar structures form by water expulsion after deposition and after the development of convolute laminations.

Deposition of most of the Jackfork sands took place from high-energy turbidity currents near the base of a submarine slope. Deposition was rapid and the resulting deposits were quick or under-consolidated. During early stages of dewatering, convolute laminations developed in response to stresses imposed by continuing sedimentation. Continued water expulsion resulted in the development of dish and pillar structures.

INTRODUCTION

As a result of the efforts of a large number of researchers the upper Paleozoic Formations of the Ouachita Mountains have been established and generally accepted as one of the classic flysch sequences of North America. The Jackfork Group (Pennsylvanian) stratigraphically overlies the Stanley Group (Mississippian) within this sequence, and the base of the Jackfork marks the beginning of a period of large-scale sand and silt sedimentation in the geosyncline.

The rhythmic interbedding of sandstone and shale, the occurrence of slumped intervals, the presence of a variety of sole marks on the sandstone beds, and the occurrence of animal feeding trails and burrows have been noted by virtually everyone who has studied the Jackfork. Based on evidence derived from studies of these features, structural and stratigraphic relationships, and comparisons with other flysch sequences, recent workers (Briggs, 1973; Chamberlain, 1971; Morris, 1971; Klein, 1966; King, 1961; Cline, 1960, and Cline and Shelburne, 1959) have concluded that the basin was an actively subsiding deep marine trough during the deposition of the Jackfork. Although there is some controversy as to the location and nature of the source area, there is general agreement that sediment transport within the trough, especially the axial portion, was from east to west.

The purpose of this paper is to present recently gathered information which can be used to better define the environments within which the Jackfork was deposited.

SEDIMENTARY STRUCTURES

Six different internal sedimentary structures have been described from the fine-grained quartzose sandstone beds which comprise a large part of the Jackfork Group. These structures can be subdivided into a group of primary structures and a group of diagenetic structures. The primary structures include plane-bed flat laminations, climbing-ripple cross-laminations, and small-scale festoon cross-laminations. Diagenetic structures present are convolute laminations, dish structures, and pillar structures.

Primary Structures

Flat laminations

Cut perpendicular to bedding, flat laminations appear as fine dark lines separated by wider, lighter-colored bands (Fig. 1a). The dark lines represent thin, parallel zones rich in argillaceous and fine organic material. The lighter-colored bands range from one to greater than ten mm in thickness and are composed of relatively clay-free, fine-grained sand. Fractures parallel to the bedding planes commonly show primary current or parting lineations. In many beds flat laminations are the only primary structure present. In beds where flat laminations are associated with the other primary structures, the flat laminations almost always occur at the base of the bed and are overlain by the other structure. In a few instances, the lowest two to five cm show massive, structureless sandstone overlain by sand showing flat laminations (Fig. 1b). Where small-scale festoon cross-laminations overlie flat laminations, the contact appears to be erosional with the festoons cutting into the flat laminations (see upper part Fig. 1a). Where climbing-ripple cross-laminations overlie flat laminations the contact is gradational and appears to be depositional rather than erosional.

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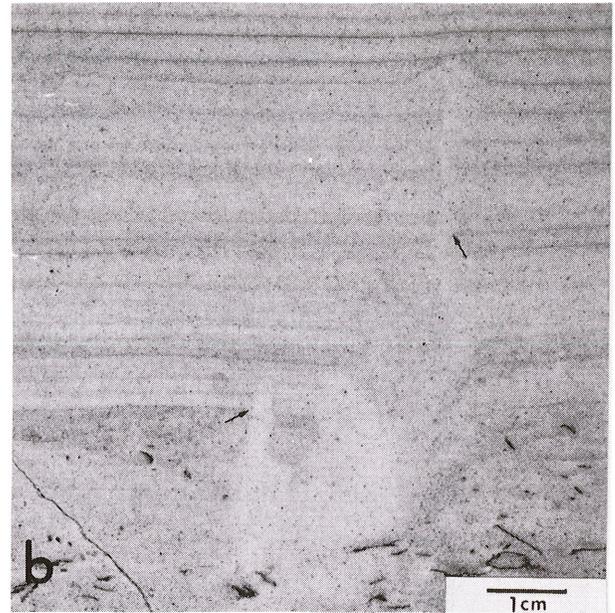
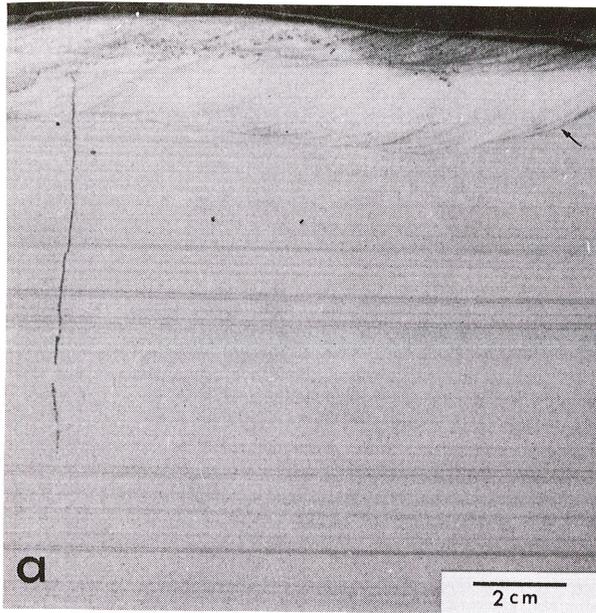


Figure 1. -- Flat laminations in fine-grained sandstones. (a) Flat laminated sand overlain by and reworked into festoon cross-laminations. Arrow indicates erosional contact at base of cross-laminations. (b) Flat laminations overlying structureless sand. Small pillars (arrows) originating within structureless zone marks paths of water escape. Dark material is plant debris.

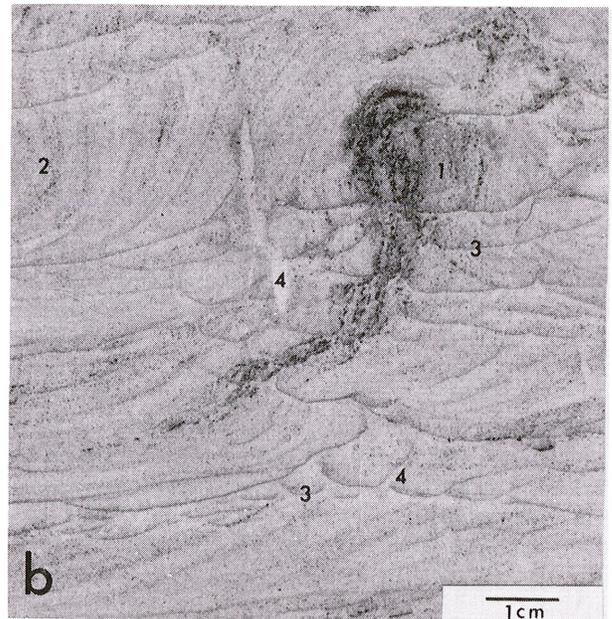


Figure 2. -- Climbing-ripple cross-laminations and convolute laminations. Both beds are underlain by flat laminated sand (not shown). (a) Sets of type A climbing-ripple cross-laminations grading upward into convolutions. Dark material is plant debris. Current was from right to left. (b) Type A climbing ripples grading up into heavily convoluted zone. The convoluted zone consists of anticlinal features (1) composed of deformed climbing-ripple cross-laminations, and synclinal features (2) composed of oversteepened (non-ripple) cross-laminations. Both the convoluted zone and the climbing-ripple zone are cut by dish (3) and pillar (4) structures.

All of the flat laminations observed in this study appear to represent deposition by high velocity currents. This is evidenced by the occurrence of parting lineations which, in conjunction with the plane bed form, indicate the upper flow regime of Simons (et al., 1965) according to Allen (1968). This conclusion is further substantiated by the superposition of cross-laminated sand of the same grain size, which must result from deposition by a current of a lower flow power (Allen, 1970a).

Festoon cross-laminations

The least common of the three primary structures is festoon-type cross-lamination. This structure is observed only in the upper three to five cm of otherwise completely flat laminated beds (Fig. 1a). The individual festoon sets generally attain a maximum thickness of about two cm.

The previously noted erosional contact between these two structures suggests that the festoons represent a reworking of the upper part of a bed originally composed entirely of flat laminations. The current directions implied by these structures are always from east to west, or parallel to the current directions implied by the other sedimentary structures. The presence of the festoons suggests that immediately after the passage of the main current, a non-sedimenting current, perhaps induced by or a residual of, the original current reworked the upper few centimeters of the previously deposited sediment.

Climbing-ripple cross-laminations

Climbing-ripple cross-lamination is probably the most common primary sedimentary structure developed in sandstones of the Jackfork. The ripples are of the type A or B₁ of Allen (1970b) which show respectively either no preservation of the stoss laminations and sharp contacts between the sets or preservation of thin stoss laminations with gradational contacts between sets. Measured normal to the base of the bed the ripples range in height from one to five cm. Almost without exception the cross-laminations become oversteepened and convoluted upwards within the bed (Figs. 2a and 2b). Cross-laminated beds are usually fairly rich in plant debris with most of it being concentrated in the lee-side laminations of the ripples (Fig. 2a). Frequently the lowermost one to eight cm of the bed will show the development of flat laminations grading up into cross-laminations.

Allen (1970b) has concluded that climbing ripples may result from either rapid deposition by rapid currents or slow deposition by slow currents. The common occurrence with upper-flow-regime flat laminations however, suggests that rapid deposition is the more likely case for the Jackfork sands.

Diagenetic Structures

Convoluted laminations

Convoluted laminations appear to develop by the progres-

sive deformation of climbing-ripple cross-laminations. Undeformed climbing ripples are commonly preserved in the lower parts of bed showing convolute laminations and the degree of deformation usually increases upwards in the bed (Figs. 2a and 2b). The usual form of the convolutions is a series of regularly spaced anticlinal forms separated by synclinal features. The cores of the anticlines are rich in medium-to coarse-grained plant debris (Fig. 2b) which originally accumulated in the lee of the now deformed ripples. The adjacent synclines are composed of cleaner, relatively plant-debris-free sand and frequently show deformed primary cross-laminated sets (not climbing ripples) developed in the lee of the anticlines (Fig. 2b). These cross-laminations dip in the same direction as the climbing-ripple cross-laminations and are generally moderately oversteepened to overturned.

The association of deformed climbing ripples in the anticlines with less deformed cross-laminations in the synclines, both showing the same current direction, indicates that the deformation took place concurrently with the deposition of the sediment. The location of the (non-ripple) cross-laminations in the lee of the anticlines suggests that their formation resulted from an interaction of the anticlinal convolution with the sedimenting current.

By knowing that the deformation took place concurrently with sediment deposition and by noting that the mode of deformation is mainly vertical, a sequence of events leading to the formation of the convolute laminations can be hypothesized. The primary assumption underlying this development is that the sediment was deposited in an underconsolidated state and that stresses arising from continued sedimentation were capable of initiating consolidation. After the deposition of a series of climbing ripples, but before the deposition of the bed ceases, the lower part of the bed begins to consolidate and expel excess pore water. This water migrates upwards through the overlying sediment and has the effect of partially liquifying the sediment and causing it to lose some of its shear strength. With continuing sedimentation, in the form of climbing ripples, the low-strength sediment deforms by bulging up in the areas ahead of the prograding ripples. These bulges form baffles in the current and cause sediment to be deposited in their lee in the form of (non-ripple) cross-laminations (see Fig. 2b). This trapped sediment places an additional shear couple on the deforming areas and aids in the growth of the bulges. Finally enough sediment is deposited over the top of the bulge to increase the confining pressure to the point where the stresses are insufficient to continue the deformation. The deformed sediment is now free to consolidate, expel water and initiate the formation of convolute laminations higher in the bed.

Dish Structure

Dish structure is one of the more common sedimentary structures in sandstones of the Jackfork and may occur either alone or in association with any of the other structures. Individual dishes are defined by the presence

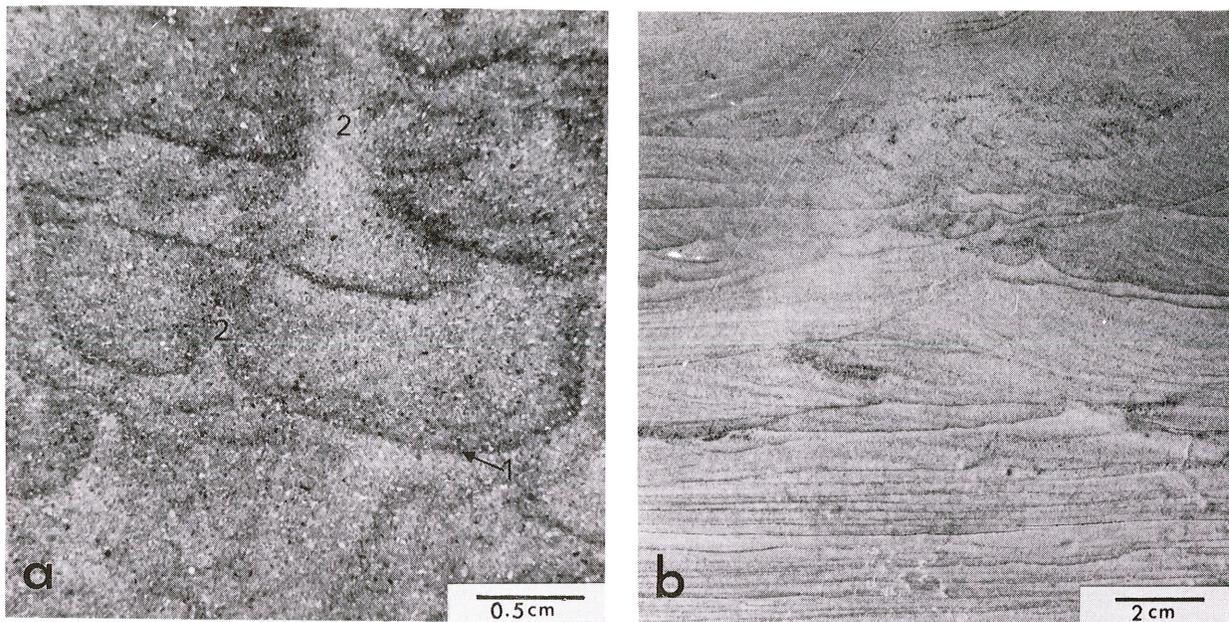


Figure 3. -- Dish and pillar structure in fine-grained sandstone. (a) Well developed dish structure defined by thin laminations of argillaceous and organic material (1) and separated by pillars (2) of light-colored relatively clean sand. Dishes are overlain by dark-colored clay-rich sand and underlain by light-colored, clay-free sand. (b) Dish structures modifying primary flat laminations in lower part of bed and cross-cutting climbing-ripple cross-laminations in upper part.

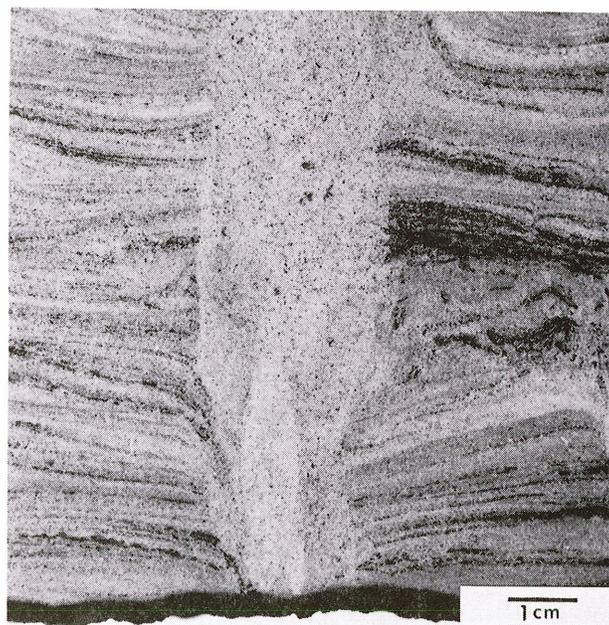


Figure 4. -- Large pillar cross-cutting climbing-ripple cross-laminations and flat laminations. Escaping water which formed this pillar originated within a thick sand bed immediately beneath the bed illustrated.

of a concave-upwards, clay- and fine-grained organic-rich lamination (Fig. 3a). This lamination is usually underlain by a zone of relatively clean, clay-free sand and overlain by a zone of clay-rich sand. The dishes may range in length from one to four cm and the lamination is usually less than one mm in thickness.

Where dishes are formed in association with primary structures they cross-cut the primary structure and therefore postdate them (Fig. 3b). Similarly, in the case of the convolute laminations the dishes cross-cut the convolutions but are themselves undeformed (Fig. 2b) and must have formed after the deformation.

Lowe and LoPiccolo (1974) and LoPiccolo and Lowe (1973) have demonstrated that dish structures form through an interaction between earlier formed structures and upward moving waters. These authors have also suggested compaction and associated water expulsion of under-consolidated sediment as probable causes for the formation of dish structure.

Pillar Structure

Two varieties of pillar structure have been described from the Jackfork. Both are characterized by relatively clean, homogeneous, vertical to subvertical sheets or columns of sand with sharp boundaries separating them from the adjacent sediment within the bed.

The most common variety is developed between the up-turned margins of adjacent dishes (Fig. 3a). This type is generally small and usually bounded on its margins by thin clay laminations. It appears to form by a simple vertical extension of the dish margins during continued water expulsion.

The second variety may or may not be associated with dish structure. It is typically larger than the first type and is not usually bounded by clay-rich laminations (Figs. 1b and 4). The larger scale and lack of clay-rich boundaries suggest a more rapid explosive dewatering than the variety developed in association with dishes.

PROCESSES AND ENVIRONMENTS OF DEPOSITION

While all of the structures have not been observed within any single bed, a composite bed can be hypothesized and used for purposes of interpretation (Fig. 5). Excluding the festoons as resulting from a secondary reworking, the composite bed has a massive, structureless base overlain by flat laminations. The flat laminations grade upwards into climbing-ripple cross-laminations which in turn become convoluted towards the top of the bed.

This sequence is suggestive of the A, B, and C divisions proposed by Bouma (1959) as representing the lower part of a normal turbidite sandstone bed. The overlying shale would represent the E division.

The absence of the D interval can be explained by reference to Allen (1970a). In this paper Allen provides a graph which plots bed form as a function of grain size and flow power. This graph is reproduced in Figure 6 with the 'deposition history' of a composite Jackfork sandstone bed indicated by the arrow from X to Y. It can be seen that the lack of a D division is a result of the limited grain size deposited and not necessarily the result of a non-turbidity current process.

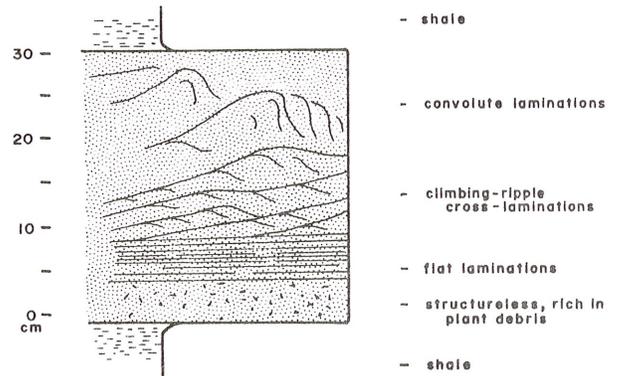


Figure 5. -- Composite bed illustrating the basic sequence of structures in sandstone beds of the Jackfork.

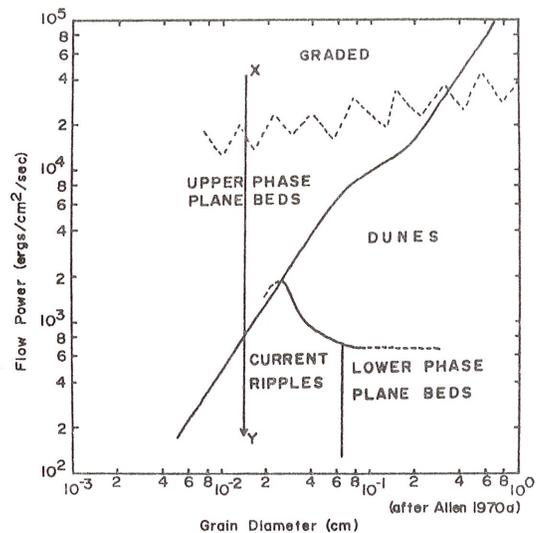


Figure 6. -- Relationship of bed form to grain size and flow power. Arrow from X to Y illustrates the general evolution of bed forms and current represented by the composite bed illustrated in Figure 5.

The absence of cut-and-fill features, coarse-grained detritus or large-scale cross-bedding together with the lack of large scale cyclicity, indicative of a system of slowly migrating environments, suggests that the most likely process of deposition would be one capable of operating in a deep

marine environment.

The conclusion that the beds do, indeed, represent deposition from turbidity currents is further substantiated by the rhythmic occurrences of the sandstones, implying an episodic rather than a continuous process.

A more detailed picture of the environment in which deposition took place emerges from a consideration of the diagenetic structures. The origin of all three of the diagenetic structures described in this paper can be explained in terms of the processes and effects of water expulsion. This water expulsion is caused by the consolidation of beds which were deposited in an underconsolidated state. In order to deposit sand-sized sediment in an underconsolidated state it appears that the rate of deposition must be fairly high. Rapid deposition would allow for a minimum of grain movement after the grain is deposited upon the bed surface. The individual grains would then be less likely to have 'found' a most stable resting position and the resulting grain framework would have a minimum number of grain-to-grain contacts and a maximum amount of pore space.

A rapid rate of deposition would imply a rapid loss of power from the sediment transporting current. The ultimate source of power for a turbidity current is derived from the down-slope component of the acceleration due to gravity acting on the excess mass of the flow relative to the surrounding water. The rapid rate of deposition would seem to imply a fairly abrupt decline in slope.

In view of the widespread distribution of convolute laminations, dish structures and pillar structures, both stratigraphically and geographically within the Jackfork,

the implication is that a large number of the sandstone beds of the Jackfork were deposited rapidly and fairly near a change in slope.

SUMMARY

Six different internal sedimentary structures have been described from sandstone beds of the Jackfork Group.

While there is nowhere developed a complete Bouma sequence, evidence indicates that the primary structures formed during deposition from turbidity currents.

The diagenetic structures developed as a result of the consolidation of originally underconsolidated sediment and indicate that the sediment was deposited rapidly.

A rapid rate of deposition implies a fairly abrupt decline in slope.

The widespread distribution of diagenetic structures in the Jackfork suggests that a large number of the Jackfork sandstones were deposited in proximity to a change in slope.

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