OUACHITA SYMPOSIUM TEXT

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1959

THE

GEOLOGY OF THE OUACHITA MOUNTAINS

A SYMPOSIUM

PUBLISHED BY THE DALLAS GEOLOGICAL SOCIETY AND THE ARDMORE GEOLOGICAL SOCIETY UNDER THE AUSPICES OF THE FIELD TRIP COMMITTEE FOR THE MEETING OF THE AMERICAN ASSOCIATION OF PETROLEUM GEOLOGISTS AND THE SOCIETY OF ECONOMIC PALEONTOLOGISTS AND MINERALOGISTS. DALLAS, TEXAS, 1959.

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A SYMPOSIUM

Edited by

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Published by
DALLAS GEOLOGICAL SOCIETY
ARDMORE GEOLOGICAL SOCIETY
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A WORD OF THANKS

This volume is the result of a belief in the value of symposium discussion of the subject area of a geological field trip. An outstanding example of this concept "The Woodbine and Adjacent Strata of the Waco Area of Central Texas--A Symposium" by Dr. Frank E. Lozo, has been the model for this symposium. Thanks are due Dr. Lozo for creating a higher precedent to follow.

The Executive Committee of the Convention asked me to organize and conduct the field trip program. This symposium-field trip concept was presented and heartily endorsed. Thanks are due this group for presenting me with an ideal occasion to present unpublished, significant geological investigations of the north Texas and southern Oklahoma areas to the geological profession.

Since guidebooks do not publish or transmit the knowledge generally expressed on field trips, symposium type field trips were viewed as highly desirable. Prior to being asked to serve on this project, this subject was discussed with W. J. Hilseweck, John C. Dunlap, William B. Heroy, Jr., and Philip F. Oetking. All were enthusiastic about both the concept and the convention to be held in Dallas. When the time came for action, this group became a natural committee ready to work together toward contributing something of value to the convention. Mr. Hilseweck suggested the Ouachita Mountains and its associated geological problems as a subject for a symposium and field trip. My personal thanks are due each of the members of this committee.

With the realization that the success of any such endeavor is dependent upon the participation of the experts of the area, Dr. L. M. Cline was asked to help organize the Ouachita Symposium and Field Trip. The thanks due Dr. Cline are evident in this publication.

The Dallas Geological Society, host and financial underwriter for the convention, has provided encouragement and supplementary financial support to the Field Trip Committee. The Ardmore Geological Society has enthusiastically contributed to this endeavor. In gratitude for such support, it is my hope that the profits from publications of this Field Trip Committee will enable these Societies to finance and publish additional geological contributions.

Thanks are due the Department of Geology at Southern Methodist University for being understanding of the time this effort has required of me.

It takes the united efforts of many individuals to bring a publication of this sort to fruition. Thanks are due all contributors to the symposium including the help of the Ardmore Geological Society on the road logs. The cooperation and effort beyond the call of duty of the typist, the draftsmen, and the printer merit appreciation and our heartfelt thanks. And last but not least, thanks are due a group of very understanding wives.

Dan E. Feray
Chairman, Field Trip Committee

March, 1959

ERRATUM

Page 73: Woodford of Arbuckle Mountains and Arkansas novaculite of Ouachita Mountains are Middle ? and Upper Devonian age not Mississippian .

CONTENTS

ILLUSTRATIONS

FOREWORD

The decision to attempt a field conference and symposium in the western Ouachita Mountains was made with the knowledge that the idea could become reality only if we received the cooperation and contributions of papers from more than a score of geologists who had done recent work in the area.

The committee enlisted the support of the Ardmore Geological Society and the Dallas Geological Society and extended invitations to possible contributors to the symposium and guidebook. We were inspired by the enthusiasm of the responses we received.

It was the intention of the committee to present accurately, and in summary form, the general outlines of the depositional and structural setting of the western Ouachitas. We believed the intention could be effectuated by (1) reviewing the history and development of geological thinking about the area; (2) presenting new and hitherto unpublished information of detailed surface and subsurface mapping, paleontology and petrography; and (3) offering the recognized authorities on the geology of the area the opportunity to amend, re-interpret or reaffirm their ideas previously in print.

Thomas B. Nolan, Director, United States Geological Survey, furnished copies of Oil and Gas Investigations Preliminary map no. 66, from which the final map showing outcrop patterns in the western Ouachitas was evolved. This map is Figure 3 of Thomas A. Hendricks' paper. Permission was granted by John T. Lonsdale, Director, Bureau of Economic Geology, The University of Texas, for publication of the map of the Ouachita Structural Belt, by Peter T. Flawn, prior to its publication by that agency. Carl C. Branson, Director, Oklahoma Geological Survey, gave permission to publish the map of the Boktukola syncline made by O. B. Shelburne. The photo-map of the western Ouachita Mountains and the McAlester Basin was reproduced from material generously furnished by Edgar Tobin Aerial Surveys.

Much research information compiled for doctoral theses at the University of Wisconsin has been used by Lewis M. Cline in connection with the guidebook discussions.

W. C. Kerr, Jr., draftsman in the research department of the Sun Oil Company, did most of the drafting on the larger maps. Nat Elliott of Mercury Printing Company, gave assistance and advice far beyond his contractual obligation in the printing, spacing and arranging of the final make-up of the symposium and guidebook. Mrs. Irving B. Hamilton prepared the final manuscripts for photo-offset printing. The excellence of her typography allowed the printing of the symposium in a manner generally restricted to more expensive printing methods.

Charlyne Feray generously contributed many hours redrafting the correlation charts in the symposium. Esther Hilseweck volunteered to transcribe the numerous editions of the road log from the field notes.

Dr. James D. Morrison of Southeastern Oklahoma College, Durant, Oklahoma, very willingly prepared the excellent article on the historical and geographical background of the area of the field trip.

The committee extends its thanks to W. W. Clawson, Senior Vice-President, Magnolia Petroleum Company, who arranged for the low-angle aerial photographs used in the symposium and guidebook. The use of the airplane, the film and the camera were furnished by that company. John Halsey of that company took the photographs and William Jenkins, also of Magnolia, prepared the lantern slides that were circulated among the affiliated societies of the A.A.P.G.

The field assistance of Allan M. Bennison, Chalmer L. Cooper, Bruce Harlton, Norman L. Johnson, Richard B. Laudon, and Hugh D. Miser was contributed at various times during the period from July to October, 1958.

In addition to his contribution to the symposium, Thomas A. Hendricks spent more than ten days in the field with Cline and others, reviewing known evidence and examining new localities. Much of his original information has been drafted for use in the guidebook. For his counsel and the contribution of his experience in the area, the committee is grateful.

Lewis M. Cline is perhaps the individual most responsible for the finished product of this symposium and guidebook. When asked to lead a field trip, write a paper for the symposium and aid in the editing of it, he promptly volunteered.

The extremely wide range of the papers in this symposium gives a fair concept of the amount of work that has been done and of projects that are underway within the Ouachita Mountains. The committee feels that this symposium is a representative progress report and brings sound concepts of the geology of the area before the profession. It is a tribute to the contributors; to their generosity, their abilities and their attitudes.

A steel worker in the British Isles, standing on the dock as the Queen Mary sailed away from her launching, was asked why he was smiling so proudly as if he owned the vessel, when he was just another of the thousands who had helped in the building. He replied, "It's an honor to have stuck rivets in such a grand ship".

Having acted as information gatherers and errand boys for this venture, your committee feels that the steel worker just about said it for us, too.

March, 1959 W. J. H.

THE CHOCTAW NATION

James D. Morrison [1]

> Serene and charming as the blissful regions of fancy, nothing here appeared to exist but what contributes to harmony.
>
> Thomas Nuttall, 1819

The land now comprising all or part of thirteen southeastern Oklahoma counties was ceded to the Choctaw Nation in 1830 "to insure to them while they shall exist as a nation and live on it" The original cession to the Choctaws included one-third of the present state of Oklahoma, extending from the Arkansas line west to the hundredth meridian and between the Red River on the south and the Arkansas River and its South Canadian branch on the north. A formal patent was issued to the Nation in 1842, signed by President John Tyler and Daniel Webster, secretary of state.

The Choctaws, however, were soon limited to the eastern third of this domain. They had no real need for all the region and made no use of the western two-thirds of the original grant. Soon after the Choctaw removal from east of the Mississippi, the Chickasaws were located among them. The Chickasaws were not satisfied to remain a minority among the Choctaws and petitioned the United States for a separate country. An agreement was therefore made in 1855 between the Choctaws and the Chickasaws, with the United States participating which removed from control of the Choctaws two thirds of their original grant. The central area was given to the Chickasaws while the more remote western region, on which some Plains Indians were living was leased to the United States so that these Indians might be left undisturbed. The line drawn between the Choctaw and Chickasaw nations began on Red River southeast of the present city of Durant, ran up the Island Bayou west and north to the source of the most easterly branch of that stream, thence northward until it intersected the South Canadian River, two miles west of the present town of Allen. The Chickasaws thus received the fertile valley of the Washita as far northwest as the ninety-eighth meridian, their western boundary, and much good bottom land on Red River.

All the area of this field conference lies within the remaining one-third of the Choctaw country, lying between the South Canadian and the Arkansas rivers on the north; the state line of Arkansas on the east, and the middle of the Red River on the south.

The area of the Choctaw Nation proper, as thus limited, was almost eleven thousand square miles, having the shape of an irregular trapezoid. The eastern boundary measured more than 120 miles and the parallel western edge some seventy miles; the greatest east-west distance is some 115 miles.

Maps in this symposium show the principal topographic features of the Choctaw country to be a double backbone of mountain ridges running east and west through the central part of the region. They are a continuation into Oklahoma from Arkansas of the Ouachita Mountains, which begin just west of Little Rock and continue to the western part of the old Choctaw Nation. The northern ridge of this double spine enters Oklahoma as Rich Mountain just west of Mena, Arkansas, and continues generally westward as the Windingstair. This is the true divide for the Choctaw country: streams to the north of it flow into the Arkansas, while surface water to the south flows to the Mississippi by way of the Red. This divide continues from the Windingstair toward the southwest until it reaches Pine Mountain some three miles northeast of Limestone Gap, near the present highway from McAlester to Atoka. The divide then veers to the northeast for a few miles but soon turns irregularly north-

[1] Dean of Instruction, Southeastern State College, Durant, Oklahoma.

west toward the South Canadian, coming very close to that stream at the northwest corner of the old Choctaw Nation. West of Pine Mountain nearly all the region is drained into Red River by way of the Boggy River and, in the southwest corner, by the Blue.

The highest point on the divide is near the Arkansas line atop Rich Mountain, 2850 feet, the altitude becoming lower toward the west. The elevation of Pine Mountain is only 1250 feet. Several of the eminences in southeastern Oklahoma, such as Rich Mountain, Cavanal Mountain, Sugar Loaf Mountain, and the Kiamichi ridge proper, rise more than two thousand feet above the surrounding valleys.

This main divide is but a few feet higher than a twin ridge, the Kiamichi Mountain, which lies to the south a few miles and parallels the Windingstair for some distance. In the valley between the ridges, the Kiamichi River flows west from its source in Arkansas to Tuskahoma (on the south side of the Potato Hills and about 15 airline miles southeast of the second day's lunch stop) almost the geographical center of the Choctaw Nation; there it breaks through the confining hills and runs southwestward and then southeastward into the Red River (see aerial photo in pocket).

North of the ridge, the Poteau River rises in Arkansas, flows westward into Oklahoma until joined by an eastward-flowing branch, the Fourche Maline, where it turns north and east to join the Arkansas River near Fort Smith. Little River rises south of the Kiamichi Ridge in Oklahoma, runs west a few miles to Nashoba, and then turns south and parallels Red River as it moves to the east across southern McCurtain County. Here Little River is augmented by two southward-flowing tributaries, Glover and Mountain Fork, the second of which has its source in Arkansas.

The lowest point in Oklahoma is where the Arkansas state line crosses Red River, the elevation being some 350 feet. All the rough country between the Kiamichi Mountain and the Red River Valley is popularly called the "Kiamichi Mountains" although the highest elevation in the region is not much more than 1250 feet.

North of Rich Mountain and the Windingstair, which was so named because the "old military road from Fort Smith to Fort Towson wound back and forth along the steep side of the mountain", is a region of sandstone hills and prairie plains. These prairies are particularly extensive along the Arkansas and in the valley of the Poteau. The valley of the Fourche Maline, just north of the Windingstair, is separated from the valley of the Arkansas by the Sans Bois Mountains, with the highest elevation in this upland, southwest of the town of McCurtain in Haskell County, being 1650 feet. East of the Sans Bois are such mountains as Cavanal near Poteau, 2400 feet, and Sugar Loaf near the Arkansas line, 2600 feet, both of which are quite imposing as they rise high above the surrounding prairies.

Early visitors were impressed by the rugged appearance of the Choctaw region. Thomas Nuttall, an English naturalist, remarked in 1819 that the "dividing ridge of Pottoe and Kiamesha" was "nearly the height of the Alleghany in Pennsylvania, very rocky, and thinly scattered with oaks and pines". Of the Kiamichi hills he said that they "strongly resemble the mountains of the Blue ridge at Harper's Ferry in Virginia". The botanist observed incorrectly that the Windingstair was as "high as any part of the Blue ridge" and mentioned two of the present scourges of the area--ticks and forest fires. Dr. Edwin James, who was with Stephen H. Long on his exploration of the Canadian and Arkansas rivers in 1820, speaking of the Ozark and Ouachita uplands in general, noted a "striking resemblance between this range and the Alleghanies" but qualified the remark by saying that in "some" particulars there was "dissimilarity".

A more prosaic early account is that of Captain John Stuart, who in 1832 had the thankless task of supervising construction of the first crude wagon road across the region from Fort Smith to Horse Prairie on Red River. He recorded that

> ". . . the land over which this road runs is for the most part, an ash colored clay, extremely poor and is timbered principally with Post Oak and Black Jack, with occasional groves of Pine and there is on the whole route something like 20

or 25 miles of Prairie, a very considerable portion of the route is covered with stone; and from where it begins. . . to the Kiamichi Prairie it is entirely a free or sandstone and from that place to Red River, twenty miles distant, is entirely a lime stone."

Stuart's road crossed the Windingstair north of Talihina and followed the valley of the Kiamichi River southward, nearly the same route used by the present Saint Louis and San Francisco Railroad line. This mountain region was little populated by the Choctaws. Many, of course, lived in the valleys and prairies which are scattered here and there among the wooded hills, but even today there are hundreds of square miles of rugged forest land with few human habitations.

There are, then, three topographical regions in the Choctaw land. There is a central upland, the highest elevations to the east near Arkansas, with a gradual slope to the west. To the north of this upland is a prairie and sandstone hills region, the plains predominant toward Arkansas and giving way to more and more timbered hills in the west. To the south of the central backbone and widening toward the west lie the prairies of the Red River valley. Toward the northwest the sandstone hills region joins the Cross Timbers, the historic dividing line between the "country suited to agriculture" and the "great prairies", to quote Captain Randolph B. Marcy, explorer of the upper Red River at a time when the myth of the Great American Desert was beginning to be questioned.

All the principal ridges of the central Ouachita Mountain area--Jackfork, Windingstair, Buffalo, Blue Bouncer, Rich, Kiamichi, Blackfork--are characterized by steep slopes and "sharp, straight barren crests" covered with rock debris. This area exhibits the roughest topography in Oklahoma.

The sandstone hills and prairie plains region in the north is not of the same geological origin as the Ouachita Mountains. Much of Pittsburg, Atoka, Coal, Hughes, Latimer, and Haskell counties, with the eastern portions of Pontotoc and Johnston counties, are sandstone hills with intervening prairies. Charles N. Gould also included in this general classification the Poteau, Sugar Loaf, and Cavanal mountains, near Arkansas in the northeast. This area also has alternate strata of sandstone and shales, but the topography is not so rough as that of the Ouachitas proper. Between the wooded hills there is much flat land and rolling prairie and the support of a fairly large population is therefore possible.

The prairie plains of the Arkansas and Poteau valleys were described in 1853 by Lieutenant A. W. Whipple, who surveyed the thirty-fifth parallel route for a possible Pacific railroad. He called it "a country of well-wooded hills, with gentle slopes and fine grassy prairie intervening". In 1849, a poorly informed Argonaut, who crossed the same area on his way to California, wrote his home town paper in Mississippi:

> "The rivers, the valleys, the prairies, and the mountains, all appear to have been planned by dame Nature to assist man in contemplating the works of Him who rules and governs all things, and preparing him for the enjoyment of that bliss which is promised only to the true and faithful beyond the grave.
>
> "Uncle Sam in giving this splendid country to the Choctaws got badly cheated, and I would not give even the small portion I have seen in my travels to this place, for all they owned on our side of the 'Big Drink'. . . "

The violent reaction of a Choctaw citizen to the statement that "Uncle Sam" had "given" the country to his Nation can well be imagined. Captain Randolph B. Marcy, who commanded the military escort for the party of which this Mississippian was a member, himself described the northern Choctaw country in his official report:

> "On departing from Fort Smith the route traverses a gently undulating district, sustaining a heavy growth of excellent timber, but occasionally interspersed with prairie lands, affording a luxuriant grass for eight months in the year, and intersected with numerous small streams flowing over a highly productive soil, thus embracing the elements of a rich and beautiful pastoral and agricultural country."

This type of surface, Marcy reported,

extended west from Fort Smith for "one hundred and eighty miles to near the 99th meridian of longitude, where the road emerges from the woodlands and enters the great plains . . . ."

South of the Ouachitas in the Choctaw land stretches another well watered and timbered rolling country, the relatively narrow valley of Red River, which ranges from ten to forty miles in width. The northern edge of this valley is a hilly, sandstone region covered by timber and is the north edge of the Gulf Coastal Plain. A strip of black waxy upland prairie, rolling or level, lies just south of these border hills; then comes a strip of sandy upland, and finally the Red River bottoms. The tributaries of the Red River--Blue, Boggy, Kiamichi, and Little Rivers--have here generally southeasterly courses, with steep mud banks and narrow valleys.

It was in the northeast prairie plains and the Red River valley that the Choctaws concentrated most thickly on their removal from Mississippi and Alabama--a natural result, since they had become largely a sedentary, agricultural people and their leaders were of mixed Indian and white blood. The rugged Ouachitas made communication difficult between these settled sections.

More than ninety per cent of the Choctaw country was originally, and remains today, a timbered region. Nearly half was oak-pine forest. This has been largely cut over in recent years but much is now becoming reforested. In McCurtain County a number of old fields have grown up in loblolly pine, but the short-leaf is the usual variety of the yellow pine. There are few areas of pure pine timber, the pine being interspersed with white oak, post oak, and blackjack oak, as well as black hickory on the uplands; and with red oak, sweet gum and sour gum, holly, ash, cedar, and some cypress in the bottoms. Shrubs common to the oak-pine forest include huckleberry, sumac, wild grape, greenbriar, dogwood, willow, and alder, with redbud scattered everywhere.

About ten per cent of the land is tallgrass prairie which is located chiefly in the Red and Arkansas valleys, in the northeast and the southwest. Here the native grasses are big and little bluestem, Indian, and switch grass; there is some grama and buffalo grass in the west. The topography of the prairie country is flat to gently rolling. Nuttall waxed poetic in one of his descriptions of the Choctaw prairie land:

> "These vast plains, beautiful almost as the fancied Elysium, were now enamelled with innumerable flowers, among the most splendid of which were the azure Larkspur, gilded Coreopsides, Rudbeckias, fragrant Phloxes, and the purple Psilotria. Serene and charming as the blissful regions of fancy, nothing here appeared to exist but what contributes to harmony."

Near the "Pottoe" the botanist "found the whole country a prairie, full of luxuriant grass about knee high . . ."; this was in the spring of 1819.

These prairies, together with the second bottoms along the streams, furnished the finest and richest farming lands. A geological survey of Bryan County in 1914, where there is a large percentage of prairie and bottom land, disclosed thirty-one different types of productive soil. In the Red River valley are alluvial, limestone prairie, and sandy soils. In the central hills and mountains the soils are poor except in the narrow stream bottoms, but good crop soils of clay, clay loam, and alluvial types characterize the Arkansas and Canadian valleys except where the sandstone hills appear. Because of the thin soils and rough surface, the greater part of the Choctaw hill and mountain area is limited to pastoral and lumbering pursuits.

Beneath the surface of this land are mineral deposits of value. Coal in particular attracted the first large concentrations of white settlement in the north and west. Nuttall and other early visitors noticed the presence of coal-bearing outcrops as they traveled through the Choctaw country in the period before the Civil War. Asphalt deposits are numerous, as are beds of pottery clay. Quartz, mica, natural gas, and limestone, suitable for road material, ballast, building stone, rock wool, and cement-making, are all present in quantity and mined in various parts of the region.

Wild game was abundant on the prairies and among the hills when the Choctaws first came. Today the only area of Oklahoma where an open deer season is allowed lies within six counties of the Choctaw domain. Buffalo were once plentiful. Peter P. Pitchlynn, Choctaw chief who met Charles Dickens on an Ohio River steamboat in 1842, invited the author "to go home with him and hunt buffaloes". La Harpe, a Frenchman who crossed the country of the Choctaws in 1719, reported killing several buffalo and a very large bear, seeing "wild cattle, . . . and a great number of wolves", and "some partridges, woodcock and plover". He also saw some Indians "busy smoking some unicorn meat" on the banks of the Canadian. La Harpe described this animal, whatever it was, thus:

> "It is an animal big as a middle-sized horse; he has hair of reddish color and the length of that of the she-goats, the legs rather thin and in the middle of the forehead a horn, without branches, of a half-a-foot long; the meat of it is very delicious. This discovery confirms that which M. de Bienville had been told of the savages that in the upper head waters of the Ouachitas River there were some unicorn."

The official Choctaw-Chickasaw exploring party which visited the country just prior to removal after 1830 reported seeing wild horses, deer, and wild turkey along the Canadian and that the headwaters of Boggy and Blue, were a fine game region. This party was led by Colonel George S. Gaines, acting for the United States government, a man in whom the Indians had the utmost confidence. His name was given to Gaines Creek after 1830. This stream, a tributary of the Canadian, was previously called the South Fork of the Canadian. A private exploring party of Choctaws in 1830, led by George W. Harkins and Robert Folsom, saw many bears and turkeys and reported that "on the west side of Kiamissa 15 or 20 miles there is buffaloe to be seen in great numbers".

The climate is temperate, the mean annual temperature about 60° F. Evidence that the climate is much the same as a century ago is indicated by a glance at the weekly weather reports published in the Choctaw Intelligencer near Fort Towson during the early 1850's. The annual rainfall is usually between thirty-five and forty-five inches, but in some years goes above fifty inches. The growing season ranges from about 210 days in the Arkansas valley to 230 along the Red. As in much of the southwest and all of the middle western United States, the most spectacular weather for the Choctaw region is furnished by a conflict between enormous air masses. Occasionally a cold, dry, and heavy air mass moves in from northern Canada to displace at the surface warm and moist air of the Gulf region. The faster this Canadian or Arctic air moves and the greater its contrast to the Gulf air, the more turbulence, heavy precipitation, and bad weather results along the border area, or front, between the air masses. The typical front moves across southeastern Oklahoma from the west or southwest toward the east and northeast. It is accompanied by high winds and tornadoes, especially in the spring (although these may occur during any month of the year), and sudden sharp drops in temperature as polar air replaces gulf or tropical air.

The timbered regions bear many scars as a result of tornadoes. Near the Arkansas line, at least, an old tornado path where the twisted and broken trees remain is locally called a "hurricane". The storm itself will likely be named a "cyclone". Gulf hurricanes do affect the Choctaw weather, but only in the form of a little rain as the storms come to an end over the dry land. The Choctaws immediately after removal were plagued by tornadoes just as the present inhabitants are. Such a storm partly destroyed the Chuahla Female Seminary near Fort Towson in the 1840's, and hail stones as large as quart cups were reported to have fallen at Armstrong Academy, north of the present Bokchito, in 1847. The Reverend Ramsey D. Potts, Baptist missionary who was principal of Armstrong Academy at the time, was said to have measured one hail stone which was "six inches in length, and about four inches in diameter". Oklahoma weather and its reporters apparently have not changed much in the past century.

The language they spoke has left many names on the land, including Oklahoma itself, which is the Choctaw word for Indian, literally "people (Okla) red (homa)". Ouachita, or Washita, is derived, through the French as the first spelling indicates, from owa (hunt) chito (big). The Choctaw word for stream was bok, which the French first pronounced bayuk and finally made into bayou. It is not uncommon to find a stream in southeastern Oklahoma called a bayou rather than a creek. Names such as Bokchito (creek, big) and Bokoshe (creek, son of, or little creek) are found on the current Oklahoma map. Tuskahoma, the capital, means "warrior, (tuska) red (homa)". Another name reminiscent of the Choctaw Indian heritage of the region is Talihina, from tali (iron) and hina (road). Others include Kinta (beaver); Alikchi (doctor); Nashoba (wolf); Kosoma (short for "isi kosoma" or "deer fetid" (goat); Tamaha (town); and Atoka (from hitoka, ball ground).

Tuskahoma, geographically and figuratively, was the heart of the old Choctaw Nation. Not only did it lie near the center of the Nation but it served for many years as the capital. The general location was selected by the Choctaws for their seat of government immediately upon removal from east of the Mississippi.

Near Tuskahoma in 1834 the Choctaws wrote the first constitution by any group, white or Indian, on Oklahoma soil. They established a government with executive legislative and judicial branches, a state police force, built schools and sent their brighter scholars east for graduate study. By this and later constitutions they governed themselves until their semi-independence ended and their people became citizens of the state of Oklahoma in 1907, by treaty. The saga of the Choctaws may well be envisioned as the 19th century version of "peaceful co-existence". Their last council house, constructed in 1883, still stands near Tuskahoma as a monument to the Choctaw Nation of Indians.

This is the land which the United States "gave" the Choctaws in 1830: parts of two river valleys separated by a rugged, timbered heartland. For the Choctaws it was literally a promised land. They hoped that it would be theirs forever, a permanent refuge from the advancing horde of whites who had overwhelmed them in the East. It would be in the Indian tongue, atukko, which may be translated "haven, retreat, or shelter".

The number of full-blood Choctaws grows smaller with each passing year, but the Choctaw blood is widely diffused throughout the population of southeastern Oklahoma. Thousands of Oklahomans are now of mixed blood as a result of more than a century and a half of intermarriage between the races. Individuals with Choctaw blood are leaders and useful citizens at all levels of society and in all walks of life throughout the state of Oklahoma, although the Choctaw blood is still concentrated in the old Nation. In these and other ways the Choctaw Nation still survives in the land which was "to inure to them while they shall exist as a nation and live on it".

OUACHITA PROBLEMS

C. W. Tomlinson[1]

Foreword

Among the geological questions concerning the Ouachita Mountains, on which contrasting opinions still can be (and are) held in good faith, two stand out because of their far-reaching significance as to the structure and stratigraphy of the region, and the orogenic mechanisms by which that structure was attained. These are: (1) the extent of low-angle thrust faulting, and (2) the source and mode of emplacement of the exotic boulders which occur most abundantly in the Johns Valley shale. Related questions include the following: (3) how sharp are the contrasts between Arbuckle and Ouachita facies? (discussed in connection with question one); (4) what is the age and correct correlation of the Crystal Mountain sandstone and older strata?; (5) what is the age of the Stanley shale and Jackfork sandstone?; (6) how should the abrupt development of the Stanley and Jackfork to great thickness be explained? (discussed in connection with question one); (7) were the masses of Caney (Delaware Creek) shale which are a ubiquitous component of the Johns Valley shale deposited in situ, or are they exotic like the associated limestone and sandstone boulders? (discussed in connection with question five); (8) how much of the overlying "Atokan" section, if any, may be of Morrowan age?

Evidence Favoring Extensive Low-Angle Thrust Faulting

The closely faulted belt. - Thrust faulting is conspicuous in the northwestern margin of the Ouachitas in Oklahoma. Sharp contrast between much of the Paleozoic section east of the main belt of faults and that west of it served to evoke the plausible idea that this faulted belt perhaps constituted the forward edge of a great thrust sheet which had moved many miles northwestward, shoving into juxtaposition two facies of rocks which may have been deposited scores (Dake, 1921) or even hundreds (Ulrich, 1927, p. 26; van der Gracht, 1931, p. 999; Harlton, 1933, p. 6) of miles apart. Differential movement to such distances has been envisaged as taking place chiefly on one or more low-angle sole faults, above which the imbricate thrusts developed (Powers, 1928, Fig. 4, p. 1039; Miser 1929, Fig. 6, p. 19 and Pl. 2; Hendricks, 1947, sheet 3, Howell, 1947, pp. 51-52).

Sharp convergences. - The stratigraphic contrast between the Ouachita and Arbuckle facies is especially sharp in Atoka County, at the western tip of the mountains. That part of the stratigraphic section (Caney shale or early maps) between the Wapanucka limestone (Morrowan) above and the Woodford formation (lower Mississippian or upper Devonian) below is less than 1200 feet thick in a well[2] three miles west of the Choctaw fault as mapped by Hendricks and others (1947) - the outermost major fault in the Ouachita border belt. Less than ten miles to the southeast, on the other side of the belt of faults, the Stanley and Jackfork groups appear in the correlated interval above the Arkansas novaculite, and attain a thickness not less than 4,000 feet in T. 1 S., R. 12 E. Within the next 20 miles to the east this increases to more than 15,000 feet (Harlton, 1938, pp. 864-886; Hendricks and others, 1947, text on sheet 1).

Farther north, the thinner section persists into the middle of the faulted belt, but its thickening is equally sudden beyond that.

[1] Geologist and independent oil producer, Ardmore, Oklahoma.

[2] Pasotex Petroleum Co. and Anderson-Prichard Oil Corp.'s No. 1 Underhill, well in the cen. SE 1/4 SW 1/4 sec. 26, T. 1 N., R. 11 E., Atoka County, Oklahoma.

At Pinetop, in sec. 4, T. 2 N., R. 15 E., strata between the Atoka (presumably post-Wapanucka) and Woodford formations as mapped by Hendricks and others (1947, sheet 1) occupy a band only 800 feet wide on the map. Three to five miles southeast, at least 3000 feet of Stanley and Jackfork strata occur in Bald Mountain, between faults which apparently cut out both the lower Stanley and the upper Jackfork.

Contrasts in facies. - Other parts of the Paleozoic section show such substantial differences in facies between one side of the faulted belt and the other, that different formation names are applied to a majority of them as between the two areas (Taff, 1902, and all later writers).

Fensters. - Substantiation of the presence of folded low-angle thrusts in the Ouachita Mountains of Oklahoma has been claimed by interpretation of an area in the Potato Hills (Tps. 2 and 3 N., R. 19-21 E., Pushmataha and Latimer Counties; Miser, 1929, pp. 18-20; Powers, 1928, Figs. 3 and 4, pp. 1038, 1039, crediting Miser; Hendricks and others, 1947, sheet 2, inset map; Miser, 1954, Miller, 1956) and one in the Choctaw anticlinorium (T. 4 and 5 S., R. 23-25 E., McCurtain County; Miser, 1929, pp. 20-21; Miser, 1954) as fensters. If true, these would demonstrate a horizontal component of movement of the overthrust mass, toward the present main trace of the underlying thrust fault, of at least 10 miles in the first case and 20 miles in the second case.

An unpublished interpretation (Arbenz, 1955; Miller, 1956b) which would tend to confirm the existence of the Potato Hills fenster claims that the strata in the northern rim of the fenster, in the overthrust sheet, display synclinal structure in beds that are upside down, and postulates that this constitutes the toe of a recumbent anticline or nappe which had its roots in the southern rim of the fenster.

Faults near Waldron, Arkansas. - Recent mapping in the Waldron quadrangle, Arkansas (Reinemund and Danilchik, 1957) portrays a diagonal fault (the Jones Creek fault) which is interpreted by the authors of the map as a tear fault constituting the eastern end of the Choctaw thrust sheet, whose northern limit is the Choctaw fault. They also show a curving transverse fault (the Johnson Creek fault), the plane of which is indicated to be nearly horizontal for several miles, and is interpreted as the base of a thrust sheet which has its northern limit at the Ti Valley fault.

Outcropping low-angle thrusts. - Two nearly horizontal faults of small displacement are known in surface exposures, both in road cuts. One is on Oklahoma State Highway 3 in the SE 1/4 sec. 14, T. 2 S., R. 11 E., just southeast of Atoka, in Black Knob ridge (for a photograph of it, see Harlton, 1953, Pl. 1, p. 782. The other is on United States Highway 71 in sec. 12, T. 5 S., R. 32 W., four miles northwest of Wickes, Polk County, Arkansas.

Boulders of tectonic origin? - It has been suggested (van der Gracht, 1931, pp. 1021, 1051; 1931a, pp. 709-714; Harlton, 1933, pp. 5-6; Howell, 1947, pp. 51-52) that at least some of the exotic masses commonly referred to as boulders, most of which are found in the Johns Valley shale, may be fault slices or constituents of fault gouge, and themselves help to substantiate the sole-fault hypothesis by indicating the presence at depth in the Ouachita area, of formations of non-Ouachita facies. Even by some who thought the boulders more probably were deposited by other methods, they have been regarded as evidence of such a buried facies, although various locations have been suggested for it (Powers, 1928; Kramer, 1933; Rea, 1947).

Relation of Ouachita to Arbuckle structure. - Records of wells drilled in north Texas indicate that rocks classified as belonging to the Ouachita facies exist there at least as far west as a linear southwest projection of the Ouachita Mountain front from Atoka, Oklahoma. Nevertheless, strata classified as of Arbuckle facies have been drilled in the Arbuckle trend 15 miles southeast of such a projection, in sec. 13, T. 5 A., R. 12 E., Bryan County, Oklahoma. Other wells afford evidence that the faults which border the Tishomingo-Belton horst-anticline, the chief structural feature of the eastern Arbuckle Mountain area, also continue southeastward beyond that projection. Outcrops of Ouachita facies and with a Ouachita trend (southwesterly) occur on Fronterhouse Creek in sec. 19, T. 3 S., R. 11 E., within four miles north of the projection (beneath Cretaceous beds) of the southeastward-trending fault zone (Boggy Depot fault) which constitutes the northern border of that horst.

The two features are almost at right angles to each other. Their exact relationship is concealed under the edge of the Gulf coastal plain (cf. Hendricks, 1940, p. 2147). Some think that the Ouachita frontal fault curves back sharply into a reentrant, crossing the Tishomingo anticline farther east and curving westward again at its southern margin. If Ouachita thrusting took place later than the Arbuckle orogeny, that would be the more likely situation. If, on the other hand, the Ouachita frontal fault continues straight southwestward to a termination against the Boggy Depot fault, then movement on the latter must have taken place later than on the former; and such later uplift may have permitted erosional removal from the eastern end of the horst, of a Ouachita thrust sheet or sheets which once covered it.

In either event, these structural relationships can be explained by postulating that there has been substantial northwestward thrusting of the Ouachita front here - either alongside of a pre-existing horst (or arch) of Arbuckle-type rocks, or over one which was later re-exposed by erosion prior to its reburial under Comanchean sediments.

Evidence Tending to Minimize Low-Angle Thrust Faulting

Steep angles on major thrust faults. No sole fault proven. - Even by those who believe that low-angle thrusting has been very extensive, the major thrust faults are acknowledged to reach the surface at steep angles - commonly 60° to 80° and rarely less than 45° from horizontal (Miser and Purdue, 1929, Pl. 15 and Fig. 6, p. 138; Powers, 1928, Fig. 4, p. 1039; Miser, 1929, p. 18 and Fig. 6, p. 19; Hendricks and others, 1947, sheet 3). No sole fault has been demonstrated by drilling, nor have seismic data on the region been published, adequate to establish the presence or absence of such a fault.

None of the geologists who have spent most time in the Ouachitas (e.g., Miser, Hendricks, Harlton, Cline) has endorsed in print the extreme hypothesis that the entire body of sediments in the range has been imported by thrusting from a site outside its present borders.

Published cross-sections of McCurtain and southern Leflore counties, Oklahoma (Honess, 1923, Pl. 1; Miser, 1929, Fig. 4, p. 16) and of the DeQueen and Caddo Gap quadrangles, Arkansas, (Miser and Purdue, 1929, Pls. 15, 16; Miser, 1929, Figs. 2, 3, 5, pp. 13-15, 17) display fan folding rather than consistent thrusting or overturning in one direction (see Fig. 1). In the first-mentioned area, and also in the Hot Springs quadrangle, Arkansas (Purdue and Miser, 1923, sections B-B', C-C', D-D'), both northward and southward thrust faults are shown. In the first two areas just mentioned, the authors show flanking folds overturned outward on either side of major anticlines (cf. Tomlinson and White, 1935; Misch and Oles, 1957, pp. 1901, 1902). Folds overturned southward are especially common in the southern part of the Ouachita region (Purdue, 1909, Fig. 2, p. 43, and Fig. 5, p. 55; Honess, 1923, Pls. 1, p. 82; Purdue and Miser, 1923, sections A-A', B-B', C-C'; Miser and Purdue, Pl. 15, sections A-A', B-B', C-C'; Tomlinson and White, 1935; Misch and Oles, 1957, p. 1901).

Southeastward thrusting exists in the southeast limb of the Kiamichi anticline in T. 1 N., R. 18 E., Pushmataha County, Oklahoma (Cline, map, this volume). Farther south in the same limb, the Jackfork sandstone is overturned to the southeast.

Alternative explanations for the "fensters". The supposed fenster in McCurtain County has been proved non-existent by more detailed later studies (Pitt, 1955; Tomlinson and Pitt, 1955). It is a practically unbroken anticlinorium, with no major faults at its borders.

For the feature described as a fenster in the Potato Hills, an alternative explanation has been offered (Kramer, 1933, Fig. 3, secs. d and e; Misch and Oles, 1957, p.1901): broken anticlines facing each other across the faulted area; though Kramer's hypothesis of pre-Stanley orogeny seems unnecessary (see Fig. 2). Detailed mapping there (as yet unpublished: Misch and Oles, Allan P. Bennison, Tomlinson) seems to establish that the southern border fault of the supposed window does not curve smoothly into the northern

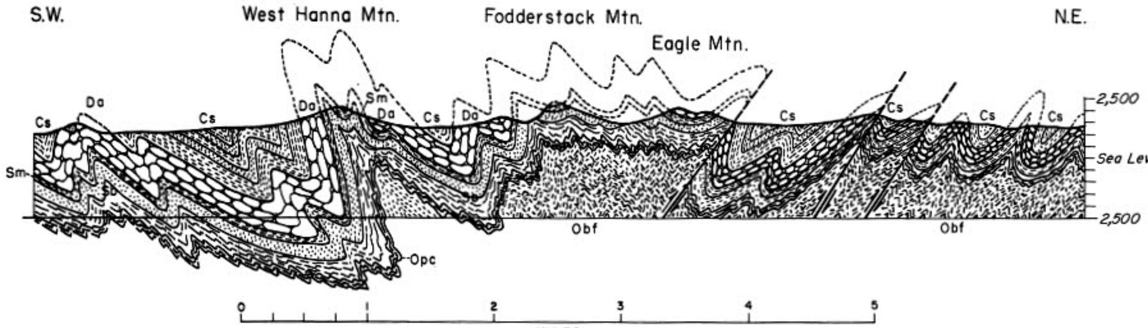


Figure A.—Structure section through several mountain ridges near Shady, DeQueen quadrangle, Arkansas. Shows a fan-shaped arrangement of folds. (After Miser and Purdue, U. S. Geol. Survey Bull. 808.) **Cs,** Stanley shale; **Da,** Arkansas novaculite; **Sm,** Missouri Mountain slate; **Sb,** Blaylock sandstone; **Opc,** Polk Creek shale; **Obf,** Bigfork chert.

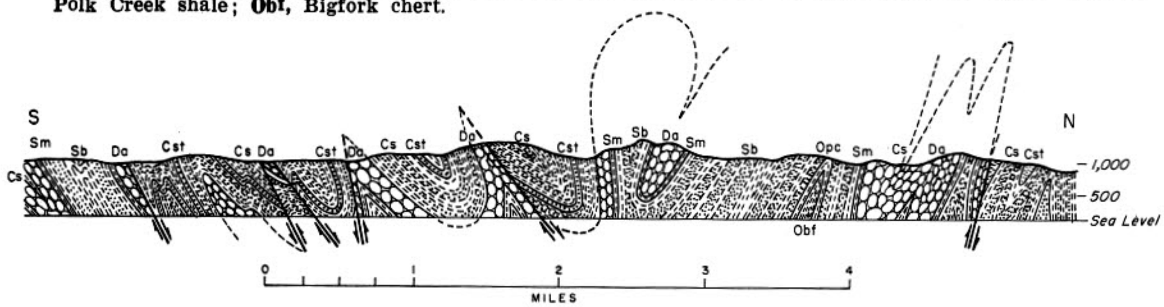


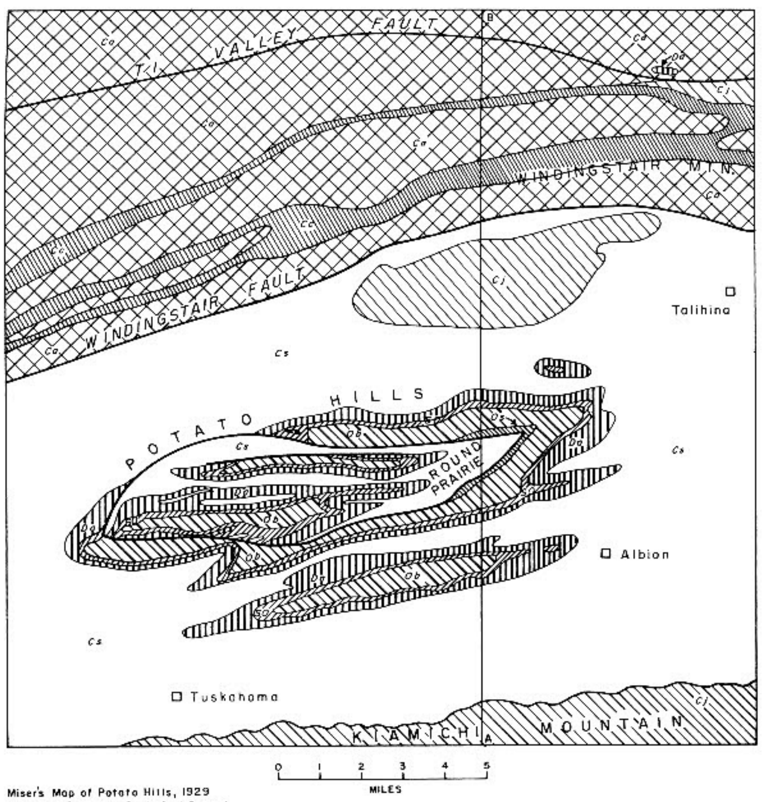
Figure B.—Structure section across several mountain ridges in northern McCurtain County, Okla. Shows fan-shaped arrangement of folds. (After Honess, Oklahoma Geol. Survey Bull. 32.) **Cs,** Stanley shale; **Cst,** Tuff near base of Stanley shale; **Da,** Arkansas novaculite; **Sm,** Missouri Mountain shale; **Sb,** Blaylock sandstone; **Opc,** Polk Creek shale; **Obf,** Bigfork chert.

Fig. 1.--Structure sections showing fan folding in Ouachita Mountains, Oklahoma and Arkansas. After Miser, 1929.

one at both ends as heretofore drawn (Miser, 1929, Fig. 6, p. 19; Hendricks and others, 1947, sheet 2, inset map). On the contrary, they do not seem to meet on the west (Misch and Oles, 1957, p. 1901); and the northern fault, swinging southeastward, ends against the other at nearly a right angle, while the southern fault continues northeastward beyond that junction (see Fig. 3; also Roe, 1955).

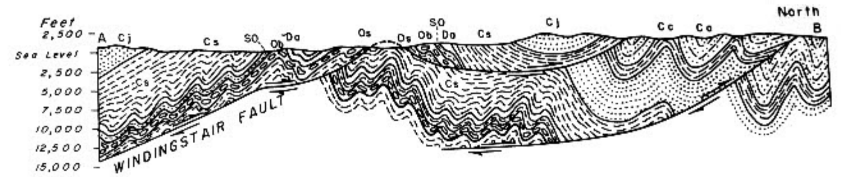
The areal geologic map of the Potato Hills suggests a possible analogy to the "rabbit ears" anticlinoria of the Ardmore district (Tomlinson, 1952, pp. 1826-28; Figs. 2, 8). In the Ouachita region we have two major shale sequences, each some thousands of feet thick. The lower one comprises the Mazarn and Womble shales, and the upper is the Stanley shale. Between them lie the relatively brittle Bigfork chert and Arkansas novaculite, with the intervening Polk Creek and Missouri Mountain formations, usually sketched here as a thousand feet or less in aggregate thickness for the four formations (Fig. 2). Anticlinorial folding rather than a simple broad anticline would be a natural result of crustal compression involving such a section. Suppose compressional stresses continued or were resumed after the initial development of such an anticlinorium, perhaps when the uplift had been partially stripped by erosion of its overburden of still younger formations; and that there exists below the Mazarn shale (as is probable), a thick section of strata which on the average possess greater competence than the shales above mentioned. This competent section would have been forced up in the anticlinorial area, in the early stages of the orogeny, to a position only a few thousand feet below the topographic surface. With further compression, what would be more natural than for the outer folds, flanking the core of the structural high, to do most of the later yielding? And for the brittle chert-novaculite section, between the great shales, to break near the crests of those folds and be thrust inward over the core area to some extent?

The two high Bigfork ridges which border the "fenster" on the north and south meet in the northeast corner of the hills, giving the appearance of a forked anticline with each of

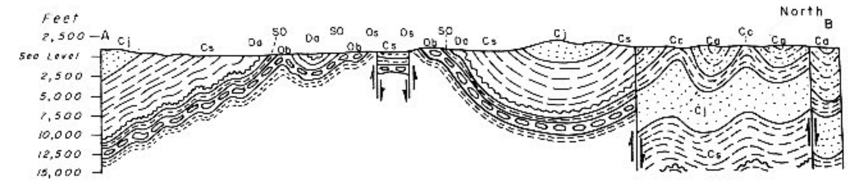


Miser's Map of Potato Hills, 1929
(Courtesy Oklahoma Geological Survey)

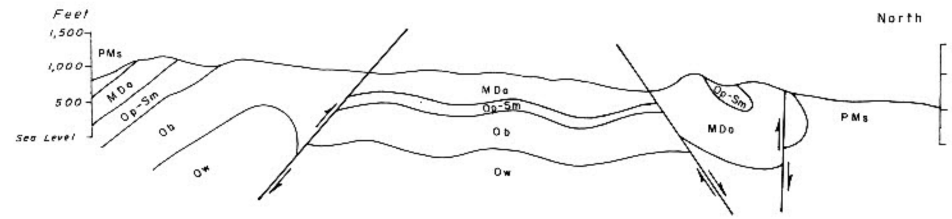
Ca, Atoka formation (Carboniferous); **Cc,** Caney shale (Carboniferous); **Cj,** Jackfork sandstone (Carboniferous); **Cs,** Stanley shale (Carboniferous); **Da,** Arkansas novaculite (Devonian and Devonian?); **SO,** Missouri Mountain shale (Silurian) and Polk Creek shale (Ordovician); **Ob,** Bigfork chert (Ordovician); **Os,** Stringtown shale (Ordovician).



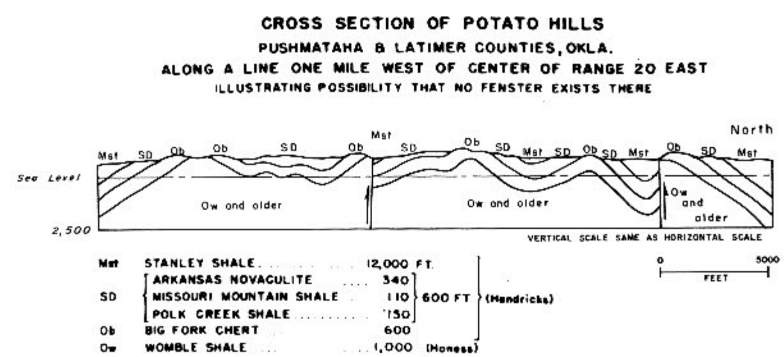
Miser's section, 1929. (Courtesy Okla. Geol. Survey.)



Kramer's section, 1933. (Courtesy Journal of Geology.)



One of Miller's sections, 1955. (Courtesy B.W.Miller.)



Tomlinson, 1956.

Fig. 2. Various interpretations of Potato Hills structure.

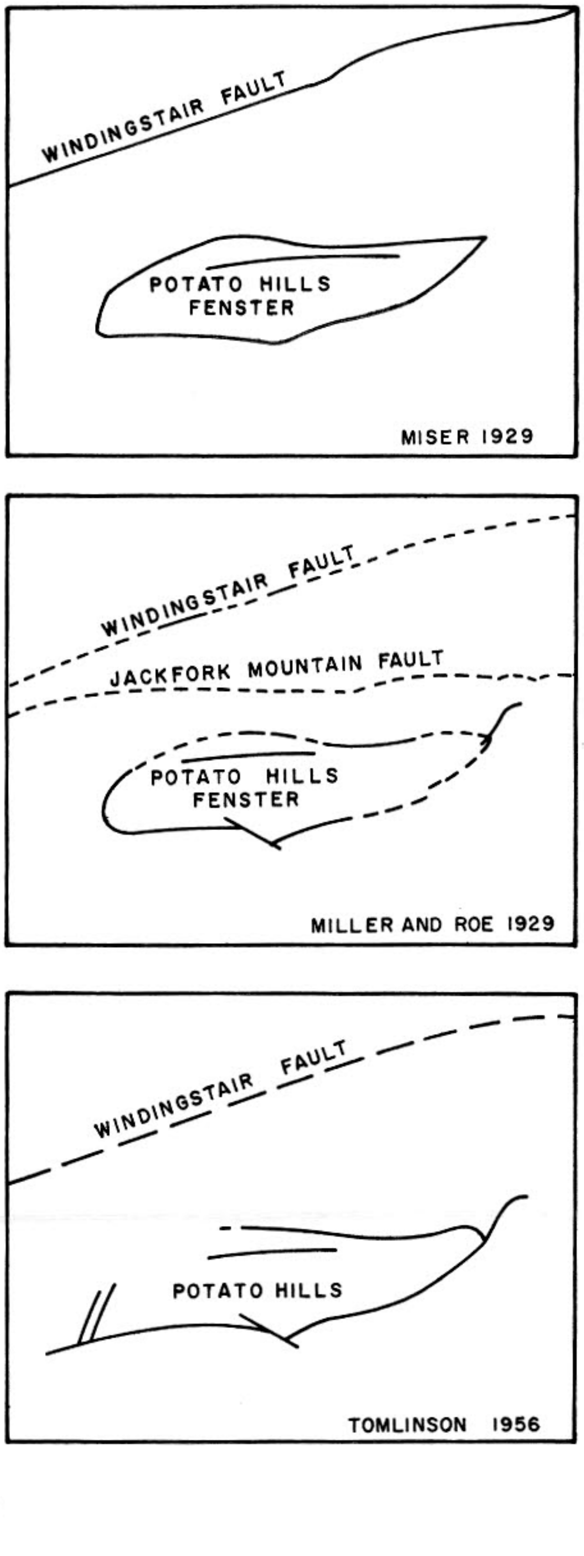


Fig. 3.-- Major fault patterns in the Potato Hills--three interpretations.

the two forks downfaulted toward the other. Only the latter feature is unusual, as forking anticlines are not uncommon. The extension of the southern fault past the northern one, which ends against it, suggests that movement on the longer fault, and upthrusting of the south rim, continued somewhat later than on the northern fault and rim anticline. This impression is strengthened by the fact that the formations in this northern rim ridge are overturned in their eastern portion, where the south rim has crowded against them, - but are not overturned in the western half of the north ridge (Roe, 1955; Tomlinson, 1956).

Difficulties with the fenster hypothesis for the Potato Hills. - Careful examination of the supposed toe of a recumbent nappe, on the north margin of the Potato Hills, has convinced at least one highly competent surface geologist (Bennison, 1957) that the strata reported to be upside down and in synclinal (recumbent anticlinal) position there (Arbenz, 1955; Miller, 1956b) are actually right side up. If so, they afford no support to the fenster hypothesis. The structure in question is confined to a very small area near the western tip of the Arkansas novaculite outcrop, and probably is set apart from the rest of that outcrop by small cross-faults.

No obvious limits exist for the postulated nappe. Along nearly the whole north rim of the Potato Hills, the Stanley shale borders the novaculite in normal stratigraphic sequence. Therefore the recumbent nappe had to be portrayed (Miller, 1956b) as extending beyond the Hills to a conjectural strike fault (interpreted as an extension of the Jackfork Mountain fault) in the Stanley shale belt north of the Hills. This was suggested as the farther side of the downfolded portion of the nappe. No east or west limits were indicated for it. They too should be represented by faults. If the nappe were coextensive to the west with the Jackfork Mountain fault as mapped by Miser (1954), it would be at least 60 miles long, and its forward edge nearly everywhere would be composed of Stanley shale. Yet no roots for it have been suggested except in the Potato Hills.

In the concededly autochthonous area within the supposed fenster in the Potato Hills are a number of close folds very similar in type to those which constitute the rim. Though not quite so high structurally, the inner anticlines bring to the surface the same sequence of formations as those in the supposedly allochthonous rim, from the Arkansas novaculite to and including Bigfork chert. This is perfectly normal if all the folds, inner and outer alike, are members of an autochthonous anticlinorium. But if the rim anticlines are part of a thrust sheet which extends (as claimed) the length of the Windingstair fault and has here moved at least ten miles toward the trace of that fault, is it not a remarkable coincidence that in this fenster, the only place in the 125-mile length of that thrust sheet where the autochthonous rocks beneath it can be seen, these rocks are identical with those in the overlying sheet? For the Potato Hills are also the only area in which these rocks appear at the surface in the supposed thrust sheet, in all its length.

Facies changes not all sharp. - The facies change across the belt of imbricate faulting is much less pronounced in some formations than in others (Harlton, 1953; Misch and Oles, 1957, p. 1904; cf. Hendricks, 1958, p. 2762). The eastward change in the Simpson group consists chiefly in an increase in the proportion of shale and siltstone and a relative decrease in limestone - normal changes in passing from foreland to geosynclinal facies. If the upper part of the Arbuckle group is represented, as has been suspected (Purdue, 1909, p. 30; Miser and Purdue, 1929, pp. 23, 24, 127), the change in it is similar. Electric logs of the Womble, Bigfork, Polk Creek, Missouri Mountain, Arkansas novaculite and basal Stanley formations are surprisingly similar to those of the Bromide, Viola, Sylvan (for Polk Creek and Missouri Mountain), Woodford, and Caney formations, respectively. The similarity of Caney to basal Stanley in this respect must be discounted as produced at different times, however, if the true position of the Caney in the Ouachita region is above the Jackfork. (See infra, discussion of age of Stanley and Jackfork).

The Bigfork contains more chert and less limestone than the Viola; but part of the chert in the Bigfork is accounted for by secondary silicification (Harlton, 1953, pp. 783-785), as clearly shown by fossil shells in which carbonate has been replaced by chert.

There is no appreciable difference between the Polk Creek shale and the graptolitic, bituminous lower member of the Sylvan (Harlton, 1953, p. 787). In some places in the Ouachita Mountains, the equivalent of the upper, greenish clay shale member of the Sylvan has been mapped in the base of the overlying Missouri Mountain, or in the Blaylock where the latter is differentiated. It becomes somewhat more indurated in the "core area" north of Broken Bow, where it might be described as platy claystone. Similarly, the Arkansas novaculite is practically indistinguishable from the Woodford chert of the Ardmore district (cf. Harlton, p. 790).

Little convergence in many formations. Thickening of Stanley-Jackfork section not confined to the border belt. Rate of convergence in it not excessive for a geosynclinal border area. - Except for the Stanley-Jackfork section, no extraordinary change in thickness is known to occur between the two sides of the main belt of thrust faulting, in the formations whose equivalents crop out on both sides of that belt. The rapid southeastward thickening of the Stanley-Jackfork section, as noted above, begins in the imbricate faulted belt; but most of it takes place south and east of that belt, in the less faulted central part of the Ouachita region. The thickening is no more rapid than in other formations in unfaulted areas similarly situated at the border of a geosyncline. For example, in the unfaulted area west of the Ouachita front, the Atoka formation thickens eastward from 2400 feet to 7800 feet in 12 miles.[3] East of the Criner Hills, the Pennsylvanian section of the Ardmore basin thickens within seven miles from 4000 feet in the Pleasant Hill syncline to 16,000 feet around Lake Murray. This is largely due to onlap of younger strata toward the Hills (the margin of the basin), which is not clearly demonstrable for the Stanley-Jackfork section, but it is at as great a rate as any change in the thickness of the latter. In both examples the distance has been lessened by crustal shortening, with resulting steep dips. No major thrusting is known to be involved in the Ardmore case.

The thrusts as a product of close compression of a geosyncline in situ. - The thickening of the Stanley-Jackfork is associated

[3] From sec. 35, T. 1 N., R. 9 E. (G. L. Reasor's #1 Woodford-Orr well) to sec. 26, T. 1 N., R. 11 E. Pasotex and Anderson-Prichard's # 1 Underhill well).

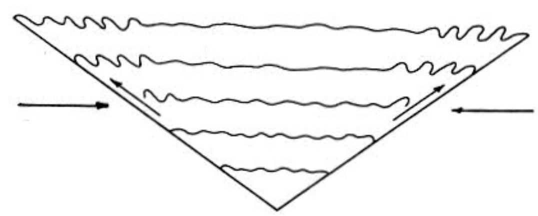
with the steep marginal slopes of the Ouachita geosyncline. Both this thickening and the observed facies changes in other formations doubtless now appear to take place somewhat more abruptly than they did prior to the Ouachita orogeny, because the imbricate faulting of the border belt, and also the inner folds, represent crustal shortening. Although the geosyncline has been compressed, and its contents may have bulged over its borders to some extent, there is no convincing evidence that it has been moved substantially from its original position. Two writers (Ulrich, 1927, p. 26; van der Gracht, 1931, p. 999) have postulated that the entire Ouachita range is now north of the original position of the geosyncline from which it was squeezed.

That concept has always been difficult to reconcile with the fact that the major thrust faults mapped in Oklahoma fail to extend far into Arkansas (Branner and others, 1929; Croneis, 1930, Pls. 1-A, 1-B). Reinemund and Danilchik (1957) recently have tentatively shown some of them as extending some 40 miles east of the State line, though they find them too obscure to map with solid lines along most of that distance, and do not indicate great stratigraphic displacement on them. None has yet been mapped along the easternmost 80 miles of the north border of the Ouachitas; and Misch and Oles (1958, p. 2773) state that they fail to find the bordering (Choctaw) fault in the easternmost 50 miles of its supposed trace in Oklahoma.

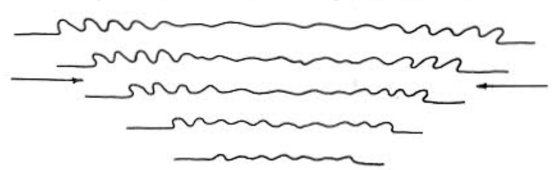
Croneis' (1930, pp. 165, 166) graphic description of the growth of the (autochthonous) Ouachita Mountains out of the Ouachita geosyncline by the closing of a vise upon it remains a valid picture. It is not likely that while one part of the range grew in this manner, another part, continuous in structure and stratigraphy with the first, was shoved in from a wholly separate region far to the south. (cf. van der Gracht, 1931, pp. 1019-1021).

Any "sole faulting" which may be present therefore probably had its roots within the present geosynclinal area. At most, it probably is limited to the marginal portions of that area; and it may not exist at all. For the Caddo Gap and DeQueen quadrangles in Arkansas, Miser and Purdue (1929, p. 124) wrote that "the faults were produced by the breaking and overthrusting of strata in closely compressed anticlines". This is probably true also of the major faults in Oklahoma other than those in the imbricate border belt; and may be true of some of them also - or even, in a large sense, of that belt as a whole. (cf. Misch and Oles, 1957, p. 1902).

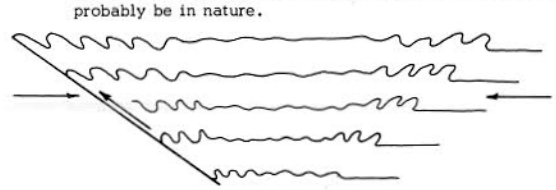
The autochthonous portrait of the Ouachitas, including their fan folding, is wholly consistent with Chamberlin's wedge theory of diastrophism. The presence of outward thrusting on the outer (convex) side of the curved portion of the range in Oklahoma, and its absence on the straighter north border of the Arkansas Ouachitas, likewise conform to his ideas (Chamberlin, 1925, esp. pp. 766-773, Figs. 7-11, same pp.). Chamberlin's sketches, in fact illustrate well the alternative concepts of Ouachita structure (Fig. 4).



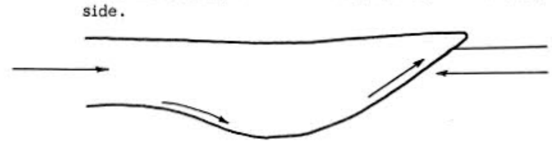
A. - Diagrammatical representation of a wedge bordered by shear planes which sharply delimit the block. These are planes of accommodation between the deformed wedge and the less deformed regions adjoining.



B. - Wedge shape by folding alone. As the folding dies out downward the folded zone becomes narrower. The folds are diagrammatic, making the case simpler than it would probably be in nature.



C. - A wedge bordered sharply on one side by shear planes while the folding dies out more gradually on the other side.



D. - Similar to Fig. C, but with more pronounced shearing. This grades into a one-sided orogeny.

Fig. 4. -- R. T. Chamberlin's sketches of wedge diastrophism (Courtesy Journal of Geology)

Contrasts with major low-angle thrust sheets elsewhere. - The great faults in the interior of the Ouachita Mountains (Boktukola and Octavia faults), and the inner ones of the border belt (Windingstair and Jackfork Mountain faults) are bordered on the upthrow side by thick bodies of Stanley shale, in most places thrust against synclinal or homoclinal masses of Jackfork sandstone. This is a perfectly logical relationship for the broken crest of a tight fold with incompetent shales exposed in its core, but it is an extraordinary thing for shales to constitute the moving prow of a great low-angle thrust sheet. Contrast this with the great section of Precambrian quartzites above the Lewis overthrust in Glacier National Park, or the massive Paleozoic limestones of Heart Mountain and Sheep Mountain in Wyoming, or the competent Paleozoics of Cumberland Mountain above the Pine Mountain thrust in Kentucky, Virginia, and Tennessee. These may ride on lubricant shales, but the thrust sheets include as their basal member no such thickness of shale as the 10,000-foot Stanley. The incompetent "lubricant" is more apt to be ridden over by the thrust sheet than to be carried along intact as its basal member.

Contrast also the relatively flat-lying position of those thrust masses with the sharp folds and steep dips of the Ouachitas. Mapping by L. M. Cline seems to indicate that the Octavia fault in places shows little or no stratigraphic displacement (unlike the Rocky Mountain thrusts above mentioned).

The great folds of the central Ouachita region, such as the Kiamichi anticline, afford no evidence that they lie over a sole thrust.

The small visible low-angle thrusts are superficial. - Neither of the two low-angle faults mentioned above, seen in Ouachita road cuts, has a measurable displacement as great as 50 feet. Both involve competent chert masses at the border of structurally high areas, thrust out over incompetent shales at levels now topographically above adjacent valleys. Neither demonstrates thrust movement in excess of an amount that might readily have been generated in the adjacent, narrow local uplift. Unlike the hypothetical sole faults at depth, these are superficial features. (cf. Misch and Oles, 1958, p. 2780).

Evidence from gravity surveys. - Gravity surveys (Woollard, 1956; Exploration Surveys, Inc., 1950) show a strong negative anomaly in the Ouachita region, which contrasts sharply with positive anomalies in the Arbuckle and Wichita Mountains, where basement rocks reach the surface. The negative anomaly in the Ouachitas may mean that the basement is deeply buried here. If so, it renders less plausible the idea that the exposed rocks of this region constitute an imported thrust plate, for they lie over a geosynclinal basin now: and this, more likely than any other, is apt to be their native basin. The deepest (central) part of the anomaly lies in that part of the mountain region where surface structure and stratigraphy suggest that the sedimentary fill should be thickest - in Pushmataha and southern LeFlore counties, Oklahoma.

Some of the boulder beds are no "friction carpet". - The argument that the exotic boulders in the Johns Valley shale may constitute a sort of "friction carpet" (Howell, 1947, p. 52; cf. van der Gracht, 1931, pp. 1021, 1051, and 1931a, pp. 709-714) along thrust planes has some plausibility for localities where the boulder-bearing shales lie adjacent to a fault trace. But it is almost inconceivable where they are far from such a trace, and constitute simply a central member of an unbroken stratigraphic section 15,000 feet thick in a continuous outcrop band 25 miles long, as in the Lynn Mountain syncline south of Clayton, Oklahoma (Cline, 1956c, p. 104; cf. Rea, 1947, p. 48). According to Cline the Johns Valley shale, with its included boulder beds, lies conformably between the Jackfork sandstone below and the Atoka formation above, with no sign of strike faulting at its borders. Undisturbed parallel bedding characterizes the shales in which the boulders are embedded, except for normal depositional variations within a foot or two of some of the exotic masses.

Doubts concerning the faults near Waldron. - Both the Jones Creek fault and the Johnson Creek fault in the Waldron quadrangle (Reinemund and Danilchik, 1957) are under a handicap like that of the Glover Creek fault (now believed to be non-existent) in McCurtain County, Oklahoma, in that their presence is based on inference rather than observation. They map almost the entire length of the Jones Creek fault as hidden by alluvium. A single sandstone member of the Atoka is shown to be displaced about 300 feet by it. Greater displacement elsewhere is possible, judging from regional relationships. The authors describe it as a tear fault and show it with steep dip (their section C-C'), but offer no evidence to establish the presence of a low-angle fault extending westward from it beneath the Choctaw anticline. They indicate little displacement on the Choctaw fault north of that fold. (cf. Misch and Oles, 1958, p. 2773). Most of its trace in this quadrangle, as mapped, is also concealed under alluvium.

The authors admit that there is little evidence of faulting along the supposed trace of the low-angle Johnson Creek fault, which is shown as following the base of the Atoka around the east end of the Black Fork syncline. One of their reasons for sketching a fault here is that their mapping indicates that in another locality there exist several thousand feet of Atoka strata older than the lowest Atoka beds which here lie on Johns Valley shale. The other reason they state is that structures in older rocks east of that syncline do not seem to extend into it. Situations similar to both of these could have been brought about by unconformity; and the second feature might even be ascribed, at least in part, to easier yielding to stress in the Johns Valley shale than in the more competent Atoka.

The age of the supposedly older Atoka strata at the north side of the Black Fork syncline farther west should be carefully checked in view of the difficulty of distinguishing one part of the Atoka from another, the possibility of repetition of strata by unrecognized folding or strike faulting, and the extremely rapid changes in thickness which are known to characterize the Atoka even in unfaulted areas.

Misch and Oles (1958, p. 2782), while accepting Reinemund and Danilchik's mapping of the Johnson Creek fault, would explain it by "disharmonic separation" - sliding of younger beds over older without involving an imported thrust plate in the usual sense.

The authors of the Waldron map express no doubt about the presence of great thrust sheets there. They speak of them as though their existence was taken for granted rather than as a hypothetical interpretation requiring careful proof. To a geologist more skeptical on that matter, in the absence of convincing local evidence of faulting in either place, it would seem just as permissible to sketch a steep thrust fault with its trace parallel to bedding farther west in the north limb of the Black Fork syncline, as to draw a horizontal fault around the east end of that fold. The former interpretation would remove the need for considering the Atoka sediments north of it to be older than those to the south; and there is far more of proven precedent in the region for steep strike faults, than for the flat fault.

The clash between Arbuckle and Ouachita trends. - That some northwestward thrusting took place in the Ouachita frontal region near Atoka is acknowledged. How great the horizontal component of this thrusting was, is open to debate. The Arbuckle facies of rocks here may have extended farther east before that thrusting, than it did to north or south of the Tishomingo horst. This would tend to reduce one's estimate of the amount of thrusting needed to bring about the present situation. Misch and Oles (1958, pp. 2781-2782) recently have suggested that "strike-slip faulting along the Arbuckle trend could have produced a similar pattern".

Origin of the Exotic Boulders

Mode of emplacement. - Among the processes by which it has been suggested that the "Caney boulders" could have been scattered into their present positions are the following: (1) Transportation by ice floes (Ulrich, 1927, pp. 34-36, 44; Powers, 1928, p. 1046; Rea, 1947, p. 49) or (2) icebergs (Miser, 1929, p. 30). (3) Direct deposition by glaciers, as till, et cetera. (4) Submarine landslides or (5) mudflows (Kramer, 1933, pp. 590, 614-615) or (6) turbidity currents (Cline, oral comment to Tomlinson). (7) Thrust faulting (implied by Howell, 1947, p. 51). (8) Of the great exotic slabs in Johns Valley itself, some have thought they might be tops of buried hills, in place (M. P. White, oral comment to Tomlinson; cf. Kramer, 1933, p. 615).

Several of the authors cited above have discussed more than one of these possibilities at considerable length, pro and con. The most plausible cases seem to have been made for method (1) and for some combination of (4), (5), and (6), which grade into each other. It is not clear that either of these two choices can be wholly ruled out on the basis of evidence so far described. There is more cogent adverse evidence against (2), (3), (7), and (8).

Source areas. - Drifting of shore ice might permit distribution to a greater distance, and in more varied directions, than slippage down a depositional slope. Source areas for the boulders have been suggested as lying to the west (Ulrich, 1927, pp. 9-10), south (Powers, 1928, pp. 1044-1045), north and northwest (Kramer, 1933, pp. 611-614), and from a combination of these directions (Rea, 1947, p. 49). Curiously, no present uplift now exposing strata represented by the boulders seems to fill the bill for all of them. Certain formations (e. g., Sycamore limestone) which now crop out only in the Arbuckle region are represented, but also some (e.g., St. Clair limestone and coarse sandstone of St. Peter type) which are more widespread in the Ozark region, and occur in the Arbuckle Mountains only near the eastern border of the exposed portion of that uplift (Ham, 1957). Others, such as Cotter dolomite and Boone chert (Rea, 1947, p. 49), occur only in the Ozarks, and have been recognized only in the easternmost (Arkansas) exposures of "Caney" boulder beds. It seems almost certain, therefore, that the boulders came either from more than one source area bordering (or within) the Ouachita geosyncline, or from one so extensive that outcrops of rocks of both Arbuckle and Ozark types occurred within it.

It is difficult (Powers, 1928, pp. 1044-1045; Miser, 1929, p. 30) to consider the present exposed area of the Arbuckle Mountains as a source for boulders embedded in typical fossiliferous Mississippian Caney shale, like those at several excellent exposures east of the Kiamichi River (Cline, 1956c, p. 103; cf. Girty, 1927), because that range almost certainly was not yet in existence when that shale was deposited. This shale is exposed along most of the borders of the Arbuckle Mountains, but no exotic boulders have been found in it there. The same negative evicence tends to exclude as a possible source the Criner Hills and Muenster Arch uplifts which parallel the Arbuckles on the southwest; although Caney outcrops there are limited to the east border of the Criner Hills.

This argument applies less effectively to the exotics in the upper part of the Johns Valley shale, for that part (Harlton's Round Prairie formation as redefined by him in 1959) is believed to be of early Pennsylvanian age; and the Criner Hills shed boulder conglomerates in Morrowan time. Local conglomerates or scattered pebbles occur in the Wapanucka limestone on the northeast flank of the Arbuckles (Morgan, 1924, p. 59), and a part of that region probably was exposed to erosion in Atoka time. A thin breccia and other evidence of unconformity have been described (Wallis, 1915) in the Wapanucka near the town of that name. Both boulders and pebble conglomerate occur in shales which have been correlated with the Wapanucka near the northern margin of the Oklahoma Ouachitas (Powers, 1928, pp. 1042-1044), and pebbles are found in the limestone itself along their western margin.

If all of the Johns Valley shale ("Ouachita Caney" of Ulrich; Round Prairie formation of Harlton, 1934; Johns Valley shale and overlying Round Prairie formation of Harlton, 1959) were of Morrowan age, and the extensive masses of Mississippian Caney shale included in the lower part of this formation were all exotic like the limestone boulders, as was long contended (cf. Ulrich, 1927, pp. 6, 20-22; White, 1934; Harlton, 1934, pp. 1020, 1022, 1039, 1042, Figs. 1, 2; Harlton, 1938, pp. 854, 889-900, Fig 1; Miser, 1934, pp. 971, 974, 999-1000; Hendricks, 1940, p. 2146; Hendricks and others, 1947) a major source of the exotic in the present area of the Arbuckle Mountains would be only a little less plausible: for it would still be difficult to understand why none of them came to rest in the strata of Johns Valley age (e.g., Wapanucka limestone) which outcrop on the northeast flank of the Arbuckles, in an almost continuous band between that area and the Ouachita Mountains.

One point of some weight favors Powers' (1928, pp. 1044-1046) concept that an uplift may have existed in Caney time in an area south of the western portion of the present Ouachita Mountains, in trend with the Arbuckle Mountains but southeast of the point where that range disappears under Cretaceous sediments of the Gulf coastal plain. That is the fact that by far the largest exotic masses known in the Caney (lower Johns Valley) shale of the Ouachitas lie in Johns Valley itself, within 20 miles of the Cretaceous border. This area is also near parts of Kramer's hypothetical "Bengalia" (Kramer, 1933, Fig. 2, p. 602) buried hills, supposed by him to have been in Caney time a chain of islands, now hidden by thrust sheets, within the north and west border of the present Oklahoma Ouachitas; but these would be less likely to have furnished a source for the boulders of Sycamore limestone which are found in Johns Valley. No typical Sycamore lithology is found in the northeastern margin of the Arbuckle Mountains, though it is well developed along their southern margin and the western part of their north border. A different facies (Ahloso or Mayes) appears to the northeast, followed by the Boone formation of the Ozarks.

Somewhere within the area where the Sycamore facies was deposited, outcrops of it must have been exposed to erosion shortly thereafter, to provide the boulders in question: unless they are actually not exotics, but summits of buried hills. The latter hypothesis was given a color of respectability by an apparent geographical distribution of exotic masses in Johns Valley in stratigraphic order--older rocks toward the margins of the synclinal bowl of shale, younger ones farther in. But the exotics are too widely scattered and their arrangement too irregular to permit assurance that this is more than coincidence. Besides, it seems probable that the same general processes which laid down the exotics in the Johns Valley shale elsewhere, account also for those in Johns Valley; and no other area of them exhibits such an approach to systematic arrangement. And it is reported that a hole was drilled through the largest exotic mass (Viola) in Johns Valley, at the instance of Sidney Powers, finding it only a few feet thick and passing into rocks of Ouachita facies below.

The "Arbuckle orogeny", the major mountain-building for the western and southern parts of the range at least, did not begin to yield known boulder conglomerates before Desmoinesian time, and was not over till well into Virgilian time. But the possibility cannot be ruled out that slight uplift may have occurred in or just prior to Caney time in interior areas of the range, or in an area to the southeast of it (now under Cretaceous cover), sufficient to permit erosion of the Sycamore limestone or even of older strata.

Recent studies of cross-bedding in the Atoka and of flow casts in the Jackfork (Cline, 1958) indicate that the dominant direction of flow in parts of the Ouachita trough during parts of Jackfork and Atoka time was from the east. No such data have yet been reported from the Johns Valley shale, between those two formations. Curiously, east and southeast are the only directions from which no author has seriously suggested that the exotic boulders are apt to have come; though movement from the Ozarks might well have been southwestward as well as southward.

At present, it appears that both the location of the source area or areas from which the boulders came, and the manner of their transportation, are still in doubt. It seems likely, however, that a closer approach to solution of both questions may be achieved by more thorough analysis of the character of the boulders and their relation to the enclosing shale.

Problems of Age of Strata

Crystal Mountain sandstone and older beds. - The oldest identifiable faunas that have been reported from the Ouachita region of Oklahoma are from the upper part of the Womble shale. These are of approximately the same age as the Bromide formation of the Arbuckle Mountains. From Arkansas, the oldest fauna reported is lower in the section, from the lower part of the Mazarn shale (Miser and Purdue, 1929, pp. 27-28). The latter consists wholly of graptolites which have been classified as of Beekmantown (early Ordovician) age (E. O. Ulrich, cited by Miser and Purdue, 1929, p. 28; Elias, 1956). Collections made in 1953-1956 by Wm. D. Pitt and M. K. Elias from Miser and Purdue's locality near Norman (formerly called Womble), Arkansas, have included only poorly preserved specimens.

Chiefly on the strength of that faunal assignment, it has been supposed (Purdue, 1909, p. 30; Honess, 1923, Pl. 1; Miser and Purdue, 1929, p. 24) that the Collier shale might be of Cambrian age. Questionable support is given to this inference by the local presence of a conglomerate or breccia at the top of the Collier (Purdue, 1909, pp. 30-32) or base of the Crystal Mountain (Miser and Purdue, 1929, pp. 24-25; Pitt, 1955, pp. 19-21), but this may be intraformational. As seen in Oklahoma, most of the fragments in it are angular; and in both Arkansas and Oklahoma they consist of but two types of rock: chert and limestone of the same character as the subjacent limestone member of the Collier shale. Patches of shale-pebble conglomerate in Arkansas (pitt, Tomlinson, Cline, personal observations, 1953-1957, not yet published) higher in the Crystal Mountain are little more convincing as evidence of emergence, or of a systemic boundary.

If the lower Mazarn is indeed of basal Ordovician age, then the Womble shale, Blakely sandstone, and Mazarn shale represent not only all of the Simpson group of the Arbuckle Mountains, but much of the Arbuckle group as well; and even the basal (upper Cambrian) part of the Arbuckle group may be represented in the Collier and Lukfata formations. Yet a suspicion persists among some geologists that the Crystal Mountain sandstone may represent one of the sandstones in the Simpson group. It is of the same "St. Peter type", with well-rounded, frosted sand grains. Of course, similar sands could have developed in Cambrian time in the Ouachita region, even though not elsewhere.[4]

If the Crystal Mountain should prove to be Chazyan or younger, then the Ouachita correlatives of the Arbuckle group and even those of the lower Simpson may not be exposed anywhere. There is no evidence that the Timbered Hills group (upper Cambrian) is represented in any outcrops in the Ouachita Mountains. A thickened Cambrian section may well exist here, perhaps of geosynclinal facies, and possibly including Middle and Lower Cambrian strata as well as Upper. That could go a long way toward explaining the deep negative gravity anomaly in the Ouachita region.

Stanley shale and Jackfork sandstone. The age of these two thick formations has been a subject of debate ever since Taff first named them in 1902. Because masses of Ordovician rock (not then recognized as a separate system from the Silurian) were found above them, in what has more recently been called Johns Valley shale, Taff listed these formations as of Silurian age. The pre-Mississippian masses soon were recognized as exotics. In 1904 Ulrich (cited by White, 1937, p. 45) discovered fragments of Carboniferous plants in the upper Stanley or lower Jackfork. Purdue (1909, p. 40), called the Stanley Carboniferous, and White (cited by Girty, 1909, p. 8) tentatively assigned its meager flora to the upper Mississippian or lower Pennsylvanian. Honess, in 1923 (pp. 174-178) cited meager faunal and floral evidence, interpreted by Ulrich and White respectively, for calling the Stanley Mississippian or early Pennsylvanian, and classified the Jackfork as Pennsylvanian (ibid., p. 202).

Honess (1924, pp. 14-18) also found a Morrowan fauna near the middle of the section to which the name Jackfork had been applied. Miser (1926, 1954), with approval of Honess (Miser and Honess, 1927, p. 23) applied the name "Atoka" to the sediments above that fauna in the area (Boktukola syncline) where the fauna was found; but did not separate those younger strata from the Jackfork elsewhere on the same maps. Cline (1956, 1958) and Shelburne (1959) have now separated the two in much of southeastern Oklahoma, but the task is not yet completed. It is possible that in Arkansas also some Morrow (or even Atoka) sediments have been mapped as Jackfork (e.g., southeast of Hollis, Arkansas: Griley, 1956; cf. the probable Morrow fauna studied by Girty from the "Jackfork" near Pulaski, Arkansas: Miser, 1934, pp. 989-991.

Girty (1909) classified as Mississippian the fauna of the Caney shale overlying the Jackfork; but Ulrich (1927, pp. 36-48; cf. Miser and Purdue, 1929, pp. 74-75) contended that the Mississippian (Caney) fossils found above the Jackfork in Johns Valley and elsewhere were exotic, like the boulders of older limestones which were found near those fossils: that they were originally deposited elsewhere and later transported by floating ice to their present locations. That interpretation permitted him to class the Jackfork as Pennsylvanian, as White (1934, 1937) and Schuchert (cited by Miser and Purdue, 1929, pp. 68-69) were inclined to do. In 1937 White described a number of new floral collections by Miser and others from the Jackfork and upper Stanley, and referred "not only the Jackfork to the Pennsylvanian, but also the upper half and, by implication, the non-plant-bearing lower half of the Stanley as well". (White, 1937, p. 45).

Miser for a time opposed that concept (Miser, 1926; Ulrich, 1927, pp. 36-47; Miser and Purdue, 1929, pp. 74-75), but yielded to it later (Miser, 1934, pp. 999-1000; Miser, 1954). Girty (cited by Miser and Honess, 1927, pp. 23-24; and by Miser and Purdue,

[4] Editor's note: See Ham, Fig. 2, this symposium.

1929, pp. 72-74), however, apparently continued steadfast in his belief that the invertebrate faunas indicated Mississippian age for both Stanley and Jackfork, and that the overlying Mississippian Caney faunas were in place.

Harlton (1934) considered both Stanley and Jackfork "post-Mississippian" (ibid., pp. 1027-1028); but suggested placing them, with the overlying Johns Valley shale, in a separate system such as Schuchert (1924) had proposed, to be called the Bendian. Cooper (1945, pp. 396-397) called them "post-Chester-pre-Pottsville", but left the question open as to whether they should be classed as Mississippian or as Pennsylvanian, or part one and part the other.

Prior to 1950, the Caney shale had been found above the Jackfork (excluding Atoka beds mapped as Jackfork) or associated with boulder beds deemed to be younger than Jackfork, in many places in the northern and western parts of the Ouachita Mountains in Oklahoma (cf. Ulrich's list of localities, Ulrich, 1927, pp. 10-12), but nowhere southeast of the Kiamichi River. It so happened that in all of those places the known extent of Mississippian Caney shale was rather limited. They were either near the center of synclinal basins as in Johns Valley and Prairie Mountain, or adjacent to faults, as in the shale bank at the hairpin turn on Oklahoma Highway 2 north of Clayton (SE 1/4 sec. 3, T. 3 N., R. 15 E). The fossils themselves all appear to be in their original matrix, showing little sign of wear; and many of them are too fragile to have been redeposited undamaged apart from the enclosing shale. The idea therefore was presented that these bodies of Caney shale, rather than only their contained fossils, were themselves great exotic boulders or tectonic slivers (Miser, 1934, pp. 999-1000; Harlton, 1934, p. 1028; Cooper, 1945, p. 390).

But the larger of them are a mile or more in extent (e.g., in Johns Valley, - cf. Harlton, 1959). Some of them are relatively undeformed over many acres. It was difficult to conceive how any of the various methods of transportation (see preceding pages) conceived for the limestone boulders could have moved such masses of shale intact for any such distance (many miles) as would have to have been traversed to bring them from any known or imaginable locality of more extensive Caney shale outcrops.

Hass (1950) described conodont faunas from the lower Stanley which he construed as demonstrating Mississippian age.

In the last few years Cline (1956) has discovered a series of exposures in the Lynn Mountain syncline, south of Clayton and east of the Kiamichi River, which he believes clearly prove that Miser was right in his early opposition to the exotic concept for these shale bodies or their contained Caney fossils. Here is a group of outcrops of Johns Valley shale scattered miles apart along a single continuous outcrop band 25 miles long between the same parallel formations—Jackfork sandstone (Harlton's Game Refuge formation, uppermost Jackfork - Harlton, 1959, p.134) on one side and Atoka strata (or Harlton's Round Prairie formation, treated by Cline as basal Atoka) on the other. In all of the shale outcrops the bedding of the shale is essentially parallel to that of the adjacent formations above and below. And exotic boulders of pre-Stanley limestones, sandstones and cherts are imbedded in the fossiliferous Mississippian shale. It is obviously a shale which was deposited as mud on the sandstones which underlie it: not a group of exotic masses of shale brought into their present stratigraphic relationships as intact m masses of shale which was first deposited elsewhere at an earlier date and then moved bodily like the limestone boulders. Thus Cline has established that the Jackfork, as well as the Stanley, is older than the Caney shale, and of Mississippian age. He has now shown these outcrops to more than a score of geologists, and all of them seem to agree with this conclusion. This problem, therefore, now appears to be near the same solution which was stated by Girty (1927) thirty years ago; - that the "Caney" shale of the Ouachita region (later called Johns Valley shale - Johns Valley shale and Round Prairie formation of Harlton, 1959) is Mississippian in its lower part and Pennsylvanian in its upper part; and that the Stanley and Jackfork are both of Mississippian age. If Cline is right, Taff and Girty were also right (1902, 1909) in treating the Caney shale of the Arbuckle region as one and the same formation as the Caney shale of the Ouachitas. Taff's type locality for the Caney in Johns Valley may still be valid, provided the Pennsylvanian upper portion of what was then called Caney be excluded to conform to modern usage of that name as describing Mississippian strata only.

Ulrich had earlier (Ulrich, 1927, pp. 40, 42) noted Mississippian Caney fossils in the shale matrix (lower Johns Valley shale) of the boulder-bearing bed in Johns Valley (T. 1 S., R. 16 E.), but considered them to be imported "remanie" in spite of the unanimous opinion of other observers (including Miser) that they were indigenous to the containing shale. This is an isolated locality in the heart of a synclinal bowl; with structural and stratigraphic relationships, especially to younger strata, less convincing than in the new outcrop band discovered by Cline.

One rather weak support heretofore for the concept that the Jackfork might be younger than the bodies of Mississippian shale which had been found overlying it, but until recently only in relatively small scattered masses, lay in the fact that the gross stratigraphic position of the Stanley-Jackfork section in the Ouachitas corresponds to that of the Goddard-Springer section in the Arbuckles. Both lay above Arkansas novaculite (late Devonian and/or early Mississippian) or its approximate correlative, the Woodford chert, and beneath Morrowan strata. Both included a lower body of shale with minor sandstone, and an upper unit containing more sandstone. Both were thousands of feet in thickness. These facts made it difficult to accept the idea, now pretty well proven, that the two sections were laid down in different epochs of time, and that neither is represented by any great thickness of strata in the area where the other is best developed.

But is that so surprising? Perhaps the rapid development of the Ouachita geosyncline in Stanley-Jackfork time, in itself made unnecessary great contemporaneous geosynclinal development in the nearby Arbuckle retion. And the revival of the Arbuckle geosyncline as the Ardmore-Anadarko basin in Goddard and Springer time may reciprocally have given cause for a slackening of the pace of similar processes in the Ouachita area. Perhaps one geosyncline at a time was enough for Oklahoma.

In recent years data have accumulated to show that the amount of sandstone in the Springer diminishes eastward, instead of increasing to the east as would be expected if it were approaching the composition of the Jackfork. And the presence of only a thin section in the Ouachita region to represent the 4,000-foot Goddard-Springer section in the Ardmore Basin is no more surprising than its equally thin representation along the northeast flank of the Arbuckle Mountains. Indeed, it simplifies isopach maps to have this section thin in the two adjacent areas (the Oklahoma Coal Basin and the Ouachitas); and avoids the awkward portrayal of contemporaneous thick sedimentation in two troughs (Ardmore-Anadarko and Ouachita geosynclines), one of which terminated against the other at right angles.

The similarity in electric logs between the Bromide-to-Caney section of the Ardmore Basin and the McAlester coal basin, and the Womble-to-basal-Stanley section of the western Ouachitas, still tempts a student of those logs to see in the basal Stanley, rather than above the Jackfork, a correlative of the Caney shale. But although the electric-log similarities are substantiated by fossil evidence as indicating true correlations for the <u>pre</u>-Stanley strata, no Caney-type faunas have been found in the lower Stanley. The similarity of its electric log pattern to that of the Caney farther west may signify no more than the fact that both are shales. One would not be mistaken for the other in outcrop; whereas the Caney shale in the Johns Valley formation is indistinguishable from the Caney of the Arbuckle Mountains, both lithologically and faunally.

<u>Morrowan strata in the Ouachitas</u>. - The Wapanucka limestone itself is mapped in the northwestern border of the Ouachita Mountains. Eastward along the north border it thins and appears to interdigitate with shale and sandstone into which it finally disappears north of Talihina, Oklahoma, where it appears to be represented in the <u>upper</u> part of the Johns Valley shale. (cf. Miser, cited by Ulrich, 1927, pp. 36-27; Miser, 1934, pp. 1000, 1001). In the interior portions of the Oklahoma Ouachitas west of the Kiamichi River, Morrowan fossils (including goniatites) have been collected in the upper part of the Johns Valley shale (Girty, 1927; Ulrich, 1927, pp. 19-21; Harlton, 1933, pp. 4-5; Harlton, 1959, p.137) - the part to which Harlton (1959, p.137) now limits his Round Prairie formation.

Honess (1924, pp. 14-20) found a good Morrowan molluscan fauna in a thin sandstone farther east, in the Boktukola syncline. Elias, Cline and others have collected from this member at and near Honess' locality, and agree with this age assignment (personal communications to Tomlinson; Cline, 1956, pp. 100, 105-106). What appears to be the same zone is represented by abundant crinoid fragments, with more meager molluscan material, on the Indian Road on Kiamichi Mountain at the southwest corner of Township 2 North, Range 22 East. This is classified by Cline and Moretti (1956, p. 4, item 107) as basal Atoka. It may be a part of the Round Prairie formation (restricted).

Drilling in the Arkansas Valley synclinorium in western Arkansas has established the fact that the Morrowan strata of the Ozarks thicken and become more dominantly clastic as they pass southward under that valley. A similar southward change is noted in Oklahoma (Laudon, 1958, Figs. 12-14). In the deepest part of the synclinorium, no drilling has yet penetrated below the Atoka. This thickening toward the Ouachitas, and the accompanying increase in sand and shale of Morrowan age, justifies a suspicion that a much greater thickness of sediments in the Ouachita region may belong to this series than the relatively thin zones in which Morrowan faunas have so far been found. The lower part of the thick overlying section of sands and shales ascribed to the Atoka formation in the inner Ouachitas of Oklahoma may be of Morrowan age, though indistinguishable lithologically from the true equivalent of the type Atoka. And it appears that fossiliferous Morrowan strata have been mapped as Jackfork in central Arkansas (Miser, 1934, pp. 989-991, quoting Girty).

In many parts of the Ouachitas, existing published maps (Miser, 1926, 1954; Branner, 1929) include with the Jackfork both Morrowan and Atokan strata. Current studies are gradually segregating Atoka from Jackfork in these areas. The upper limit of the Morrowan strata which probably lie between true Jackfork and true Atoka is difficult to define. Their lower limit is also uncertain in much of the region, where the Johns Valley (or Caney) shale has not yet been identified.

References

Arbenz, Kaspar, 1955, Oral communication to C. W. Tomlinson.

Bennison, Allan P., 1957, Letter to C. W. Tomlinson dated January 6, 1957.

Branner, G. C., and others, 1929, Geologic map of Arkansas: Arkansas Geol. Survey.

Chamberlin, R. T., 1925, The wedge theory of diastrophism: Jour. Geology, vol. 33, pp. 755-792.

Cline, L. M., 1956a, Regional stratigraphy and history of Ouachita Mountain area (abs.), Amer. Assoc. Petroleum Geologists Bull., vol. 40, no. 12, pp. 427, 428.

Cline, L. M., 1956b, Mississippian-Pennsylvanian stratigraphy of central Ouachita Mountains, Oklahoma: Ardmore Geol. Soc., Guidebook to Ouachita Mountains field conference, southeastern Oklahoma, pp. 47-63.

Cline, L. M., 1956c, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains, Oklahoma: Tulsa Geol. Soc. Digest, vol. 24, pp. 100-106.

Cline, L. M., 1958, Oral communication to C. W. Tomlinson.

Cline, L. M., and Moretti, Frank, 1956, Two measured sections of Jackfork group in southeastern Oklahoma: Oklahoma Geol. Survey, Circular 41, pp. 1-20.

Cooper, C. L., 1945, Age relations of Stanley and Jackfork formations of Oklahoma and Arkansas: Jour. Geology, vol. 53, pp. 390-397.

Croneis, Carey, 1930, Geology of the Arkansas Paleozoic area: Arkansas Geol. Survey Bull., vol. 3.

Dake, C. L., 1921, The problem of the St. Peter sandstone: Missouri School Mines and Metall. Bull., vol. 6, no. 1, p. 55.

Elias, M. K., 1956, Oral communication to C. W. Tomlinson.

Exploration Surveys Inc., Dallas, Texas, 1950, Regional gravity map of Texas and Oklahoma (privately distributed)

Girty, G. H., 1909, Fauna of the Caney shale of Oklahoma: United States Geol. Survey Bull., 377.

Girty, G. H., 1927, Age of Caney, Jackfork, and Stanley formations: Oklahoma Geol. Survey Bull., vol. 44 (Miser and Honess), pp. 23, 24.

Goldstein, August, Jr., and Hendricks, T. A., 1953, Siliceous sediments of Ouachita facies in Oklahoma: Geol. Soc. America Bull., vol. 64, pp. 421-442.

Griley, H. L., 1956, Letters to C. W. Tomlinson dated Sept. 11, Sept. 24, and Oct. 16, 1956.

Ham, William E., 1957, Oral communication to C. W. Tomlinson.

Harlton, B. H., 1933, Review of Kramer's "Boulders of Bengalia": Amer. Assoc. Petroleum Geologists Bull., vol. 28, pp. 1538, 1539.

Harlton, B. H., 1934, Carboniferous stratigraphy of the Ouachitas with special study of the Bendian: Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 1018-1049.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 22, pp. 852-914.

Harlton, B. H., 1947, The siliceous shale members of the Stanley-Jackfork deposits in the Ouachita Mountains, Oklahoma, with notes on the "Johns Valley" shale: Tulsa Geol. Soc., Guidebook to field conference in western part of Ouachita Mountains in Oklahoma, pp. 41-46.

Harlton, B. H., 1953, Ouachita chert facies, southeastern Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 37, pp. 778-796.

Harlton, B. H., 1959, Age classification of the upper Pushmataha series in the Ouachita Mountains: In press, this symposium.

Hass, W. H., 1950, Age of lower part of Stanley shale: Amer. Assoc. Petroleum Geologists Bull., vol. 34, pp. 1578-1584.

Hendricks, T. A., 1940, Structural interpretation of recent gravity observations in southeastern Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 24, pp. 2143-2149.

Hendricks, T. A., 1958, Discussion of Misch and Oles' "Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: preliminary report": Amer. Assoc. Petroleum Geologists Bull., vol. 42, pp. 2757-2765.

Hendricks, T. A., and others, 1937, Geology of Black Knob Ridge, Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 21, pp. 1-26.

Hendricks, T. A., and others, 1947, Geology of the western part of the Ouachita Mountains of Oklahoma: United States Geol. Survey Prelim. Map 66, three sheets including text.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey Bull., vol. 32.

Honess, C. W., 1924, Geology of southern LeFlore and northwestern McCurtain counties, Oklahoma: Oklahoma Geol. Survey, Circular 3.

Honess, C. W., 1930, Oil and gas in Atoka, Pushmataha, McCurtain, Bryan, and Choctaw counties, Oklahoma: Oklahoma Geol. Survey, Bull. 40, vol. 3, pp. 83-108.

Howell, J. V., 1947, The origin of imbricate structure (with accompanying glossary): Tulsa Geol. Soc., Guidebook to field conference in western part of Ouachita Mountains in Oklahoma, pp. 50-53.

Kramer, Wm. B., 1923, Boulders from Bengalia: Jour. Geology, vol. 41, pp. 590-621.

Laudon, Richard B., 1958, Chesterian and Morrowan Rocks in the McAlester Basin of Oklahoma: Oklahoma Geol. Survey, Circular 46.

Miller, B. W., 1956a, The geology of the Potato Hills, Pushmataha and Latimer counties, Oklahoma (abs.): Tulsa Geol. Soc. Digest, vol. 24, p. 66.

Miller, B. W., 1956b, Oral presentation of same subject to Tulsa Geological Society, March 19, 1956.

Misch, Peter and Oles, K. F., 1957, Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: preliminary report: Amer. Assoc. Petroleum Geologists Bull., vol. 41, pp. 1899-1905.

Misch, Peter, and Oles, K. F., 1958, Reply to T. A. Hendricks' "Discussion of Misch and Oles' 'Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: preliminary report' ": Amer. Assoc. Petroleum Geologists Bull., vol. 42, pp. 2765-2783.

Miser, H. D., 1921, Llanoria: Amer. Jour. Sci., vol. 2, pp. 61-89.

Miser, H. D., 1926, Geologic map of Oklahoma: United States Geol. Survey.

Miser, H. D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geol. Survey Bull., vol. 50.

Miser, H. D., 1934, Carboniferous rocks of Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 972-981.

Miser, H. D., 1954, Geologic map of Oklahoma: United States Geol. Survey and Oklahoma Geol. Survey.

Miser, H. D., and Honess, C. W., 1927, Age relations of the Carboniferous rocks of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geol. Survey Bull., vol. 44.

Miser, H. D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles, Arkansas: United States Geol. Survey Bull., vol. 808.

Morgan, Geo. D., 1924, Geology of the Stonewall quadrangle, Oklahoma: Oklahoma Geol. Survey Bull., vol. 2.

Pitt, William D., 1955, Geology of the core of the Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Circular 34.

Powers, Sidney, 1928, Age of the folding of the Oklahoma Mountains: Geol. Soc. America Bull., vol. 39, pp. 1031-1072.

Purdue, A. H., 1909, The slates of Arkansas: Arkansas Geol. Survey.

Purdue, A. H., and Miser, H. D., 1923, Hot Springs Folio, Arkansas: United States Geol. Survey, Geol. Atlas, Folio 215.

Rea, H. C., 1947, The erratic boulders in the Johns Valley shale of the western Ouachita Mountains, Oklahoma: Tulsa Geol. Soc., Guidebook to field conference in western part of Ouachita Mountains in Oklahoma, pp. 47-49.

Reinemund, J. A., and Danilchik, Walter, 1957, Preliminary geologic map of the Waldron quadrangle and adjacent areas, Scott County, Arkansas: United States Geol. Survey (and Arkansas Geol. Survey), Oil and Gas Inv. Map OM 192.

Taff, J. A., 1902, Description of the Atoka quadrangle: United States Geol. Survey, Geol. Atlas, Folio 79.

Tomlinson, C. W., 1952, Odd Geologic Structures of Southern Oklahoma: Bull. Amer. Assoc. Petroleum Geologists, vol. 36, pp. 1820-1840.

Tomlinson, C. W., 1956, unpublished maps.

Tomlinson, C. W., and Pitt, William D., 1955, Recent studies in the Ouachita Mountains: Tulsa Geol. Soc. Digest, vol. 23, pp. 89-97.

Tomlinson, C. W., and White, Maynard, 1935, Bucher's laws applied to Arbuckle and Ouachita Mountains (abs.): Geol. Soc. America, Proc. for 1934, p. 118.

Ulrich, E. O., 1927, Fossiliferous boulders in the Ouachita "Caney" shale and the age of the shale containing them: Oklahoma Geol. Survey Bull., vol. 45.

van der Gracht, W. A. J. M. van Waterschoot, 1931, Permo-Carboniferous orogeny in south central United States: Amer. Assoc. Petroleum Geologists Bull., vol. 15, pp. 991-1057.

van der Gracht, W. A. J. M. van Waterschoot, 1931b, The pre-Carboniferous exotic boulders in the so-called "Caney shale" in the northwestern front of the Ouachita Mountains of Oklahoma: Jour. Geology, vol. 39, pp. 697-714.

Wallis, B. F., 1915, The geology and economic value of the Wapanucka limestone of Oklahoma: Oklahoma Geol. Survey Bull., vol. 23.

White, David, 1934, Age of Jackfork and Stanley formations of Ouachita geosyncline, Arkansas and Oklahoma, as indicated by plants: Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 1010-1017.

White, David, 1937, Fossil plants from the Stanley shale and Jackfork sandstone in southeastern Oklahoma and western Arkansas: United States Geol. Survey Prof. Paper 186, pp. 43-68.

Woollard, George P., 1956, Bouguer gravity anomaly map of the United States: Unpublished.

THE OUACHITA STRUCTURAL BELT [1]

Peter T. Flawn [2]

Introduction

The Ouachita system is a belt of Paleozoic rocks and structures which borders the southern part of the North American continent much in the same way that the Appalachian system margins the eastern edge of the continent. This prism of deformed rocks has been called variously the Ouachita structural belt, the Ouachita foldbelt, the Ouachita-Marathon foldbelt, and the Llanoria structural belt. The folded and faulted rocks extend for a distance of more than 1,300 miles from a point in southwestern Alabama through a sinuous course into northern Mexico; furthermore, it seems certain that the belt extends farther eastward, possibly into northern Florida, and farther southward into Coahuila and Chihuahua, Mexico, where it is lost in a confusion of Laramide structures of Late Cretaceous--Early Tertiary age (Plate I). In all of its more than 1,300-mile course, the Ouachita structural belt is exposed for a strike length of less than 300 miles. The major exposure for which the belt is named is in the Ouachita Mountains of western Arkansas and eastern Oklahoma, where east-west-trending folded and thrust-faulted Paleozoic rocks of Ouachita facies, in some areas showing varying degrees of very weak to low-grade metamorphism, are exposed in an area about 200 miles long and 50 miles wide. In Texas, the Ouachita belt is buried for nearly all of its 900-mile length; it comes to view only in far west Texas where a structural salient of folded and thrust-faulted rocks of Ouachita facies is exposed in the Marathon uplift and again briefly in the Solitario uplift. In central Texas, steeply dipping rocks of the front of the belt are exposed in the Turkey Bend area of Lake Travis (Barnes, 1948). In north-central and northeastern Mexico there are widely separated exposures of Pennsylvanian (?) and Permian sediments of geosynclinal (Ouachita?) facies in the Sierra del Cuervo, Chihuahua (De Cserna and Diaz G., 1956), Las Delicias--Acatita area of Coahuila (Kelly, 1936; R. E. King, 1944), and the Cd. Victoria area of Tamaulipas (Heim, 1940; Humphrey and Diaz G., 1953; Bodenlos and others, 1956); it is possible that these rocks are part of the Ouachita belt but correlations cannot yet be made with confidence.

Since 1954, the Ouachita structural belt has been the subject of a research project of the Bureau of Economic Geology of The University of Texas. The study involves investigations of the stratigraphy and petrography of rocks encountered in wells penetrating the belt and the immediately adjacent foreland. This paper presents some of the results of this project, but the study is not yet complete and the structural interpretations herein are preliminary and subject to modification as more well data become available. Needless to say, the vast amount of descriptive data on which these conclusions are based cannot be included in a symposium volume. The writer has presented a summary of the structure of the Ouachita belt in Texas in an earlier paper (Flawn, 1958). Short papers reporting on the results of similar studies have recently been published (Masson, 1955; Woods, 1956). The writer gratefully acknowledges the helpful criticism of August Goldstein, Jr., Philip B. King, Virgil E. Barnes, and John T. Lonsdale.

Many articles on problems of the Ouachita structural belt have been written during the last three decades; unfortunately

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there has been confusion in the meaning of several commonly used terms. In this paper the term Ouachita facies simply identifies rocks that are lithologically and faunally similar to those exposed in the Ouachita Mountains; it carries no connotations of metamorphism. Some Ouachita facies rocks are metamorphosed and some are not. The Ouachita structural belt is the name given to the belt of deformed rocks--it includes deformed sedimentary rocks of Ouachita facies, metamorphosed rocks, and in some areas deformed rocks of foreland facies.

Tectonics of the Ouachita Belt

The western segment.--The southwest-striking axes of the Ouachita Mountains disappear beneath the coastal plain overlap in southern Oklahoma; some of the most formidable problems in interpretation of the structural relations of the belt occur in this area of southern Oklahoma where the southwest-striking Ouachita axes apparently intersect the southeast-trending Arbuckle element. Subsurface data show that the Ouachita chains are deflected around the Arbuckle element in a sharp structural recess, and there are reasons to suspect major tear-faulting along the northeast side of the Arbuckle mass, the Ouachita rocks having moved northwestward. Possibly in the subsurface to the southeast, Ouachita facies rocks have been thrust over the plunging Arbuckle element, as suggested by van der Gracht (1931, p. 1000). Although there are many unsolved problems in southern Oklahoma, it is clear that the Ouachita belt strikes southward into Texas, arcs sharply around the Llano uplift in central Texas, and trends westward into southwest Texas. The general course of the belt in Texas was first determined by Sellards (1931) and Miser and Sellards (1931). In Uvalde, Kinney, Maverick, and Zavala counties, there is another problem area where the well-defined zones mapped in the subsurface belt to the east and north (Plate I) are lost in a large area characterized by poor well control and profuse intrusion of Cretaceous-Tertiary igneous rocks. Westward the course of the Ouachita belt is defined in an area of adequate well control (Val Verde and Terrell counties) and by the exposures of the Marathon and Solitario uplifts. Farther south in Mexico, the Sierra del Cuervo, Las Delicias--Acatita, and Cd. Victoria exposures and a few wells offer tantalizing glimpses of Paleozoic rocks of geosynclinal facies that may well be part of the Ouachita belt.

In all of the 900-mile length of the belt in Texas, there are less than 325 wells that have penetrated rocks of the structural belt and sedimentary rocks of the immediately adjacent foreland shelf and basins. Such sparse well control means that only major features of the belt can be recognized with any confidence.

The western segment of the belt is geologically divisible into two parts: (1) the northern limb extends roughly north-south from southeastern Oklahoma to south-central Texas and is clearly a buried extension of the Ouachita Mountains; Ouachita Mountain stratigraphic terminology can be applied to unmetamorphosed beds of the frontal zone; (2) the southwestern limb in southwest Texas and Mexico shows a number of structural and lithologic differences from the northern limb; Ouachita Mountain terminology cannot be applied with confidence, and in part of the area the frontal zone of the belt is very narrow or missing completely in the subcrop (Plate I). Moreover, orogenic activity in this part of the belt seems to have occurred later or persisted longer than in the area to the north.

The northern limb of the belt in Texas can be divided into a frontal zone of unmetamorphosed folded and thrust-faulted rocks, a zone of incipiently to very weakly metamorphosed rocks, and a zone of highly sheared low-grade metamorphic rocks (Plate I). The frontal zone, which is comparable to the Valley and Ridge province of the Appalachian system in scale and in structures, is composed mostly of essentially unmetamorphosed dark shales and sandstones of the Stanley formation;[3] sporadic cores and a history of excessive hole deviation in many wells indicate that in many areas these beds are tilted at high

[3] In only one well in the subsurface Ouachita belt in Texas was Jackfork sandstone identified, namely, Cox No. 1 Leslie in Fannin County.

angles. In some areas along the front of the belt (Grayson and Collin counties, Ellis County, and Bell, Coryell, and McLennan counties) the pre-Cretaceous subcrop is composed of pre-Stanley Ouachita facies rocks, mostly of Bigfork and Womble lithologies but also including Arkansas novaculite and Missouri Mountain shale. In two of these areas (Grayson-Collin counties and Bell County) wells have passed through older Ouachita facies rocks into younger rocks of foreland facies, and it seems reasonably well established, therefore, that these early Paleozoic Ouachita facies rocks have been raised along thrust faults. In areas where well control is sufficiently dense to locate these frontal overthrusts accurately, minimum displacements are on the order of 5 to 6 miles. In central Texas where the Ouachita belt has been crushed against the Llano buttress, foreland facies rocks on the southeast side of the Llano uplift have been deformed and are part of the structural belt.

East of the frontal zone is a belt composed of dark clastic rocks showing incipient to very weak metamorphism; these rocks are angular poorly sorted carbonaceous micaceous sandstones, graywackes, and silty slaty shales. This zone widens southward in the area where the Ouachita belt wraps around the Llano buttress, and in this area rocks of the frontal zone also show incipient metamorphism.[4] No stratigraphic name can be applied to these dark clastic rocks at this time--possibly they are a near-source facies of lower Paleozoic Ouachita facies rocks.

Farther eastward in this area the pre-Mesozoic subcrop is composed of very highly sheared phyllites, slates, and metaquartzites; according to the writer's interpretation, these rocks have been thrust westward over the unmetamorphosed and very weakly metamorphosed rocks of the belt. This contact is a major structural discontinuity; it is named the Luling overthrust front from the area where these highly sheared rocks were first penetrated by the drill, and it may be analagous to the Blue Ridge front of the Appalachian area. The Luling front is not a simple overthrust; more likely it is a very complex zone of multiple overthrusts and strongly overturned folds, possibly even with development of nappes. In south-central Texas there is in the subcrop a narrow zone south of the Luling overthrust front which is distinguished by intensely deformed dark slates, altered greenstones, and partly mylonitized granitic rocks; the extreme deformation and the mylonitic rocks suggest that this zone may be the sole along which the Luling front advanced.

Immediately west of the Llano uplift along the southern margin of the Kerr basin there is evidence to suggest that the Ouachita belt has been thrust northward or northwestward over the foreland basin sequence of the Kerr basin much in the same way that the Ouachita rocks to the north have been thrust over the sedimentary rocks of the Fort Worth basin. Both Stanley and pre-Stanley lithologies are present in this area and both show incipient to weak metamorphism along the very front of the belt.

It has not been possible to map the Ouachita belt through Medina, Uvalde, Kinney, Maverick, and Zavala counties, although the few pre-Cretaceous wells in the area show the presence of the same types of rocks that are present in the belt to the north and east; this is due in part to poor well control but there seems also to be a structural break or discontinuity in the course of the Ouachita chains in this area. Undoubtedly the profusion of intrusive basic igneous rocks of Cretaceous-Tertiary age in this same area is a reflection of some deep Paleozoic structural condition.

West of this area, from Kinney County to Brewster County, the course of the belt can be fairly well located. Essentially unmetamorphosed Ouachita facies (Marathon facies) rocks of the frontal zone of the belt occur in a great structural salient that is partly exposed in the Marathon and Solitario uplifts. Although the rocks of the Marathon basin are remarkably similar in facies to Ouachita Mountain rocks, the sequence in

[4] Units are defined on the basis of lithology; metamorphism may transgress lithologic boundaries. It is important to bear in mind that lithologic units and metamorphic zones are separate and distinct entities.

the northern part of the Marathon area contains more carbonate rocks than the Ouachita Mountain section; apparently the northern part of the Marathon area was close to the foreland and during intervals of Paleozoic time a foreland environment prevailed. The Ouachita or Marathon facies rocks in the Marathon basin have been displaced northward along a great frontal overthrust (Dugout Creek overthrust) which has a minimum displacement of 6 or 8 miles (P. B. King, 1937, p. 170; and personal communication, 1957). Incipiently to weakly metamorphosed Marathon basin units (Tesnus and Marathon formations) are present in the subsurface in southern Terrell County. Highly sheared low-grade metamorphic rocks, including fine-grained schists, phyllites, marbles, and metaquartzites, are exposed in the Sierra del Carmen in Coahuila, Mexico, to the south (Flawn and Maxwell, 1958). To the east in Val Verde and Kinney counties the frontal zone so well displayed in the Marathon basin is either absent or very narrow in the subcrop. Highly sheared fine-grained schists, phyllites, marbles, and metaquartzites similar to those exposed in the Sierra del Carmen are bordered on the north by foreland sediments of the Val Verde basin. In this same area there is an ill-defined area of structurally high metavolcanic rocks, probably Precambrian in age, mantled by Paleozoic carbonate rocks (probably early Paleozoic) (Plate I). Two hypotheses are presented: (1) frontal zone rocks similar to those exposed in the Marathon basin occur south of the Precambrian (?) high and have been overridden by a plate of highly sheared low-grade metamorphic rocks, or (2) the frontal zone of the geosyncline was never developed in this area because of the presence of a persistently high area of Precambrian (?) rocks--this area was a southward-projecting cusp of the foreland shelf. The highly sheared low-grade metamorphic rocks encountered in subsurface and exposed in the Sierra del Carmen are similar both in lithology and type of metamorphism to the rocks which are present eastward to the south of the Luling overthrust front.

In Chihuahua, Mexico, southwest of the Marathon basin, Pennsylvanian (?) and Permian flysch, sheared and incipiently to very weakly metamorphosed, is exposed in the Sierra del Cuervo; a neighboring exposure in the Placer de Guadalupe area is composed of Pennsylvanian (?) and Permian sediments showing variable incipient to weak metamorphism which seem to be of foreland facies or at least seem to belong on the foreland side of the structural belt. There are thick Pennsylvanian (?) and Permian sediments of geosynclinal facies southeast of the Sierra del Cuervo in the Las Delicias--Acatita area. Southeastward along the strike of the Sierra Madre in the Cd. Victoria area there are exposures of Mississippian-Permian flysch in fault contact with garnetiferous gneisses and dark schists of unknown age. In northeastern Coahuila the No. 2-A Peyotes well penetrated highly sheared low-grade metamorphic rocks; in central Coahuila, the No. 1 Barril Viejo well (also known as No. 2 San Marcos) penetrated granodiorite southeast of a reported exposure of "granite" in Potrero de la Mula (Kellum and others, 1936). To the east in Nuevo Leon, the No. 101 Chapa well encountered highly sheared slates and metagraywackes beneath Jurassic beds. There is evidence to indicate the presence of a granitic terrane within this Paleozoic tectonic land of northern Mexico; this was recognized earlier by Kellum and others (1936) during a study of the Coahuila peninsula, a Mesozoic structural feature. They say (1936, pp. 977-978):

> "....in the south this landmass (Coahuila peninsula) was formed by marine Permian sediments and lavas, by Permian or post-Permian intrusive granites and granodiorites, and by a series of phyllites, quartzites, slates, shales and conglomerates of Paleozoic age, probably in part Permian; ....in central Coahuila it is composed of granitic rocks and associated dikes and quartz veins;.... in northern Coahuila it is composed of pre-Cambrian mica schist." [5]

The evidence pointing to the existence of this granitic terrane is summed up as follows: (1) there are granitic outcrops in Potrero de

[5] The "pre-Cambrian mica schist" referred to is the schist exposed in the Sierra del Carmen (Flawn and Maxwell, 1958) and may be Paleozoic or Precambrian.

la Mula.[6] (2) granodiorite was penetrated in the No. 1 Barril Viejo, (3) there are thick Jurassic (and Triassic?) sequences of arkose in eastern Coahuila and in the subsurface in Nuevo Leon, (4) there are fragments of cataclastically altered granite in conglomerates beneath Jurassic beds in the Humble No. 1 Bandera County School Land well in Maverick County, Texas, (5) Miller (1955) deduced the presence of a granitic source area in northern Mexico for the late Permian or Triassic Pierce Canyon redbeds of the Delaware basin of Texas and New Mexico. The age of the granitic rocks is unknown--if they are related to the granite intrusions of the Las Delicias--Acatita area, the intrusion is very late Permian or post-Permian; on the other hand, if the numerous granite pebbles and cobbles in the Haymond boulder bed of the Marathon basin were derived from an uplift of this terrane, the granite is pre-Atokan.

Connection of the Ouachita structures exposed in the Marathon basin with exposures of Paleozoic rocks in Mexico has been discussed by Kellum and others (1936), R.E. King (1944), Eardley (1951), P. B. King (1951), and Diaz G. (1956). Inspection of the map, the fact that there are Permian flysch sequences in Mexico, and the fact that deformation in Mexico was very late Permian indicate that a simple projection of the Ouachita front southward or southwestward is not a sound interpretation. Currently available data do not permit mapping of the pre-Mesozoic subcrop in northern Mexico; in the writer's opinion, however, it is reasonable to assume that the Paleozoic tectonic sediments in the area, the metamorphic rocks, and the granitic rocks are part of the same Paleozoic system that margins the craton to the north--the Ouachita system.

The eastern segment.--East of the Ouachita Mountains the Ouachita structural belt disappears beneath the thick prism of sediments in the Mississippian embayment. A number of wells in southeast Mississippi and southwest Alabama have encountered metasedimentary rocks and indicate that the belt trends southeastward beneath the Mississippi embayment from its last known exposures in central Arkansas. Very little original work on this area has been done in connection with this project; this phase is still in the data-gathering stage. Most of what is known about the area has come from the work of Applin (1951), Bridge and Berdan (1950), and Rice (1957). A preliminary interpretation from their data suggests that highly sheared metasedimentary rocks have been thrust over Pennsylvanian (Atoka) and Ordovician (Knox) beds along the southwestern margin of the Black Warrior basin.

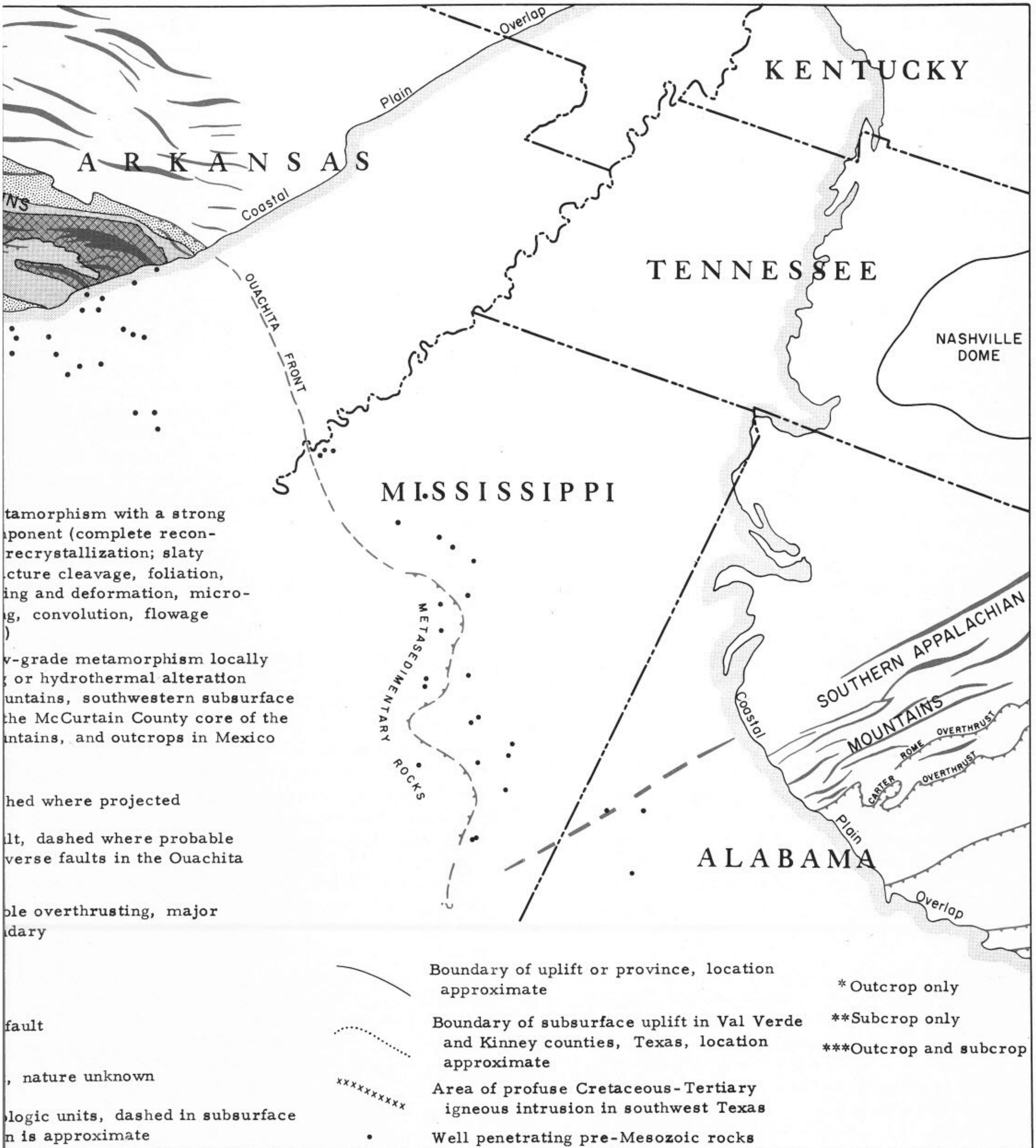
In this area the Ouachita belt is trending southeast and as projected intersects the southwest-plunging Appalachian structures in much the same way that it appears to intersect the Arbuckle trend in southern Oklahoma. This suggests that perhaps the Ouachita structures are sharply recessed around the plunging Appalachian elements and continue southeastward. The work of Applin (1951) and Bridge and Berdan (1950) shows that there are early and middle Paleozoic sedimentary rocks in northwestern and north-central Florida (Suwanee River basin of Braunstein, 1958)--possibly the Ouachita belt passes southwest of this area.

Foreland Elements

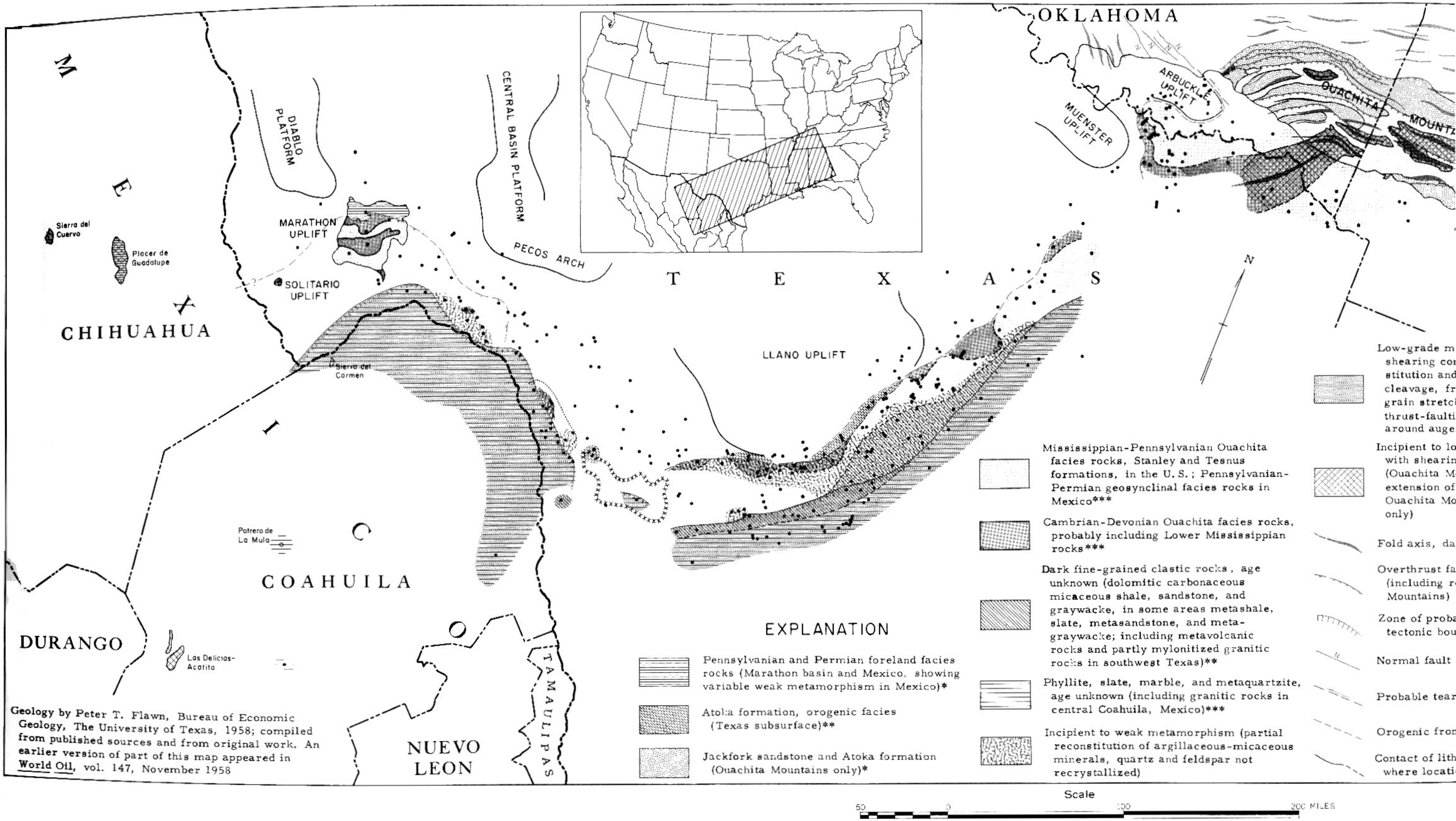
Foreland basins.--Study of the stratigraphic record and structures of the frontal basins of the Ouachita belt provides a great deal of information useful in reconstructing the history and development of the Ouachita system. The lower Paleozoic stratigraphic record shows that the early Paleozoic foreland of the Ouachita geosyncline was a more or less stable shelf characterized by deposition of thick carbonates and fine clastics. The abundant siliceous material in the Devonian foreland is interpreted as a reflection of distant volcanic activity in the mobile area to the south and east. Mississippian shales indicate continued fine clastic deposition in the foreland area while the geosyncline was advancing closer toward the craton. Pennsylvanian (Atoka and early Strawn) clastic wedges in the Black Warrior, McAlester-Arkansas, Fort Worth, and Kerr basins record strong foreland downwarps in

[6] There is a question as to whether the granitic rocks in Potrero de la Mula are pre-Lower Cretaceous (probably pre-Mesozoic) in age or post-Cretaceous intrusions (Teodoro Diaz G., personal communication, 1958).

Plate I



KENTUCKY
ARKANSAS
TENNESSEE
MISSISSIPPI
ALABAMA
NASHVILLE DOME
Overlap
Plain
Coastal
OUACHITA FRONT
METASEDIMENTARY ROCKS
SOUTHERN APPALACHIAN MOUNTAINS
OVERTHRUST
ROME OVERTHRUST
CARTER
Coastal
Plain
Overlap
tamorphism with a strong
ponent (complete recon-
recrystallization; slaty
cture cleavage, foliation,
ing and deformation, micro-
g, convolution, flowage
)
v-grade metamorphism locally
or hydrothermal alteration
untains, southwestern subsurface
the McCurtain County core of the
ntains, and outcrops in Mexico
hed where projected
lt, dashed where probable
verse faults in the Ouachita
ble overthrusting, major
dary
fault
, nature unknown
logic units, dashed in subsurface
n is approximate
Boundary of uplift or province, location approximate
Boundary of subsurface uplift in Val Verde and Kinney counties, Texas, location approximate
Area of profuse Cretaceous-Tertiary igneous intrusion in southwest Texas
Well penetrating pre-Mesozoic rocks
* Outcrop only
** Subcrop only
*** Outcrop and subcrop



OUACHITA STRUCTURAL BELT
(Preliminary Map)

response to the thrust of the mobile belt against the foreland; most of the later Strawn-Canyon rocks are post-orogenic products derived from the uplifted geosynclinal rocks. The major foreland basin downwarp in the Val Verde basin to the west was post-Strawn and possibly mainly Wolfcamp in age (Adams and others, 1952; Frenzel, 1957).

One of the most interesting and significant foreland basin units is the Atoka formation; this name has been applied to sedimentary units from Mississippi to Texas which show marked facies differences. Studies of the Atoka in connection with this project have been restricted to petrographic examinations of Atoka sandstones and shales close to the Ouachita orogenic front and, locally, beneath overthrust Ouachita facies plates. The writer does not feel qualified to discuss the regional stratigraphy of the Atoka or problems of correlation; it should be noted, however, that in some areas close to the orogenic front the Atoka sandstones contain abundant rock fragments derived from the uplifted Ouachita blocks to the south and east--these include fragments of phyllite, slate, metaquartzite, metashale, shale, chert, and composite strained, stretched, and crushed quartz grains. Grains of the distinctive cherts of the Arkansas novaculite and Bigfork chert can be recognized in these Atoka sandstones. In attempting to map the subsurface relationships of the Ouachita belt and the adjacent foreland basin rocks, the writer has found it convenient to refer to this lithology as "the orogenic facies of the Atoka" (Plate I).

Foreland uplifts marginal to the Ouachita structural belt are of two types: (1) large, more or less domical uplifts with a history of positive tendency throughout the Paleozoic (Ozark uplift,[7] Llano uplift) and (2) elongate faulted uplifts of late Paleozoic age (Arbuckle, Muenster, and Fort Stockton or Pecos uplifts). The large persistently positive features antedate the formation of the Ouachita system; they are ancient fundamental elements of the North American craton and to a great extent influenced the course of the Paleozoic Ouachita system, which was a craton-margin system. The second type of foreland uplift, the elongate, faulted uplifts, formed in response to the orogenic thrust of the Ouachita belt against the craton (or, in the case of the Arbuckle element, to forces generated in an intra-cratonic geosyncline), and most probably their form and trend were controlled by older Precambrian structures (Flawn, 1956, pp. 69-71)--the orogenic forces of the Ouachita belt were resolved along these pre-existing zones of crustal weakness.

Development of the Geosyncline

Whereas few geologists doubt the formation of a geosyncline along the course of what is now the Ouachita structural belt in Mississippian-Pennsylvanian time, some have challenged the idea that the area was a geosyncline during early and middle Paleozoic time (Barton, 1945; Eardley, 1951; Harlton, 1953). Their objections seem to arise chiefly from the fact that the early and middle Paleozoic sequences exposed in the Ouachita Mountains and Marathon basin are not particularly thick and that early Paleozoic sedimentary rocks in the foreland basins are in fact much thicker than rocks of the same period in the so-called Ouachita geosyncline. The difference in facies between the Ouachita Mountain and the Marathon basin rocks and the rocks of the foreland basins is explained on the grounds of distance from source area, the muddy and sandy Ouachita facies are supposed to differ from the clean carbonate rocks of the foreland basins only because of proximity to a southern source of clastics.

The writer does not subscribe to this view. There is evidence that the early and middle Paleozoic rocks of the Ouachita system were deposited in a completely different tectonic environment from that of the foreland basins; early and middle Paleozoic rocks of Ouachita facies differ from common near-shore facies in the presence of substantial quantities of siliceous sediments and dark graptolitic shales. Conglomerate and beds containing exotic boulders attest to sporadic early Paleozoic movements within the belt. From regional studies of the belt it is clear that the pre-Mississippian sequences exposed in the Ouachita Mountains

[7] Possibly including the Hunton arch in pre-Mississippian time.

and Marathon basin are merely the foreland edges of sedimentary prisms which thicken to the south and east. Rigid application of the criterion of excessive thickness in recognition of geosynclinal sediments does not take into account the possibility that a geosynclinal trench may not develop as a single basin but may consist of two or more troughs separated by more or less continuous or en echelon submarine welts; in such a situation the trough nearest the active source area receives the great load of rapidly deposited clastic sediments while the frontal trough receives only fine clastics and possibly siliceous sediments. There is, moreover, convincing faunal evidence that the early Paleozoic Ouachita belt was the locus of a biofacies distinct from that of the foreland (Wilson, 1954a, 1954b). But probably the most convincing argument for an early Paleozoic Ouachita geosyncline is the simple fact of congruence of a linear arcuate late Paleozoic geosyncline more than 1,300 miles long with a very distinctive early and middle Paleozoic Ouachita (and Marathon) facies. Although the axis of late Paleozoic geosynclinal deposition was shifted closer to the foreland than the axis of maximum deposition of earlier Paleozoic time, a history of Paleozoic mobility along a linear zone peripheral to and south of the old Precambrian craton seems well demonstrated. The course of the Ouachita structural belt as we now see it is a reflection of the shape of the old stable craton against which it was deformed; the orogenic front is recessed around the unyielding or buttressing elements of the craton and shows strong salients in the "bays" of the old craton.

Age of the Deformation

Information on the age of the deformation of the Ouachita belt is given in the structures of the deformed rocks themselves and in the sedimentary sequences in the frontal basins. Thick clastic deposits of Atoka age occur in the Black Warrior, McAlester-Arkansas, Fort Worth, and Kerr basins and indicate upwarps of the Ouachita belt south and east of these basins in Middle Pennsylvanian time; in the Val Verde basin the thick clastic section is post-Strawn--probably late Pennsylvanian and early Permian in age--suggesting a later uplift of the Ouachita belt in this area or at least a later downwarp in the frontal basin.

In the Ouachita Mountains, beds of Atoka age are involved in the frontal thrust faults of the structural belt; in Grayson County, Texas, lower Paleozoic Ouachita facies rocks are thrust over Strawn beds; in the Marathon basin area late Pennsylvanian and perhaps Wolfcamp beds are affected by thrust-faulting; in Mexico late Pennsylvanian (?) and Permian beds are strongly deformed (in the Las Delicias--Acatita area these deformed beds include Capitan and possibly Ochoa equivalents). In the northern and eastern parts of the belt in the Ouachita Mountain area the deformation can be stratigraphically dated only as Des Moines or post-Des Moines, that is, post-Middle Pennsylvanian (Hendricks, personal communication, 1958). Miser (1934, p. 1009) dated the orogeny in this area as late Pennsylvanian. However, van der Gracht (1931), who presented an amazingly complete picture of Ouachita tectonics when only limited subsurface data were available, concluded that the major period of Ouachita folding was Pennsylvanian and the final overthrusting was Permian. The interpretation of Permian age of the thrusting is based on the presence of supposedly Ouachita-derived Permian sediments in the Wichita Mountains (van der Gracht, 1931, pp. 1014, 1028) and a study of joints by Melton (1930). This is rather indirect evidence. The Ouachita-derived Permian sediments could have resulted from Permian epeirogeny in the Ouachitas. More recent investigators in the Ouachita Mountain area (Hendricks and others, 1947; Misch and Oles, 1957; Reinemund and Danilchek, 1957) have not discussed in published sources the age of the deformation. Eardley (1951, p. 213), summing up various opinions, stated that the age of the main Ouachita (Mountain) orogeny cannot be fixed until better evidence is at hand. In the Marathon basin, P. B. King (1937; 1951, p. 151) was able to date the orogeny rather closely as late Pennsylvanian; more recent work indicates that movements may have persisted into early Wolfcamp time. The situation in Mexico is much different because the deformation there is very late Permian and there are Permian flysch sequences.

It is concluded that the final orogenic pulses in the Ouachita system in the United States were late Pennsylvanian and early

Permian (Wolfcamp) and that in Mexico there was late Permian orogenic activity.

Whatever the age of the final orogenic movements in the Ouachita system, the thick sequences of Mississippian (and possibly early Pennsylvanian) flysch in both the Ouachita Mountains and Marathon basin (Stanley and Tesnus formations), and succeeding molasse deposits, attest to powerful orogenic movements throughout the course of the geosyncline in Mississippian and early Pennsylvanian time. In 1950, following detailed petrographic work on the older rocks of the Ouachita Mountains, Goldstein (unpublished manuscript) concluded that in the Ouachita Mountain area a major epeirogeny occurred near the end of the Ordovician and extended well into Silurian time; he referred to this movement as a Taconic epeirogeny. Later he (Goldstein, 1958) developed strong evidence of an early Paleozoic geosynclinal cycle in the Ouachita Mountains. P. B. King (1951, p. 150) also pointed out that there are indications of earlier (Taconic ?) movements in the Ouachita Mountains. In the Marathon area, Ordovician conglomerates are evidence of early Paleozoic movements in the southwestern part of the belt (P. B. King, 1937, pp. 37, 47).

Comparison with the Appalachian System

Beyond the fact that the Appalachian and Ouachita systems are both late Paleozoic orogenic belts marginal to an older North American craton, and that they are characterized by similar structures, there is an interesting similarity in scale.

The major structural divisions of the Appalachian system (the frontal Valley and Ridge province, the overthrust and metamorphosed Blue Ridge province, and the extensively intruded and metamorphosed Piedmont province) are familar to most geologists. In scale and in pattern the divisions recognizable in the buried Ouachita belt are similar to those of the Appalachian system. The frontal zone of the Ouachita belt is about 40 miles wide and is composed of essentially unmetamorphosed folded and thrust-faulted beds; the Luling overthrust front may be compared to the Blue Ridge front, and the phyllite-slate-metaquartzite sequence with its imposed slaty and fracture cleavage resembles Blue Ridge province rocks. In the Appalachians, the Blue Ridge province is, like the Valley and Ridge province, about 40 miles wide, but only a 15-mile width of this belt has been found with the drill in the Ouachita belt because of the increasing depth of burial to the south and east. The foundation of the coastal plain of Texas and Louisiana may well be a deeply buried Ouachita "piedmont".

Inspection of the frontal structures of the well-exposed Appalachian system will illustrate the hazards of attempting to project the conclusions of areal stratigraphic studies too far. In western Virginia and eastern West Virginia, the Appalachian frontal zone consists of autochthonous folded beds cut by high-angle reverse faults, but not far to the south in Kentucky and Tennessee the Appalachian front has been displaced over the foreland along the Pine Mountain overthrust. This demonstrates the need for caution in extending the results of studies in either the Ouachita Mountains or Marathon basin too far into the subsurface.

References

Adams, J. E., Frenzel, H. N., and others, 1952, Marathon basin, Brewster and Pecos counties, Trans-Pecos Texas: West Texas Geol. Soc. Field Trip Guidebook, 78 pp.

Applin, P. L., 1951, Preliminary report on buried pre-Mesozoic rocks in Florida and adjacent states: U. S. Geol. Survey Circ. 91, 28 pp.

Barnes, V. E., 1948, Ouachita facies in central Texas: Univ. Texas, Bur. Econ. Geol. Rept. Inv. No. 2, 12 pp.

Barton, J. M., 1945, Pre-Permian axes of maximum deposition: Amer. Assoc. Petroleum Geologists Bull., vol. 29, pp. 1336-1348.

Bodenlos, A. J., and others, 1956, Estratigrafia del Cenozoico y del Mesozoico a lo largo de la Carretera entre Reynosa, Tamps. y Mexico, D. F. Tectonica de la Sierra Madre Oriental. Vulcanismo en el Valle de Mexico. Excursiones A-14 y C-16, Congreso Geologico Inter. XX, 323 pp.

Braunstein, Jules, 1958, Eastern Gulf Coast oil and gas geology: World Oil, vol. 146, no. 7, pp. 133-139.

Bridge, Josiah, and Berdan, J. M., 1950, Preliminary correlation of Paleozoic rocks from test wells in Florida and adjacent parts of Georgia and Alabama: U. S. Geol. Survey unpublished report.

De Cserna, Z., and Diaz G., Teodoro, 1956, Estratigrafia Mesozoico y tectonica de la Sierra de Chihuahua, etc.: Excursion A-13, Congreso Geologico Inter. XX, 120 pp.

Diaz G., Teodoro, 1956, Generalidades sobre la geologia del norte de Mexico, in Estratigrafia Mesozoica y tectonica de la Sierra de Chihuahua, etc. Excursion A-13, Congreso Geologico Inter. XX, 120 pp.

Eardley, A. J., 1951, Structural geology of North America: Harper Bros., 624 pp.

Flawn, P. T., 1956, Basement rocks of Texas and southeast New Mexico: Univ. Texas Pub. 5605, 261 pp.

Flawn, P. T., 1958, The subsurface Ouachita structural belt in Texas: World Oil, Nov., vol. 147, no. 6, pp. 141 ff.

Flawn, P. T., and Maxwell, R. A., 1958, Metamorphic rocks in the Sierra del Carmen, Coahuila, Mexico: Amer. Assoc. Petroleum Geologists Bull., vol. 42, pp. 2245-2249.

Frenzel, H. N., 1957, Introduction, in Wolfcamp of the Glass Mountains and the Permian basin: Soc. Econ. Paleont. and Min., Permian Basin Section, Field Trip Guidebook, pp. 1-3.

Goldstein, August, Jr., 1958, Paleozoic sandstones from the Ouachita Mountains of Oklahoma and Arkansas: Amer. Assoc. Petroleum Geologists, symposium volume, pp. 97-117.

Harlton, B. H., 1953, Ouachita chert facies in southeastern Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 37, pp. 778-796.

Heim, Arnold, 1940, The front ranges of the Sierra Madre Oreintal, Mexico between Cd. Victoria and Tamazunchale: Ecologae Geol. Helvetiae, vol. 33, pp. 313-352.

Hendricks, T. A., and others, 1947, Geology of the western part of the Ouachita Mountains in Oklahoma: U. S. Geol. Survey Oil and Gas Inv. Map No. 66.

Humphrey, W. E., and Diaz G., Teodoro, 1953, A.M.G.P. Primera Convencion, Mexico, D. F., Febrero 9 al 15.

Kellum, L. B., Imlay, R. W., and Kane, W. G., 1936, Evolution of the Coahuila peninsula, Mexico: Geol. Soc. America Bull., vol. 47, pp. 969-1009.

Kelly, W. A., 1936, Geology of the mountains bordering the valleys of Acatita and Las Delicias: Geol. Soc. America Bull., vol. 47, pp. 1009-1035.

King, P. B., 1937, Geology of the Marathon area: U. S. Geol. Survey Prof. Paper 187, 148 pp.

King, P. B., 1951, Tectonics of middle North America: Princeton Univ. Press, 203 pp.

King, R. E., 1944, Geology and paleontology of the Permian area northwest of Las Delicias, southwestern Coahuila, Mexico: Geol. Soc. America Special Paper 52, 130 pp.

Masson, P. H., 1955, Metamorphism in the Llanoria structural belt (abs.): Geol. Soc. America Bull., vol. 66, p. 1593.

Masson, P. H., and Sellards, E. H., 1931, Pre-Cretaceous rocks found in wells in the Gulf Coastal Plain south of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 15, pp. 801-818.

Melton, F. A., 1930, Age of Ouachita orogeny and its tectonic effects: Amer. Assoc. Petroleum Geologists Bull., vol. 14, pp. 57-72.

Miller, D. F., 1955, Petrology of Pierce Canyon redbeds, Delaware basin, Texas and New Mexico: Doctoral Dissertation, The University of Texas, 148 pp.

Misch, Peter, and Oles, K. F., 1957, Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: Amer. Assoc. Petroleum Geologists Bull., vol. 41, pp. 1899-1905.

Miser, H. D., 1934, Carboniferous rocks of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 971-1009.

Reinemund, J. A., and Danilchek, W., 1957, Preliminary geologic map of the Waldron quadrangle and adjacent areas, Scott County, Arkansas: U. S. Geol. Survey Oil and Gas Inv. Map OM-192.

Rice, Elmer, 1957, Unpublished maps.

Sellards, E. H., 1931, Rocks underlying Cretaceous in the Balcones fault zone area of central Texas: Amer. Assoc. Petroleum Geologists Bull., vol. 15, pp. 819-827.

van der Gracht, W. A. J. M. van Waterschoot, 1931, Permo-Carboniferous orogeny in south-central United States: Amer. Assoc. Petroleum Geologists Bull., vol. 15, pp. 991-1057.

Wilson, J. L., 1954a, Ordovician stratigraphy in the Marathon folded belt: Amer. Assoc. Petroleum Geologists Bull., vol. 38, pp. 2455-2475.

Wilson, J. L., 1954b, Late Cambrian and early Ordovician trilobites from the Marathon uplift, Texas: Jour. Paleont., vol. 28, pp. 249-285.

Woods, R. D., 1956, The northern structural rim of the Gulf basin: Trans. Gulf Coast Assoc. Geol. Societies, vol. VI, pp. 3-11.

STRUCTURE AND VEIN QUARTZ OF THE OUACHITA MOUNTAINS OF OKLAHOMA AND ARKANSAS [1]

Hugh D. Miser [2]

Introduction

The structure and the vein quartz of the Ouachita Mountains are closely related in origin and they in turn are related to the other geologic features of the region. They thus constitute an important chapter on the geology of the Ouachita Mountains.

The exposed sedimentary rocks are 25,000 feet thick; they range in age from Cambrian to Pennsylvanian; and they were folded, faulted, and metamorphosed during the Pennsylvanian period. The deposits of vein quartz, asphaltite, and certain metalliferous minerals were formed during the same period. Igneous rocks - some of probable Cretaceous age and some of possible Ordovician age - are exposed in a few small areas.

Structure

The present discussion of the structure of the Ouachita Mountains is abstracted mainly from my paper of 1929 on this subject which was published as Bulletin 50 of the Oklahoma Geological Survey. This discussion is, however, brought up to date by including the subsequent studies of many geologists.

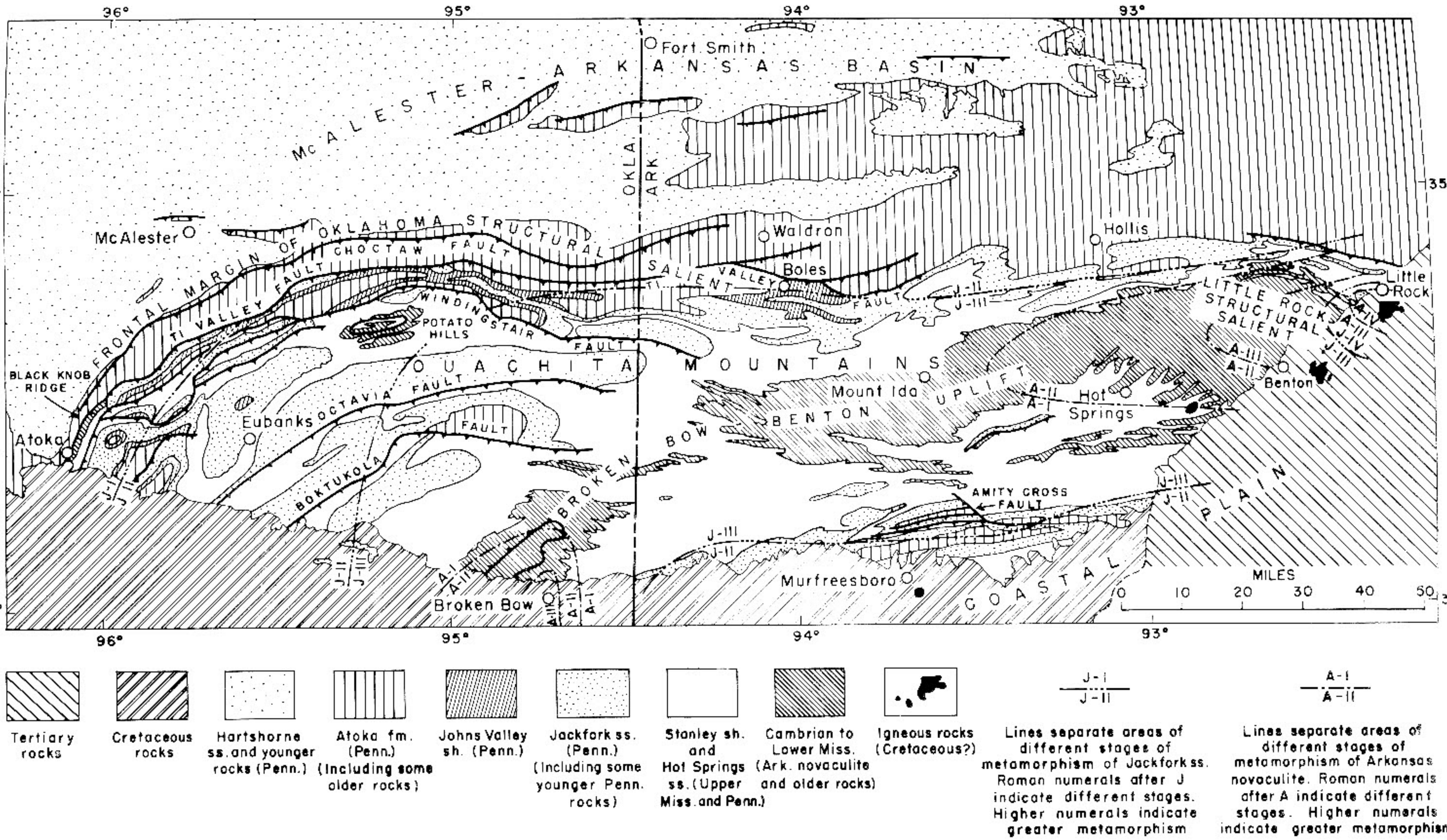
The paper of 1929 gave the first published description of the structure of the entire Ouachita Mountain region. It was based on information that I had obtained during all or parts of 14 field seasons beginning in 1907, and on the work of other geologists including Griswold (1892), Ashley (1897), Drake (1897), Taff (1902), Wallis (1915), Honess (1923, 1924), Purdue (1909), and Purdue and Miser (1923). From 1923 to 1925 I compiled, for the State geologic maps of Oklahoma (Miser, 1926) and Arkansas (Branner, 1929), all available maps including much unpublished mapping by Taff, Purdue, and me. Also, at this time I compiled a geologic map of the Ouachita Mountains which showed clearly the major structural features including the Oklahoma structural salient. The recognition of this salient led promptly to my interpretation of its forward thrusting. Next, there followed in 1927 my mapping and discovery of the window in the Potato Hills in Oklahoma. In December of that year a paper on the structure of the Ouachita Mountains was presented by me at the Cleveland, Ohio, meeting of the Geological Society of America. My conclusions stated there included discussion of the window in the Potato Hills and of a probable window near Broken Bow, McCurtain County, Oklahoma. The interpretation of a window in that county was based on Oklahoma Geological Survey Bulletin 32 by Honess. He was at the meeting and commented favorably about my conclusions.

The geologic map (Fig. 1) herewith includes not only the mapping that was compiled in 1923-1925 for the Arkansas part of the Ouachita Mountains but also the mapping of Reed and Wells (1936) and of Reinemund and Danilchik (1957). It also follows the second geologic map of Oklahoma (Miser, 1954) which contains the mapping of Hendricks, Gardner, Knechtel, and Averitt (1947) and the mapping of many oil companies.

The rock strata throughout the Ouachita Mountains have been deformed by folding

[1] Publication authorized by the Director, U. S. Geological Survey.

[2] United States Geological Survey, Washington, D. C.



H. D. Miser   Fig. 1   Geologic map of Ouachita Mountains. Also shows progressive metamorphism of Jackfork sandstone and Arkansas novaculite.

and faulting. The faults are chiefly reverse and thrust faults. The individual folds range from open to closely compressed anticlines and synclines. Also, there are compound folds - anticlinoria and synclinoria - some of which are normal fan folds and some are inverted fan folds.

The principal anticlinal fold of the Ouachita Mountains extends from Benton, Arkansas, in a west-southwest direction to Broken Bow, Oklahoma, and is essentially a compound anticlinorium consisting of several anticlinoria. It is here named the Broken Bow - Benton uplift. Along its higher parts, strata of Devonian, Silurian, Ordovician, and Cambrian age are exposed and these are succeeded on the flanks of the fold by strata of Pennsylvanian and Mississippian age. A smaller anticlinorium is in the Potato Hills where strata of Devonian, Silurian, and Ordovician age are exposed in a window and in the overthrust sheet. A belt of outcrop which may represent one limb of a smaller fold is at Black Knob Ridge in Oklahoma where the same rocks are exposed on the northwest edge of an overthrust sheet (Hendricks, Knechtel, and Bridge, 1937; Hendricks, Gardner, Knechtel, Averitt, 1947). Several small faulted exposures of strata of Devonian and Mississippian age overlain by Pennsylvanian strata lie in the frontal or northwestern part of the mountains in Oklahoma.

A dominant feature of the Ouachita Mountains is the Oklahoma structural salient which occupies all the Oklahoma part of the mountains and extends into western Arkansas. A second and smaller like feature is the Little Rock structural salient at the east end of the mountains.

The frontal marginal belt of the Oklahoma structural salient forms a great arc about 200 miles long. It trends northeasterly and then easterly along the northern edge of the mountains in Oklahoma and next passes southeasterly transverse to the mountains in western Arkansas from Boles to the Amity cross fault (see Figs. 1 and 2). The Choctaw fault lies at the north edge of the mountains in Oklahoma. Other long, generally concentric and parallel faults are farther south in that State; and several structural blocks, each separated by a major fault and each being more or less distinctive in the structure and stratigraphy of the rock formations, have been recognized and describe(Hendricks (1947). Concerning these struc tural blocks Hendricks states, "Geologica the area between the Choctaw and Ti Valle faults may be considered the frontal part (the Ouachita Mountains, for this area exhi its many features that indicate a transitior from the geology of the McAlester coal bas and the Arbuckle Mountains to that of the c tral part of the Ouachita Mountains. Structurally, this frontal part of the Ouachita Mountains is similar to the central part of the mountains, and is unlike the McAlester coal basin and Arbuckle Mountain regions. Stratigraphically, it is similar to the coal basin and Arbuckle Mountain regions and is almost completely unlike the central part of the Ouachita Mountains."

The rock sequence that is south of the Ti Valley fault is commonly called the Ouachita Mountain facies. The rocks of this facies seem to have been thrust north-northwestward a distance of 20 miles or more. This estimate of the thrusting was obtained by measuring the distance from the Ti Valley fault to a projected line that represents a west-southwesterly continuation of the structural axes of the region immediately north of Mount Ida, Arkansas, (Fig. 2). Such a continuation of these axes would pass through or near Eubanks, Oklahoma. The distance from the Ti Valley fault to Eubanks is about 20 miles. The asphaltite deposits (Ham, 1956), the oil seeps, and the few known small oil pools of the Ouachita Mountains are found in Oklahoma as far as 20 to 25 miles southeast of the Ti Valley fault. They thus are restricted to the northwest part of the Oklahoma structural salient. My opinion is that these hydrocarbons did not come from the Ouachita Mountain facies but came from rocks like the Arbuckle Mountain facies over which the Ouachita rocks have been thrust a distance of about 20 miles (Miser, 1934a). If the hydrocarbons were indigenous in the Ouachita facies, some indications of them should have been preserved and noted farther south and east in the Ouachita Mountains of Oklahoma and Arkansas.

The discovery of the window in the Potato Hills in 1927 established the presence of a low-angle overthrust planes in the Oklahoma structural salient. Since then, Hendricks, Gardner, and Knechtel (1947, Prelim. Map 66, sheet 1) have mapped a klippe of the

Ti Valley fault in Oklahoma, and Reinemund and Danilchik (1957) have mapped low-angle faults in the Waldron area, Arkansas. The window fault in the Potato Hills is interpreted by me to represent a southward and somewhat folded continuation of the plane of the Windingstair fault which follows the south base of Windingstair Mountain, 3 miles north of the Potato Hills. The northward thrust of the rocks above the window fault seems to be 3 miles or more. The rock formations exposed inside the window and in the thrust sheet are the Stanley shale, Arkansas novaculite, Missouri Mountain shale, Polk Creek shale, Bigfork chert, and Womble shale. A layer of conglomerate, generally less than 1 foot thick, occurs at or near the base of the Stanley shale in the window; it was not observed by me in the Stanley shale outside the window. Otherwise the rock strata in the window and those of the thrust sheet are the same. The anticlines inside the window are more closely compressed than those outside the window. During my mapping of the Potato Hills in 1927 I considered the possibility of a faulted inverted fanfold there, but the facts, as I saw them, indicated the existence of a window. Others who have worked in the Potato Hills and who follow the window interpretation include Miller (1956), Arbenz (1956), and company geologists, one of whom did the mapping in the Potato Hills as shown on the 1954 edition of the Oklahoma geologic map.

My interpretation of the window near Broken Bow, in northern McCurtain County, is that the window fault is a southward and slightly folded continuation of the plane of the Boktukola fault (Figs. 1 and 2). This interpretation, first presented publicly in a paper in 1927, was first published in 1929 (Oklahoma Geol. Survey Bull. 50). Later it was included on the second geologic map of Oklahoma (Miser, 1954), which represents my present opinion on the subject. It is based on the structural and stratigraphic data presented by Honess in his monumental contribution to the geology of the Ouachita Mountains (Honess, 1923). He points out that the deformed strata in the central part of the pre-Carboniferous area in McCurtain County do not partake of the folds of the rocks that surround the central area. To explain these differences in structure I interpret the central area to be a window through the overthrust sheet of the Boktukola fault. My estimate of the thrusting of this sheet is 5 miles and possibly 8 miles. These distances are the apparent amount of displacement of the Atoka formation, Jackfork sandstone, and Stanley shale, which are exposed along the trace of the Boktukola fault.

The low-angle thrusting just described supports the conclusion of Powers (1928), that the faulting in the Ouachitas included "notable overthrusting of the sheet (Decken) type". His outline map of the Ouachita Mountains in Oklahoma and one of his structure sections show the window in the Potato Hills (Powers, 1928, pp. 1038-1039).

The Little Rock structural salient at the east end of the Ouachita Mountains shows an arcuate arrangement of the structural trends in the region between Little Rock and Hot Springs (Figs. 1 and 2). Apparently, the great part of the salient lies beneath the Cretaceous, Tertiary, and Quaternary strata of the Coastal Plain.

Relation of structure to gravity

A gravity minimum of great regional significance coincides with the combined area of the Ouachita Mountains and the adjacent McAlester-Arkansas basin (Lyons, 1950; Cook, 1956). This gravity feature culminates, as stated by Cook, "in a tremendous gravity minimum over the Ouachita Mountains with a closure of about 70 milligals (Lyons, 1950). * * * This anomaly constitutes the strongest gravity minimum in the Midcontinent region." The area of greatest minimum within the minus 100-milligal contour (Lyons, 1950, map on pp. 34-35) includes the Potato Hills and the adjacent region. The minimum seems to be due to a combination of several factors including (a) the thick sequence of Pennsylvanian and Mississippian strata in the area of closure, (b) the relatively thick sections of shale in the pre-Mississippian rocks of the Ouachita facies, and (c) the thickening of the rock section by the piling up of overthrust sheets.

Relation of structure to exotic boulder beds

Exotic boulder beds are found in an arcuate belt 125 miles long and 17 miles wide along the northern part of the Oklahoma structural salient. These boulder beds are mainly in the Johns Valley shale (Ulrich, 1927; Powers, 1928; Harlton, 1934, 1938, 1947; Miser, 1934; Moore, 1934; Cooper 1945; Hendricks and others, 1947; Rea, 1947; Reinemund and Danilchik, 1957) but other beds are in the Stanley shale (Harlton, 1938, 1947) and Jackfork sandstone (Harlton 1938, 1947; Hendricks and others, 1947). The exotic boulders are chiefly masses of limestone of Ordovician to Mississippian age and masses of shale of Ordovician and Mississippian age. Some areas of shale of Mississippian age have been interpreted by Cline (1956a) as beds in their normal stratigraphic position between strata that have undergone marked changes in depositional facies. The exotic boulders have been interpreted by me as having a local source from fault scarps along the north side of the Johns Valley sea. Some of the source areas may now lie between the Choctaw and Ti Valley faults but some of the source areas seem to have been overridden by the thrust sheets of the Oklahoma structural salient. The boulders in the Johns Valley shale at and near Boles, Arkansas, resemble rocks that are exposed in the Ozark region and those in Oklahoma resemble rocks that are now exposed between the Choctaw and Ti Valley faults and in the Arbuckle Mountains. My present opinion is that the boulders were transported by submarine landslips from local fault scarps but that some of the largest masses of limestone (Mississippian and older) and of Caney shale (Mississippian) on and near the Ti Valley fault may have been derived directly by overthrusting from parent beds.

A shale that contains a few blocks of limestone, some of Early Mississippian age and others of Late Mississippian age (Mackenzie Gordon, oral communications, 1956, 1958) is exposed in the Little Rock structural salient. The exposure is at the south base of Forked Mountain, 5 miles east of Hollis, Arkansas, and 45 miles west of Little Rock, Arkansas. These blocks may lie in a fault zone or in an exotic boulder bed in or below the Jackfork sandstone which forms Forked Mountain. Factors that are involved in offering these two suggestions include the Pennsylvanian age (White, 1934, 1937 and Girty in Miser, 1934) of the Jackfork sandstone and the Pennsylvanian and Mississippian age (White, 1934, 1937; Hass, 1950, 1956) of the underlying Stanley shale. A solution of the stratigraphy and structure of the Forked Mountain area awaits geologic mapping there.

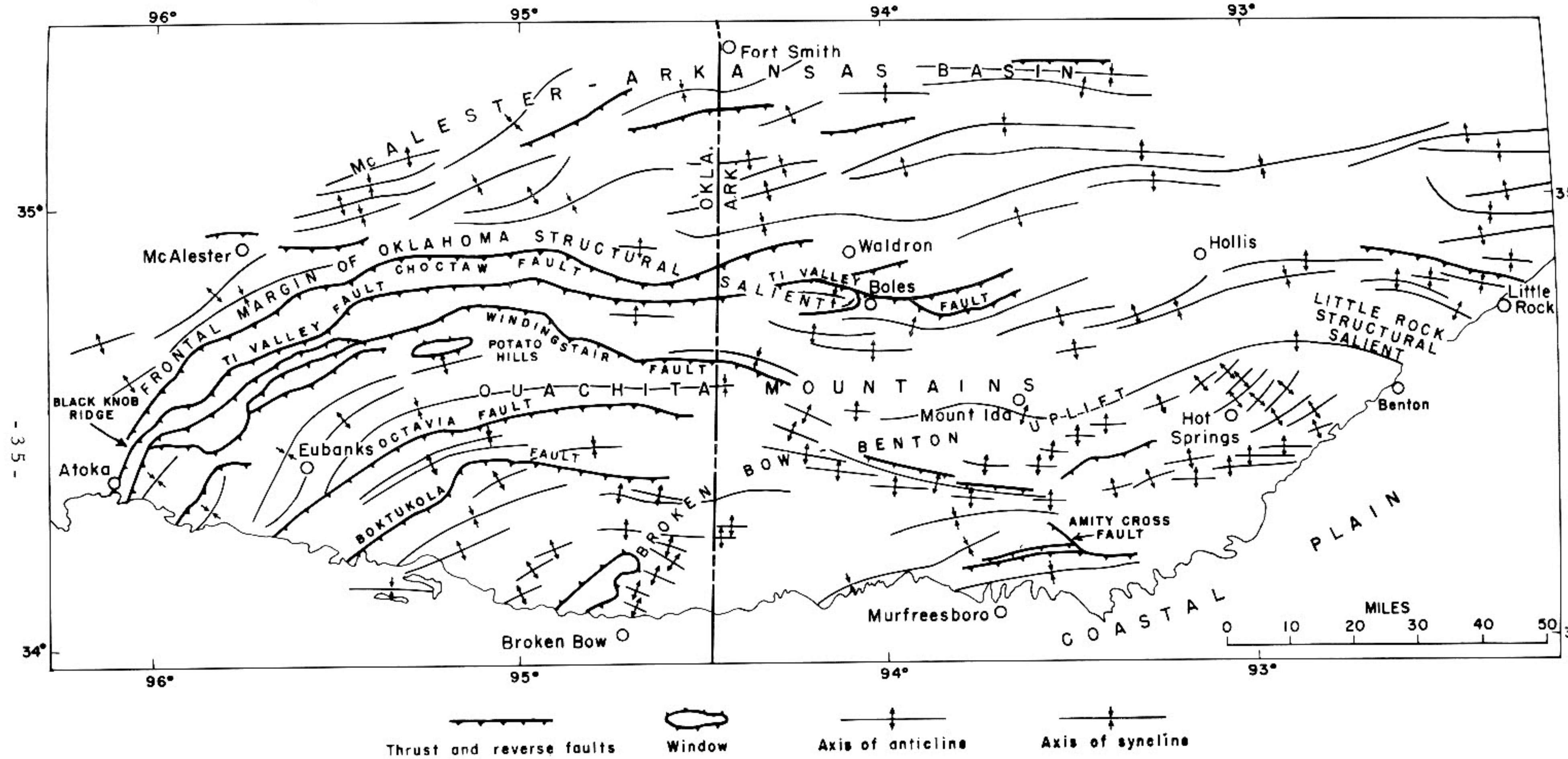
Times of deformation

The boulder-forming orogeny to account for the faulting in Johns Valley time (of Morrow age) was the earliest of several mountain-making movements that took place in the Ouachita region in Pennsylvanian time. Deformation after Johns Valley time and before Atoka time (also Early Pennsylvanian) seems to be indicated by a basal conglomerate of the Atoka formation in the Ouachita Mountains and by chert conglomerates (Melton, 1930) in the Atoka formation. Further deformation including thrust faulting probably took place coincident with movements in the Arbuckle Mountains in Middle and Late Pennsylvanian time (Ham, 1954; Arbenz, 1956). The Tishomingo anticline of the Arbuckle Mountains is believed by Ham (1956) to have moved northwestward between the Washita Valley and Reagan faults, probably as a result of stress transmitted from the Ouachita Mountains. Melton (1930) and van der Gracht (1931) suggest overthrusting of the rocks of the Ouachita Mountains in the Permian period.

Relation of structure to metamorphism

The rock strata in the central belt of the Broken Bow-Benton uplift show the effects of low-grade metamorphism that was produced by dynamic movement and by heat from depth of burial. The arching to form this uplift took place after the major part of the deformation (the close folding and the overthrusting) and after most of the metamorphism of the rock strata of the Ouachita geosyncline (see Figs. 1 and 2).

In the Broken Bow-Benton uplift the basal part of the Stanley shale (Pennsylvanian and Mississippian) and the older shales (Devonian to Cambrian) have been changed in most areas to slates, some of which show differently colored ribbon-like bands on cleavage surfaces that cut across the bedding. In addition, the Crystal Mountain sandstone (Ordovician?), the Blakely sandstone (Ordovician), the Blaylock sandstone (Silurian), and the sandstones of the Stanley shale (Mississippian and Pennsylvanian) and



H. D. Miser Fig. 2 Structure map of Ouachita Mountains the Jackfork sandstone (Pennsylvanian) have been metamorphosed. A noteworthy feature is that the most altered rocks of the Ouachita Mountains occur in an area north of Broken Bow, Oklahoma, and in a large area southwest of Little Rock, Arkansas. Here the slates are more schistose than elsewhere and the Arkansas novaculite resembles fine-grained quartzite. Furthermore, in the Little Rock area the Jackfork sandstone is quartzite and the Stanley shale as a whole is more slaty, is harder, and is blacker than it is in exposures elsewhere. Rocks of the same age and same original character as those of the Broken Bow-Benton uplift are exposed in Black Knob Ridge at the west end of the Ouachita Mountains and in the Potato Hills in the northwest part of the mountains. In these two areas the rocks have not been metamorphosed like those of the Broken Bow-Benton uplift; the shales have not been changed to slates nor have the sandstones been changed to quartzites.

To add to our knowledge of the observed field relations just described Robert P. Bryson of the United States Geological Survey made a petrographic study of specimens that were collected by me from the Jackfork sandstone and Arkansas novaculite at numerous localities in all parts of the Ouachita Mountains. The study disclosed (1) that a wide belt of metamorphosed rocks occupies the central part of the Broken Bow-Benton uplift, (2) that the metamorphism reaches a maximum near Broken Bow, Oklahoma, and Little Rock, Arkansas, and (3) that the metamorphism decreases both northward and southward from the Broken Bow-Benton uplift (see Fig. 1).

The progressive metamorphism of the Jackfork sandstone has resulted in marked changes, namely, (a) the formation of less regular boundaries between the sand grains, (b) the formation of sutures along the contacts between the grains, and (c) the transformation of the interstitial material into coarser-grained material, with the development of new quartz grains, mica, and chlorite. These changes have thus transformed the texture of the Jackfork sandstone to the mosaic texture of quartzite in parts of the area. The specimens of sandstone from the different localities have been classified by Bryson into four stages or groups (designated by Roman numerals) on the basis of the progressive changes recognized during the microscopic examination of thin sections of the specimens. A brief description of these four stages follows:

I. Rounded sand grains; grain boundaries showing no effect on individual grains of their impingement on adjacent grains.

II. Rounded to angular sand grains; some grain boundaries modified by impingement on adjacent sand grains; some small amount of recrystallization of interstitial material to make new quartz, chlorite, and mica.

III. Angular sand grains; many grain boundaries modified and sutures developed on contacts between (a) adjacent original sand grains, (b) new grains of quartz formed by the partial recrystallization of the interstitial material, and (c) original and new grains.

IV. Mosaic of angular grains with sutured boundaries, resulting from modification of the original sand grains and recrystallization of the interstitial material to make new quartz, chlorite, and mica.

Progressive metamorphism of the Arkansas novaculite has destroyed original sedimentary textures and increased the grain size. It has caused the disappearance of radiolaria, detrital quartz grains, and grains of rhombohedral calcite and chalcedony, present in the original novaculite. In addition, dynamic metamorphism has produced fracturing and shearing and preferred orientation of the grains in some of the specimens. The novaculite specimens have been classified by Bryson into four stages (designated by Roman numerals). A description of the four stages is here given:

I. Fine-grained (less than 0.01 mm.) novaculite; variable grain size; radiolaria, chalcedonic grains, detrital quartz sand grains, and rhombohedral grains of calcite are preserved in specimens from many localities; quartz veins may be present.

II. Fine-to medium-grained novaculite; radiolaria and chalcedonic grains not preserved; detrital quartz grains show boundary changes; rhombohedral grains or casts of calcite may remain; quartz veins common.

III. Medium-grained novaculite; detrital quartz and rhombohedral grains or casts not preserved; quartz veins common.

IV. Coarse-grained (more than 0.03 mm.) novaculite.

The oilstone quarries in the Arkansas novaculite near Hot Springs, Arkansas, are in an area where the stage of metamorphism falls in stage II (Fig. 1). Another area showing this stage of metamorphism of the novaculite is a few miles north of Broken Bow, Oklahoma.

Several notable contributions to the petrography and metamorphism of the rocks of the Ouachita Mountains have been made in recent years. These include papers by Goldstein and Reno (1952), Goldstein and Hendricks (1953), Goldstein (1955), Goldstein and Flawn (1958), and Weaver (1958).

Vein quartz

The pursuit of my hobby - the collecting of Arkansas quartz crystals since 1907 - aroused my interest in their geologic story, and I presented in 1943 some observations and conclusions on the quartz crystals and veins and on the relations of the quartz veins to the structure, metamorphism, and metalliferous deposits of the Ouachita Mountains (Miser, 1943).

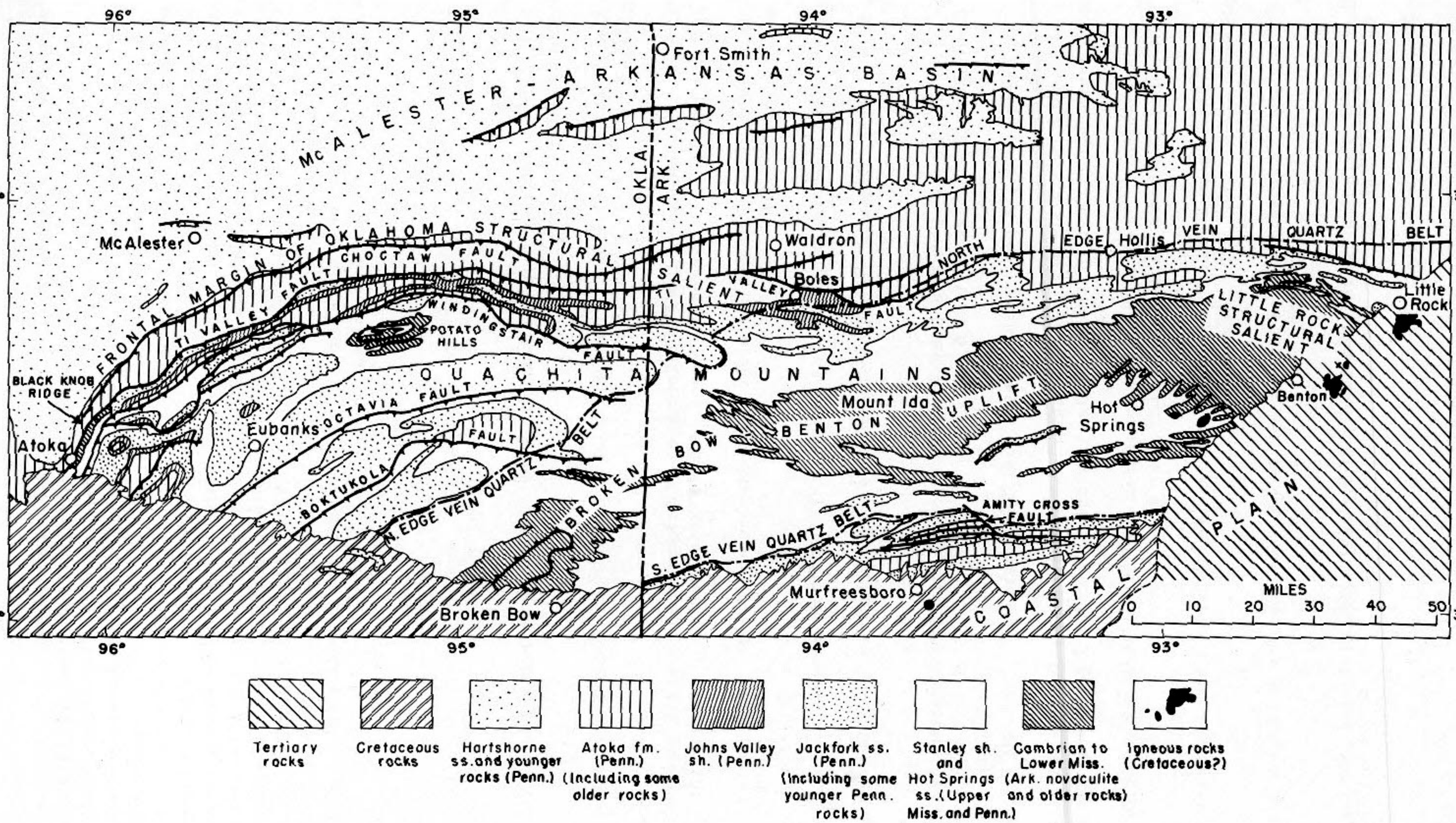
Most of the veins and crystals are restricted to a belt, 30 to 40 miles wide, extending in a west-southwesterly direction from Little Rock, Arkansas, to Broken Bow, northern McCurtain County, Oklahoma, a distance of about 150 miles (see Fig. 3). I do not know how far the vein-bearing rocks extend eastward beyond Little Rock, owing to their concealment by Cretaceous and Tertiary strata, but the vein-bearing rocks extend southwestward beneath Cretaceous strata in southeastern Oklahoma and northeastern Texas, as shown by cuttings from wells.

The veins reach a width of 30 feet in Arkansas and 100 feet in Oklahoma and are most numerous in the central part of the vein quartz belt where they occur in shale, slate, sandstone, and other rocks. But along and near the borders of the vein quartz belt the veins are usually confined to sandstone beds that lie between thick beds of shale. The veins were formed by the filling of open fissures and do not show evidence of much replacement of wall rock.

Associated minerals include calcite (as veins and tabular and rhombohedral crystals), siderite, chlorite, adularia, orthoclase, and dickite. Outside the vein quartz belt dickite, paper-thin quartz veinlets, and small quartz crystals are widely distributed along and near reverse and thrust faults; some occurrences of these minerals along and near such faults are in the Oklahoma-Arkansas coal field.

The vein quartz, the quartz crystals, and the associated minerals of the Ouachita Mountains are hydrothermal and they were formed during the closing stage of Pennsylvanian orogeny (Honess, 1932; Miser, 1943; Engel, 1946, 1952). The time of quartz deposition is inferred largely from the structural relations of the veins to the enclosing rocks and from the deformation of the veins themselves. The deformation of the rocks of the Ouachita Mountains began early in the Pennsylvanian period and continued to the late part of the period. Quartz veins follow faults, fractures, and bedding planes; and vein-filled fractures cut across folds and across slaty cleavage. The veins are not folded but have been faulted, fractured, and crushed at numerous places.

The emplacement of the quartz veins thus seems to have taken place after the more intense deformation of the rocks and after the formation of the slates in the central part of the Broken Bow-Benton uplift. But their emplacement seems to have preceded or accompanied moderate deformation. The suggestions are offered (1) that the regional arching to form the Broken Bow-Benton uplift was the final stage of the Pennsylvanian orogeny and (2) that the arching developed tensional fractures to serve as suitable sites for the quartz veins. Such upwarping during the period of quartz deposition would have produced brecciation and fractures in the quartz that would be healed



H. D. Miser Fig. 3, Geologic map of Ouachita Mountains showing area of occurence of vein quartz partly or wholly by a later deposition of quartz.

The belt of vein quartz thus coincides with the Broken Bow-Benton uplift. Also, the belt coincides with the belt of metamorphosed rocks on that uplift. The widest quartz veins in Arkansas are west of Little Rock and the widest veins in Oklahoma are north of Broken Bow.

Deposits of antimony, lead, copper, zinc, and mercury, are found at places in the Ouachita Mountains, and are associated with a small amount of vein quartz. Some geologists have connected the origin of these deposits with the probable Cretaceous igneous rocks of Arkansas (Hess, 1908; Branner, 1932) but others, among whom I may be counted, believe that these metalliferous deposits and their associated minerals are related to the time of Pennsylvanian structural deformation (Honess, 1932; Stearn, 1935; Reed and Wells, 1938; Gallagher, 1942; Miser, 1943). This conclusion indicates that these metalliferous deposits were formed at the time of the extensive vein quartz deposition.

References

Arbenz, J. K., 1956, Tectonic map of Oklahoma showing surface structural features, Oklahoma Geol. Survey GM-3.

Ardmore Geological Society, 1956, Guide Book, Ouachita Mountain Field Conference, Southeastern Oklahoma.

Ashley, G. H., 1897, Geology of the Paleozoic area of Arkansas south of the Novaculite region, with introduction by J. C. Branner, Proc., Am. Philos. Soc., vol. 36, pp. 217-318.

Bokman, John, 1953, Lithology and petrology of the Stanley and Jackfork formations, Jour. Geology, vol. 61, pp. 152-170.

Branner, G. C., 1929, Geologic map of Arkansas, Arkansas Geol. Survey.

Branner, G. C., 1932, Cinnabar in southwestern Arkansas, Arkansas Geol. Survey Inf. Circ. 2, p. 28.

Cline, L. M., 1955, Regional stratigraphy and history of the Ouachita Mountain area, Oklahoma City Geol. Soc. Shale Shaker, vol. 6, no. 3.

Cline, L. M., 1956, ibid., Amer. Assoc. Petroleum Geologists Bull., vol. 40, no. 2, pp. 427-428.

Cline, L. M., 1956a, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains, Oklahoma, Tulsa Geological Society Digest, vol. 24, pp. 100-106.

Cline, L. M. and Moretti, Frank, 1956, Two measured sections of Jackfork group in southeastern Oklahoma, Oklahoma Geol. Survey, Circular 41.

Cook, K. L., 1956, Regional gravity survey of northeastern Oklahoma and southeastern Kansas, Geophysics, vol. 21, no. 1, pp. 88-106.

Cooper, C. L., 1945, Age relations of Stanley and Jackfork formations of Oklahoma and Arkansas, Jour. Geol., vol. 53, pp. 390-397.

Drake, N. F., 1897, A geological reconnaissance of the coal fields of the Indian Territory, Proc. Am. Philos. Soc., vol. 36, pp. 326-419.

Engel, A. E. J., 1946, The quartz crystal deposits of western Arkansas, Econ. Geology, vol. 41, pp. 598-618.

Engel, A. E. J., 1952, Quartz crystal deposits of western Arkansas, U. S. Geol. Survey Bull. 973, pp. 173-260.

Gallagher, David, 1942, Quicksilver deposits near the Little Missouri River, Pike County, Arkansas, U. S. Geol. Survey Bull. 936-H, pp. 189-219.

Goldstein, August, Jr., 1955, Ouachita facies of Texas and Oklahoma, Tulsa Geological Society Digest, vol. 23, pp. 107-115.

Goldstein, August, Jr., and Reno, D. H., 1952, Petrography and metamorphism of sediments of Ouachita facies, Amer. Assoc. Petroleum Geologists, vol. 36, no. 12, pp. 2275-2290.

Goldstein, August, Jr., and Hendricks, T. A., 1953, Siliceous sediments of the Ouachita facies in Oklahoma, Geol. Soc. America Bull., vol. 64, pp. 421-442.

Goldstein, August, Jr., and Flawn, P. T., 1958, Oil and gas possibilities of Ouachita structural belt in Texas and Oklahoma, Amer. Assoc. Petroleum Geologists Bull., vol. 42, no. 4, pp. 876-881.

Griswold, L. S., 1892, Whetstones and the novaculites of Arkansas, Arkansas Geol. Survey Ann. Rept. for 1890, vol. 3.

Ham, W. E., 1954, Collings Ranch conglomerate, Late Pennsylvanian, in Arbuckle Mountains, Oklahoma, Amer. Assoc. Petroleum Geologist Bull., vol. 38, no. 9, pp. 2035-2045.

Ham, W. E., 1955, Structural geology of Arbuckle Mountain region, Oklahoma City Geol. Soc. Shale Shaker, vol. 6, no. 3.

Ham, W. E., 1956, ibid., Amer. Assoc. Petroleum Geologists Bull., vol. 40, no. 2, pp. 425-426.

Ham, W. E., 1956, Asphaltite in the Ouachita Mountains, Oklahoma Geol. Survey, Mineral Report 30.

Harlton, B. H., 1934, Carboniferous stratigraphy of the Ouachitas with special study of the Bendian, Amer. Assoc. Petroleum Geologists Bull., vol. 18, no. 8, pp. 1018-1049.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains, Amer. Assoc. Petroleum Geologists Bull., vol. 22, no. 7, pp. 852-914.

Harlton, B. H., 1947, The siliceous shale members of the Stanley-Jackfork deposits in the Ouachita Mountains, Oklahoma, with notes on the "Johns Valley" shale, Tulsa Geological Society, Guidebook to field conference in the western part of the Ouachita Mountains in Oklahoma, May 8, 9, and 10, 1947.

Hass, W. H., 1950, Age of lower part of Stanley shale, Amer. Assoc. Petroleum Geologists Bull., vol. 34, no. 7, pp. 1578-1584.

Hass, W. H., 1953, Conodonts of the Barnett Formation of Texas, U. S. Geol. Survey Prof. Paper 243-F, p. 72.

Hass, W. H., 1956, Conodonts from the Arkansas novaculite, Stanley shale, and Jackfork sandstone, Ardmore Geol. Soc., Guidebook, Ouachita Mountain field conference, southeastern Oklahoma, pp. 27-33.

Hendricks, T. A., 1947, Geology of the western part of the Ouachita Mountains in Oklahoma, U. S. Geol. Survey Prelim. Map 66, Oil and Gas Inv. Ser. (text).

Hendricks, T. A., Knechtel, M. M., and Bridge, Josiah, 1937, Geology of Black Knob Ridge, Oklahoma, Amer. Assoc. Petroleum Geologists Bull., vol. 21, no. 1, pp. 1-29.

Hendricks, T. A., Gardner, L. S., Knechtel, M. M., and Averitt, Paul, 1947, Geology of the western part of the Ouachita Mountains of Oklahoma, U. S. Geol. Survey Prelim. Map 66, Oil and Gas Inv. Ser.

Hess, F. L., 1908, The Arkansas antimony deposits, U. S. Geol. Survey Bull. 340, pp. 251-252.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma, Oklahoma Geol. Survey Bull., vol. 32.

Honess, C. W., 1924, Geology of southern Le Flore and northwestern McCurtain counties, Oklahoma, Bur. Geol., Circ. 3, Norman, Oklahoma.

Honess, C. W., 1927, Geology of Atoka, Pushmataha, McCurtain, Bryan, and Choctaw Counties, Oklahoma, Oklahoma Geol. Survey Bull., vol. 40-R.

Honess, C. W., Communication of May 15, 1932, quoted in Branner, G. C., 1932, Cinnabar in southwestern Arkansas, Arkansas Geol. Survey Inf. Circ. 2, p. 28.

Howell, J. V., 1947, The origin of imbricate structure, Tulsa Geological Society, Guidebook to field conference in the western part of the Ouachita Mountains in Oklahoma, May 8, 9, and 10, 1947, pp. 50-52.

Howell, J. V., 1947a, Glossary of structural terms especially applicable to Ouachita type of folding, Tulsa Geological Society, Guidebook to field conference in the western part of the Ouachita Mountains in Oklahoma, May 8, 9, and 10, 1947, pp. 52-53.

Kansas Geological Society, 1931, Guide Book, Fifth Annual Field Conference, Aug. 30 to Sept. 5, 1931.

Lyons, P. L., 1950, A gravity map of the United States, Tulsa Geol. Soc. Digest, vol. 18, pp. 33-43.

Melton, F. A., 1930, Age of the Ouachita orogeny and its tectonic effects, Amer. Assoc. Petroleum Geologists Bull., vol. 14, no. 1, pp. 57-72.

Miller, B. W., 1956, The Geology of the Potato Hills, Pushmataha and Latimer counties, Oklahoma, Tulsa Geological Society Digest, vol. 24, p. 66.

Misch, Peter, and Oles, K. F., 1957, Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: preliminary report, Amer. Assoc. Petroleum Geologists Bull., vol. 41, no. 8, pp. 1899-1905.

Miser, H. D., 1926, Geologic map of Oklahoma, U. S. Geol. Survey.

Miser, H. D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas, Oklahoma Geol. Survey Bull., vol. 50.

Miser, H. D., 1934, Carboniferous rocks of Ouachita Mountains, Amer. Assoc. Petroleum Geologists Bull., vol. 18, no. 8, pp. 971-1009.

Miser, H. D., 1934a, Relation of Ouachita belt of Paleozoic rocks to oil and gas fields of Mid-Continent region, Amer. Assoc. Petroleum Geologists Bull., vol. 18, no. 8, pp. 1059-1077.

Miser, H. D., 1943, Quartz veins in the Ouachita Mountains of Arkansas and Oklahoma, their relations to structure, metamorphism, and metalliferous deposits, Econ. Geol., vol. 38, no. 2, pp. 91-118.

Miser, H. D., 1954, Geologic Map of Oklahoma, U. S. Geol. Survey and Oklahoma Geol. Survey.

Miser, H. D., and Honess, C. W., 1927, Age Relations of the Carboniferous rocks of the Ouachita Mountains of Oklahoma and Arkansas, Oklahoma Geol. Survey Bull., vol. 44.

Miser, H. D., and Purdue, A. H., 1929, Geology of the De Queen and Caddo Gap quadrangles, Arkansas, U. S. Geol. Survey Bull. 808.

Moore, R. C., 1934, The origin and age of the boulder-bearing Johns Valley shale in the Ouachita Mountains of Arkansas and Oklahoma, Amer. Jour. Sci., vol. 27, pp. 432-453.

Oklahoma Academy of Science and Sigma Gamma Epsilon (Gamma Chapter), 1952, Road log, Geological field trip in the eastern part of the Ouachita Mountains in Oklahoma, April 26 and 27, 1952.

Oklahoma City Geological Society, 1950, Guide Book, Field conference in the eastern part of the Ouachita Mountains in Oklahoma, Nov. 4 and 5, 1950.

Pitt, W. D., 1955, Geology of the core of the Ouachita Mountains of Oklahoma, Oklahoma Geol. Survey, Circ. 34.

Powers, Sidney, 1928, Age of the folding of the Oklahoma Mountains, Geol. Soc. America Bull., vol. 39, pp. 1031-1072.

Purdue, A. H., 1909, The slates of Arkansas, Arkansas Geol. Survey.

Purdue, A. H., and Miser, H. D., 1923, Description of the Hot Springs district, U. S. Geol. Survey Geol. Atlas U. S., Hot Springs Folio 215.

Rea, H. C., 1947, The erratic boulders in the Johns Valley shale of the western Ouachita Mountains, Oklahoma, Tulsa Geological Society, Guidebook to field conference in the western part of the Ouachita Mountains in Oklahoma, May 8, 9, and 10, 1947, pp. 47-49.

Reed, J. C., and Wells, F. G., 1938, Geology and ore deposits of the southwestern Arkansas Quicksilver district, U. S. Geol. Survey Bull. 886, pp. 51-53.

Reinemund, J. A., and Danilchik, Walter, 1957, Preliminary geologic map of the Waldron Quadrangle and adjacent areas, Scott County, Arkansas, U. S. Geol. Survey Map OM-192, Oil and Gas Inv. Ser.

Stearn, N. H., 1935, Discussion of Tech. Pub. 612, Amer. Inst. Mining and Met. Eng., Trans., vol. 115, pp. 244-246.

Taff, J. A., 1902, Description of the Atoka Quadrangle, U. S. Geol. Survey Geol. Atlas, U. S., Atoka Folio 79.

Tomlinson, C. W., and Pitt, W. D., 1955, Recent studies in the Ouachita Mountains, Tulsa Geological Society Digest, vol. 33, pp. 89-97.

Tulsa Geological Society, 1947, Guidebook, Field conference in the western part of the Ouachita Mountains in Oklahoma, May 8, 9, and 10, 1947.

Ulrich, E. O., 1927, Fossiliferous boulders in the Ouachita "Caney" shale and the age of the shale containing them, Oklahoma Geol. Survey Bull., vol. 45.

van der Gracht, W. A. J. M. van Waterschoot, 1931, Permo-Carboniferous orogeny in South-Central United States, Amer. Assoc. Petroleum Geologists Bull., vol. 15, no. 9, pp. 991-1057.

Wallis, B. F., 1915, The geology and economic value of the Wapanucka limestone of Oklahoma, Oklahoma Geol. Survey Bull., vol. 23.

Weaver, C. E., 1958, Geologic interpretation of argillaceous sediments. Part II. Clay petrology of Upper Mississippian-Lower Pennsylvanian sediments of central United States, Amer. Assoc. Petroleum Geologists Bull., vol. 42, no. 2, pp. 272-309.

White, David, 1934, Age of Jackfork and Stanley formations of Ouachita geosyncline, Arkansas and Oklahoma, as indicated by plants, Amer. Assoc. Petroleum Geologists Bull., vol. 18, no. 8, pp. 1010-1017.

White, David, 1937, Fossil plants from the Stanley shale and Jackfork sandstone in southeastern Oklahoma and western Arkansas, U. S. Geol. Survey Prof. Paper 186-C, pp. 43-67.

STRUCTURE OF THE FRONTAL BELT OF THE OUACHITA MOUNTAINS

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Abstract

The structure of the frontal belt of the Ouachitas is dominated by faulting. The faulting in general consists of a complex set of reverse faults roughly parallel with the Ouachita Mountain front and a related set of cross faults. Most of the reverse faults appear to dip at high angle and to have a horizontal component of movement that is of the order of magnitude of the vertical movement. Others, such as the Pine Mountain and Windingstair faults appear to have had greater horizontal movement than vertical movement, and at least locally appear to dip at moderate angles. The Ti Valley fault and some minor faults appear to have a low angle of dip. The cross faults are of two types. In the northern part of the area they are characterized by strike-slip movement. South of the Windingstair fault the movement was dominantly upward. The minimum amount of movement on the reverse faults in the frontal belt appears to have been in excess of 50 miles with the greatest part of that movement concentrated on the Ti Valley, Windingstair, and Pine Mountain faults. More or less simultaneous deformation seems to have occurred in an extreme frontal block, the block between the Ti Valley and Windingstair faults, and the block south of the Windingstair fault, with the deformation culminating in movement on the Ti Valley and Windingstair faults. Several lines of evidence suggest that the direction of movement was generally northward and that greater movement occurred in the eastern part of the area than in the western part. Incompetent shale zones constituted gliding planes along which thrust movement took place, with the principal ones being the Womble shale, Springer formation, Caney shale, Stanley shale, and Johns Valley shale. One can postulate from indirect evidence the existence of an early period of faulting along the north margin of a late Mississippian - early Pennsylvanian geosyncline. However, the structural development of the frontal Ouachitas started in Atoka time and continued until middle Pennsylvanian time and possibly as late as early Permian time.

Introduction

The structure of a deformed geosyncline is made up of many interrelated features. When one discusses the structure of only part of a deformed geosyncline it must be realized that, even if accurate in itself, the discussion cannot be comprehensive. This is true of the Ouachita Mountains. A discussion of the structure of the frontal belt may not satisfy a student of the southern part of the Ouachitas. A discussion including the southern Ouachitas with both the frontal belt and the known subsurface would still leave much to be desired by the theorist who is aware of the possible significance of the deeper part of the Ouachita structural complex buried beneath the Gulf Coastal plain and not yet opened to study by drilling. Therefore, I wish to emphasize that this discussion deals with the structure of the frontal belt of the Ouachita Mountains in Oklahoma and is in no sense a complete discussion of the Ouachita Mountain structural province. The discussion will attempt to analyze the known features of the structure, particularly those shown on United States Geological Survey Oil and Gas Investigations Preliminary Map 66 (Hendricks and others, 1947), Figs. 3 and 4, and to deduce the structural history from that analysis.

Discussion

Dip of faults. - A fundamental question is that of the dip of the fault planes. This is

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a particularly difficult question because exposed fault planes are rare and dips must be inferred. Evidence indicates that many of the reverse faults dip at high angles. The general tendency of the faults to maintain relatively straight traces parallel to belts of outcrop of beds dipping at high angle suggests that the faults themselves dip at high angles. At a number of places in the belt adjacent to the Choctaw fault, belts of outcrop of Wapanucka limestone dipping at high angle are terminated by a fault only at a point where a change of strike and dip occurs. These changes in dip are such that an anticlinal nose is present in a bed that terminates against a fault in the updip direction, as is the case in the termination of the Wapanucka limestone against the Choctaw fault near the center of T. 1 N., R. 12 E. On the other hand, a plunging synclinal trough is present at a number of places where the bed terminates against a fault situated on the downdip side, such as the westward termination of the Wapanucka limestone in secs. 1 and 11, T. 2 N., R. 13 E. This suggests that the fault dips at an angle approximately equal to that of the bed whose outcrop it parallels.

In sec. 1, T. 2 N., R. 17 E., field evidence indicates that the reverse fault near the axis of the Jackfork Mountain syncline dips at an angle of 45° - 60°. The base of the Jackfork sandstone has been moved downdip a slightly greater distance by erosion of the upthrown side than has the associated reverse fault. Since the base of the Jackfork dips about 45° the fault must dip at a slightly higher angle.

In addition, the segment of northward dipping but essentially vertical Jackfork sandstone that lies south of the reverse fault is about 1000 feet wide on the downthrown side and is absent on the upthrown side. This also would be consistent with a south dip on the fault of more than 45° and definitely less than 60°.

Mapping of the quarry at Stringtown in 1935 showed that a minor fault exposed on the floor of the quarry passed into its wall at each end. At one end the fault dipped northward at an angle of more than 70°, and at the other end it dipped southward at about the same angle.

Where the Jackfork Creek fault crosses the south line of sec. 11, T. 1 N., R. 15 E., on the south side of Jackfork Mountain beds of Jackfork sandstone dipping southward were observed north of and some 50 feet higher than southward dipping beds of Stanley shale. The two outcrops were separated by a short covered interval in which the reverse fault is present. The minimum angle of dip of the fault that would permit it to pass between the two outcrops was approximately 60°.

In contrast with the general suggestion that the reverse faults dip at angles of about 60° is the exposure of a minor fault in a road cut in the SE 1/4 sec. 15, T. 2 S., R. 11 E., for a distance of more than 100 feet. At the west end of the exposure the fault dips at an average angle of about 20° but throughout most of its exposure the fault is approximately horizontal.

From the SE 1/4 sec. 22, T. 1 N., R. 14 E., to the east side of sec. 31, T. 2 N., R. 15 E., the trace of the Windingstair fault has the form of a salient that extends about a half-mile northwestward. This salient coincides very closely with a syncline in the overthrust block in which the beds near the fault dip at angles averaging about 20°. The parallelism between the trace of the Windingstair fault and the outcrop of the contact between the Stanley shale and the Jackfork sandstone suggests that the dip of the fault is of the same order of magnitude as the dip of the contact or approximately 20°.

At two places, one in the easternmost tier of sections in Tps. 1 and 2 N., R. 13 E., and the other across R. 17 E., in Tps. 3 and 4 N., the Pine Mountain fault truncates several underlying fault blocks at angles in map plan of 45° to 70°. At these places, the dip on the Pine Mountain fault at its outcrop appears to be steep and the movement on the fault seems to be of strike-slip type. The faults and strata cannot be matched on the opposite sides of the strike-slip zones. It appears probable, therefore, that four to five miles of strata have been overridden by the eastern part of the Pine Mountain fault along each strike-slip zone and that at a relatively shallow depth the dip of the Pine Mountain fault decreases to a low angle.

In the west central part of T. 2 N., R. 14 E., an area of about three square miles is underlain by rocks of a facies unlike those present on at least three sides of the area. The rocks exposed to the south and southeast are not diagnostic and could be of the same facies. This area has been interpreted as a klippe or remnant separated by erosion from the structural block overlying the Ti Valley fault to the southeast. It is possible, however, that the faults bounding this area on its east and west sides continue southward and connect with the Ti Valley fault trace and that a salient extends about three miles northwest of the Ti Valley fault rather than that an isolated klippe is present. Regardless of which interpretation is correct, the relationships indicate that the Ti Valley fault dips at an extremely low angle at this locality.

Between Atoka, Oklahoma and Denison, Texas (Fig. 1) a reentrant extends southeastward in the frontal fault of the Ouachita structural belt (Hendricks, 1940). The central part of this reentrant has basement rocks present beneath Cretaceous beds of the Coastal Plain. The basement uplift is flanked by strata of the facies characteristic of the Arbuckle Mountains. Overlying a curving fault are strata of Ouachita facies. A reentrant of this magnitude indicates very strongly that the fault dips at a low angle and that it is curved or folded.

In the Waldron quadrangle of western Arkansas, Reinemund and Danilchik (Reinemund and Danilchik, 1957) have mapped the Johnson Creek fault as a low-angle fault approximately parallel to the bedding of Atoka strata in the Black Fork syncline, which rests upon the fault. Their mapping shows the fault as present on three sides of the syncline. The fault is identified by the truncation of sandstone beds in both the overlying and underlying blocks. On the east end of the syncline, the fault appears to be essentially parallel to the bedding of the overlying and underlying formations, which dip at angles of 20 to 25 degrees.

H. D. Miser (Miser, 1929) has mapped the Potato Hills area of Oklahoma as an intricately folded area encircled by the trace of a thrust fault or window. B. W. Miller (Miller, 1956) reported that detailed mapping showed that Miser's interpretation is consistent with field relationships. However, the trace of the fault surrounding the window cannot be mapped for a distance of about two miles in the northwestern part of the Potato Hills where Stanley shale is present on both sides of its supposed trace. It is possible that the overriding block may have been breached by erosion at that locality and the Potato Hills may be a deep embayment in a major fault. Either of these interpretations would be consistent with low-angle faulting accompanied by folding of the fault plane. Misch and Oles (Misch and Oles, 1957) have recently questioned this interpretation and have suggested that the actual situation may be that the Potato Hills area is an anticlinorial uplift. Other workers who have done very detailed mapping believe that the Potato Hills comprise a faulted anticline. Arguments based on theoretical considerations have been advanced by proponents of each idea. In general these arguments are based on interpretations of features such as those listed below:

1. Variations in the tightness of folds from the inner part of the window to the outer part.

2. Variations in inclination of minor faults and axial planes of folds.

3. Breaching by minor faults outside the window of the folds that represent the maximum elongation of the Potato Hills.

4. Assumed matching of fold axes from the inner to the outer part of the window.

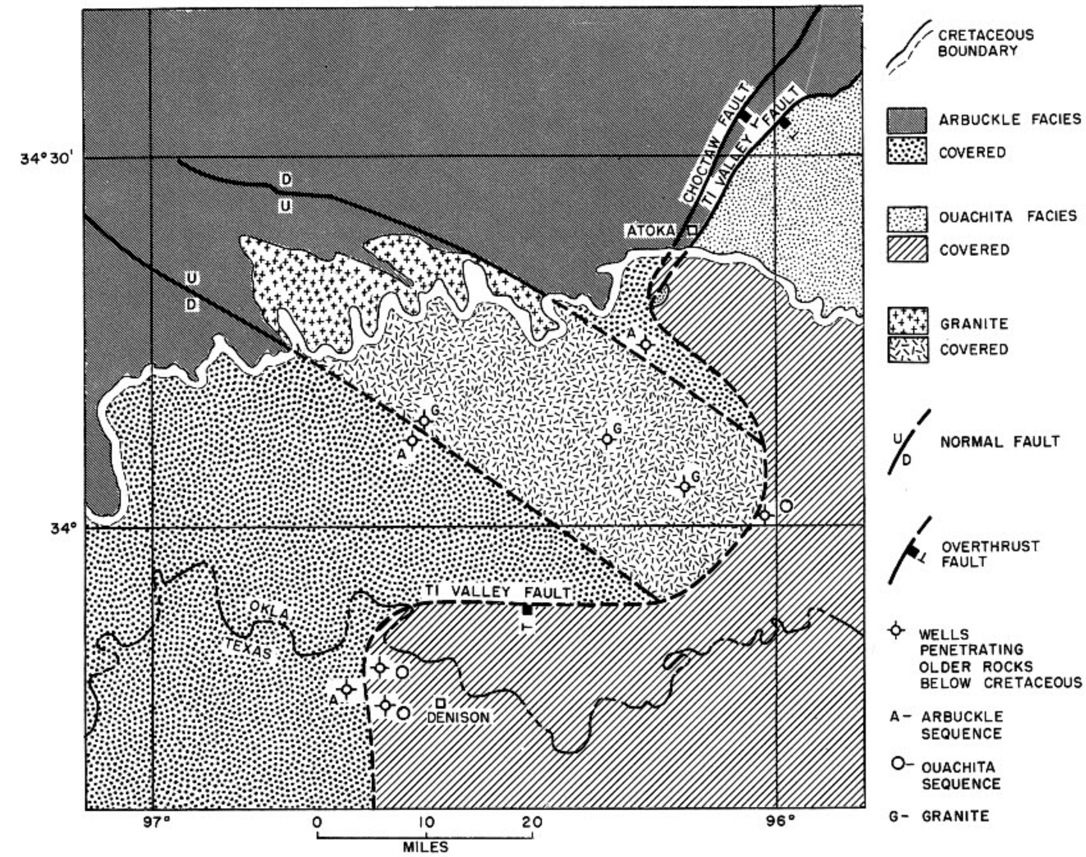
5. Minor depositional differences between strata inside the window and the same horizons outside the window.

Some of the arguments are very intriguing, but no conclusive evidence rules out Miser's original interpretation of the presence of a window.

Nature of the cross faults. - The cross faults appear to be of two general types. South of the Windingstair fault they seem to have undergone primarily vertical movement, whereas north of the Windingstair fault they are of strike-slip type.

A cross fault with about 500 feet of throw cuts the axis of a syncline near the common corners of Tps. 1 and 2 N., Rs. 14 and 15 E. The syncline is essentially symmetrical and the axis of the syncline is not displaced appreciably by the cross fault. This suggests that the movement on the cross fault was essentially vertical.

FIG. 1 SKETCH MAP OF REENTRANT IN TI VALLEY FAULT BETWEEN ATOKA AND DENISON



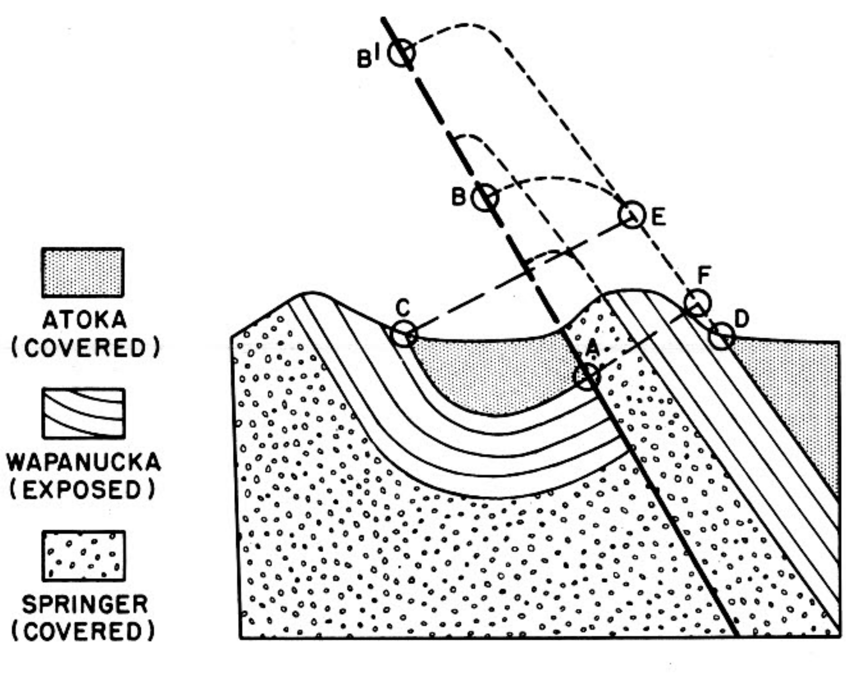


FIG. 2 DISPLACEMENT RELATIONSHIPS ON A REVERSE FAULT

The outcrop pattern in the pie-shaped fault block in secs. 9 and 16, T. 1 N., R. 15 E., could only have been formed by uplift of the pie-shaped wedge followed by erosion that caused downdip migration of the outcrops in the up-faulted block. The absence of crumpling and distortion in the beds involved in the entire complex of cross faults that both trend and intersect at various angles in the northeast corner of T. 1 N., R. 15 E., suggests that the movement on many of the faults was essentially normal to the bedding and that little internal compression occurred in the blocks.

The fact that the principal cross faults are terminated sharply by an underlying reverse fault rather than passing into a fold suggests that the movement on the cross faults is a function of differential movement of individual segments in the overriding block of the reverse fault. A considerable range is represented in the age of the beds at the base of the blocks bounded by cross faults, both between adjacent blocks and within a single block. This suggests that the underlying reverse fault cut the bedding at an angle and that some differential strike-slip movement occurred so that blocks having beds of different age at the base are now opposite each other. In some instances, however, the reverse faults appear to have overridden an uneven surface and a part of the movement on the cross faults seems to have occurred roughly normal to the plane of the reverse fault as an adjustment to that uneven surface.

The cross faults that lie north of the Windingstair fault in Tps. 12 and 13 E., are of a different nature. Two sets of small cross faults in secs. 2 and 10, T. 2 N., R. 13 E., reveal fracturing and strike-slip offset of competent beds. The small cross fault that cuts Stanley and Jackfork strata in sec. 1, T. 1 S., R. 12 E., passes southeastward into an S-fold. This suggests that the movement was partly strike-slip.

At the cross fault in secs. 20 and 29, T. 1 N., R. 13 E., the beds on the east side appear to have been moved relatively northward. One basis for this opinion is the fact that the Atoka formation overlaps the Johns Valley shale and onto the upper part of the Jackfork sandstone in southward dipping beds on both sides of the fault but about a mile farther north on the east side. Since this situation is unique to these localities it appears logical to use it as an indication of the amount of offset. In addition, an anticlinal axis on the west side of the fault lies along the prolongation of a synclinal axis on the east side and a syncline farther north on the west side is adjacent to a belt of southward dipping beds that could represent the south limb of an anticline. These relationships could not have resulted from vertical movement on the fault but are consistent with a northward movement of the east side of about a mile.

The cross fault that extends southward from sec. 13, T. 1 N., R. 13 E., to sec. 36 has on its west side a syncline involving more than 5000 feet of beds from lower Jackfork to Atoka age on its north flank and overturned Atoka beds on its south flank. No evidence of faulting could be detected for a distance of 2 1/2 miles. On the east, however, three thrust slices involving Jackfork, Johns Valley shale, and Atoka beds are present. The beds dip southward in the southernmost two of the fault slices and presumably are slices faulted off the south limb of an anticline that has been overridden. The northernmost belt of Atoka beds may lie on the north limb of the syncline that is present west of the fault. It appears that compressive forces west of the fault may have developed an overturned syncline while on the east side even greater compression ruptured both the syncline and the anticline which originally lay south of the fault.

Amount of movement on faults. - Questions are frequently raised as to the amount of movement that has taken place along any one fault plane and the total amount of movement in the deformation of the Ouachita Mountains. Only under the most unusual circumstances would it be possible to determine the absolute amount of movement on a fault. In the case of reverse faulting, whether it be low-angle thrusting or high-angle slice faulting, the determination of total movement is particularly difficult, as it is seldom possible to match any unit in the overriding block with the same unit at an initially contiguous point in the block that has been overridden. Even the amount of apparent stratigraphic displacement on a reverse fault cannot be assumed to represent a minimum amount of movement on the fault unless it is measured between units that are actually touching across the fault. When identical horizons can be established only at some distance from the fault on its two sides, errors may be introduced due to undetected changes in dip in the unidentified beds. In Fig. 2 the actual displacement on the fault may be the distance A-B or A-B' or some greater or lesser distance. The actual stratigraphic displacement is the distance A-F. The apparent stratigraphic displacement is the distance C-E. This apparent stratigraphic displacement is greater than the minimum displacement required to explain the known relationships because it includes the two belts of unobserved dip reversal between C and A and B and D. Many of the faults in the block southeast of the Choctaw fault locally truncate minor synclines in the lower block and minor anticlines in the upper block. Therefore, it seems probable that the conditions shown in Fig. 2 apply in general to these faults. If so, the minimum displacement on each slice fault from the Choctaw fault southeastward to the Pine Mountain fault is of the order of magnitude of a half mile reduced by a small amount to provide for the change in dip in unexposed beds near the fault.

On this basis of estimation, the apparent minimum cumulative stratigraphic displacement on the Choctaw fault and the slice faults between the Choctaw and Katy Club faults is about three miles.

The Katy Club fault overrides or merges with two faults north of it in part of the area. The apparent stratigraphic displacement on each is 1500 to 2000 feet, so the minimum movement on the Katy Club fault combined with the two that it overrides appears to be about a mile.

The cumulative apparent stratigraphic displacement on the various slice faults between the Katy Club fault and the Pine Mountain fault is about 3 1/2 miles.

The minimum cumulative displacement on the Pine Mountain fault appears to be about 10 miles on the basis of the fault blocks overridden by it in the two belts of pronounced strike-slip movement.

The cumulative apparent stratigraphic displacement on the slice faults between the Pine Mountain fault and the Ti Valley fault is about three miles.

The klippe or salient of the Ti Valley fault in the central part of T. 2 N., R. 14 E., indicates a minimum displacement of about four miles. The pronounced change in facies that takes place across the Ti Valley fault and the amount of local curvature of the fault trace indicate that the movement is considerable. The Caney and Springer formations are present north of the Ti Valley fault and beds equivalent in age to the part of the Johns Valley shale from which a Morrow fauna has been identified are absent, whereas the Stanley, Jackfork, and Johns Valley formations are present in the same interval south of the fault. In addition, this sequence of beds expands from a thickness of about 1500 feet north of the fault to more than 10,000 feet within a short distance south of it. These pronounced changes suggest that the movement is great. The work of Maher and Lantz (Maher and Lantz, 1953) north of the Ouachita Mountains in Arkansas reveals a similar change in strata in this general zone that takes place in a distance of about 50 miles in a relatively unfaulted region. However, the control for that change is a set of wells aligned at a small angle to the most likely trend of depositional strike. If one assumes no change in the depositional strike in the set of wells, the amount of change that takes place across the Ti Valley fault can be inferred to take place in Arkansas in a distance as little as 20 miles. The possibility of change in depositional strike and in rates of maximum depositional change between the frontal Ouachitas and the area studied in Arkansas cause this figure to indicate only the order of magnitude of foreshortening. The depth and width of the reentrant in the Ti Valley fault south of Atoka suggest possible movement of the order of magnitude of 20 miles regardless of what the direction of movement may be.

The structure of the belt between the Ti Valley and Windingstair faults is so complex that estimates of displacement are especially difficult. The only criterion that can be applied is that the cumulative stratigraphic displacement is of the order of magnitude of four miles. Attention is called to the fact that this does not take into account any foreshortening by folding, which is indicated to be considerable in this belt by the presence of a number of overturned synclines.

The width of blocks that are overridden by the Windingstair fault reaches approximately two miles and the apparent stratigraphic displacement across it at many places is about a mile. Changes in the thickness and character of the Jackfork sandstone across the Windingstair fault are no greater than changes in the Jackfork that take place in a distance of a few miles in areas both north and south of the fault. On the basis of these criteria, the minimum displacement across the Windingstair at its outcrop is considered to be of the order of magnitude of three miles.

The cumulative stratigraphic displacement on the Weathers, Jackfork Mountain and Jackfork Creek faults is about 2 1/2 miles. In addition, throughout part of its course the Jackfork Creek fault completely overrides the south limb of the Jackfork Mountain syncline, which is about a mile wide. Also, in secs. 29 and 30, T. 3 N., R. 19 E., two slivers of Arkansas novaculite have been dragged upward on the Jackfork Creek fault and now lie between beds of Stanley shale and about 10,000 feet higher in the section. This suggests that the movement there was at least 2 miles. Therefore, the minimum cumulative movement on this group of faults appears to be at least 4 1/2 miles. In the event that this group of faults merges at depth with the Windingstair fault the minimum cumulative movement on the sole fault would be 7 1/2 miles. A figure of this magnitude is consistent with the minimum displacement that would be required if this sole fault encircles the window that H. D. Miser (Miser, 1929) has postulated in the Potato Hills.

On the basis of the evidence available in the frontal part of the Ouachita Mountains, a minimum cumulative total northward movement on reverse faults in excess of 50 miles can be estimated, as shown in the following table:

AMOUNT OF MOVEMENT ON FAULTS

FAULT	MINIMUM DISPLACEMENT IN MILES	BASIS OF ESTIMATE
Choctaw	0.5	Stratigraphic displacement
Choctaw to Katy Club	2.5	Cumulative stratigraphic displacement
Katy Club	1.0	Stratigraphic displacement, blocks overridden, facies change
Katy Club to Pine Mountain	3.5	Cumulative stratigraphic displacement
Pine Mountain	10.0	Blocks overridden
Pine Mountain to Ti Valley	4.0	Cumulative stratigraphic displacement
Ti Valley	20.0 ±	Facies change, klippe, reentrant between Atoka and Denison
Ti Valley to Windingstair	4.0	Cumulative stratigraphic displacement
Windingstair	3.0 ±	Stratigraphic displacement, blocks overridden, possible window
Windingstair to Jackfork Creek	4.5	Stratigraphic displacement, blocks overridden, dragged slivers
	53.0	

It is interesting to note that Miser (Miser, 1929) estimated, on the basis of the arcuate protrusion of the Oklahoma salient of the Ouachitas, that crustal shortening of an order of magnitude of 70 miles took place.

Sequence of faulting. - Let us consider the sequence of faulting. How does one determine such a sequence? In this area the evidence is confined to the truncation of an older fault by a younger fault, the overriding of older fault blocks and segments of folds by a younger fault, and the offsetting of an older fault by a younger fault. The application of these criteria reveal that three definite zones of faulting are present in the frontal Ouachitas. These are the zone from the Choctaw fault to the Ti Valley fault, the zone between the Ti Valley and Windingstair faults, and the zone south of the Windingstair fault.

The Choctaw fault constitutes the front of the Ouachita Mountains from the Coastal Plain south of Atoka to western Arkansas. It separates steeply northward dipping beds of Atoka age on the north from steeply southward dipping beds of Springer, Wapanucka and Atoka age, generally from 1000 to 5000 feet lower in the stratigraphic column, on the south. The relationship is that of a major anticline or anticlinal zone that has been faulted with the south side moving upward and overriding the crest. South of the Choctaw fault is a belt of imbricate structure (Howell, 1947) consisting of a series of fault slices in which older beds on the south side of each fault are in contact with younger beds on the north. It appears from these relationships that these represent repetitive slices faulted off from and moved upward on the south flank of a major anticlinal uplift which was complicated by only minor folding.

I know of no way to determine the relative ages of these slices. It may be speculated that the most southerly may have formed first and the next fault to the north may have developed after the mass added to the overriding block by the vertical movement had become great enough to cause the transmission of force through the fault plane and into the lower block. However, the Choctaw fault may have formed first and as the mass of the overriding block became too great to be moved the overriding block may have ruptured and the next fault to the south may have formed. On the basis of observed relationships the Choctaw fault, Katy Club fault, and their related slice faults must be considered pene-contemporaneous.

The Pine Mountain fault overrides three fault blocks that lie northwest of it in the eastern part of R. 13 E., and four fault blocks of the Choctaw and Katy Club group that lie north of it in Rs. 16 and 17 E. This indicates that the Pine Mountain fault is younger than the Choctaw and Katy Club blocks. Several faults in the overlying block terminate against the Pine Mountain fault both at their east end and west end. However, the overlying faults tend to parallel the Pine Mountain fault but with a lesser curvature. This suggests that the various fault slices were formed before the Pine Mountain fault suffered northward movement on the east side of the tear zones and that their planes were bent by that movement. These relationships indicate that the Pine Mountain fault is the youngest of a series of slice faults of which it is the most northerly member. The Ti Valley fault overrode the Pine Mountain fault and several of the faults south of the Pine Mountain fault and is therefore considered younger than all of the faults that lie north of it. The belt between the Ti Valley fault and the Windingstair fault is broken by a complex set of reverse faults that generally trend parallel with the Ti Valley fault and a set of tear faults on each of which the east side appears to have moved northward. Several of these faults are terminated by the Ti Valley fault although several of them either terminate or offset reverse faults in the belt. On the other hand, overlying reverse faults cut off each of the tear faults at their south end. These relationships suggest that the faults in the belt south of the Ti Valley fault developed successively from the Ti Valley fault southward to the Windingstair fault. Since both the Windingstair fault and the Ti Valley fault appear to be younger than the faults between them, it seems probable that movement on the Ti Valley fault took place throughout the entire period of deformation of the belt south of it and ended at approximately the time of movement on the Windingstair fault.

The structural pattern south of the Windingstair fault is markedly different from that to the north. It is characterized by widely spaced reverse faults developed in the thick and relatively incompetent Stanley shale with most of the fault blocks including a belt of outcrop of the thick and relatively competent

Jackfork sandstone which is broken by numerous cross faults on which the movement appears to have been largely upward. The cross faults appear to be adjustment faults that developed in the competent beds of a thrust block that either moved over an irregular surface or suffered differential horizontal movement or some combination thereof. In that event the movement on a reverse fault and on the cross faults in the overlying block was contemporaneous. On the basis of that premise, the termination on the south by the Jackfork Mountain fault of all except two of the cross faults in the block above the Weathers fault from the western part of R. 18 E., across R. 17 E., indicates that movement started on the Weathers fault in that belt and was followed by the development of the Jackfork Mountain fault. The Jackfork Creek fault cuts off the Jackfork Mountain fault in the western part of R. 17 E., and therefore appears to be younger. However, from that point westward the cross faults of the Weathers fault block offset the Jackfork Creek fault, which indicates that generally in that part of the area the cross faults to the north and the related Weathers fault are younger than the Jackfork Mountain and Jackfork Creek faults. These relationships indicate that movement either started earlier on the eastern part of the Weathers fault than on its western part or recurrent movement took place on the western part. The Weathers fault is offset by nearly all of the cross faults in the block above the Windingstair fault, whereas they are terminated by the Windingstair fault, thus indicating that the Windingstair fault is younger. That the cross faults are contemporaneous with or younger than the underlying reverse faults is indicated by the fact that at no place is the underlying fault offset by a cross fault. On the contrary, at the few places where the relationship is apparent the cross fault terminates against the underlying fault.

A summary of the information provided by consideration of the relative ages of the faults of the frontal part of the Ouachitas is as follows:

Northern Part of Area

1. Faulting in northernmost block
2. (a) Faulting in Pine Mountain block
   (b) Strike-slip and overthrusting Pine Mountain fault
3. Deformation of Ti Valley-Windingstair block
4. Ti Valley fault

Southern Part of Area

1. Faulting north to south in eastern part
2. (a) Faulting south to north in western part
   (b) Renewed faulting in eastern part
3. Deformation of Ti Valley-Windingstair block
4. Windingstair fault

In each of these parts of the area stages 2 and 3 may have been essentially contemporaneous. The overall picture that seems to emerge is the more or less simultaneous deformation of a frontal block, a block between the Ti Valley and Windingstair faults, and the block south of the Windingstair fault with the entire period of deformation culminating in the final movement on the Ti Valley and Windingstair faults.

Direction of movement. - The general configuration of the Oklahoma structural salient suggests that the strata within the salient have moved northward or north-northwestward (Miser, 1929). Some details of the structure in the frontal part of the Ouachitas tend to confirm this.

In general the reverse faults, anticlines, and synclines trend east-west in the eastern part of the frontal belt. In the western part they swing to a trend of west-southwest while the actual front of the structural belt swings even more or to a south-southwest trend. Southward-dipping sequences of beds that represent either the south limb of anticlines or the north limb of synclines are many times more common than northward dipping sequences. Where northward dipping sequences are present they are almost invariably nearly vertical or overturned so that the apparent dip is southward. The axial planes of both anticlines and synclines are almost without exception inclined toward the south. All of these features indicate overthrusting from the south as was pointed out by C. W. Honess in 1924 (Honess, 1924).

It has been previously pointed out that on tear faults with pronounced strike-slip movement the movement appears to have been northward or north-northwestward on the east side, which is toward the center of the salient. In the two parts of the Pine Mountain fault where a strike-slip component appears important similar northward movement on the east side apparently occurred.

Two northeastward plunging anticline and syncline fold couples are present in Black Knob Ridge in the extreme southwestern part of the Ouachitas. These fold couples overlie less pronounced but similar plunging curvatures of the underlying Ti Valley fault. These fold couples appear to have been produced by northward movement in the overthrust or eastern fault block. The strata in Black Knob Ridge consist of a thin sequence of incompetent Womble shale above the Ti Valley fault, overlain by about 1500 feet of Ordovician to Mississippian beds of which about 75 per cent are relatively competent, in turn overlain by the incompetent Stanley shale. During the early part of the movement, the beds in the overthrust block probably lay parallel with the fault plane and had, therefore, the form of an incipient plunging anticline-syncline couple. As a northward component of force on the east side was transmitted along the bedding of the competent units, it is probable that at the point of change in strike the force tended to move the east flank of the incipient anticline northward and produce drag on the west flank, thus accentuating the incipient anticline and syncline and producing the existing fold couples. Such movements produced, between the fault plane and base of the lowest competent bed in the fold couple, a zone of reduced pressure into which the incompetent Womble shale has been squeezed. The Womble shale is highly contorted and slickensided in those areas, although elsewhere in Black Knob Ridge it shows minor contortion.

Several minor reverse faults break the strata in Black Knob Ridge near Stringtown. These faults trend eastward and the strata on the south side have been thrust upward in relation to the strata on the north side. The faults lie inside an arc formed by the Ti Valley fault. These minor faults appear to have been produced by a component of force transmitted northward on the east side of the Ti Valley fault along the strike of the competent beds. The block overridden by the Ti Valley fault is believed to have formed a buttress that retarded the northward force and caused these relief faults to form.

All of these features indicate that the movement in the deformation of the frontal belt of the Ouachitas in Oklahoma was essentially northward, and that the central part of the belt suffered more movement than the western part.

Period of faulting during deposition. In considering faulting in the Ouachita Mountains, one must also take into account some remarkable depositional features. Beds of shale that contain boulders of limestone, sandstone, and shale that are derived from formations of Ordovician to Mississippian age of a facies dissimilar to that present in normal stratigraphic sequence below the boulder bearing shales themselves are known to be present at two horizons in the Stanley-Jackfork sequence and several horizons in the Johns Valley shale (Harlton, 1938, and Hendricks, 1947). The boulders in the Johns Valley shale are of a facies characteristic of the Arbuckle Mountains, the McAlester Basin, and the Boston Mountains. They are not similar to beds of the same age in the Ouachita Mountains south of the Ti Valley fault. It appears probable, therefore, that these boulders were derived from outcrops that lay west and north of their place of deposition. The deposition of these boulders took place during several stages of Upper Mississippian and Lower Pennsylvanian time (Harlton, 1934, and 1938, Cline, 1956). The exotics range from granules up to blocks several hundred feet long. Sidney Powers and M. G. Cheney (Powers, 1928, Cheney, 1929) inferred that these boulders were derived from rubble produced by faulting along the north margin of a geosyncline in which the extremely thick Stanley-Jackfork-Johns Valley sequence was deposited. It seems probable that at least local reverse faulting from the northwest and north along the margin of the late Mississippian and early Pennsylvanian geosyncline of the Ouachita Mountains represents the first stage of faulting in the Ouachita Mountains. However, this faulting should be considered to be a part of the depositional history of the Ouachitas rather than the structural history. A purely diagrammatic representation of this faulting is shown on Preliminary Map No. 66 (Hendricks, 1947) as the "Powers Fault".

Gliding planes. - The reverse faults in the frontal part of the Ouachita Mountains tend to follow the bedding of a preferred weak shale zone in each of several belts.

The oldest zone of weakness is the Womble shale, in which the Ti Valley fault at Black Knob Ridge and the window fault in the Potato Hills are located. The Springer shale is the gliding zone in most of the belt adjacent to the Choctaw fault. Between the Pine Mountain and Ti Valley faults the Caney shale appears to have been a primary zone of weakness and the Atoka formation a secondary zone higher in the section. From the Ti Valley to the Windingstair fault the zones of gliding seem to be about equally divided between the weak Johns Valley shale and shales of the overlying Atoka formation, with the older Stanley shale having provided the gliding plane at a few places. South of the Windingstair fault the Stanley shale, beneath a thick competent sequence of the Jackfork sandstone, is the preferred zone of gliding with the Johns Valley shale constituting a secondary zone higher in the section. The only apparently significant feature of these relationships is that in any one belt the reverse faulting tends to follow the weakest beds. This vertical preference of weak beds suggests strongly that the forces that produced the deformation were greater in the horizontal direction than in the vertical direction, whereas the fact that along its strike any one fault breaks across competent beds to a higher or lower incompetent unit which it then follows indicates that a lesser vertical component of force was important.

Age of deformation. - Evidence existing within and adjacent to the Ouachita Mountains provides some indication of the time of the earlier stages of structural movement in the deformation of the Ouachita geosyncline.

It has previously been pointed out that boulder-bearing beds in the Stanley-Jackfork and Johns Valley shales suggest that southward thrusting occurred, at least at intervals in both time and space, along the north margin of the basin of Mississippian-lower Pennsylvanian (Morrowan) deposition.

The Atoka formation overlaps the Johns Valley shale and some of the upper part of the Jackfork sandstone at some localities in the western part of the Ouachita Mountains. Also, coarse conglomerates are present in Atoka beds west of the Ti Valley fault near Atoka. The pebbles in these conglomerates were derived from rocks of a facies characteristic of the belt in the frontal part of the Ouachita Mountains north of the Ti Valley fault. These facts indicate that an uplift was present within the area of the present Ouachita Mountains and southeast of Black Knob Ridge in Atoka time.

The absence of strata younger than Atoka in even the extreme frontal part of the Ouachita Mountains and the thickening of Desmoinesian strata into the McAlester Basin (Hendricks, 1937, Hendricks and Parks, 1937, 1950) indicate that the Ouachita Mountains suffered regional uplift in late Atoka time that continued concurrently with the progressive downwarping of the Demoinesian basin to the north. Depositional features of the Hartshorne sandstone and overlying formations indicate that the south limit of their deposition was near the present south side of the McAlester Basin. Locally derived intraformational conglomerates in Desmoinesian beds on folds in the Arkansas coal field that are related to the Ouachita structural province indicate that those structures were developing to a minor degree in Desmoinesian time. Conglomerates in beds of Desmoinesian age as young as the Thurman sandstone (Morgan, 1924, Knechtel, 1937) offer additional evidence of uplift in the western part of the Ouachita Mountains. The involvement of the Boggy shale (Clawson, 1928) and possibly beds as young as Thurman (Hendricks, 1937) in folds related to the Ouachita Mountain deformation indicates that structural movement in the Ouachita Mountains continued at least as late as Middle Desmoinesian time.

On the basis of jointing in the Seminole uplift interpreted as related to Ouachita Mountain deformation and on the basis of depositional and stratigraphic data from areas in north-central Texas and central Oklahoma, Melton and others (Melton, 1930, Cheney, 1929, van der Gracht, 1931) have inferred that pulses of Ouachita Mountain orogeny occurred in Middle and late Pennsylvanian time and as late as early Permian time.

Conclusion

What is the picture that emerges from consideration of the various structural features of the frontal part of the Ouachita Mountains?

A geosyncline was warped downward so

sharply in late Mississippian and early Pennsylvanian time that its north margin was faulted upward to the south and supplied exotic detritus to the basin.

Downwarping continued until a thick sequence of late Mississippian and early Pennsylvanian beds were deposited, although local areas in the western part were structurally high and suffered erosion.

General compression occurred in Desmoinesian time concurrent with downwarping and deposition in the McAlester Basin. Successive pulsations of uplift and deformation took place until late Pennsylvanian or early Permian time, when the deformation of the Ouachita geosyncline was essentially completed. Faulting occurred during this time in a series of stages that probably were more or less continuous, culminating in the major thrusting of the Ti Valley and Windingstair faults.

The cumulative north-northwestward movement in the frontal part of the Ouachita Mountains was of the order of magnitude of 50 miles.

The basic answers to the nature of the forces that deformed the Ouachita geosyncline are to be found in data concealed beneath deposits of the Gulf Coastal Plain.

References

Cheney, M. G., 1929, History of the Carboniferous sediments of the Mid-Continent oil field, Amer. Assoc. Petroleum Geologists Bull., vol. 13, pp. 557-594.

Clawson, W. W., Jr., 1928, Oil and gas geology of Coal and Pittsburg counties, Oklahoma Geol. Survey Bull., vol. 40-JJ.

Cline, L. M., 1956, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains in Oklahoma, Tulsa Geol. Soc. Digest, vol. 24, pp. 100-106.

Harlton, B. H., 1934, Carboniferous stratigraphy of the Ouachitas with special study of the Bendian, Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 1018-1049.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains, Amer. Assoc. Petroleum Geologists Bull., vol. 22, pp. 852-914.

Hendricks, T. A., 1937, Geology and fuel resources of the southern part of the Oklahoma coal field; Part 1, The McAlester district, U. S. Geol. Survey Bull., 874-A; Part 4, The Howe-Wilburton district, U. S. Geol. Survey Bull., 874-D.

Hendricks, T. A., 1940, Structural interpretation of recent gravity observations in southeastern Oklahoma, Amer. Assoc. Petroleum Geologists Bull., vol. 24, pp. 2143-2149.

Hendricks, T. A., and Parks, Bryan, 1937, Geology and mineral resources of the western part of the Arkansas coal field, U. S. Geol. Survey Bull., 847-E.

Hendricks, T. A., and Parks, Bryan, 1950, Geology of the Fort Smith district, Arkansas, U. S. Geol. Survey Prof. Paper 221-E.

Hendricks, T. A., and Knechtel, M. M., and Bridge, Josiah, 1937, Geology of Black Knob Ridge, Oklahoma, Amer. Assoc. Petroleum Geologists Bull., vol. 21, pp. 1-26.

Hendricks, T. A., and others, Geology of the western part of the Ouachita Mountains in Oklahoma, U. S. Geol. Survey Oil and Gas Inv. Ser. Preliminary Map 66.

Honess, C. W., 1924, Geology of southern LeFlore and northwestern McCurtain counties, Oklahoma, Oklahoma Bur. Geol., Circular 3.

Howell, J. V., 1947, The origin of imbricate structure, Tulsa Geol. Soc., Guidebook to field conference in the western part of the Ouachita Mountains of Oklahoma, pp. 51-53.

Knechtel, M. M., 1937, Geology and fuel resources of the southern part of the Oklahoma coal field, Part 2, The Lehigh district, U. S. Geol. Survey Bull. 874-B.

Maher, J. C., and Lantz, R. J., 1953, Correlation of pre-Atoka rocks in the Arkansas Valley, Arkansas, U. S. Geol. Survey Oil and Gas Inv. Chart OC 51.

Melton, F. A., 1930, Age of the Ouachita orogeny and its tectonic effects, Amer. Assoc. Petroleum Geologists Bull., vol. 14, pp. 57-72.

Miller, B. W., 1956, The geology of the Potato Hills, Pushmataha and Latimer counties, Oklahoma, Tulsa Geol. Soc. Digest, vol. 24, p. 66.

Miser, H. D., 1926, Geologic map of Oklahoma, U. S. Geol. Survey.

Miser, H. D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas, Oklahoma Geol. Survey Bull., vol. 50.

Miser, H. D., 1934, Relation of Ouachita belt of Paleozoic rocks to oil and gas fields of Mid-Continent region, Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 1059-1077.

Misch, Peter, and Oles, K. F., 1957, Interpretation of Ouachita Mountains Oklahoma as autochthonous folded belt: Preliminary report, Amer. Assoc. Petroleum Geologists Bull., vol. 41, pp. 1899-1905.

Morgan, G. D., 1924, Geology of the Stonewall quadrangle, Oklahoma, Oklahoma Bur. Geol. Bull., vol. 2.

Powers, Sidney, 1928, Age of the folding of the Oklahoma Mountains, Geol. Soc. America Bull., vol. 39, pp. 1031-1072.

Taff, J. A., 1902, Description of the Atoka quadrangle, U. S. Geol. Survey, Geologic Atlas, Folio 79.

van der Gracht, W. A. J. M. van Waterschoot, 1931, Permo-Carboniferous orogeny in south-central United States, Amer. Assoc. Petroleum Geologists Bull., vol. 15, pp. 991-1057.

OIL AND GAS POSSIBILITIES OF THE OUACHITA PROVINCE

J. V. Howell and Paul L. Lyons [1]

Introduction

Known geologic and geophysical data can be combined to give an approximate interpretation of the Ouachita province. The occurrence of oil, gas and asphalt within the area is widespread. Also, these hydrocarbons are found in rocks of the same age as those which are highly productive in adjacent areas. These factors combine to indicate that the region must be considered to have distinct possibilities of commercial accumulation of oil and gas.

The difficulty in evaluating these possibilities by the drill rises from the differences of opinion extant regarding the surface geology of the area. Varying interpretations of the complex expression of the widespread surface faults and occasional folds have shrouded the deeper structure in mystery and speculation.

The following assembly of geological and geophysical factual data permit certain conclusions to be drawn; they may help lead to the ultimate proper interpretation of the province.

Geophysics

Gravity. - A regional Bouguer gravity map of the Ouachitas is shown on the accompanying map pocket. A truly great negative anomaly is evident, with the lowest points reaching values lower than minus 110 milligals. The central part of the gravity minimum is centered along the common line of townships 1 N. and 1 S., ranges 20 and 21 E. It portrays a deep sedimentary basin. Significantly, the center of the minimum coincides with the possible area of thickest sediments. The two outstanding features are as follows: (1) the reduced gradients in the vicinity of the gravity trough; this indicates great depth for the causative anomaly; (2) the independence of the gravity contours in relation to such surface features as the Potato Hills, the Choctaw anticline and the Ouachita anticline, which extend the province into Arkansas. This establishes these features, wherein older rocks are exposed at the surface, as entirely different from nearby uplifts such as the Arbuckles, the Wichitas and the Osage Island, for these latter features are obvious uplifts of the basement rocks and are in each case reflected by gravity maxima.

Calculations from a north-south gravity profile along the 95th meridian reveal the following: if a maximum density contrast of 0.1 is assumed, with the sediments being that much lighter than basement rocks, a seven mile depth to basement in the deepest part of the sedimentary prism indicated will only account for about one-third of the observed anomaly.

Such a contrast is not warranted, evidently, since the shales, in bulk, may readily reach the density of granitic or heavier rocks at even moderate depths. The best assumption is that the earth's crust, above the Mohorovicic discontinuity, has folded downward into the mantle. If a normal crustal depth of 10 miles is assumed for northeastern Oklahoma, then a possible 10 mile downfold, or extension of 20 miles of the crust into the mantle, is indicated by the gravity profile. The basement is probably roughly conformable to this downfold. This downfold, if only 20 miles, is still 10 miles short of the maximum crustal thickness ever

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observed, which is in the neighborhood of the Alps. In a broad way, we can speculate from this gravity evidence that a moderately deep prism of sediments, mostly Pennsylvanian and Mississippian in age, accumulated in a local crustal depression, the Ouachitas, which is a unique part of an enormously long trough. Isostatic rebound may have uplifted the incompetent shale series so that they literally flowed outward from the uplift in intricate and imbricate thrust patterns. The area is still out of isostatic equilibrium, provided the shales are not "light" shales, since the isostatic correction is from plus 20 to plus 30 milligals over the area; this leaves 80 to 90 milligals minimum anomaly (isostatic) for the central part of the feature.

Magnetics. - A complete magnetic picture is not available. The magnetic maps examined show the following features:
(1) deep seated anomalies which do not coincide with the surface features, i. e., the Potato Hills are a magnetic minimum;
(2) alignments of elongate anomalies with certain faults.

In tandem with the gravity data, these permit the following conclusions: (1) the "basement" of the area is highly irregular, mostly rather deep, and it contains significant uplifts; (2) some of the folding and faulting of the area was accompanied by intrusion of igneous rocks containing magnetite.

The assignment of depths to magnetic anomalies offers the first departure toward geophysical solution of the attitude of the lower strata and basement of the area.

Seismic. - Modern methods make it possible to obtain seismic data in the Ouachitas. Local structure may be mapped at depth if extreme care and control are used.

It is obvious that a great service to geology would be rendered if a group of commercial seismic parties could be collected on a given day to record a deep refraction profile over the entire mountain mass. This would undoubtedly result in extremely valuable and concise data regarding the crustal thickness, the roots of the mountains, and the depths to basement along the profile.

Evidences of oil and gas

Evidence pointing toward possible production of oil and gas in the Ouachita Mountain area is by no means conclusive, but it appears sufficient to justify further study an future development. Such evidence is both direct and inferred.

Direct evidence. - (1) Veins of grahami and seepage of liquid asphalt in beds of Stanley, Jackfork and Bigfork (Viola) age. A list of known occurrences is appended, and these localities are noted on the geological map. The wide distribution, both stratigrap ically and geographically, shows convincingl that conditions of occurrence of oil are general throughout the area. (2) The occurrence at many points, particularly along major faults, of Caney shale and Woodford, unaltered and of a type normal to oil fields of the Arbuckle region is a strong argument for favorable conditions. (3) Although the carbon ratios within the Ouachitas are not know because of absence of coal beds, two lines of evidence suggest that they are not overly high (a) The shales show little evidence of dynamic metamorphism, i. e., are not slates, except locally. (b) It has been shown (Hendricks, T. A., Amer. Assoc. Petroleum Geologists Bull., vol. 19, no. 7, 1935, p. 945) that carbon ratios are high in closely folded regions but are lower wherever stresses have been relieved by faulting. Folding is minor in most of the Ouachita area, and faulting is predominant. (4) At several localities within the mountains oil in small amounts has been foun by drilling. This is especially true in McGee Valley and at Redden. The oil is of normal gravity (40-42 degrees Be.). Gas does not accompany this oil, perhaps because of the shallow depth (200 to 400 feet), although some other reasons might also be advanced. Gas occurs in fair amount in Potato Hills (sec. 30, T. 3 N., R. 21 E.), in a well in T. 3 N., R. 22 E., southeast of Talihina, and near Jumbo, in sec. 9, T. 2 S., R. 15 E (5) If our assumptions are correct, the beds below 2500 to 5000 feet will be found relatively undisturbed over large areas and metamorphism in the lower rocks may be expected to be inconsiderable. (6) Southeast of Stringtown in the NE 1/4 of sec. 19, T. 1 S., R. 12 E., the uppermost ten feet of Viola, immediately beneath the Sylvan shale, is impregnated with asphalt. The present structure, a steep eastward dip away from a fault scarp, is not sufficient to have caused this relationship, but such accumulation could well have occurred prior to the faulting.

Thus, under the following conditions, i.e., relatively undisturbed beds folded instead of faulted at moderate depths beneath the highly disturbed surface beds, closed folds in this lower zone may well be expected to produce if and when they can be found.

TABLE 1. ASPHALT DEPOSITS, OUACHITA MOUNTAINS

Locality	Nature of Deposits	Formation
S. line sec. 16, T. 1 S., R. 12 E.	Liquid seep	Viola and Sylvan contact
NE 1/4 sec. 19, T. 1 S., R. 12 E.	Asphalt	Top of Viola
SW 1/4 sec. 8, T. 1 S., R. 13 E.	Asphalt	Stanley sandstone
NE 1/4 sec. 25, T. 1 S., R. 13 E.	Pimroy mine	Stanley ss. and sh.
NE 1/4 sec. 13, T. 1 S., R. 14 E.	Asphalt	Stanley sandstone
Cen. E. line SW 1/4 sec. 16, T. 1 S., R.15 E.	Grahamite	Stanley shale
Cen. S. line SW 1/4 sec. 21, T. 1 S., R. 15 E.	Grahamite	Stanley shale
Cen. NW 1/4 sec. 28, T. 1 S., R. 15 E.	Grahamite veins	Stanley shale
SE 1/4 sec. 29, T. 1 S., R. 15 E.	Liquid seep	Stanley sh. and ss.
SW 1/4 sec. 12, T. 2 S., R. 11 E.	Asphalt impregnation	Stringtown and Viola
S 1/2 sec. 13, T. 2 S., R. 13 E.	Asphalt	Stanley sandstone
E 1/2 sec. 26, T. 2 S., R. 13 E.	Asphalt	Stanley sandstone
Cen. sec. 35, T. 2 S., R. 13 E.	Asphalt	Stanley shale
NE 1/4 NW 1/4 sec. 3, T. 2 S., R. 14 E.	Grahamite	Stanley shale
NW 1/4 NE 1/4 sec. 6, T. 2 S., R. 15 E.	Grahamite	Stanley shale
W 1/2 sec. 28, T. 2 S., R. 15 E.	Jumbo grahamite veins	Stanley sandstone
SW 1/4 NW 1/4 sec. 20, T. 7 S., R. 24 E.	Asphalt	Trinity sandstone
SW 1/4 sec. 23, T. 1 N., R. 14 E.	Grahamite	Stanley shale
SW 1/4 sec. 28, T. 1 N., R. 14 E.	Asphalt	Stanley sandstone
Sec. 25, T. 1 N., R. 15 E.	Shallow oil well	Stanley sandstone
Sec. 36, T. 2 N., R. 14 E.	Spring	Jackfork sandstone
Sec. 28, T. 2 N., R. 15 E.	Several 1-barrel wells	Stanley sandstone
SE cor. sec. 6, T. 2 N., R. 16 E.	Several 1-barrel wells	Stanley sandstone

TABLE 1. (continued)

Locality	Nature of Deposits	Formation
Sec. 7, T. 2 N., R. 16 E.	Oil spring near group of shallow oil wells	Probably Jackfork sandstone
NE 1/4 sec. 9, T. 2 N., R. 18 E.	Asphaltic vein	Stanley sandstone
SE 1/4 sec. 1, T. 2 N., R. 19 E.	Asphalt veins	Siliceous ls. in crest of small folds; Potato Hills
NE 1/4 sec. 2, T. 2 N., R. 19 E.	Asphalt veins	Siliceous ls. and slaty sh.; Potato Hills
SE 1/4 sec. 24, T. 3 N., R. 26 E.	Asphalt	Fine-grained argillaceous ss.; probably Jackfork
1 mile E. of Page, SW slope Black Fork Mountain on J. E. McClure estate; 300 yds. N. of K. C. S. tracks		
Arkansas; NW end of Fourche Mountain about 10 miles E. of state line	Grahamite	

TABLE 2. IMPORTANT OIL TEST WELLS IN OUACHITA AREA

1. Southwest Exploration Co. # 1A Hochman. Loc. - NE 1/4 SE 1/4 sec. 16, T. 2 N., R. 14 E., Pittsburg Co., Oklahoma Elev. 775 feet; total depth 8744 feet; completed February, 1955.

2. Northern Ordnance # 1 Fulton Ranch. Loc. - Sec. 19, T. 1 S., R. 12 E.; elev. 865 feet, total depth 7007; completed Nov. 19, 1944.

3. E. P. Halliburton # 1 Bagnell. Loc. - SW 1/4 SW 1/4 NE 1/4 NE 1/4 sec. 30, T. 3 S., R. 15 E. Total depth 6006 in Stanley; Stanley at surface. Completed Nov. 6, 1937.

4. Max Pray # 1 Wyrick. Loc. - SW 1/4 NW 1/4 SE 1/4 sec. 26, T. 1 N., R. 14 E. Completed Sept., 1958.

   Stanley sandstone and shales to depth of 8200 ft.
   Arkansas novaculite (Woodford) 8335 to 8400 ft.
   Polk Creek (Sylvan) 8400 to 8420 ft.
   Bigfork (Viola) cherty limestones 8420 to 9098 ft.
   Womble (Simpson) shales 9098 to 12,088 ft. (total depth).

   Dipmeter surveys to depth of 8420 ft.: SE dips to 3795 ft., NW dips to 4150 ft., SE dips to 6056 ft., SW dips to 6922 ft., SE dips to 7424 ft., NW dips to 7952, and SW dips to 8420 ft.

   DST: 3676-86 show of oil; 3497-3521 show gas; 8431-9852 est. 300,000 CF gas; 8293-8303 est. 200,000 CF gas.

TABLE 2. (continued)

5. Sikes Burkhalter # 2 Denton-Perrin. Loc. - SW 1/4 SE 1/4 SE 1/4 sec. 9, T. 2 S., R. 15 E. Completed Nov. 18, 1958. Stanley shale from surface to 5480 ft. (TD). Gas shows in 15 sandstones between 2331 and 4680. 5 1/2-inch casing set at 5479; sandstones at 4680-5381 perforated; est. 260 CF gas after Riverfrac. D. and A.

Reference

Lyons, Paul L., Gravity map of the United States, Tulsa Geol. Soc. Digest, 1951.

THE AGE OF MINERALIZATION IN THE OUACHITA MOUNTAINS OF ARKANSAS AND OKLAHOMA

B. J. Scull [1]

Introduction

Sulfides of mercury, antimony, lead, zinc, copper; oxides of manganese and titanium; and diamonds, barite and quartz comprise the economic mineral suite of the Ouachita province. The mode of occurrence of each of these minerals is limited to certain environments; some have a restricted geographic range while others occur in small deposits throughout the province. The accompanying map shows the locations of the various types of mineral deposits. In only a few of the many deposits are the ore minerals found in commercial quantities.

The ages of the mineralized deposits of the Ouachita Mountain area have been ascribed in part to the Middle or Late Pennsylvanian and in part to the early Upper Cretaceous. The cinnabar, antimony, lead, zinc, copper, manganese and quartz deposits have been classed as Paleozoic (Miser, 1917; Honess, 1923; Miser and Purdue, 1929; Stearn, 1936; Reed and Wells, 1938; Gallagher, 1942; Miser, 1943; Engel, 1951). The titanium, diamond and barite deposits have been classed as Mesozoic (Williams, 1891; Miser, 1914; Ross, Miser and Stephenson, 1929; Ross, 1941; Holbrook, 1947; Scull, 1956).

The data obtained in the study of the barite deposits of Arkansas (Scull, 1956) indicate that barite is the common denominator in all the mineral deposits except perhaps the quartz crystals. The presence of barite in most of the mineral deposits and igneous rocks along with other features and certain inferences indicate that the mineralized belt of the Ouachita Mountains is an early Upper Cretaceous epithermal metallogenetic province.

At present, investigation of the cinnabar, lead-zinc-copper and antimony deposits is difficult because most of the mine shafts are caved and full of water. Fortunately these deposits were well documented earlier in the reports of the United States and state geological surveys. Collections of specimens were obtained from local residents as well as from veins for comparison with the published data. Parts of the diamond area (tourist attraction), some manganese deposits, and most of the titanium and barite deposits are accessible for detailed study.

Each of the types of mineral deposits are briefly described in the following paragraphs. The references to the literature were selected partly as being most informative and partly because they contain rather inclusive bibliographies.

Diamonds

The diamond-bearing peridotite in Pike County, Arkansas, is nearly identical petrographically with the diamond-bearing rocks of South Africa (Miser and Ross, 1923). The storied "yellow and blue grounds" are common features of both areas. The Arkansas peridotite occurs as massive intrusions, dikes, breccias and fine-grained tuffs. Quartz and barite veins cut the intrusive peridotite and the brecciated peridotite. The intrusives and breccias contain xenolithic Jackfork sandstone and Arkansas novaculite showing that typical Ouachita sediments were stoped during emplacement. Miser and Ross (1923) give a resume of literature dating back to 1842.

The age of the peridotite emplacement can be ascertained more readily than the age

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of any other mineralized deposit in the southern Ouachita Mountain area. The intrusions and eruptions occurred after Trinity deposition (Lower Cretaceous) and were concluded before the end of Tokio deposition (early Upper Cretaceous) (Miser and Purdue, 1929). The Trinity is devoid of pyroclastic material and is intruded by the peridotite. The overlying Woodbine and Tokio formations contain appreciable pyroclastic or erosion derived peridotitic material near these diamond-bearing rocks. Elsewhere these formations are characterized by syenitic and phonolitic pyroclastic components (Ross, Miser, and Stephenson, 1929). Concentrations of tuff and ash indicate eruptive centers near Nashville and Lockesburg (Plate 1). The tuffs are similar in composition to the Magnet Cove intrusives.

Cinnabar

The Arkansas Cinnabar district is within the Athens Plateau, the southern segment of the Ouachita Mountain physiographic unit, and is located about nine miles north of the peridotite area. The cinnabar deposition was controlled by fracture patterns and intergranular porosity in the sandstones of the Jackfork and Stanley formations. A resume of the earlier reports on this district is given in Reed and Wells (1938). The individual deposits and regional setting are described in reports by Stearn (1936), Reed and Wells (1938) and Gallagher (1942). Stearn, and Reed and Wells present analyses to show that mineralization occurred during major tectonic activity--the Ouachita orogeny. Their strongest evidence is that there are no adjacent Cretaceous igneous rocks, and that mineralized slickensides and fractured and polished cinnabar are present in the ore bodies. All of the reports cited above agree that the mineralizing solutions egressed from the parent material along the planes of the major thrusts in the area.

Stibnite is associated with the cinnabar, and quartz is the chief gangue material. Some coarsely crystalline barite and minor amounts of dolomite and siderite are also present as gangue. Reed and Wells (1938) list all the hypogene and supergene minerals that have been recognized in these deposits.

Antimony

The antimony deposits are located near Gillham about 20 miles west of the cinnabar district. Stibnite is the chief antimony mineral and is associated with lead, zinc and copper sulfides in comby quartz veins. Most of the veins are concordant with the enclosing shales and sandstones of the Stanley formation. The most complete descriptions of these deposits are in the reports of Wait (1880), Hess (1908), and Miser and Purdue (1929).

Lead-Zinc-Copper

The lead-zinc-copper deposits are scattered throughout the southern Ouachita Mountain area. These deposits occur in quartz veins emplaced in fractured sediments ranging from Collier shale to Jackfork sandstone. Descriptions of these deposits are given in reports of Comstock (1888), Bain (1901), Honess (1923), and Miser and Purdue (1929). For two reasons the report of Honess is of particular importance to this discussion of the age of mineralization. Honess describes quartz veins in McCurtain County, Oklahoma, that contain potash feldspar as an essential component, and are thus pegmatite veins near the mineralized areas. In the description of the Eades Mine, Honess (Pt. 2) notes the presence of barite gangue in the deposit.

Before continuing this brief review of the various type of deposits, the distinct relationship of the deposits discussed above should be emphasized. Cinnabar and stibnite are restricted to epithermal deposits. The occurrence of stibnite in the cinnabar district indicates a common origin with the antimony deposits farther west. The occurrence of lead, zinc, and copper sulfides in the antimony belt indicates that the lead-zinc-copper deposits have a common origin with the antimony and cinnabar deposits. The paragenetic sequence of sphalerite-chalcopyrite-galena is epithermal. Orthoclase (adularia)-quartz veins are epithermal indicators. Comby quartz and barite are typically epithermal minerals.

As suggested in previous reports the sulfide deposits of the Ouachita Mountains belong to a single metalliferous epoch. The character of the sulfides and associated nonmetals show that this province is epithermal - an important factor in determining the age of these deposits.

Manganese

The manganese deposits are confined to the upper and lower members of the Arkansas novaculite. The ore minerals are manganese oxides concentrated in fractures and porous leached zones. Many of the deposits have been described in detail by Penrose (1891) and Miser (1917).

The upper member of the novaculite contains isolated crystals of manganiferous calcite which through leaching supplied some of the manganese in the ore deposits. There was insufficient carbonate to have been the sole source of the manganese. Current evidence tends to force the conclusion that the novaculite is a syngenetic deposit and not a silica replaced limestone. The novaculite accumulated in a basin which many writers term as "starved". As such, this basin received the ultimate in weathering products. Manganese is one of the ultimate products of laterization. Minute amounts of manganese are present in unaltered massive beds of the upper and lower novaculite members. Much, if not all, of the manganese in the zones of concentration was leached from the novaculite by migrating solutions.

Some of the manganese deposits may have been formed by deposition from percolating meteoric water but most of it was concentrated by telethermal solutions and is genetically related to the barite of the region. Psilomelane which contains essential barium is present in several deposits. The vast area invaded by the barium-bearing solutions will be discussed later. The inter-relationship of the barite and manganese is demonstrated by their stratigraphic relationships. The manganese occurs in the lower and upper novaculite and barite occurs in the middle novaculite and the lower Stanley. Their respective sites of deposition are shown on the location map (Plate 1).

Titanium

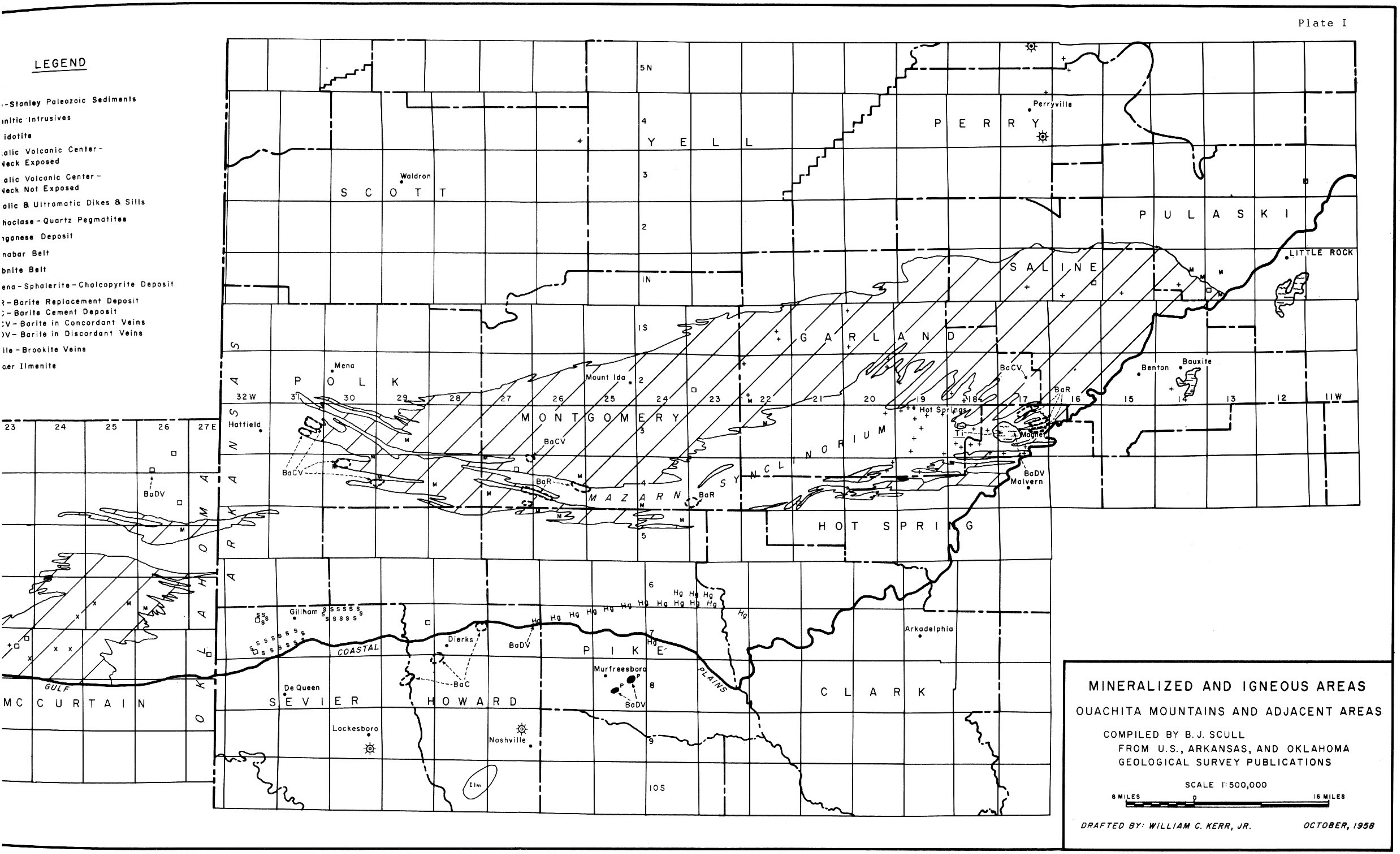
The titanium deposits are an integral part of the Magnet Cove intrusives complex. The titanium minerals are sphene (titanite), taeniolite (titanium mica), ilmenite, perovskite ($CaTiO_3$), brookite and rutile (TiO_2). The polymorphs brookite and rutile are the only titanium minerals present in potentially commercial amounts.

The igneous rocks of the Cove complex are nepheline and leucite syenites, fine-grained syenite equivalents, and medium to coarse-grained melanocratic rocks. Ultramafic dikes intrude the coarser rocks. These igneous rocks invaded folded Paleozoic sediments. The sediments in immediate contact with the intrusives were metamorphosed and now stand as a rim around the igneous area. This topography is the reason for the term "cove".

Williams (1891) wrote one of the few classics in American petrography after his study of this area. Specifics on the brookite and rutile deposits have been presented by Ross (1941), Holbrook (1947), and Fryklund and Holbrook (1950). A recent and accurate map of the igneous rocks of the Magnet Cove complex has been published by the United States Geological Survey (Erickson and Blade, 1956).

The rutile deposits are restricted to the igneous rocks and the brookite deposits are essentially restricted to the metamorphosed sediments. The rutile occurs in feldspar-carbonate veins cutting phonolite. Four carbonate-feldspar assemblages are known and probably represent four phases of vein formation. The brookite was formed in quartz veins most of which are contained in the Arkansas novaculite.

Sinter or "tufa" deposits within the Cove are cut by barite veinlets, and, of more significance, the feldspars of much of the igneous rock are saturated with barium. Potash feldspars can accept up to four percent BaO in the ionic groupings without disrupting the space lattice. The ionic radius of barium ($1.46A^o$) is larger than potassium ($1.31A^o$) so substitution occurs mostly at high temperatures where ionic bonding is less rigid. The association of titanium with barium-rich feldspars is an important facet to the fixing of the age of the various mineralized deposits of the Ouachita area.



MINERALIZED AND IGNEOUS AREAS
OUACHITA MOUNTAINS AND ADJACENT AREAS
COMPILED BY B. J. SCULL
FROM U.S., ARKANSAS, AND OKLAHOMA GEOLOGICAL SURVEY PUBLICATIONS
SCALE 1:500,000
8 MILES 0 16 MILES
DRAFTED BY: WILLIAM C. KERR, JR. OCTOBER, 1958

Barite

The barite deposits in the southern Ouachita area occur as four distinct types: replacement deposits, cement in sandstone and gravel, concordant veins, and discordant veins. The discordant veins are mentioned above as occurring in the peridotite area, in the Magnet Cove complex, in the cinnabar deposits and in the lead-zinc-copper deposits. The concordant veins are found in the thin-bedded middle member of the Arkansas novaculite. These veins were emplaced with only minor disturbance of the wall rock. Metasomatic alteration is lacking, although some features in the silicic shales could be interpreted as reconstitution.

The cementing barite is found in the sands and the basal Pike gravel of the Trinity formation. The rate of deposition was slow enough to allow barite crystals three or four inches long to develop. In many such crystals the sand grains are "floating". Much of the barite was deposited as penetrating crystals, so that aggregates weathered out of the host rock have the rosette configuration which characterizes the barite roses of the Permian Garber sandstone in central Oklahoma. There is nothing, however, to indicate a genetic relationship between the Garber and the Trinity barite concentrations.

Replacement barite deposits occur near the extremities of the Mazarn synclinorium which extends west from Magnet Cove for about 60 miles. The barite is restricted to the lowermost part of the Stanley formation; over 95% of the replacement is confined to synclines. In a few deposits faults blocked fluid migration which resulted in local barite concentrations. In all the larger deposits concentration zones were controlled by the availability of capillary openings for fluid migration.

The Chamberlin Creek synclinal barite deposit, at the eastern end of the Mazarn synclinorium, is the source of about 25% of the world's production of the mineral. Both the Baroid and Magcobar companies operate mines in this deposit. The syncline is truncated at the western end by the Magnet Cove igneous complex. Detailed petrographic studies, trace element data, and titanium and strontium content show that the barite replacing the Stanley units was deposited from fluids derived from the adjacent igneous rocks.

The igneous source of the western replacement deposits is not definitely located but the stratigraphy, structure, ore types, trace elements (expecially titanium), and replacement mode in these deposits are identical with those of the deposits associated with Magnet Cove. Igneous dikes are present in the general vicinity of the western deposits (Plate 1). The weight of evidence indicates that all the replacement deposits are consanguineous.

The Igneous Provenance

In order to show that the mineralized deposits of the Ouachita Mountains and the flanking physiographic provinces were formed within a single effusive epoch, it is necessary to prove beyond reasonable doubt that the post-Atoka igneous rocks belong to the same magmatic suite, that epithermal deposits of Pennsylvanian age could not have survived the post-Atoka pre-Tokio erosion intervals, and that the igneous rocks and mineral deposits are genetically related.

The igneous rocks of the Ouachita and adjacent provinces crop out in widely scattered but small areas. Igneous rocks have been recorded in an area extending northward from the Pike County peridotites to Scott County and eastward from the peridotite to the syenites near Little Rock. The area of verified exposures of igneous material occupies the belt between the Coastal Plain and the Arkansas River. A rather complete compilation of known igneous localities can be made from the reports of Williams (1891), Miser and Purdue (1929), Croneis (1930), Fryklund and Holbrook (1950), and Erickson and Blade (1956).

The syenites south and east of Little Rock, east of Benton, in Magnet Cove, at Potash Sulfur Springs and in the volcanic breccias near Perry are associated with and essentially connected by swarms of sills and dikes. The mineralogical composition and attendant rock names of these sills and dikes are calculated to send geologists scrambling for the nearest copy of Johannsen.

Monchiquite, fourchite, ouachitite, tinguaite, and phonolite are the more common rocks of this dike-sill complex. Tinguaite is a porphyritic phonolite dike, perhaps an unnecessary name but entrenched in the literature. Phonolite is the fine-grained to aphanitic equivalent of aegerine-nepheline syenite. Monchiquite is porphyritic dike material with olivine and pyroxene in a glassy or analcite groundmass. Fourchite has little or no olivine and ouachitite has essential biotite as well as pyroxene phenocrysts. Basaltic hornblende, pyrite, sphene and apatite are the more common accessory minerals in these mafic rocks and calcite is ordinarily present in major porportions. The tinguaites and phonolites are of course directly related to the syenites, nepheline syenites and leucite syenites of the various plutonic localities in the Ouachita area. The monchiquite group can be related to the unusual Magnet Cove suite of melanocratic phanerites and the peridotite in Pike County. The melanocratic phaneritic rocks in Magnet Cove include shonkinite (augite and orthoclase), ijolite (diopside and nepheline), melteigite (diopside, biotite and nepheline), jacupirangite (magnetite and pyroxene) and melagabbro. The chief rock variety in the peridotite area is kimberlite (olivine, diopside and phlogopite).

The post-Atoka igneous rocks form a sub-silicic suite, which in places is aluminum deficient and in places barium and strontium rich. All of these rocks are comparatively rich in carbonate, titanium and potassium. The most spectacular correlation feature is the presence of perovskite ($CaTiO_3$) in the kimberlite of Pike County and in the Magnet Cove suite. One nicety of correlation is available because of the abundance of titanium in the igneous suite. The existence of a syenite-phonolite volcanic center near Nashville as ascertained by Ross, Miser and Stephenson (1929) has been mentioned above. A few miles south of this eruptive center in the near-shore or beach sediments of the Tokio (Cretaceous) are found concentrations of ilmenite (Holbrook, 1948). The ilmenite is restricted to the Tokio and was derived from pyroclastics.

The igneous rocks under discussion undoubtedly were formed in various phases or waves but they are all products of a silica deficient, titanium-potassium rich magmatic source, and were emplaced after Trinity and before Brownstown time in the early Upper Cretaceous.

Structure and Depth Control of Mineralization

The character of the economic mineral deposits was reviewed in the preceding sections. These deposits are epithermal or telethermal in every known system of ore deposit classification. The early Upper Cretaceous igneous suite is the only known igneous source possible for these deposits. The ore deposits are found in sedimentary rocks as young as Atokan and therefore cannot have been derived from the igneous activity that produced the Middle Mississippian Hatton tuffs in the lower part of the Stanley formation.

The readily available Cretaceous magmatic source for the sulfides has been discounted by some geologists because of slickensides and other movement phenomena associated with the mineralized zones, although structural displacements of even inches are capable of producing slickensides. The available facts are explicitly against a Paleozoic age for the epithermal mineral suite.

Cinnabar is one of the most diagnostic epithermal minerals. It ordinarily is found associated with thermal spring deposits but has a known extreme depth range of about 3000 feet. Most cinnabar has been produced from depths less than 500 feet although the famous Almaden, Spain production extends below 1100 feet. Barite and comby quartz are typically epithermal minerals. Stibnite is formed principally in the epithermal zone but is not uncommon in mesothermal deposits. Although the exact depth range of epithermal and mesothermal veins are subject to a number of variables, the extremes approach 5000 feet for epithermal deposits and 10,000 feet for mesothermal deposits.

The cinnabar in Arkansas was deposited in fractures associated with the Cowhide thrust and subsequent cross faults. The ore was emplaced in the Stanley and Jackfork sandstones. The antimony deposits were formed in the Stanley formation. The replacement barite was deposited in the basal Stanley units and the concordant vein barite was emplaced in the Middle Arkansas novaculite. Comby quartz is the common gangue of all the sulfide deposits of the region. Conservatively, 6000 feet of Stanley had to be removed before the novaculite was exposed, 5000 to 6000 feet of Jackfork had to be removed before the Stanley was exposed, and an extreme minimum of 6000 feet (maximum 14,000 feet) of Atoka had to be removed before the Jackfork was exposed.

The initial spasms of the Ouachita orogeny are reflected in the change in sedimentation from novaculite to the Stanley sands and shales and the conglomerates in the Stanley. Pulsations of major import did not occur until middle Atoka time (Hendricks and Parks, 1950), and certainly no major uplift occurred until after Middle Pennsylvanian time. The Middle Pennsylvanian Hartshorne through Boggy formations in the Arkansas Valley and McAlester basin are essentially devoid of coarse clastics.

Since major erosional stripping did not occur in the Ouachita Mountain area until after Middle Pennsylvanian, it does not seem logical that 20,000 feet of sediments could have been removed in the sites of epithermal deposition prior to the cessation of the Ouachita orogeny in late Pennsylvanian or early Permian time.

Ruling out epithermal vein deposits in association with Ouachita orogeny necessitates explaining mineralized slickensides and fault "horses". Evidence is accumulating to show that the Ouachita area has been tectonically active in the post-orogenic era. Tanner (1954, 1956), following the suggestions of Melton (1947), has shown that the Ouachita Mountain area has been subjected to over 3500 feet of post-orogenic uplift and implies that 10,000 feet of post-Pennsylvanian uplift may be provable.

Vestal (1950) points out that several thousand feet of Permian beds were deposited less than 20 miles south from the present Atoka exposures in the Ouachita Mountains. Several thousands of feet of Jurassic and Lower Cretaceous sediments were deposited on the Permian sequence. During the erosional period preceding the deposition of the Upper Cretaceous sediments the Lower Cretaceous and Jurassic units were truncated progressively northward. The total truncation may exceed 10,000 feet (Weeks, 1939).

The amount of relative uplift in the Ouachita area accompanying this erosion cycle, calculated by Melton's strike-overlap method, is no less than 3000 feet. Movement of this magnitude is sufficient to reactivate slippage along fault planes developed during the Ouachita orogeny. These later movements probably were tension rather than compression release. The Ouachita structural segment, in the manner of a massif, was the fulcrum of Gulf Coast subsidence. The post-Atoka pre-Woodbine subsidence aggregated over 10,000 feet. This sagging caused severe tension against the massif and the resulting zone of weakness controlled the loci of magmatic invasion. Minor structural adjustments along existing features within the Ouachita segment are to be expected.

Less profound but more indicative evidence of Cretaceous rather than Late Paleozoic age is the nature of deposition of the sulfide-bearing veins. These veins, in many of the deposits, filled open fissures and fractures. In places, the openings were not filled but were lined with comby quartz. The lack of wall rock alteration implies open systems for the solutions from which the ore and gangue were precipitated. It is unlikely that open fractures would survive erosion cycles during which over 20,000 feet of overburden were removed.

The sites of the epithermal sulfide precipitation were near-surface during the Cretaceous igneous activity and were buried under 10,000 feet or more of Ouachita sediments during the Ouachita orogeny. The slickensides and other movement features associated with the ore minerals were produced during response to a major pre-Upper Cretaceous structural adjustment of features formed during the Ouachita orogeny.

Relationship of the Ore Minerals and Igneous Rocks

Geologic age and chemical affinities are ordinarily used to define a metallogenetic epoch. The age of the Pike County peridotite is definitely established as post-Trinity and pre-Brownstown - early Upper Cretaceous. The chemical and mineralogical composition of the other igneous rocks of the region show that they are related to the peridotite and are early Upper Cretaceous in age.

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The diamonds in the peridotite and the rutile and brookite of Magnet Cove are integral components of the host rocks and are effectively dated. The replacement barite deposits are chemically and genetically related to the Magnet Cove igneous complex and therefore are of the same age as the igneous suite.

The relationship of the replacement barite to the vein barite and to the barite cement in the Trinity beds is indicated by similar trace element suites. The discordant vein barite in the peridotite and syenite is late stage igneous as is the replacement barite. The concordant vein barite in the novaculite and the barite cement in the Trinity are not in direct contact with igneous rocks therefore they cannot be classified as to stage.

The section above was devoted to establishing an early Upper Cretaceous age for the various sulfide deposits in the Ouachita Mountains. The age and epithermal character of these deposits and the presence of barite gangue in several of them indicate that they were derived from the Upper Cretaceous magmatic suite.

The extensive presence of quartz gangue with sulfides derived from a sub-silicic magma may seem anomalous, but it is not. Late stage quartz veins are found in the peridotite and syenite areas. This sufficiency of silica resulted from assimilation of Arkansas novaculite, Jackfork sandstone, and, perhaps, older sandstones and cherts. The titanium deposits of Magnet Cove offer an example of the potency of the magmatic emanations. The solutions carrying titanium from the crystallizing magma leached silica from the Arkansas novaculite and deposited it as the quartz of the brookite-bearing veins. The silica for the quartz of the sulfide deposits was obtained from the country rock by the migrating epithermal solutions.

Relating the manganese deposits to the Upper Cretaceous activity can be based only on the association with the barite veins in the Arkansas novaculite and the replacement barite in the basal Stanley. Nearly all of the manganese was derived from the country rock and only the concentration is attributable to the Upper Cretaceous activity. The manganese is concentrated in fractures and porous zones. The open fractures are probably products of unloading and the porous zones are products of leaching within the zone of ground water activity. As in the sulfide deposits, several thousands of feet of overlying sediments had to be removed before the sites of concentration were available for the type of deposition that occurred. The only indications that the manganese concentrations are intimately associated with Upper Cretaceous metallogenetic epoch are the presence of psilomelane in the deposits and the occurrence of these deposits stratigraphically immediately above or below the barite-bearing zones.

Summary

The Ouachita Mountain region was intruded by sub-silicic magmas along zones of weakness created by interplay of the Gulf Coast subsidence and the Ouachita segment stability during early Upper Cretaceous time. Some of the magmatic material stoped through the overlying sediments so that volcanic eruptions occurred. The igneous rocks, differentiated from the magmas, constitute a nepheline syenite-peridotite suite containing various malanocratic rock types. Solutions from the effusive bodies deposited in the country rocks several hundred thousand tons of barite, several thousand tons of rutile-brookite, and a few tons of cinnabar, stibnite, galena, chalcopyrite and sphalerite. None of the known igneous rocks crystallized at depths greater than 3000 feet and the ore minerals form a typical epithermal suite.

Titanium and potassium are the chief indicators of the interrelationship of the sub-silicic igneous rocks and barium is the chief linkage between the igneous rocks and the ore minerals.

References

Bain, H. F., 1901, Preliminary report on the lead and zinc deposits of the Ozark Region, U. S. Geol. Survey Twenty-Second Ann. Report, Pt. 2.

Comstock, T. B., 1888, Report on preliminary examination of the geology of western central Arkansas, Ark. Geol. Survey Ann. Report for 1888, Vol. 1.

Croneis, Carey, 1930, Geology of Arkansas Paleozoic area with especial reference to oil and gas possibilities, Arkansas Geol. Survey Bull., vol. 3.

Engel, A. E. J., 1951, Quartz crystal deposits of western Arkansas, U. S. Geol. Survey Bull. 973-E.

Erickson, R. L., and Blade, L. V., 1956, Map of bedrock geology of Magnet Cove igneous area Hot Spring County, Arkansas, U. S. Geol. Survey Min. Inv. Map MF 53.

Fryklund, V. C. and Holbrook, D. F., 1950, Titanium ore deposits of Hot Spring County, Arkansas, Ark. Geol. and Conv. Com. Bull., vol. 16.

Gallagher, David, 1942, Quicksilver deposits near the Little Missouri River Pike County, Arkansas, U. S. Geol. Survey Bull. 936-H.

Hendricks, T. A. and Parks, Bryan, 1950, Geology of the Fort Smith district Arkansas, U. S. Geol. Survey Prof. Paper 221-E.

Hess, F. L., 1908, Antimony deposits of Arkansas, U. S. Geol. Survey Bull. 340, pp. 241-252.

Holbrook, D. F., 1947, A brookite deposit in Hot Spring County, Arkansas, Ark. Geol. and Conv. Com. Bull., vol. 11.

Holbrook, D. F., 1948, Titanium in southern Howard County, Arkansas, Ark. Geol. and Conv. Com. Bull., vol. 16.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma, Okla. Geol. Survey Bull., vol. 32.

Melton, Frank A., 1947, Onlap and strike-overlap, Amer. Assoc. Petroleum Geologists Bull., vol. 31, pp. 1868-1878.

Miser, H. D., 1914, New areas of diamond-bearing peridotite in Arkansas, U. S. Geol. Survey Bull. 540.

Miser, H. D., 1917, Manganese deposits of the Caddo Gap and DeQueen quadrangles, Arkansas, U. S. Geol. Survey Bull. 660.

Miser, H. D. and Ross, C. S., 1923, Diamond-bearing peridotite in Pike County, Arkansas, U. S. Geol. Survey Bull. 735.

Miser, H. D. and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles, Arkansas, U. S. Geol. Survey Bull. 808.

Miser, H. D., 1943, Quartz veins in the Ouachita Mountains of Arkansas and Oklahoma, their relation to structure, metamorphism and metalliferous deposits, Econ. Geol., vol. 38, pp. 91-118.

Penrose, R. A. F., Jr., 1891, Manganese - Its uses, ores, and deposits, Ark. Geol. Survey, Ann. Report for 1890, vol. 1.

Reed, J. C., and Wells, F. G., 1938, Geology and ore deposits of the southwestern Arkansas quicksilver district, U. S. Geol. Survey Bull. 886-C.

Ross, C. S., Miser, H. D. and Stephenson, L. W., 1929, Water-laid volcanic rocks of early Upper Cretaceous age in southwestern Arkansas, southeastern Oklahoma, and northeastern Texas, U. S. Geol. Survey Prof. Paper 154.

Ross, C. S., 1941, Occurrence and origin of the Titanium deposits of Nelson and Amherst counties, Virginia, U. S. Geol. Survey Prof. Paper 198.

Scull, B. J., 1956, Origin and occurrence of barite in Arkansas, Doctoral dissertation, University of Oklahoma. Also, Ark. Geol. and Conv. Com., Inf. Circ. 18 (1958).

Stearn, N. H., 1936, The cinnabar deposits in southwestern Arkansas, Econ. Geol. vol. 31, no. 1, pp. 1-28.

Tanner, W. F., 1954, Tectonic and paleogeographic inferences from low-dip unconformities, Amer. Assoc. Petroleum Geologists Bull., vol. 38, pp. 886-899.

Tanner, W. F., 1956, Geology of Seminole County, Oklahoma, Oklahoma Geol. Survey Bull., vol. 74.

Vestal, Jack H., 1950, Petroleum geology of the Smackover formation of southern Arkansas, Ark. Geol. and Conv. Com., Inf. Circ. 14.

Wait, C. E., 1880, The antimony deposits of Arkansas, Am. Inst. Min. Eng. Trans., vol. 8, pp. 42-52.

Weeks, Warren B., 1938, South Arkansas stratigraphy with emphasis on the older Coastal Plain beds, Amer. Assoc. Petroleum Geologists Bull., vol. 22, pt. 11, pp. 953-983.

Williams, J. F., 1891, The igneous rocks of Arkansas, Arkansas Geol. Surv., Annual Rept. for 1890, vol. 2.

CORRELATION OF PRE-STANLEY STRATA IN THE ARBUCKLE-OUACHITA MOUNTAIN REGIONS [1]

William E. Ham [2]

Introduction

The pre-Stanley rocks concerned with in this paper are of Devonian and older age. They include a sequence of Devonian, Silurian, and Ordovician strata with an exposed thickness of 3,800 to 6,300 feet in the Ouachita Mountains, and a sequence of Devonian, Silurian, Ordovician, and Upper Cambrian strata 6,800 to 11,500 feet thick in the Arbuckle Mountains.

The Ouachita Mountains have the form of a belt approximately 200 miles long, almost equally divided between Oklahoma and Arkansas, that extends from Little Rock westward to Atoka. About three-fourths of the pre-Stanley outcrops are in Arkansas, chiefly in a nearly continuous strip 100 miles long in the Mena-Mount Ida-Hot Springs-Little Rock area. The remaining one-fourth is in Oklahoma, chiefly in McCurtain County north of Broken Bow, in the Potato Hills area southwest of Talihina, and in Black Knob Ridge at Atoka. The outcrop areas are rather small but are widely distributed, giving a reasonable measure of geographic control and showing that the sequence thins westward while undergoing no appreciable change in facies in the 200-mile length of the Ouachita Mountain belt.

The sequence in the Ouachita Mountains extends from the top of the Arkansas novaculite down to the oldest exposed strata, Collier and Lukfata formations, here interpreted as of Early Ordovician age. The rocks are characterized by black shales, which predominate in the lower part of the sequence and are interbedded with other strata in the upper part. Fossils are sparsely distributed, graptolites and conodonts being the principal faunal elements. The rocks are only moderately well exposed, and structural complications are locally severe, so that stratigraphic relations and thicknesses are incompletely known.

The much smaller outcrops of the Arbuckle Mountain region cover about 900 square miles in a triangular area whose southeastern point lies barely 20 miles west of the Ouachita Mountains at Atoka. This is near the eastern edge of a broad region known as the "Arbuckle Mountain facies", which consists mostly of normal marine limestone, dolomite, sandstone, and shale, and extends generally westward and southward several hundred miles into Texas. The stratigraphic sequence includes all beds from the top of the Woodford formation to the base of the Reagan sandstone, and hence includes all strata of middle and early Paleozoic age down to the top of the Precambrian. Carbonate rocks predominate, and locally they contain abundant marine fossils such as brachiopods, trilobites, cephalopods, corals, and sponges. The term "Arbuckle facies" as used by geologists in the past refers to such carbonate rocks and their accompanying green shales and quartzose sandstones. This facies is built up mostly of biochemical products in a carbonate rock environment, and as such it contrasts markedly with the black shales and predominant clastic rocks of equivalent age in the Ouachita Mountains.

It is the purpose of this paper to (a) establish with available control the correlation of all pre-Stanley stratigraphic units, primarily by a review of the literature, (b) discuss the contrasting geosynclinal aspects

[1] Publication authorized by the director, Oklahoma Geological Survey.

[2] Oklahoma Geological Survey.

of the Arbuckle and Ouachita regions, (c) estimate probable depth to Precambrian rocks in McCurtain County, Oklahoma, and (d) show that maximum facies differences are restricted to rocks of the pre-Trenton Middle and Early Ordovician, and to the Silurian and Early Devonian, while intervening and closing stratigraphic units extend through both regions without appreciable change in thickness and with slight, but certainly non-discriminating, change in facies.

The conclusion is reached that although certain facies differences are indeed real, they are not yet sufficiently well established in space to compel acceptance of the concept of large-scale thrusting. Conversely, if a well is drilled in the frontal belt of the Ouachita Mountains and encounters black shale facies in the Womble and Mazarn formations, then the contrast between them and their nearby Simpson and Arbuckle equivalents will be so great that sole thrusting will have been established. A second approach, tried but not yet attained, would be the drilling of a well through the Ouachita rocks and finding beneath them a thick sequence of Arbuckle carbonates, or possibly Simpson green shales and bioclastic limestones interbedded with petroliferous quartzose sandstones.

In the preparation of this paper I have depended heavily on published accounts, but in addition I am indebted to William D. Pitt for furnishing composite stratigraphic sections of Ouachita rocks in McCurtain County, Oklahoma, and in southwestern Arkansas. Thomas W. Amsden kindly gave composite sections of Hunton strata in the Arbuckle Mountains and discussed with me the problems of correlating them with the Blaylock and Missouri Mountain formations. A new approach to correlating the Early Ordovician formations by graptolites and associated fossils was given to me by W. B. N. Berry in discussions during July, 1955. His familiarity with the graptolite faunas has made possible a more precise correlation of upper and middle Arbuckle limestones with black shales of the Womble and Mazarn formations.

Stratigraphic Classification

The stratigraphic successions in the Arbuckle and Ouachita Mountains are given in Table 1, and are summarized graphically in terms of principal facies in the cross section of Fig. 1. Arbuckle Mountain strata below the Sycamore have a maximum thickness of 11,500 feet on the southwest flank of the Arbuckle anticline, in the southwestern part of the outcrop region. Equivalent strata are 6,800 feet thick in the eastern part of the Arbuckle Mountains, the loss of section being mostly the result of thinning of individual formations in a less rapidly subsiding shelf environment rather than loss by erosion at unconformities. Geosynclinal aspects are restricted entirely to rocks of Ordovician and Late Cambrian age, which are 10,800 feet thick in the west compared with 6,000 feet in the east, giving a west-east thickness ratio of nearly 2 to 1.

A direct comparison can not be made with the succession in the Ouachita Mountains because, as here interpreted, strata of Cambrian age are not exposed in that region. Pre-Stanley rocks of Ordovician (Early, Middle, and Late), Silurian, and Devonian age range in thickness from 6,300 feet in Arkansas to 3,800 feet in McCurtain County, Oklahoma. The westward loss in thickness results from the wedge-out of Blakely sandstone as well as from marked thinning in the Mazarn, Womble, and Blaylock formations. This thinning trend westward toward the edge of the Ouachitas at Black Knob Ridge is confirmed by the complete absence of Blaylock sandstone (Fig. 1), but can not be demonstrated for the Mazarn and Womble, as they are concealed or inadequately exposed.

In the Ouachita Mountains there are no Ordovician geosynclinal thicknesses to compare with those in the Arbuckle Mountains. Rather the pre-Stanley Ouachita rocks were deposited in a shallow offshore basin or shelf that was steadily being depressed from Early Ordovician through Late Devonian time. Both the rate of subsidence and the influx of clastic sediments were greater in Arkansas, yielding a thicker section of sedimentary rocks. In Oklahoma the shelf was sinking more slowly and was receiving a greatly reduced volume of sandstone and shale.

Correlation

Devonian and older rocks in the two regions are dated and correlated by normal marine invertebrates and by conodonts. As

System	Series	Group	ARBUCKLE MOUNTAINS	OUACHITA MOUNTAINS
Mississippian	Lower		Sycamore-Caney	Stanley
Mississippian	Middle ? and Upper		Woodford (C)	Arkansas novaculite: upper div. (U) middle div. (C) lower div.-Pinetop chert (U)
Devonian	Lower	Hunton group	Frisco (Oriskanian) (M) Haragan-Bois d'Arc (Helderbergian) (M)	Absent?
Silurian	Lower Middle Upper	Hunton group	Henryhouse (M) ------?------	Missouri Mountain (U)
Silurian	Lower Middle Upper	Hunton group	Chimneyhill (M) ------?------	Blaylock (G)
Ordovician	Upper		Sylvan (G)	Polk Creek (G)
Ordovician	Upper		Fernvale (M)	Bigfork (GM)
Ordovician	Middle		Viola (MG)	Bigfork (GM)
Ordovician	Middle	Simpson group	Bromide (M) Tulip Creek (M) McLish (M) Oil Creek (M) Joins (M)	Womble (G)
Ordovician	Lower	Arbuckle group, Upper	West Spring Creek (MG)	Blakely (G)
Ordovician	Lower	Arbuckle group, Upper	Kindblade (M)	Mazarn (G)
Ordovician	Lower	Arbuckle group, Middle	Cool Creek (M)	Crystal Mountain (U) Collier (U) Lukfata (U)
Ordovician	Lower	Arbuckle group, Middle	McKenzie Hill (M)------?------	
Cambrian	Upper	Arbuckle group, Lower	Butterly (M) Signal Mountain (M) Royer (U) Fort Sill (M)	not exposed
Cambrian	Upper	Tmbrd Hills group	Honey Creek (M) Reagan (U)	not exposed
Precambrian			granite and rhyolite	(M) typical marine fauna (G) graptolites (C) conodonts (U) unfossiliferous or virtually so

TABLE 1. PRE-STANLEY CLASSIFICATION AND CORRELATION IN ARBUCKLE AND OUACHITA MOUNTAINS

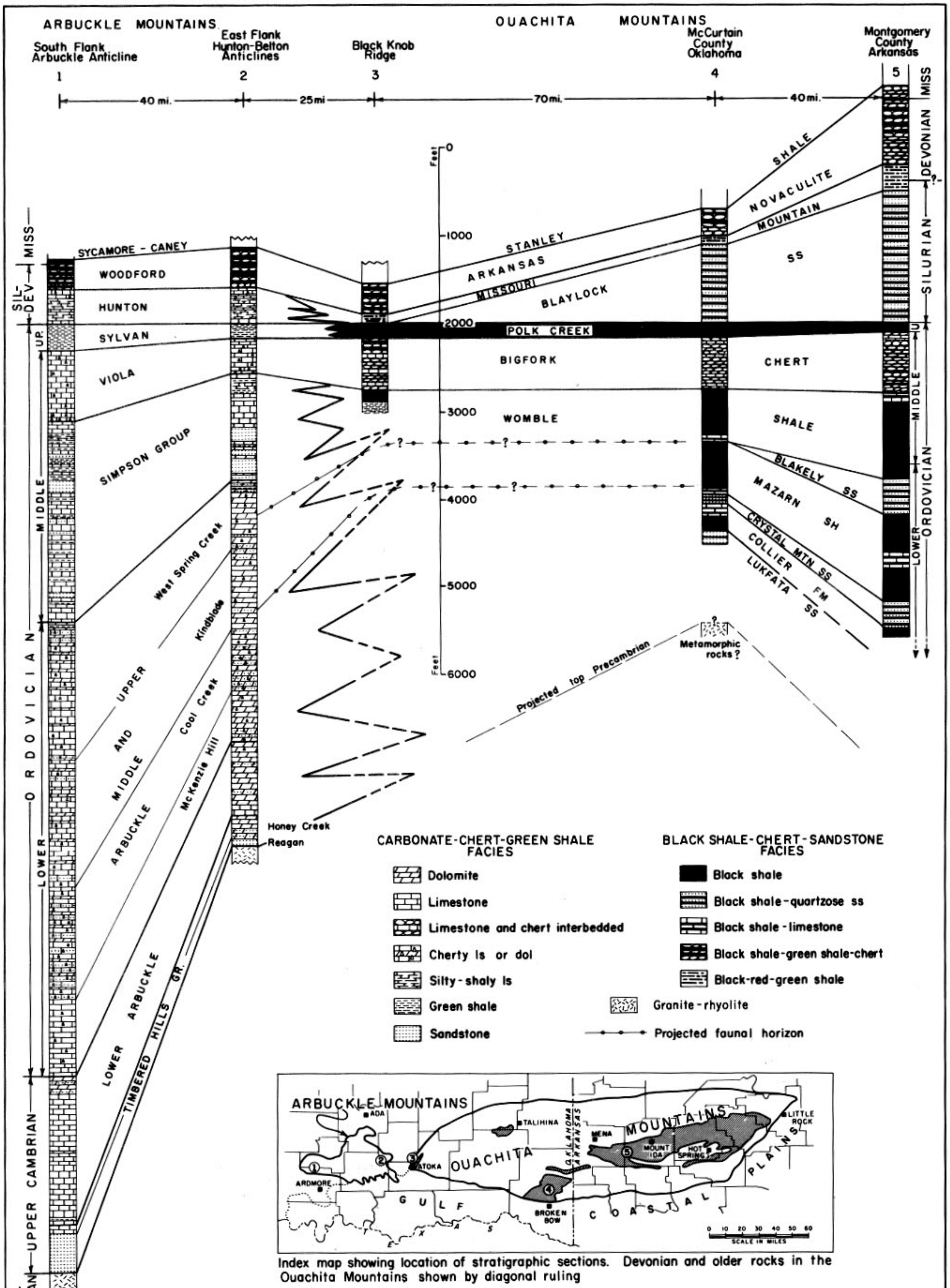

FIGURE 1. CORRELATION AND FACIES OF PRE-STANLEY ROCKS IN OUACHITA-ARBUCKLE MOUNTAIN REGIONS

shown in Table 1, conodonts have proved most valuable for the Devonian black shales, whereas graptolites in black shales of Ordovician and Silurian age from the Ouachita Mountains are most valuable for correlating with the richly fossiliferous limestones of the Arbuckle Mountains. Reasonable correlations can be made for all major stratigraphic units except (a) the Hunton limestone, whose probable equivalents in the Ouachitas are mostly unfossiliferous, and (b) the unfossiliferous basal few hundred feet below the Mazarn shale.

Woodford formation and Arkansas novaculite. The youngest rocks here considered are of Devonian and earliest Mississippian age. Through a study of conodonts Hass has shown that the middle division of the Arkansas novaculite of the Ouachita Mountains is essentially contemporaneous with the Woodford formation of the Arbuckle Mountains.

According to the interpretations of Haas (1951), based on the most exhaustive research yet done, the Arkansas novaculite at Caddo Gap, Arkansas, is 940 feet thick. The lower 785 feet or approximately 84 percent of the formation is of Devonian age, and the upper 155 feet or 16 percent is of Early Mississippian age. Most of the conodonts have been obtained from black shales in the middle division of the formation, which is 350 feet thick. The lower 320 feet of the middle division is of Late Devonian age, whereas the upper 30 feet is of Early Mississippian (Kinderhookian) age. The lower division is 465 feet thick and consists of unfossiliferous massive novaculite or chert; it is believed by Hass to be of Onondagan (early Middle Devonian) age. The upper division of the novaculite is 125 feet thick at Caddo Gap and likewise is sparingly fossiliferous, yet Hass cites good evidence for believing it to be late Kinderhookian or Osagean.

Hass (1956, p. 27-29) also has shown that the four conodont zones which characterize the middle division of the Arkansas novaculite are represented in the Woodford formation, indicating their age equivalence. The lower three of these are of Upper Devonian age, and each has been found in the Woodford at several localities in the Arbuckle Mountains. The top conodont zone is early Mississippian, and equivalents of this zone have been found in the top foot or two of the Woodford on Henryhouse Creek and near Ada, in the Arbuckle region. The Woodford formation in most parts of the Arbuckle Mountains is 300-400 feet thick and is entirely similar lithologically to the middle division of the Arkansas novaculite, as it consists of black shale, siliceous shale, and beds of chert.

The Arkansas novaculite extends from Arkansas into McCurtain County and into the frontal belt of the Ouachita Mountains, changing westward in thickness and locally in character. In Black Knob Ridge the formation is 360 feet thick (Hendricks, et al, 1937, p. 25). The lower 130 feet is massive and thin-bedded gray chert or novaculite considered to be the lower division of the formation. The upper 230 feet is shale and novaculite of the middle division, considered to be equivalent to the Woodford formation. The upper division, prominent in Arkansas, is not recognized. From the exposures in Black Knob Ridge it is clear that both the lower and middle divisions of the novaculite are thinner at the western margin of the Ouachita Mountains, having in general less than half their thickness of Caddo Gap, Arkansas. Such westward thinning also characterizes the pre-novaculite formations.

There is no major discontinuity between the Arkansas novaculite and the Woodford formation, either in terms of age or lithology, and the two formations are here considered to have been deposited as a blanket extending continuously within and between the Ouachita and Arbuckle Mountains, as well as extending northward and westward as a black shale facies over many hundred thousand square miles.

The principal unsolved problem concerns the lower division of the Arkansas novaculite and its equivalents, if any, in the Woodford formation. This unit in the Ouachita Mountains is almost wholly unfossiliferous. The nearest approach to an answer may be found in the Pinetop chert of Hendricks (1947, sheet 1), which crops out in a ridge in secs. 3 and 4, T. 2 N., R. 15 E. It contains some sparsely fossiliferous limestone beds and lenses, and is correlated with the lower division of the novaculite. T. W. Amsden (personal communication) has found post-Hunton carbonate rocks of somewhat similar character locally in the Arbuckle Mountains, and pending further faunal studies he has indicated a possible correlation between them and the

Pinetop chert.

Devonian strata of the Hunton group. Approximately the upper half of the Hunton group in the Arbuckle Mountains is of Devonian age. Through the early studies of Reeds (1911) and the more recent and much more comprehensive studies by Amsden, a sequence of fossiliferous limestones and marlstones locally 400 feet thick has been shown to be of Early Devonian age. According to Amsden (1957, 1958a, 1958b) the Haragan and Bois d'Arc formations are Helderbergian, whereas the Frisco formation at the top of the Hunton group is Deerparkian, close in age to the Oriskany sandstone.

This sequence has no lithologic counterpart in the Ouachita Mountains. If the lower division of the Arkansas novaculite is of Onondagan age as suggested by Hass, it is possible that Oriskanian and Helderbergian strata are absent, even in Arkansas where the stratigraphic succession is thickest.

Until better paleontologic control is available, possibly from studies of spores and pollen, little can be said about Early Devonian rocks in the Ouachitas. An unconformity at the base of the novaculite, corresponding to the well-known unconformity at the base of the Woodford, may have resulted in local removal of older beds. Reliable evidence for such an unconformity is not known in the Ouachita Mountains, however, and it seems more likely that Early Devonian strata are represented, either in the siliceous sediments low in the Arkansas novaculite or in the variegated shales of the Missouri Mountain formation. Although generally considered Silurian on the basis of stratigraphic position and a few poorly preserved fossils (Hendricks, 1947, sheet 1), the Missouri Mountain is not yet accurately dated; and the unconformity at its base, which in Black Knob Ridge results in removal of the Polk Creek shale (Hendricks, 1947, sheet 1), may mark the beginning of Devonian time.

Henryhouse-Chimneyhill and Missouri Mountain-Blaylock. Limestones and marlstones of undoubted Silurian age make up the lower part of the Hunton group in the Arbuckle Mountains. They have a composite thickness of approximately 350 feet, but owing to erosion beneath several unconformities within the group this composite thickness is nowhere attained at one locality. Locally all Silurian strata are absent, and in two areas they are built up to a maximum thickness of about 300 feet. The Chimneyhill and Henryhouse formations that make up this sequence are at some places richly fossiliferous, and according to Amsden (1957) they are probably of Early (Alexandrian), Middle (Niagaran), and early Late Silurian age.

Silurian faunas in the Arbuckle Mountains include probably at least a hundred species of brachiopods, trilobites, corals, sponges, Bryozoa, cephalopods, gastropods, graptolites, and pelecypods. These faunas are unknown in the Ouachita Mountains, where presumably equivalent beds are composed of sandstone interbedded with black and yellowish-gray shale of the Blaylock formation.

In Oklahoma the Blaylock sandstone is present only in McCurtain County, where it is 800-900 feet thick and consists of fine-grained chloritic quartzite interbedded with black, blue, and gray shale (Honess, 1923, pp. 87-99; Pitt, 1958, personal communication). A few worm trails cover some of the lower sandstone beds, but otherwise the formation is unfossiliferous. No Blaylock sandstone occurs in the frontal belt of the Ouachita Mountains, and it is clear that the formation disappears northward and westward both in Oklahoma and in Arkansas.

The Blaylock reaches its maximum thickness of 1,500 feet in the southern part of the Ouachita Mountains in Arkansas, and in that area it consists of angular fine-grained quartz and feldspar in sandstone layers interbedded with black and gray shale (Miser and Purdue, 1929, pp. 42-44). The only fossils useful for dating the Blaylock are graptolites found in shale in the lower part of the formation at one locality on Blaylock Mountain. From the collection made there by Miser, Ulrich identified seven species of graptolites, in the genera *Monograptus*, *Dimorphograptus*, *Gladiograptus*, and *Dictyonema*, by which he made a correlation with the Birkhill shales of Scotland, considered in Great Britain to be the base of the Silurian system (Miser and Purdue, 1929, p. 45). The Birkhill graptolite fauna is otherwise unknown in America.

Because most of the Blaylock formation is unfossiliferous, there remains a great question as to its equivalents in the Hunton group. Where present in maximum development there is no direct evidence for unconformity within the Blaylock, and presumably it is wholly of Silurian age, equivalent to the Chimneyhill and perhaps to the Henryhouse.

Overlying the Blaylock is the Missouri Mountain formation, which is everywhere present at its expected stratigraphic position in the Ouachita Mountains. In distribution it is similar to the Arkansas novaculite, as both of them are widespread blanket deposits that cover extensive tracts. They are completely unlike the previously deposited Blaylock sandstone, whose occurrence depends mostly on rate of basin sinking and the supply of clastic sediments from a southeastern source.

The Missouri Mountain formation ranges in thickness and character as follows: 50 to 300 feet of black, red, and green slates or shales in Arkansas (Miser and Purdue, 1929, pp. 45-49); 60 to 100 feet of black and gray slaty shale interbedded with thin greenish-gray sandstone in McCurtain County, Oklahoma (Honess, 1923, pp. 104-109); and 110 feet of green siliceous shale and interbedded chert in the Black Knob Ridge area (Hendricks, 1947, sheet 1). All strata are unfossiliferous except "a few conodonts, spicules, and fragments of brachiopods..., together with one bryozoan" from the upper part of the formation at Black Knob Ridge (Hendricks, 1947). The Missouri Mountain clastic sediments thus contain some faunal elements, but so far there is insufficient evidence to establish a reliable age. Possibly the variegated shales are equivalent in part to the silty and shaly limestone of the Henryhouse, and possibly they are younger than Silurian.

Regardless of the lack of accurate correlations, it is clear that the calcareous fossiliferous rocks low in the Hunton group are not faunally or lithologically represented in the Ouachita Mountains, and that contrasting lithologies and sedimentary environments characterize the Arbuckle and Ouachita provinces during Silurian time.

Sylvan-Polk Creek. In Arkansas and in McCurtain County, Oklahoma, the Polk Creek is mostly black shale or slate approximately 100 feet thick. It is described as "...coal-black, graphitic, firmly indurated but soft slate and shale..." by Honess (1923, p. 81), and "...black, fissile, and carbonaceous..." by Miser and Purdue (1929, p. 40). Where present in Black Knob Ridge, i.e., not removed by pre-Missouri Mountain erosion, it is at least 139 feet thick and consists of hard black paper shale that grades upward into soft brown platy shale (Hendricks, 1947, sheet 1).

In the Arbuckle province, the corresponding Sylvan shale in mostly 150 to 300 feet thick, thickening progressively westward, and consists of grayish-green shale that is calcareous in the lower part. In addition to a well-known fauna of graptolites, Wilson (1958) has recently discovered chitinozoans and hystrichosphaerids in this formation.

If all the stratigraphic units of the Arbuckle and Ouachita Mountains were as distinctive and as certainly correlatable as the Sylvan and Polk Creek formations, the only problems remaining would be those of structural interpretation. Both the black shales and slates of the Polk Creek and the green shales of the Sylvan have yielded an Upper Ordovician graptolite fauna that establishes the two formations as equivalents. When Decker published his comprehensive report describing nine species in five genera, including the guide fossil Dicellograptus complanatus, he cautiously wrote "... as nearly all graptolites of the Sylvan shale have been found in the Polk Creek shale of Arkansas, the practical equivalence of large parts of these formations seems to be established..." (Decker, 1935, p. 699). Thus from their contemporaneity and the knowledge that the same general range of thickness is shown for each formation in each province, it is perfectly clear that during Late Ordovician time a single deposit of shale extended uninterruptedly across southern Oklahoma and Arkansas, changing westward from black to green. Substantially the same deposit extends even farther northward as the Sylvan shale in northeastern Oklahoma (Huffman, 1953; Decker and Huffman, 1953), the Cason shale of northern Arkansas, and the Maquoketa shale of Iowa, Missouri, Illinois, Minnesota, and Wisconsin. The Polk Creek-Sylvan-Cason-Maquoketa sequence is almost as widely distributed as the black shales in the Arkansas novaculite, Woodford, Chattanooga, New Albany, Antrim, and Ohio shale. For each of these periods of time, uniform environmental conditions and a slow steady rate of subsidence characterized regions nearly continental in extent. Against this background it is hardly conceivable that the Arbuckle and Ouachita provinces occupied separate basins during the Late Ordovician and Late Devonian---certainly during these epochs there was no barrier between them.

Viola-Bigfork. That the Viola limestone of the Arbuckle Mountains is equivalent to most or all of the Bigfork chert in the Ouachita Mountains is now better established than the early works of Decker (1936a, 1936b) would indicate. Of the more recent workers both Hendricks and Harlton correlate these units unquestionably. In Black Knob Ridge, according to Hendricks (1947, sheet 1) "The fauna of the Bigfork is composed of graptolites, small brachiopods, trilobites, conodonts, spicules of various sorts, and small bodies which may be radiolarians. The graptolites occur at several horizons throughout the formation, and for the most part they are conspecific with those found in the Viola limestone of the Arbuckle region." An even clearer concept is given by Harlton (1953, p. 785), who wrote: "The fauna of the Bigfork chert at Black Knob Ridge is decisively conspecific with that found in the Viola limestone of the Arbuckle region. The fauna consists of trilobites, brachiopods, conodonts, and abundant graptolites. From a faunistic viewpoint it is clear that the chronological ages of the Bigfork and the Viola are identical. It can logically be held that these deposits are synchronous and their origin was in the same sea."

The fossils indicate a Middle Ordovician (Trenton) age in the Arbuckle Mountains and at Black Knob Ridge, and Ulrich likewise gave a Trenton age from graptolites in the Bigfork in Arkansas (Miser and Purdue, 1929, pp. 38-39). Decker(1936a, 1936b) has correlated the lower part of the Viola with the upper parts of the Womble in Arkansas and the Stringtown shale (now called Womble) at Black Knob Ridge, but the more recently expressed views of Hendricks and Harlton are probably correct and are accepted by the present writer.

The Viola-Bigfork stratigraphic unit is a valuable datum, ranging in thickness from 400 to 800 feet over thousands of square miles, and changing but slightly in lithology from black and gray chert and cherty limestone in the core area of the Ouachita province, to gray chert and limestone at Black Knob Ridge, and to gray limestone containing disseminated silica and nodules of chert in the Arbuckle province. The statement of Harlton that the two formations are deposits in the same sea is fully acceptable. The only notable change in lithology, other than the westward loss in percentage of silica, is the westward disappearance of black chert and black shale interbeds.

Both in Arkansas and in McCurtain County, Oklahoma, the Bigfork formation is about 600-700 feet thick, as near as can be estimated, and consists of coal-black chert and gray to black limestone or cherty limestone, together with thin interbedded layers of black siliceous and carbonaceous shale (Miser and Purdue, 1929, pp. 36-37; Honess, 1923, pp. 70-79). In the detailed section measured by Honess (1923, pp. 74-79), 287 feet of the Bigfork is so well exposed that a bed-by-bed description is given, and of this thickness 90 feet or approximately one-third of the exposed sequence is limestone or cherty limestone.

In Black Knob Ridge the formation is consistently about 600 feet thick. A section measured at Grant's Gap shows that gray bedded chert and limestone or cherty limestone constitutes approximately the lower 400 feet, in a ratio of about 80 percent chert to 20 percent limestone, and that nearly 200 feet at the top is composed of black and dark brown laminated chert containing interbedded black shale (Hendricks, Knechtel, and Bridge, 1937, p. 23).

The corresponding Viola limestone ranges in thickness from 400 feet in the eastern part of the Arbuckle Mountain region to 800 feet in the southwestern part. Wengerd (1948) has published a detailed account of this formation, showing that (1) it thickens regularly toward the southwest, and (2) the percentage of insoluble residue, which is mostly chert, disseminated silica, and quartz sand grains, changes from 5-10 percent in the east to about 15 percent in the west. Fine- and medium-grained limestone, much of it richly fossiliferous, makes up most of the formation. The basal 50 to 100 feet of the Viola in most of the outcrop sections is composed of laminated limestone containing approximately 25 percent disseminated silica. It weathers by leaching of carbonate minerals to a highly porous tripoli rock. This distinctive unit contains numerous graptolites and specimens of the lace-collar trilobite Cryptolithoides. It rests with abrupt contact on fine-grained limestone of the Bromide formation, which contains no chert or disseminated silica, practically no graptolites, and no Cryptolithoides. All these relations indicate that an unconformity of some magnitude is present between the Viola and Bromide formations.

The Fernvale limestone, presumably of Late Ordovician age, lies above the Viola limestone and below the Sylvan shale in the Arbuckle Mountains. It has a maximum thickness of about 65 feet at Lawrence, where the coarse-grained limestone is well exposed in quarries of the Ideal Cement Company. The beds contain a wealth of fossils, chiefly brachiopods and trilobites. This fauna has not been investigated thoroughly, and correlation of the Fernvale into the Ouachita sequence has not been demonstrated. As no coarsely crystalline bioclastic limestone of the Fernvale type is present between the Bigfork and Polk Creek formations, the Fernvale horizon is either absent or is represented by a completely different lithology.

Simpson-upper Womble. The major contrasts in facies between the Arbuckle province and the Ouachita province are in rocks of pre-Trenton age. In the Arbuckles most pre-Viola strata are limestones and dolomites, together with some green shales and sandstones, whereas in the Ouachitas most pre-Bigfork strata are black shales and sandstones, together with a poor representation of carbonate rocks. A fair measure of paleontologic control is available for correlating between the two provinces, but much remains to be learned. It is unfortunate that exposures in the Ouachita Mountains are not geographically better distributed, for aside from short sections of the Womble shale in Black Knob Ridge and the Potato Hills, there are no outcrops of pre-Trenton beds closer to the Arbuckle Mountains than central McCurtain County, an airline distance of 100 miles.

In the Ouachita Mountains the Womble shale underlies the Bigfork chert, apparently with a conformable contact, and overlies the Blakely sandstone, or the Mazarn shale where the Blakely is absent. The Womble in Arkansas is mainly black shale about 1,000 feet thick, containing thin sandstone beds throughout and, in the upper part, some beds of black chert as well as sporadic lenses of black limestone (Miser and Purdue, 1929, p. 32). Thirty-three species of graptolites from excellent exposures of the formation at Crystal Springs, Arkansas, were identified by Ulrich and assigned a Normanskill or Late Chazyan age (Miser and Purdue, 1929, pp. 34-35). This fauna has been found at several stratigraphic positions of the Womble in Arkansas, indicating that much of it, probably all but the basal part, is Middle Ordovician.

No fossils have been found in corresponding beds of McCurtain County, Oklahoma, where strata above the Mazarn and below the Bigfork are called Womble and are estimated to be about 600 feet thick (Pitt, 1958, personal communication). Most of the formation consists of poorly exposed black fissile shale and siltstone, but the basal 66 feet is well exposed and is built up of limestone and silty limestone interbedded with black shale and siltstone (Pitt, 1955, pp. 24-25).

The upper 260 feet of the Womble is exposed in Black Knob Ridge. It consists of an upper unit of black and brown partly cherty shale 80-95 feet thick, and a lower unit with an exposed thickness of 166 feet that is made up of brown to green clay shale containing thin beds and lenses of sandstone (Hendricks, 1947, sheet 1). The green shales and rounded-frosted-pitted quartz grains in the sandstone of this area are like those of the Simpson group, but the black shales do not occur in any Simpson formations of the Arbuckle Mountains.

The Bromide formation at the top of the Simpson group is one of the most fossiliferous stratigraphic units of the Arbuckle province. It consists of light gray limestone, green shale, and quartzose sandstone 300-450 feet thick, and contains an especially prolific brachiopod fauna. Of all the excellent exposures of the Bromide throughout the Arbuckle province, graptolites occur at only one locality, Rock Crossing in the Criner Hills, where four species were found in the upper 25 feet of the formation. They are Diplograptus maxwelli, Dicellograptus gurleyi, and two species of Dictyonema (Decker, 1943, pp. 1388-92). The first two species occur also in the Viola limestone. The genus Dictyonema has not been recorded from the Viola, but was identified by Ulrich in the Womble at Crystal Springs, Arkansas (Miser and Purdue, 1929, p. 34).

Decker made an exceptionally strong point of the occurrence of Diplograptus maxwelli and Dicellograptus gurleyi both in the lower Viola and upper Bromide, and chiefly on this basis he correlated all the Bromide with the Trenton (Decker, 1952, p. 135 ff). This correlation is in disagreement with the views of other writers. The exhaustive study by Cooper of Chayzan and related brachiopods shows a faunal discontinuity between Viola and Bromide strata, which is expectable from the physical relations cited above, and it gives a classification of the Bromide as pre-Trenton and post-Chazyan, close in age to what earlier geologists called Black River (Cooper, 1956, p. 123, chart 1).

From the sharply contrasting nature of the Bromide and Womble faunas, great difficulties can be expected in an attempted cor relation between them. The graptolite-bearing shales of the Womble do not contain the brachiopods of the Bromide, and there are not enough graptolites in the Bromide to establish a faunal tie. The best approach to date is that of Harlton (1953, p. 780), who states that the conodonts and ostracodes of the Womble at Black Knob Ridge are characteristic Black River or Bromide types. Harlton correlated the upper Womble with the Bromide formation, and this view is accepted by the present writer.

Ouachita equivalents of the pre-Bromide formations of the Simpson group have not been established through faunal control. The Tulip Creek, McLish, and Oil Creek formations make up two-thirds of the Simpson group in the Arbuckle Mountains, and their fossiliferous limestones and shales have yielded a brachiopod fauna of early Middle Ordovician age (Cooper, 1956, chart 1). Graptolites have not been found in this sequence, and thus there is no direct means of correlation with the graptolite shales of the middle Womble. The exhaustive work on ostracodes by Harris (1957) has established well-marked zones of Chazyan and Black River age for all formations of the Simpson group; and if these ostracode zones can be found in the Womble, a satisfactory correlation probably can be made.

The discovery of a Lower Ordovician graptolite fauna at one locality near the base of the Womble shale in Arkansas (Miser and Purdue, 1929, p. 33) has considerable value in offering a possible correlation, for the Simpson group is considered to be entirely of post-Canadian (Lower Ordovician) age. The position of early Middle Ordovician rocks thus would be middle Womble, which presumably includes quivalents of the Tulip Creek, McLish, Oil Creek, and probably Joins formations.

Upper Arbuckle and lower Womble-Blakely-Mazarn. Early Ordovician strata of the Arbuckle-Ouachita provinces are exposed in widely separated areas, and lithologically they are the most unlike of all the pre-Stanley formations. The section of massive carbonate rocks 5,200 feet thick, that makes up the Lower Ordovician part of the Arbuckle group, is completely unlike the corresponding section of black shales and sandstones of the Ouachita Mountains. Despite the lithologic dissimilarity, the assemblages of fossils are useful for making general correlations.

The very thick Arbuckle group of the Arbuckle Mountains as shown in Fig. 1 is divided into three roughly equal divisions--upper, middle, and lower. The upper division is discussed in this section. It embraces approximately the upper half of Early Ordovician time, and consists of the West Spring Creek and Kindblade formations (Ham, 1955). Each is composed of limestone and dolomite of shallow-water marine origin. The deposits are fossiliferous, containing brachiopods and gastropods at many horizons, as well as a few restricted zones of trilobites, cephalopods, sponges, and graptolites. In combined thickness the two formations range from 3,000 feet in the southwestern part of the Arbuckle Mountains to 1,700 feet in the eastern part, changing eastward from limestone into dolomite. The sponge *Archaeoscyphia* and the brachiopod *Tritoechia* characterize the lower part of the sequence, and the brachiopods *Pomatotrema*, *Syntrophopsis*, *Polytoechia*, and *Diparelasma*, together with two species of *Didymograptus*, characterize the upper part. Horn-shaped opercula of the gastropod genus *Ceratopea* are restricted to the Kindblade and Cool Creek formations, in which they occur from bottom to top in six well-defined zones (Yochelson and Bridge, 1957). None of these fossils occurs in the Ouachita facies, except *Didymograptus*.

The upper division of the Arbuckle group is here considered to be equivalent to the lower part of the Womble, all the Blakely sandstone, and all the Mazarn shale, as the available faunal evidence indicates for these beds a middle and late Early Ordovician age. The critical information regarding these strata is in Arkansas, where the Womble and Mazarn are chiefly black shale units each about 1,000 feet thick, separated in the central part of the outcrop region by the Blakely sandstone, which has a maximum thickness of 400 feet and consists of sandstone interbedded with black shale in the ratio of 1:3 (Miser and Purdue, 1929, pp. 25-33). Each of the three formations contains a graptolite fauna of Canadian or Beekmantown age. The faunal lists of E. O. Ulrich, cited in Miser and Purdue, have been reviewed by W. B. N. Berry and put in terms of more detailed modern knowledge about stratigraphic sequence and age.

In the correlation of this sequence with a part of the Arbuckle group, the upper limit is established by a collection of graptolites from beds near the base of the Womble shale in the Hot Springs district, the fauna of nine species being correlated with that of a part of the Beekmantown of New York, and thus of Early Ordovician age (Miser and Purdue, 1929, p. 33). In the Ouachita province this is the highest known occurrence of Lower Ordovician fossils, and it serves to correlate the lowest part of the Womble with the upper part of the West Spring Creek formation of the Arbuckle Mountains.

The next lower faunal control consists of graptolites found in shale near the middle of the Blakely sandstone in Blakely Mountain, in the Hot Springs district. The collection of 23 species is characterized by the genera Didymograptus, Phyllograptus, Tetragraptus, and Glossograptus (Miser and Purdue, 1929, pp. 30-31), and the species list according to Berry indicates Late Canadian age. In terms of Arbuckle rocks, this horizon probably is represented in the lower half of the West Spring Creek formation.

The most significant graptolite fauna, and the lowest fossil zone found thus far in the Ouachita Mountains, is from the lower part of the Mazarn shale. It was collected at a locality two miles northeast of Womble, Arkansas, and contains twelve species of graptolites in the genera Tetragraptus and Didymograptus (Miser and Purdue, 1929, pp. 27-28). One of the species is Tetragraptus fruticosus, which with the associated forms definitely marks a well-developed Middle Canadian zone 250 feet thick in the Marathon limestone, just above the Monument Springs dolomite, in the Marathon region of southwest Texas (Berry, personal communication, 1955). Although this Tetragraptus fauna has never been found in the Arbuckle limestone, two fossils from the Monument Springs dolomite in Texas are so well known stratigraphically in the Middle Canadian part of the Arbuckle group that a correlation can be made with considerable assurance. A sponge identified either as Calathium or Archaeoscyphia, and the brachiopod Diaphelasma, occur in the Monument Springs (Bridge, in Sellards et al., 1932, p. 234; and Cloud and Barnes, 1948, p. 66). In the Arbuckle and Wichita Mountains, Diaphelasma occurs only in the upper third of the Cool Creek formation, and Archaeoscyphia is restricted to the upper part of the Cool Creek and lower third of the Kindblade. From these occurrences it is clear that the Monument Springs dolomite member of the Marathon limestone is equivalent to beds in the upper Cool Creek and lower Kindblade of Oklahoma, and that the Tetragraptus fruticosus fauna of the Mazarn and Marathon formations lies stratigraphically in the middle part of the Kindblade formation. Thus the lower part of the Mazarn shale is equivalent to strata just above the middle of the Early Ordovician part of the Arbuckle limestone.

The Blakely sandstone disappears westward in Arkansas and likewise is absent in McCurtain County. In the same areas the Mazarn also becomes thinner, ranging from about 1,000 feet in Arkansas (Miser and Purdue, 1929, p. 26) to about 600 feet in McCurtain County (Pitt, 1955, p. 23).

Older strata. In the Arbuckle and Wichita Mountains, the pre-Kindblade stratigraphic section is mostly fossiliferous and is well exposed down to Precambrian igneous rocks. The section is divided into middle Arbuckle strata, comprising the Cool Creek and McKenzie Hill limestones and dolomites, of Early Canadian age; lower Arbuckle limestones and dolomites of Late Cambrian age; and the Timbered Hills group, also of Late Cambrian age, made up of the Reagan sandstone overlain by the Honey Creek formation. In present knowledge none of these units can be correlated with certainty into the Ouachita province, for there the sandstones, black shales, and thin limestones are unfossiliferous.

Below the Mazarn is the Crystal Mountain sandstone, about 300 feet thick in western Arkansas (Pitt, 1958, personal communication) and not more than 100 feet thick in McCurtain County (Pitt, 1955, p. 21). A thickness of 850 feet has been reported for the formation in Arkansas by Miser and Purdue (1929, p. 24). It is composed mostly of fine-grained and medium-grained sandstone, cemented either by silica or by calcium carbonate, and is interbedded with thin layers of black shale. A basal conglomerate 5-10 feet thick, made up of limestone and chert pebbles in a calcareous sandstone matrix, occurs both in Arkansas and in McCurtain County.

Pitt (1955, p. 18) has compared the Crystal Mountain sandstone "...with the McLish sandstone of the Simpson group because both are pure quartz sandstone and both appear to be in about the same stratigraphic position." Evidence from fossils as cited above does not support this comparison, but indicates instead that the Crystal Mountain formation is pre-Womble and pre-middle Canadian in age, in conformance with the opinion of Miser and Purdue (1929, p. 25). As the Tetragraptus fruticosus fauna is in the lower part of the Mazarn, just above the Crystal Mountain, and as the T. fruticosus zone equivalent probably lies in the middle Kindblade, it follows that the Crystal Mountain equivalent is Lower Kindblade or Cool Creek, low in the middle Canadian part of Early Ordovician time.

Below the Crystal Mountain sandstone is the Collier formation, made up principally of black shale and fine-grained dark limestone. This is the oldest formation exposed in Arkansas. According to Miser and Purdue (1929, p. 23), "The thickness that is exposed in the Caddo Gap quadrangle is 200 feet, but much more, probably several hundred feet, is exposed in the Mount Ida quadrangle, on the north."

In McCurtain County the thickness of the Collier is about 300 feet, divided into an upper limestone member 50-150 feet thick, consisting of silty fine-grained bluish-gray limestone interbedded with thin layers of sandstone and black shale, and a lower shale member 180 feet thick. These figures and a detailed measured section of the limestone member are given by Pitt (1955, pp. 15-18). It is the writer's observation that some of the carbonate beds in the Collier are composed of dolomite rather than limestone, and that the rocks of this sequence begin to approach the normal lithology of the Arbuckle group, except that the black shale interbeds and the high silt content of the carbonate rocks are lacking in the Arbuckle province.

In the absence of fossils, the correlation of the Collier can only be surmised. A Cambrian age was suggested by Miser and Purdue (1925, p. 25), but the writer believes that the formation is Early Ordovician, probably equivalent to the lower part of the Cool Creek formation of the Arbuckle group.

According to the account by Pitt (1955, pp. 13-15), the oldest formation exposed in the Ouachita Mountains is the Lukfata sandstone. So far as now known, it crops out only in central McCurtain County, where it lies under the Collier shale and has an exposed thickness of 150 feet. The formation consists mostly of black fissile shale and thin-bedded limestone in the lower part, and of fine-grained sandstone interbedded with black and green shale in the upper part. No fossils have been found in it, but the writer believes on the basis of stratigraphic position that the Lukfata sandstone is near the base of the Lower Ordovician, probably correlating with a part of the McKenzie Hill formation of the Arbuckle Mountains.

An Interpretation of Depth to Precambrian in McCurtain County

Under the stratigraphic concepts outlined above, the view has been expressed that even the oldest rocks--those of the Collier and Lukfata formations--are of Early Ordovician age and are probably to be correlated with the middle division or early Canadian part of the Arbuckle group. It is further shown that the base of the Womble is approximately equivalent to the top of the Arbuckle group. With the excellent exposures in the Arbuckle Mountains available for thickness comparisons, it is possible to extrapolate the top of the Precambrian from the Arbuckle province into the Ouachita province at the outcrop locality of the Lukfata sandstone in McCurtain County. The extrapolation can be only approximate, and it is based further on the assumptions that (a) no major unconformities shorten the section beneath the Lukfata, and (b) the unexposed rocks have essentially the same sedimentation rates as the exposed rocks.

As deduced from the stratigraphic occurrence of faunal zones of Early Ordovician age, the base of the Womble shale is approximately equivalent to the top of the West Spring Creek formation, and the base of the Mazarn shale is approximately equivalent to the middle Kindblade. The thickness for this dated interval of the Arbuckle limestone, in the eastern part of the Arbuckle Mountains, is about 1,500 feet, and the corresponding interval in McCurtain County is about 600 feet. The thickness ratio is 2 1/2 : 1. In the eastern Arbuckles the remaining section down to the top of the Precambrian has a thickness of 2,500 feet. To extrapolate from this thickness, the top of the Precambrian in McCurtain County would be 1,000 feet stratigraphically below the base of the Mazarn. A section of 600-foot thickness already is exposed in the Collier and Lukfata formations, leaving under this concept an unexposed thickness of only 400 feet. Presumably these concealed rocks would be entirely of Late Cambrian age.

Approached from a different viewpoint, the extrapolated unexposed thickness has the same order of magnitude. If the Lukfata sandstone is at or near the base of the Lower Ordovician, as suggested above, the 1,200-foot thickness of Cambrian rocks in the eastern part of the Arbuckles can be projected in pre-Lukfata terms; and at the ratio of 2 1/2 : 1 the estimated depth to Precambrian beneath the Lukfata in McCurtain County is about 500 feet.

Some support for the thickness ratios used above is given by the thickness of all Ordovician rocks in the two areas, the ratio being 4,800 : 2,500 or about 2 : 1. This difference would not have material effect on the extrapolations. A further complication is the knowledge that the Precambrian surface is highly irregular in some parts of the Arbuckle Mountains, having a known relief of about 950 feet (Ham, 1949, p. 47). Some allowance must be made for such irregularity, and in Fig. 1 the pre-Lukfata thickness is shown as approximately 1,000 feet.

The writer knows of no basis for assuming that the basement rocks are necessarily granites. Granites and rhyolitic lavas of Precambrian age crop out in the Arbuckle Mountains; but an elongate tract of Precambrian schists, quartzite, metagraywacke, and meta-arkose is known in subsurface along Red River westward from Love County (Flawn, 1956, Plate I). This tract may extend eastward into the Ouachita province, and, if the basement in McCurtain County consists of layered metamorphic rocks or layered volcanic rocks, gravity and seismic readings in that area would be difficult to interpret.

Summary and Conclusions

Stratigraphic correlations and regional interpretations of pre-Stanley rocks of the Arbuckle-Ouachita provinces have been advanced remarkably in the half-century during which these regions have been investigated. In perspective view, it appears reasonable that there are no major differences in age or in lithology during Viola-Bigfork, Sylvan-Polk Creek, and Woodford-Arkansas novaculite times; and that within these time periods, entirely similar sediments were deposited in open seaways that connected the Arbuckle and Ouachita provinces. Moreover, it is equally reasonable that considerably different sediments were being deposited simultaneously in the two provinces, but at different rates, first during Lower and early Middle Ordovician time and later during Silurian and Early Devonian time. At such times the respective provinces were (a) either widely separated, the change in facies between them taking place gradually, in which case large-scale thrusting is called for, or (b) the sediments grade into each other within a short distance, the facies change being possibly aided by a barrier of unknown character and position, in which case thrusting is not needed.

From field occurrences alone it does not appear possible to solve this problem, for there are insufficient exposures of beds in the critical frontal belt where the Arbuckle and Ouachita facies are closest. Available for contrast in outcrops of Black Knob Ridge are only the Missouri Mountain shale and the upper part of the Womble formation. If the lower Womble and the Mazarn of that area can be shown by drilling to be dominantly black shales, it would appear unreasonable to the writer that these strata could grade so quickly westward into the Arbuckle facies, and the concept of thrusting should be invoked. If instead the Womble and Mazarn were found to be interfingering or interbedded green shales, dolomites, and limestones, then a gradation into Arbuckle facies will have been demonstrated. A test probably less than 1,000 feet deep, drilled on the Womble outcrop just off Oklahoma Highway 3, one mile southeast of Atoka, would have a fair chance to solve this fascinating enigma.

Another concept that is being clarified is the unequal geosynclinal behavior of pre-Stanley rocks in the Ouachitas compared with those in the Arbuckles, the sediments in the two areas reflecting unequal rates of accumulation during specified times. In pre-Silurian time, by far the thicker section was laid down in the Arbuckle province, where the Ordovician-Upper Cambrian section of predominantly carbonate rocks reaches a near-geosynclinal thickness of 10,800 feet. The thickest corresponding section in the Ouachitas is mostly clastics and probably is about 5,000 feet--a figure that is only an approximation because the lowest beds, which are not exposed, must be estimated by extrapolation. In contrast, the section of Silurian and Devonian rocks has a maximum thickness of 2,700 feet in the Arkansas part of the Ouachitas, compared with a maximum thickness of 900 feet in the Arbuckles. Thus the end of Polk Creek-Sylvan time at the close of the Ordovician marks a time boundary below which maximum sinking was in the Arbuckle province, and above which maximum sinking was in the Ouachita province.

The difference in subsidence behavior for the two regions also is shown through thicknesses given by rock systems (Fig. 1). In Arkansas the sediment thicknesses are approximately 3,600 feet for Ordovician time, 1,800 for Silurian, and 900 for Devonian, and the Ordovician : Silurian : Devonian thickness ratio is 4 : 2 : 1. These data bear no similarity to those in the west, for in the thickest corresponding section of the Arbuckle Mountains the Ordovician : Silurian : Devonian thicknesses are 8,500 : 350 : 800 feet, in a ratio of approximately 25 : 1 : 2.

The Arbuckle province thus consists almost exclusively of an Ordovician geosynclinal tract. The sea floor of this province during Ordovician time was sinking so rapidly that it was literally falling, whereas in the Silurian and Devonian it was by comparison just barely sinking, and received thin layers of sediments. The dissimilar Ouachita province sank progressively and at a reasonably constant rate not only in Ordovician time but in Silurian and Devonian time as well.

Another regional feature worthy of discussion is that the pre-Stanley rocks thin markedly eastward in the Arbuckles and markedly westward in the Ouachitas, as though they were merging toward a common shelf between them (Fig. 1). In the Ouachita province the thinning is due partly to a slower rate of subsidence, but in considerable part it is also due to the loss of clastic sediments away from the source area, which lay presumably in Llanoria to the southeast. Both the Blaylock and Blakely sandstones follow this pattern, as they are thickest in the east and disappear westward. The kind of sediment deposited at any given time, however, does not change materially in character from west to east, as near as can be determined from outcrop control.

The Arbuckle province on the other hand is characterized, during Cambrian-Ordovician time, by a modest change in lithology from limestones in the west to dolomites in the east, and by an accompanying change in thickness from about 7,500 feet to 4,200 feet. The Simpson group and Viola limestone similarly thin eastward, but the only pronounced change in lithology is the occurrence in the McLish formation of the algal "birdseye" limestone instead of bioclastic calcarenite (Ham, 1954; 1955, pp. 29-30). The Woodford shale does not differ significantly in thickness or character within the Arbuckle province, but the thickness of formations in the Hunton group does show a considerable range, primarily as the result of five intra-group unconformities in addition to a pronounced unconformity at the base of the overlying Woodford (Amsden, 1957, p. 6). The essential point is that pre-Hunton strata make up approximately 87 percent of the pre-Stanley stratigraphic section in the eastern part of the Arbuckles, and that these strata change slightly in lithology while undergoing a thickness loss of nearly 50 percent, evidently by less deposition on a more slowly subsiding shelf. It is not known whether this dominantly carbonate facies of the eastern Arbuckles actually abuts against dissimilar rocks in the Ouachitas near Atoka, as most of the equivalent rocks are not exposed.

Finally, the term "Ouachita geosyncline" should be used in reference to the Stanley, Jackfork, and Atoka sediments of the Ouachita province, as this sequence according to Hendricks (1947, sheet 1) has the geosynclinal thickness of about 28,000 feet and was deposited during the relatively short span of Mississippian and early Pennsylvanian time. The term "Ouachita geosyncline" should not be applied to the pre-Stanley rocks of the Ouachita province, because they are not geosynclinal in thickness, even in their thickest exposures, but were spread as a westward-tapering blanket on a moderately subsiding shelf. The general increase in thickness toward the east suggests that the Ordovician, Silurian, and Devonian sediments are expanding toward a geosynclinal site, which probably had the form of a linear trough in the foredeep of Llanoria. The sediments there presumably were deposited mainly as conglomerates, coarse sandstone, and shale with a thickness magnitude of perhaps 20,000 feet. This hypothetical trough would constitute a pre-Stanley Ouachita geosyncline, of which the exposed Devonian and older black shales and associated sediments represent deposits on a fringing shelf. Collapse, metamorphism, and uplift of this early Paleozoic geosynclinal tract is seen as the source of sediments needed for the filling of the newly developing late Paleozoic geosyncline. Obviously the keels of the two geosynclines lie in different geographic positions, the earlier being now deeply buried under a cover of Cretaceous and younger sediments in northern Louisiana and northeast Texas, and the younger coinciding approximately with the present position of the Ouachita Mountains.

References

Amsden, T. W., 1957, Stratigraphy and paleontology of the Hunton group in the Arbuckle Mountain region: Introduction to stratigraphy: Oklahoma Geol. Survey, Circular 44, 57 pp.

Amsden, T. W., 1958a, Ibid: Haragan articulate brachiopods: Oklahoma Geol. Survey, Bull. 78, pp. 1-144.

Amsden, T. W., 1958b, Ibid: Bois d'Arc articulate brachiopods: Oklahoma Geol. Survey, Bull. 82, 110 pp.

Berry, W. B. N., oral communication, July, 1955.

Cloud, P. E., Jr., and Barnes, V. E., 1948, The Ellenburger group of central Texas: Univ. of Texas, Publication 4621, 473 pp.

Cooper, G. A., 1956, Chazyan and related brachiopods, part I: Smithsonian Misc. Collections, vol. 127, 1024 pp.

Decker, C. E., 1935, Graptolites of the Sylvan shale of Oklahoma and Polk Creek shale of Arkansas: Jour. Paleontology, vol. 9, pp. 697-708.

Decker, C. E., 1936a, Some tentative correlations on the basis of graptolites of Oklahoma and Arkansas: Amer. Assoc. Petroleum Geologists Bull., vol. 20, pp. 301-311.

Decker, C. E., 1936b, Table of tentative Lower Paleozoic correlations on basis of graptolites: Amer. Assoc. Petroleum Geologists Bull., vol. 20, pp. 1252-1257.

Decker, C. E., 1943, Three more graptolites from Simpson of Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 27, pp. 1388-1392.

Decker, C. E., 1952, Stratigraphic significance of graptolites of Athens shale: Amer. Assoc. Petroleum Geologists Bull., vol. 36, pp. 1-145.

Decker, C. E., and Huffman, G. G., 1953, Sylvan graptolites in northeastern Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 37, pp. 451-452.

Flawn, P. T., 1956, Basement rocks of Texas and southeast New Mexico: Univ. of Texas, Publication 5605, 261 pp.

Ham, W. E., 1954, Algal origin of the "birdseye" limestone in the McLish formation: Oklahoma Acad. Sci., Proceedings, vol. 32, pp. 200-203.

Ham, W. E., 1949, Geology and dolomite resources, Mill Creek-Ravia area, Johnston County, Oklahoma: Oklahoma Geol. Survey, Circular 26, 104 pp.

Ham, W. E., 1955, Field conference on geology of the Arbuckle Mountain region: Oklahoma Geol. Survey, Guidebook III, 61 pp.

Harlton, B. H., 1953, Ouachita chert facies, southeastern Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 37, pp. 778-796.

Harris, R. W., 1957, Ostracoda of the Simpson group of Oklahoma: Oklahoma Geol. Survey, Bull. 75, 333 pp.

Hass, W. H., 1951, Age of Arkansas novaculite: Amer. Assoc. Petroleum Geologists Bull., vol. 35, pp. 2526-2541.

Hass, W. H., 1956, Conodonts from the Arkansas novaculite, Stanley shale, and Jackfork sandstone: Ardmore Geological Society, Guidebook, Ouachita Mountain Field Conference, pp. 25-33.

Hendricks, T. A., 1947, Geology of the western part of the Ouachita Mountains in Oklahoma: U. S. Geol. Survey, Oil and Gas Investigations, Preliminary Map 66, Sheet 1, text.

Hendricks, T. A., Knechtel, M. M., and Bridge, J., 1937, Geology of Black Knob Ridge: Amer. Assoc. Petroleum Geologists Bull., vol. 21, pp. 1-29.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains in Oklahoma: Oklahoma Geol. Survey, Bull. 32, Part 1, 278 pp.

Huffman, G. G., 1953, Sylvan shale in northeastern Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 37, pp. 447-450.

Miser, H. D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadranges, Arkansas: U. S. Geol. Survey, Bull. 808, 195 pp.

Pitt, W. D., 1955, Geology of the core of the Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Circular 34, 34 pp.

Reeds, C. A., 1911, The Hunton formation of Oklahoma: Amer. Jour. Science, 4th Ser., vol. 32, pp. 256-268.

Sellards, E. H., Adkins, W. S., and Plummer, F. B., 1932, The Geology of Texas: Univ. of Texas, Bull. 3232, vol. 1, 1007 pp.

Wengerd, S. A., 1948, Fernvale and Viola limestones of south-central Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 32, pp. 2183-2253.

Wilson, L. R., 1958, A chitinozoan faunule from the Sylvan shale of Oklahoma: Oklahoma Geol. Notes (Oklahoma Geol. Survey), vol. 18, pp. 67-71.

Yochelson, E. L., and Bridge, J., 1957, The Lower Ordovician gastropod Ceratopea: U. S. Geol. Survey, Prof. Paper 294-H, pp. 281-304.

SUMMARY DISCUSSION OF THE GEOLOGY OF THE CORE AREAS OF THE OUACHITA MOUNTAINS, ARKANSAS AND OKLAHOMA

William D. Pitt [1]

Introduction

At the request of the Ouachita Field Trip Committee I have summarized below (1) the field evidence and reasons for believing that the central (core) area of the Choctaw anticlinorium of southeastern Oklahoma is anticlinal in structure, rather than a fenster; and (2) the significant findings of recent work in the core area of Arkansas.

Core Area Of Oklahoma: Anticlinal Structure

Reasons supporting the anticlinal interpretation of the core area are listed and briefly described below; the supporting field evidence was presented during the Ardmore Geological Society's field trip to the Ouachita Mountains in 1956. These reasons and related problems are discussed in the guide book of the above-mentioned field trip and also in the 1955 Tulsa Geological Society Digest.

Flankward dip. - At many locations in the heart of the Ouachitas the rocks may be seen dipping away from the Lukfata sandstone, the oldest outcropping formation. A cursory inspection of the writer's core-area map will reveal the flankward dip of the rocks away from the core area. As Tomlinson notes (Tomlinson and Pitt, 1955, p. 92) "As seen from the air, the Crystal Mountain sandstone makes a nice series of northwestward-dipping, 'flatirons', a hogback intersected by ravines, for three miles northeastward from sec. 22, 5 S., 23 E. This sandstone is a ridge-maker along most of its strike...".

Sequence. - Clearly the oldest rocks are in the central part of the core area. This fact can be checked by walking the larger creeks that flow across the formation contacts, as shown on the map of the core area (Pitt, 1955). This relationship may be seen along Lukfata Creek in the NW 1/4 SE 1/4 sec. 17, T. 5 S., R. 24 E. By walking southward from the cement ford for about 1/2 mile, one may see the three members of the Lukfata sandstone and both members of the Collier shale; along this traverse the rock sequence dips gently southward and is almost completely exposed. Farther down this same creek the overlying Crystal Mountain sandstone, the Womble shale, the Mazarn shale and the Bigfork chert are encountered in that order, and all clearly dip away from the core. In conclusion, there is a mappable and recognizable sequence of rocks in the core area, and this sequence is clearly seen in progressing from the core area outward in any direction if one follows a stream bed that drains across the formation boundaries.

"Crystal Mountain Ridge" area, SW 1/4 sec. 27, E 1/2 sec. 8, T. 5 S., R. 23 E. - If a fenster exists in the central part or core area of the Choctaw anticlinorium, the Collier shale must be drawn as Honess mapped it: lying next to the Womble shale. The only location where Honess was sure of having mapped Collier shale is in this "Crystal Mountain Ridge" area. He expressed this conclusion as follows (Honess, 1923, p. 39):

> "The Collier shale as mapped should therefore be regarded as doubtful Collier, with the exception of that portion which lies in secs. 27 and 28, of T. 5 S., R. 23 E. In this region the thin limestones and graphitic shales positively underlie the massive conglomerate occurring at the base of the Crystal Mountain sandstone and they must,

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> therefore, be Collier in age at this place."

It is significant that this is the area where one can demonstrate that the Collier shale is present, that (1) the Collier lies in normal sequence with respect to younger formations; and (2) that here the regional dip is westward, not eastward as Honess' mapping demands. The normal sequence can be seen in the bed of Glover River (E 1/2 sec. 28, 5 S., 23 E.), just west of the Crystal Mountain ridge during low-water periods; this relationship militates against Honess' Glover fault. Honess writes that this fault

> "was drawn from the center of Sec. 28, T. 5 S., R. 23 E., northeast for 12 miles to the west range line of sec. 18, T. 4 S., R. 25 E., between the Womble schistose sandstone on the north and some black slates (Collier?) on the south. This is an assumed line of displacement, which for convenience of reference will be known as the Glover fault. Three formations are missing along this line if (sic) the correlations as made are correct..."

In seeing the normal sequence here one should begin at the base of the Crystal Mountain ridge in the SW 1/4 sec. 27, T. 5 S., R. 23 E.; there one successively encounters in going westward along the stream bed: Collier shale (lower member of the Collier shale formation), Collier limestone (member), Crystal Mountain sandstone, and the Mazarn and Womble shales; the Blakely sandstone is not present in this area. The investigator will clearly see there, as well as on the ridge, that the regional dip is westward, not eastward as Honess' mapping demands; this incorrect mapping of the regional dip in part explains Honess' having mapped (in this creek-bed location and adjacent area) the Womble lying next to the Collier shale.

Finally, in this ridge area, as was noticed by Tomlinson and others before the writer mapped the area, the Crystal Mountain sandstone that caps the ridge is obviously underlain by a shale that crops out on the east side as well as on the west side of the ridge. In fact, the Collier shale extends in continuous outcrop from the west side around the north end of the ridge, which is only a spur from the main outcrop band of Crystal Mountain sandstone, to the east side; and the base of the capping sandstone is higher on the east than on the west. The shale east of the ridge is Collier, not Mazarn: supporting the idea that the older rocks are invariably near the center of the anticlinorium, and it demonstrates the regional southwestward dip, a direction supporting an anticlinal interpretation and not in keeping with the fenster hypothesis.

Conclusion. - The core area of Oklahoma I conclude is simply the central part of the Choctaw anticlinorium. In the core area regional anticlinal dip is demonstrated, as it is in the rest of the Choctaw anticlinorium. In the core area the writer found a recognizable sequence of rocks; this sequence is clearly seen by walking flankward along stream beds, from the central or oldest part of the core area.

Crystal Mountains Area, Arkansas

In 1952 the writer examined the rocks of the core area of Arkansas in order to compare them with those of the core area in Oklahoma. C. W. Tomlinson sponsored the writer in a mapping project of the Arkansas core area, done intermittently during 1954, 1955 and 1956. The project included not only the type area of the Blakely sandstone but also that of the Collier shale, the Mazarn shale and the Crystal Mountain sandstone, the general area in or adjacent to the highest ridges of the Crystal Mountains. The writer's report, including a map of the area, was submitted to the Arkansas Geological Survey for publication; map corroboration of the writer's conclusions below must await publication of this report.

Structure. - In general, the hills in which the Crystal Mountain sandstone crops out are anticlinal and the valleys between them are synclinal. This is true generally for the Crystal Mountain sandstone areas mapped by Engel (Engel, 1946), as well as the area mapped by the writer. The only homoclinal ridges of the Crystal Mountain sandstone are adjacent to the outcrop areas of the Collier shale along upper Huddleston and Collier Creeks.

By comparing the writer's geologic map of this area with that of Miser and Purdue (Miser and Purdue, 1929), it is apparent that the writer substantiated much of their mapping.

Several disagreements, however, were found and because they are significant both structurally and stratigraphically, they are discussed below.

High Peak Ridge

Unlike Miser and Purdue, the writer mapped this ridge as an anticline rather than a homocline; the fact that High Peak Ridge (the ridge extending 3 3/4 miles westward from the center of the W 1/2 sec. 15, T. 3 S., R. 25 W.) is anticlinal is demonstrated by the many large north-dipping sandstone ledges that crop out along the north side of High Peak Ridge. Its smooth, rounded topographic appearance, as shown on the topographic map of the Glenwood Quadrangle is also strongly suggestive of an anticline, especially at its western end for there its topography typifies that of an anticlinal nose. The anticlinal structure of High Peak Ridge can also be demonstrated by walking along Montgomery Creek where it crosses the western nose of High Peak Ridge in the W 1/2 sec. 15, T. 3 S., R. 25 W. Here north dip along the north side of the structure and south dip along its south side can be seen clearly. Finally, the anticlinal structure is further substantiated at its southeast nose in the SW 1/4 sec. 21, T. 3 S., R. 24 W. Here Collier Creek crosses the east nose of High Peak anticline at the outcrop belt of the basal massive sandstone of the Crystal Mountain sandstone formation; here in fact the basal massive sandstone forms an anticlinal arch that is clearly visible at the stream's edge. This sandstone can be walked continuously around the east nose of the anticline, from the north slope of High Peak across Collier Creek, along the east wall of its canyon, and back into the south (locally overturned) flank of the anticline.

Stratigraphic significance. - Recognition of the anticlinal structure of High Peak Ridge means also the recognition of true Collier shale near its central part. Slightly more than 100 feet of dark fissile shale, calcareous siltstone and some sandstone are exposed there; especially noteworthy of the Collier shale is the lack of limestone beds that were considered to be typical of the Collier shale in its type area; mapping High Peak Ridge as an anticline rather than a south-dipping homocline means that the limestone-bearing shale north of High Peak Ridge is Mazarn shale, not Collier shale as Miser and Purdue mapped it. Limestones are present in the Mazarn south of the ridge too; and also in Oklahoma on the north flank of the Choctaw anticlinorium near Government Spring, west of State Highway 21. In the central part of High Peak anticline an almost complete section of the Crystal Mountain sandstone is exposed; here the formation clearly does not aggregate more than 450 feet, rather than the earlier estimate of 850 feet.

Wheeler Mountain Ridge Area

Wheeler Mountain Ridge is the name assigned to the ridge that extends eastward more than nine miles from the SW 1/4 of sec. 9, T. 3 S., R. 25 W. to and beyond the eastern edge of the "Crystal Mountains", and includes "Mount Ida" and "Logan Gap Mountain". As originally mapped by Miser and Purdue the shale that lay north of this ridge was mapped as Collier except that part of the shale outcrop band that lies north of Logan Gap Mountain. In other words, according to this interpretation the sandstone found along this ridge overlies the shale north of it and dips southward on the north side of the ridge except north of Logan Gap Mountain. The writer is in fundamental disagreement with this interpretation, believing that Wheeler Mountain is an anticline, so that the sandstone found along it dips under the shale that lies north of the ridge. It is important to clear up this point because if Wheeler Mountain Ridge is anticlinal many square miles north and northwest of the ridge must be Mazarn shale, rather than Collier shale as shown on the state geologic map of Arkansas, published in 1929. Reasons that support the belief that this ridge is an anticline are enumerated below.

Proof of anticlinal structure. - The chief difference between Miser and Purdue's map and that of the writer is that the writer believes that the regional dip along the north side of this ridge is northward, that the sandstone found along the ridge underlies rather than overlies the shale that crops out in the north-bounding valley. North dip along the north side of the ridge is clearly visible at a number of outcrops, especially in sections 9 and 10, T. 3 S., R. 25 W.

Several lines of evidence support the

conclusion that Wheeler Mountain Ridge locally is overturned. One place that this overturning is believed to be demonstrated is in the SE 1/4 sec. 8, T. 3 S., R. 24 W.; here drag folds on a road cut are interpreted as being normal and therefore an indication of overturned beds. An even more convincing example of overturning along the north side of this ridge is seen at an outcrop near the crest of the ridge in the NE 1/4 SE 1/4 SW 1/4 sec. 10, T. 3 S., R. 24 W. Here a key part of the crest of the Wheeler Mountain anticline is clearly visible: steeply north-dipping sandstone beds bend around an arch to become overturned north of it.

Small anticlinal ridges within the shale valley north of Wheeler Mountain Ridge is also supporting evidence for the belief that the shale in this valley is Mazarn shale rather than Collier, overlying rather than underlying the Crystal Mountain sandstone. If the oblong sandstone outcrops in this area are Crystal Mountain sandstone, the shale adjacent to them must be Mazarn shale because the sandstone outcrop areas are clearly anticlinal. Larger anticlinal ridges that border this valley on the north also support the belief that the shale in the valley is Mazarn if the sandstone that caps these ridges is Crystal Mountain sandstone. Some of these ridges were mapped by Engel (Engel, A.E.J., 1946), who also identified the sandstone found on them as Crystal Mountain sandstone.

Mapping Project North Of Wheeler Mountain

In order to see whether or not the area north of Wheeler Mountain contained outcrops of Collier shale, C. W. Tomlinson of Ardmore, Oklahoma, gave his help and financial support to four Oklahoma University graduate students who mapped in 1958 the supposed outcrop area of Collier shale that lay north and northwest of Wheeler Mountain. These graduate students are: Messrs. Richard Cohoon, Harry Lee, Glenn Robb and John Watson. Their mapping now is largely completed and a joint report probably will be published next year. We can now make several general conclusions regarding the structure and stratigraphy of this area of supposed Collier shale that lies north of Wheeler Mountain:

(1) The area is largely composed of rocks younger than the Crystal Mountain sandstone.
(2) Collier shale outcrop areas were not discovered.
(3) Crystal Mountain sandstone crops out in the eastern and central part of the area of supposed Collier shale.
(4) Every patch of Crystal Mountain sandstone west of the Mount Ida-Norman highway is anticlinal within itself.
(5) The sandstones that form the ridges are successively younger as one moves from the east to the west end of the area of supposed Collier shale.
(6) There is no large-scale faulting in the area.
(7) Overturning is in general toward the south.

References

Ardmore Geological Society, 1956, Ouachita Mountain Field Conference of Southeastern Oklahoma.

Engel, A. E. J., 1946, The quartz crystal deposits of western Arkansas: Econ. Geol., vol. 41, no. 6, pp. 598-618.

Honess, Charles W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey Bull. 32.

Miser, H. D. and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles, Arkansas: United States Geol. Survey Bull. 808.

Miser, H. D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geol. Survey Bull. 50, 30 pp., 7 figs., 3 pls. inc. map.

Miser, H. D., 1929, Geologic map of Arkansas.

Pitt, William D., 1955, Geology of the core of the Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey Circ. 34, 33 pp.

Tomlinson, C. W., and Pitt, William D., 1955, Recent studies in the Ouachita Mountains: Tulsa Geol. Soc. Digest, vol. 23, pp. 89-97.

CORRELATION OF LOWER PALEOZOIC FORMATIONS OF THE ARBUCKLE AND OUACHITA AREAS AS INDICATED BY GRAPTOLITE ZONES

Charles E. Decker [1]

Foreword

In 1941 Charles E. Decker published a paper entitled, "Arbuckle Formations in the Ouachita Mountain Region" in the annual report of the Oklahoma Academy of Science. The paper included a chart showing correlations of the Lower Paleozoic formations of the Arbuckle Mountain area with their equivalents in the Ouachita Mountains. This chart was modified and brought down to date in August, 1958, by Dr. Decker shortly before he passed away, and it is included as Table I in this paper, with the salient stratigraphic zones noted.

Dr. Decker wrote the final draft of this manuscript in longhand and submitted it July 28, 1958. It was transcribed and returned to him. The corrected copy, with notations in his own handwriting, was received August 16, 1958.

Cambrian System

Lukfata sandstone. - The oldest rocks exposed in the Choctaw anticlinorium in the core of the Ouachita Mountains in McCurtain County, Oklahoma, comprise 145 feet of sandstone, shale and limestone. William D. Pitt (1955, p. 11, Fig. 2, p. 13) named these beds the Lukfata sandstone, doubtless because of a prominent sandstone member at the top. The Lukfata formation is, without doubt, the partial equivalent of the Upper Cambrian Reagan sandstone of the Arbuckle area.

Collier shale. - Purdue (1909, p. 557) gave the name Collier to some cherty shales underlying the Crystal Mountain sandstone on Collier Creek, Montgomery County, Arkansas. Miser and Purdue (1929, pp. 23, 24) listed its thickness as 200 feet in the Çaddo Gap quadrangle of Arkansas, but state that it is several hundred feet thicker in the Mount Ida quadrangle to the north. Although it has not been found to be fossiliferous, it has been assigned to the Upper Cambrian on the basis of stratigraphic position.

Pitt (1955, pp. 15-18) divided the Collier of McCurtain County, Oklahoma, into two members, a lower shale member which is 180 feet thick, and an upper limestone member 140 feet thick.

Ordovician System

Crystal Mountain sandstone. - The Crystal Mountain sandstone (named by Purdue, 1909, p. 557) attains a thickness of 850 feet (Miser, 1917, pp. 67, 68) in the Arkansas portion of the Ouachitas, where it includes a conglomerate at the base and appears to be unconformable on the Collier shale. On the basis of stratigraphic position the formation has previously been assigned a Lower Ordovician age. [2]

Because of similar lithologic characteristics, Pitt (1955, pp. 18, 19) compared the Crystal Mountain of the Ouachitas with the McLish sandstone of the Simpson group of the Arbuckles saying "both are pure sandstones and appear to be in about the same stratigraphic position". This correlation is certainly incorrect because the McLish is in the middle portion of the Simpson group and

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[2] See Ham, this symposium, Table 1.

is Burgen or St. Peter in age (Chazy). The Oil Creek and Joins formations intervene between the McLish and the top of the Arbuckle limestone in the Arbuckle Mountain area. The Crystal Mountain sandstone belongs (stratigraphically) far below the top of the Arbuckle, opposite the McKenzie Hill formation near the base of the Ordovician.

Mazarn shale. - Miser and Purdue (1929, p. 26) give the thickness of Mazarn shale as about 1000 feet in Arkansas, whereas Pitt (1955, p. 23) has given an approximate thickness of 600 feet for this formation in the Oklahoma Ouachitas. Pitt incorrectly correlates the Mazarn with the "upper part of the Arbuckle limestone and the lower part of the Simpson group", whereas the extensive Mazarn graptolite fauna contains several Lower Ordovician extensiform Didymograptus and many species of Tetragraptus and Phyllograptus, all of which occur below the Didymograptus protobifidus zone which lies 808 feet below the top of the Arbuckle limestone. Thus the Mazarn is older than the Didymograptus protobifidus zone in the lowest part of the West Spring Creek formation, and parts of the Kindblade and Cool Creek formations.

Blakely sandstone. - The Blakely sandstone of the Caddo Gap area, Arkansas, consists of 400 feet of ridge-forming sandstones and interbedded shales. The formation extends westward into Oklahoma but has not yielded identifiable fossils. In the Blakely Mountain area of the Hot Springs district an outcrop has questionably been assigned to the Blakely formation and shales in the midst of this outcrop have yielded a graptolite fauna of Middle Deepkill age, so it may represent a zone equivalent to part of the upper Mazarn elsewhere.

Didymograptus protobifidus zone. - The diagnostic graptolite Didymograptus protobifidus occurs in a zone about 808 feet below the top of the West Spring Creek formation of the Arbuckle group in exposures along the United States Highway 77 at a locality 2 1/4 miles north of Springer, Oklahoma. Elsewhere this diagnostic species occurs in great profusion in the old crusher quarry at Big Canyon in the Arbuckle Mountains, in limestones at the northwest edge of the Wichita Mountains, in the Smithville limestone in northeastern Arkansas, and in the Marathon

SYSTEM	GROUP	REGION: ARBUCKLE	REGION: OUACHITA
SILURIAN	RANCH	HENRYHOUSE SHALE	MISSOURI MOUNTAIN SLATE
		CHIMNEYHILL LS.	BLAYLOCK SS.
ORDOVICIAN	PATTERSON	SYLVAN SHALE	SYLVAN-POLK CREEK
		FERNVALE	
		VIOLA	BIGFORK CHERT
	SIMPSON	BROMIDE	STRINGTOWN-WOMBLE (TRENTON GRAPTOLITES)
		TULIP CREEK	
		McLISH	
		OIL CREEK	
		JOINS (DIDYMOGRAPTUS-ARTUS) DEEPKILL	
	ARBUCKLE	WEST SPRING CREEK (800' DIDYMOGRAPTUS PROTOBIFIDUS)	BLAKELY SS.
		KINDBLADE	MAZARN SHALE (TETRAGRAPTUS & PHYLLOGRAPTUS OLDER THAN DIDYMOGRAPTUS PROTOBIFIDUS)
		COOL CREEK	
		McKENZIE HILL	CRYSTAL MOUNTAIN SS.
UPPER CAMBRIAN		BUTTERLY	
		SIGNAL MOUNTAIN	
		ROYER	
		FORT SILL	
	TIMBERED HILLS	HONEY CREEK	COLLIER SHALE
		REAGAN	LUKFATA

Fig. 1. - - Correlation chart.

formation of Trans-Pecos Texas. It occurs also in Australia and in Great Britain.

Didymograptus artus zone. - Didymograptus artus, the tiny graptolite with numerous thecae (18 packed into 10 mm.), is widespread in the Arbuckle Mountains, occurring 50 to 100 feet above the base of the Joins formation (basal Simpson) where it is exposed along United States Highway 77; it is also widespread in the Black Rock formation of northeastern Arkansas. Didymograptus bifidus, a much larger associate with few thecae, occurs at the north end of the Criner Hills nine miles southwest of Ardmore, Oklahoma. It is also widespread and common in rocks of upper Beekmantown age.

Womble shale. - The next formation above the Blakely sandstone in the Ouachita Mountains is the Womble shale. The formation is about 1000 feet thick in Arkansas but is much thinner in Oklahoma. It contains several species of Trenton graptolites that are held in common with the upper Bromide formation (uppermost Simpson) and lower Viola limestone of the Arbuckle Mountains and Criner Hills. This widespread fauna extends westward from Oklahoma to Nevada, northward to Idaho and Washington, up into British Columbia, Yukon and Alaska, and thence into western China, Australia, and New Zealand. The same fauna occurs in the Appalachians throughout the full length of exposures of the Athens shale, in the Normanskill black shale of New York, thence eastward across southeastern Canada to Great Britain and then into Europe.

Bigfork chert. - A single large elongate graptolite, Diplograptus (Amplexograptus) recurrens, is common in the upper part of the Viola limestone in the Arbuckle Mountains, and in the coarsely crystalline phase of the mid-portion of the Bigfork chert in the Ouachita Mountains. Inasmuch as Diplograptus recurrens has been found in the core of a deep well in southern Michigan associated with Climacograptus lorrainensis, the upper Viola is apparently Lorraine in age.

Polk Creek shale. - The youngest Ordovician formation in the Ouachitas of Arkansas and Oklahoma is the Polk Creek shale, which is about 175 feet thick. It contains a graptolite fauna that permits correlation with the Sylvan shale of the Arbuckle Mountains. The basal Sylvan is well indurated and fissile but most of it (about 300 feet) is composed of soft shale which is abundantly fossiliferous with the following species of graptolites: Dicellograptus complanatus Lapworth and also a very narrow undescribed variety of the species, the species being very abundant; Diplograptus crassitestus Ruedemann, Climacograptus mississippiensis Ruedemann, C. putillus (Hall), C. ulrichi Ruedemann, C. tridentatus var. maximus Decker, and an added variety in Arkansas, Diplograptus calcaratus trifidus (Gurley).

Silurian System

Blaylock sandstone. - The next formation above the Polk Creek in Arkansas is the Blaylock sandstone listed by Miser and Purdue (1929, pp. 42-45) as having a conglomerate at the base, consisting mainly of sandstone, and perhaps reaching a thickness of as much as 1500 feet. At the south edge of Blaylock Mountain, along the Little Missouri River, graptolites occur in a fine-grained calcareous sandstone in the lower part of the formation. Miser and Purdue (1929. pp. 42-45) list the following species from the Blaylock at this locality: Monograptus distans (Portlock), M. gregarius (Lapworth), M. aff. M. fimbriatus (Nicholson), Dimorphograptus decussatus (Elles and Wood), and Gladiograptus perlatus Nicholson. In the late nineteen thirties H. D. Miser, John Fitts, and the writer collected the following additional species from the same general locality: Monograptus communis Lapworth, M. micronematoides, M. nilssoni (Barrande), and M. priodon (Bronn). These additional species of Monograptus are further evidence that the Blaylock sandstone belongs in the lower part of the Silurian system.

Chimneyhill limestone. - In the Arbuckle Mountains the base of the old Hunton group is the Chimneyhill limestone. The formation is thin (from 0 to 35 feet), is patchy in its distribution, is oölitic and thick-bedded in the lower part, glauconitic in the middle, and crinoidal in the upper part. Gould (1925, p. 19) listed its fossils and correlated it with the Medina of New York. Wilmarth (1938, p. 431) correlated it with the Brassfield of Ohio.

Henryhouse shale. - Above the Chimneyhill limestone in the Arbuckles is the Henryhouse shale. It was described by Reeds (1911, pp. 256-268) and again by Morgan (1924, pp. 39, 40) who listed an extensive fauna including sponges, corals, brachiopods, bryozoa, cephalopods, and trilobites. The writer (Decker, 1935, pp. 535, 536) has recorded the following graptolites from the Henryhouse: Mastigograptus beachi Decker, Thallograptus phylloides Decker, Monograptus bohemicus (Barrande), M. crinitus Wood var. exilis Decker, M. dubius (Suess), M. nilssoni (Barrande), M. scanicus Tullberg, M. tumescens Wood, M. cf. M. ultimus Perner, M. vulgaris Wood, M. (Linograptus) phillipsi Decker, M. (Linograptus) phillipsi var. multiramosus Decker, and Cyrtograptus sparsus Decker.

Morgan (1924, pp. 38-40) referred the Henryhouse to the Niagaran series of the Silurian system. M.E. Upson sent the writer for study a core taken at a depth of 12,672-12,676 feet from the Gulf-Phillips # 1 Shurtleff (Toyah Unit) well which is located in Section 20, Block C-17 Public School Land in Reeves County, Texas. The following species of graptolites were identified in this core: Monograptus dubius (Suess), M. ultimus Perner, M. vomerinus basilicus Lapworth, and M. vulgaris Wood. These species of Monograptus definitely place this zone in the Niagaran, although it had long been considered as Devonian.

Earlier William J. Hilseweck had sent in some graptolites from the Gulf Oil Corporation # 1 University "F" well in Section 22, Block 31, University Lands, Crane County, Texas, from a zone between the depths of 9328-9348 feet and the writer subsequently (Decker, 1942, pp. 857-860) described and illustrated them. They proved to be varying phases of a single species, Monograptus vomerinus (Nicholson), of Niagarian age - the zone had previously been considered as Devonian.

Al and Helen Loeblich sent the writer some graptolites from rocks dug from the drainage canal on the west side of Frisco Avenue in Blue Island, a suburb at the southwest edge of Chicago. The collection includes many specimens of a single species, Monograptus vomerinus (Nicholson) which occurs in the Niagaran rocks of that region.

Missouri Mountain slate. - The Missouri Mountain slate or shale (Purdue, 1909, p. 37) overlies the Blaylock sandstone (Possibly unconformably) and underlies the Arkansas novaculite in the Ouachita Mountains. This slate is poorly exposed in Oklahoma but at Alum Bluff near the mouth of Cedar Creek 63 feet of well exposed shale with some sandstone and a little cherty limestone may be seen. Fossils have not been found in it but it is placed in the Silurian on the basis of its stratigraphic position.

References

Cullen, John, 1917, Lime resources and industry in Oklahoma: Oklahoma Geol. Survey Bull. 26.

Decker, Charles E., 1935, Graptolites of the Sylvan shale of Oklahoma and Polk Creek shale of Arkansas: Jour. Paleontology, vol. 9, no. 8, pp. 697-708.

Decker, Charles E., 1941a, Arbuckle formations in the Ouachita Mountain region of Oklahoma: Oklahoma Acad. Sci., Annual Rept. 1941, pp. 153-155.

Decker, Charles E., 1941b, Didymograptus in North America: Jour. Paleontology, vol. 15, pp. 362-365.

Decker, Charles E., 1942, A Silurian graptolite zone in Crane County, Texas: Amer. Assoc. Petroleum Geologists Bull., vol. 26, pp. 857-860.

Decker, Charles E., 1944, Pendent graptolites of Arkansas, Oklahoma and Texas: Jour. Paleontology, vol. 18, pp. 377-386.

Decker, Charles E., 1950, Ordovician graptolites from well-core, Calhoun County, Michigan: Amer. Assoc. Petroleum Geologists Bull., vol. 34, no. 9, pp. 1903-1909.

Decker, Charles E., 1951, Small graptolite determines Ordovician horizon: Amer. Assoc. Petroleum Geologists Bull., vol. 35, no. 7, p. 1673.

Elles, Gertrude L., and Wood, Ethel M. R., 1901-1918, Monograph of British graptolites: Paleontological Soc. London, 3 vols.

Elles, Gertrude L., 1933, The lower Skiddaw slates; summary of progress: Geol. Soc. for 1932, pt. 2, pp. 94-111 (London).

Gould, Charles N., 1925, Index to the stratigraphy of Oklahoma: Oklahoma Geol. Survey Bull. 35, 115 pp.

Hendricks, T. A., Knechtel, M. M., and Bridge, Josiah, 1937, Geology of Black Knob Ridge: Amer. Assoc. Petroleum Geologists Bull., vol. 21, pp. 1-29.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey Bull. 32, pt. 1, pp. 33-109.

Huffman, George, 1958, Correlation table of formations of Arbuckle area, Ouachita Mountains, et cetera: Personally assembled.

Miser, Hugh D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles, Arkansas: United States Geol. Survey Bull. 808, pp. 22-99.

Morgan, George D., Geology of Stonewall quadrangle, Oklahoma: Oklahoma Bur. of Geology, Bull. 2, pp. 37-40.

Pitt, William D., 1955, Geology of the core of the Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Circular 34, 34 pp.

Purdue, A. H., 1909, Slates of Arkansas: Arkansas Geol. Survey.

Ruedemann, Rudolph, and Decker, Charles E., 1934, The graptolites of the Viola limestone: Jour. Paleontology, vol. 8, pp. 303-327.

Wilmarth, Grace, 1938, Lexicon of geologic names of the United States: United States Geol. Survey Bull. 896, pt. 1, p. 431.

PETROGRAPHY OF PALEOZOIC SANDSTONES FROM THE OUACHITA MOUNTAINS OF OKLAHOMA AND ARKANSAS

August Goldstein, Jr. [1]

Abstract

Sandstones of the Lukfata, Crystal Mountain, Blakely, Missouri Mountain, and Hot Springs formations consist largely of quartz and chert with only small amounts of feldspars, composite (rock) grains, and interstitial detrital argillaceous material. These sandstones are classified as quartzose sandstones or orthoquartzites. Sandstones of the Stanley formation, sandstones of the Womble formation in McCurtain County, and thin sandstone interbeds in the Arkansas novaculite contain moderate amounts of feldspars and rock grains and have an interstitial primary mud matrix; these rocks are classified as lithic graywackes. Sandstones of the Blaylock, Jackfork, and Atoka formations contain small to moderate amounts of feldspar, rock grains, and interstitial "clay"; these rocks are classified as quartzose subgraywackes or "protoquartzites".

Theoretical considerations based on sandstone types, concealed sediment thicknesses, position of major unconformities, and faunal studies suggest a double Paleozoic geosynclinal cycle in the Ouachita Mountain region, with periods of maximum geosynclinal downwarping in the Ordovician and Mississippian.

Introduction

Some sedimentologists believe (i. e., Pettijohn, 1949) that a complete petrographic analysis is necessary for thorough understanding of a stratigraphic section, and that the sandstone members are most valuable for this purpose. Very little data have been published previously concerning the petrography of the sandstones of the Ouachita Mountains. The purpose of this study is to describe petrographically the sandstones cropping out in the Ouachita Mountains of Oklahoma and western Arkansas, and to make certain interpretations in regard to the tectonic conditions under which these sandstones were deposited.

The sandstones described herein range in age from Cambro-Ordovician (?) or Ordovician to Lower Middle Pennsylvanian. Generalized columnar sections by Branson, Cline, and Ham are shown on pp. 119, 179, and 73, respectively in this volume. Sandstones of the Lukfata, Crystal Mountain, Womble, Blakely, Blaylock, Missouri Mountain, Arkansas Novaculite, Hot Springs, Stanley, Jackfork, and Atoka formations are discussed below.

Previous work. - The classic study of Honess (1923) is the only published detailed description of the petrography of sandstones of the Ouachita Mountains; his study was restricted to outcrops in McCurtain County, Oklahoma. A generalized summary of the petrography and metamorphism of sediments of Ouachita facies was written by Goldstein and Reno (1952). Numerous valuable field studies have been published; the writer has drawn liberally upon the works of Hendricks and others (1937, 1947), Harlton (1938), Miser and Purdue (1929), Honess (1923), and Pitt (1955) for the field descriptions.

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[1] Bell Oil and Gas Company, Tulsa, Oklahoma.

critically by T. A. Hendricks, P. T. Flawn, E. L. Selk, and G. J. Verville.

Lukfata Sandstone

The Lukfata sandstone formation was named by Pitt (1955); its type locality is along Lukfata Creek in Secs. 8 and 17, T. 5 S., R. 24 E., McCurtain County, Oklahoma. However, the strata which comprise the Lukfata sandstone were described originally as part of the Crystal Mountain sandstone, and were mapped as Crystal Mountain by Honess (1923).

In order to understand the stratigraphic position of the Lukfata sandstone, it is necessary to digress slightly and discuss the relation of the Crystal Mountain sandstone to adjacent formations. The type locality of the Crystal Mountain sandstone is in the Crystal Mountains of Montgomery and Garland Counties, Arkansas. In that area the stratigraphic relations are clear; the Crystal Mountain sandstone unconformably overlies the Collier shale and is in turn overlain with apparent conformity by the Mazarn shale of Lower Ordovician age (Miser and Purdue, 1929, pp. 24-25). In many places in Arkansas there is a conglomerate at the base of the Crystal Mountain sandstone where it overlies the Collier shale. No usable fossils have been reported from either the Collier shale or the Crystal Mountain sandstone, and the base of the Collier is not exposed at any place in Arkansas. Most writers give the age of the Collier as Cambrian or Cambrian (?); because of the unconformable relation between the Collier and Crystal Mountain, these same writers give the age of the Crystal Mountain as Cambro-Ordovician (?) or Ordovician (?).

Pitt's remapping of the Cove area of the Ouachita Mountains in southeastern Oklahoma (1955) led him to the conclusion that the Crystal Mountain sandstone of this area can be separated into two distinct formations, one of which underlies the Collier shale and one of which overlies the Collier shale. The older formation was named the Lukfata sandstone formation and the name Crystal Mountain sandstone was retained in a restricted sense for the younger formation. Branson (1956, p. 15) referred all three formations to the Canadian Series of the Ordovician Period, and correlated them tentatively with the Cool Creek and McKenzie Hill formations of the Arbuckle Group.

In McCurtain County, Oklahoma, Pitt (1955, p. 13-14) divided the Lukfata sandstone formation into three members: a lower member made up of interlaminated shale and thin-bedded limestone; a middle member of interbedded platy sandstone and shale; and an upper member of more massively bedded sandstone with some thin interbeds of shale. The sandstones described below are from the middle and upper members.

Petrographically, the Lukfata sandstones are mostly fine-to medium-grained, quartzose, and well-sorted. All of the rocks have undergone low-grade metamorphism. Most of the quartz grains were originally rounded to subrounded, but their outlines were modified by secondary enlargement and subsequently changed by reaction both between adjacent grains and between grains and matrix during metamorphism. The net result of this secondary silicification and later metamorphism is a jagged, serrate, irregular contact between grains which is marked by microcrystalline, recrystallized quartz and vein quartz at places. Dust-size inclusions define the original outlines of the quartz grains. Almost every quartz grain shows strain shadows and has an undulatory extinction.

In addition to quartz sand grains, these rocks contain numerous aggregates of silt- and sand-size secondary quartz crystals, dolomite rhombs, detrital grains of shale and silty shale, sericite, chlorite, and iron oxides. Virtually all of the interstitial "clay" originally present in the rock has been reconstituted into new sericite and chlorite, and the detrital shale and silty shale fragments have been altered into low grade metamorphic rocks such as low-rank argillite [2] and high-rank argillite.

[2] The term low-rank argillite is used by P. T. Flawn for those argillaceous rocks without preferred mineral orientation or slaty cleavage in which from 25 to 50 percent of the "clay" has been reconstituted; high-rank argillite is used for similar rocks in which from 50 to 100 percent of the "clay" has been reconstituted (Flawn, 1958).

Most of the carbonate minerals present originally in these sandstones have been removed by solution, and there remain rhomb-shaped cavities which are commonly lined with hematite. Where preserved, the carbonate mineral appears to be an impure dolomite, which contains some opaque inclusions and which may have been recrystallized. The individual dolomite grains are subhedral and range in size from sublithographic to coarsely crystalline. Twinning lamellae are fairly common.

Crystal Mountain Sandstone

In its type area in the Crystal Mountains of western Arkansas, the Crystal Mountain sandstone of Ordovician (?) age is about 850 feet thick. The formation consists almost entirely of sandstone, although there is a thin basal conglomerate and a small amount of shale. Miser and Purdue (1929, p. 24) state that

> "The sandstone is massive and coarse grained and is composed of well-rounded, translucent quartz grains, cemented together in some beds by calcium carbonate but in most beds by silica. The calcareous beds are brown and become rather friable on weathering, but the siliceous sandstone is light gray, is hard and dense, and breaks into large and small blocks that cover much of the surface. Joints in the sandstone are abundant. . . . Much of the sandstone is cut with a network of veins of white quartz from a fraction of an inch to several inches thick."

In McCurtain County, Oklahoma, the thickness of the Crystal Mountain sandstone was estimated by Honess (1923, p. 48), to be at least 500 feet. In the area mapped by Pitt (1955, p. 21), the thickness of the Crystal Mountain sandstone (restricted) is estimated by him to range only from 5 to 100 feet. The petrographic descriptions which follow are limited to the restricted Crystal Mountain sandstone and descriptions of rocks from questionable outcrops have been omitted.

Petrographically, the sandstones of the Crystal Mountain formation closely resemble the sandstones of the Lukfata formation. The Crystal Mountain sandstones are quartzose, well-sorted, very fine-to medium-grained sandstones cut by veins of quartz. Cementation is dominantly siliceous. In many sandstones the contacts between adjacent quartz grains are complexly sutured and interlocked. Most specimens contain small rhomb-shaped cavities lined with hydrous iron oxides; subhedral to euhedral, sublithographic to very finely crystalline rhombs of impure (Fe- and/or Mn-bearing) dolomite are preserved in a few specimens. Some sandstones near the basal conglomerate of the Crystal Mountain have calcite cement (see Pl. 1, Fig. 1); in some specimens, the calcite cement has attacked and partially replaced the peripheries of the quartz grains.

The larger quartz sand grains appear to have been rounded or subrounded at the time of deposition; secondary enlargement of the quartz grains and subsequent partial recrystallization during metamorphism has resulted in an induced angularity. Among the more conspicuous indications of low-grade metamorphism are the abundance of quartz veins, the granulation and elongation of the quartz grains, and the reconstitution of the original interstitial argillaceous material into new chlorite and sericite. Grains of feldspar are almost completely replaced by felty aggregates of sericite laths in the more completely recrystallized sandstones. The quartz grains in many specimens contain abundant opaque inclusions, and show strain shadows and undulatory extinction. Most of the quartz veins are simple fracture-fillings which are relatively free of inclusions and of coarser grain size than the sandstones into which they were intruded; veins with comb structure were noted at places.

The basal conglomerate of the Crystal Mountain sandstone forms a massive bed ranging from about 10-14 feet thick. It consists of pebbles and granules of quartz, chert, limestone, shale and argillite in a matrix of sandy limestone or calcareous sandstone (Honess, 1923, pp. 46-47 and Pitt, 1955, p. 21). In the DeQueen 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.

Blakely Sandstone

The Blakely formation of Ordovician age ranges in thickness from 0-500 feet in the central Ouachita Mountains; it is not exposed either at Black Knob Ridge or in the Potato Hills, Oklahoma. Honess (1923, p. 58) restricts the Blakely in McCurtain County, Oklahoma to some dark quartzitic sandstone 10-15 feet thick lying immediately above the Mazarn shale. These quartzitic sandstones contain abundant veinlets of milky and smoky quartz. Pitt (1955, pp. 23-24) discusses these outcrops and concludes that they do not resemble the Blakely at its type locality in Arkansas.

In western Arkansas, where the Blakely is definitely a valid formation, Miser and Purdue (1929, pp. 29-30) describe it as follows:

> "The formation consists of about 400 feet of interbedded shale and sandstone, with the shale making probably 75 percent of the whole. . . . The sandstone is in beds, most of which are not more than 10 feet thick. It is made up of medium-sized, well-rounded, translucent quartz grains, firmly cemented together in most beds by silica but in the others by calcium carbonate."

Sandstones of the Blakely formation are moderately ferruginous and argillaceous, and fine-to medium-grained. Many of the larger quartz grains appear to have been rounded to subrounded originally, but are now mostly subangular. This angularity results both from purely mechanical granulation and from reaction with the "clay" matrix during metamorphism. The "clay" matrix has been altered to intercrystallized sericite, chlorite, and microcrystalline quartz. Some specimens are cut by quartz veinlets, and some contain iron-stained rhombic cavities.

Specimens of Blakely sandstone studied by the writer contain numerous stable minor accessory ("heavy") minerals. Zircon, tourmaline, rutile, and anatase-leucoxene are fairly common. This abundance of "heavy" minerals contrasts strongly with the rarity of "heavy" minerals in the sandstones of the Crystal Mountain and Lukfata formations. Furthermore, most of the quartz grains in the Blakely show no sign of secondary enlargement before metamorphism, and most of the quartz grains are relatively free of inclusions and extinguish sharply under crossed nicols. It is not known whether these characteristics are purely local variations due to sampling or whether they are of regional significance.

Womble Shale

The Womble formation of Ordovician age ranges in thickness from 250-1000 feet; it crops out at Black Knob Ridge, in the Potato Hills, and in the central anticlinorium of the Ouachita Mountains. The base of the Womble is marked by a fault contact at all exposures near Black Knob Ridge, and the true maximum thickness of the Womble in this area is not known. Well data suggest that the Womble is quite thick, possibly several thousands of feet.

Shale is the predominant lithologic type at Black Knob Ridge, in the Potato Hills, and in western Arkansas. In McCurtain County, Oklahoma, the formation consists largely of metamorphosed argillaceous sandstone.

Womble sandstone from western Arkansas is hard, compact, fine-grained, quartzose, and bluish-green in the unweathered specimen (Miser and Purdue, 1929, p. 32). Petrographically, a typical specimen is fine- to medium-grained, quartzose and shows well-developed secondary enlargement of the quartz grains. The quartz grains were rounded to well-rounded at the time of deposition, and the original outlines of the grains are marked by rows of inclusions. The sorting of the quartz grains is only fair, and bimodal size distribution is suggested by some thin sections. Secondary enlargement of the quartz grains has interlocked them into a solid impermeable mass, but there is little obvious indication of metamorphism in the rocks that were originally free of interstitial clay. Iron oxides, potassic feldspar grains, and such resistant "heavy" minerals as zircon, tourmaline, and leucoxene are also present.

The thin Womble sandstones at Black Knob Ridge are mostly argillaceous, very poorly sorted, and very fine-grained. The angular to subrounded, fine sand-to silt-size grains of quartz, potassic feldspar, sodic plagioclase feldspar, chert, quartzite, detrital mica flakes, and detrital shale are embedded in a "clay" matrix that has been partially reconstituted into chlorite and sericite.

In McCurtain County, Oklahoma, Honess (1923, p. 62) describes the Womble sandstones as generally soft, schistose, micaceous, and fine-grained. The fresh rocks are green in color, weathering to red. Petrographically, these rocks are among the most interesting in the Ouachita Mountain region. The sandstones are fine-to medium-grained and generally poorly sorted. They range from relatively clean quartzose sandstones to sandy shales in which large detrital grains are embedded in a clay-mica matrix like plums in a pudding. Most of the sandstones appear to have been moderately argillaceous originally. All of the sandstones are definitely metamorphosed.

The lithology most characteristic of the Womble in McCurtain County is a fine-to medium-grained, chloritic and sericitic sandstone that has undergone thorough low-grade metamorphism. A photomicrograph of a thin section of this type is shown in Pl. 1, Fig. 2. The quartz grains have been stretched out and granulated in part, and the original clay-silt ("mud") matrix has been completely reconstituted into aggregates of chlorite, sericite, biotite, and quartz. The borders of the detrital sand grains have a characteristic corroded or "moth-eaten" appearance due to reaction with the "clay" of the matrix during metamorphism. The outlines of the detrital grains are blurred and obscure as a result of partial recrystallization. Quartz is the most abundant detrital grain but there are numerous polygranular or composite (rock) grains such as quartzite, chert, and argillite. Potassic and sodic plagioclase feldspars, hydrous iron oxides, and detrital micas are fairly common. Most detrital quartz grains show strain shadows and undulatory extinction under crossed nicols. Small iron-stained carbonate rhombs and such resistant "heavy" minerals as zircon, tourmaline, and leucoxene are present in some thin sections.

The Womble sandstones in McCurtain County which contained abundant original clay show even more pronounced schistosity in thin section. The original "clay" matrix is generally completely reconstituted into chlorite and sericite; the detrital grains have not only been flattened and stretched-out parallel to the schistosity, but some also appear to have been rotated so as to bring their long axes parallel to the alignment of the long axes of the micaceous minerals. Numerous flaws or micro-faults parallel to the schistosity may also be present. Some of these rocks are typical examples of the "schistose grit" of British petrographers (Harker, 1939, pp. 240-246 and Fig. 14). A photomicrograph of a thin section of one of these schistose Womble sandstones is shown in Pl. 1, Fig. 3.

Blaylock Sandstone

The Blaylock formation of Silurian age has a smaller areal extent than most of the other formations of the Ouachita Mountains. The Blaylock formation consists of sandstone, siltstone, and shale, with sandstone and siltstone usually occurring in greater amount. The formation varies greatly in both lithology and thickness within short distances; Miser and Purdue (1929, p. 44) state that in the Caddo Gap quadrangle the Blaylock increases in thickness from a feather edge to 1500 feet within a distance of three miles. Miser and Purdue (1929, p. 44) describe the Blaylock sandstone as follows:

> "The sandstone is in remarkably even-bedded layers from 1 to 6 inches thick, but in a few places the layers reach a thickness of three feet. It consists of fine angular quartz grains and a little mica with quartz as the cementing material . . . . Most of the sandstone is hard, dense, light to dark gray, laminated, and quartzitic, but some is soft and has a yellow color and splits into more or less parallel plates when struck with a hammer. Flattened clay pellets and rather fine, crooked markings that are probably worm trails are common. Joints are numerous, and many of them are filled with thin veins of white quartz."

Petrographically, the Blaylock sandstones are coarse-grained siltstones and very fine- to fine-grained quartzose sandstones that have undergone distinct low-grade metamorphism. A photomicrograph of a thin section of Blaylock siltstone is shown in Pl. 1, Fig. 4. Most sandstones are fairly well-sorted to well-sorted, moderately micaceous and slightly feldspathic, and contain abundant silt-size stable minor accessory ("heavy") minerals. Almost every thin section is cut by quartz veinlets; some are simple fissure-fillings whereas other veinlets are compound, sinuous, bifurcating, and pinch and swell within very short distances (Honess' "zones of penetration"). The latter veinlets are clear and conspicuous in transmitted light, but separate into individual quartz grains of different sizes and different optic orientations under crossed nicols and are nearly invisible. These compound veinlets may represent repeated fracturing and healing of fractures along certain zones in the rock; they typically trend approximately normal to the bedding.

Quartz predominates greatly among the detrital "light" minerals, but there are some potassic feldspars, sodic plagioclase feldspars, detrital micas, and composite (rock) grains. Most of the detrital grains have corroded, "moth-eaten" edges where the sandstone contained some original "clay"; this "clay" in the matrix is now reconstituted into anhedral green chlorite with some sericite. The abundant chlorite acts as a cementing material in many specimens; where original "clay" was lacking the cement is siliceous and the quartz grains are interlocked into the texture of a quartzite. Inclusions of vermicular chlorite or chlorite laths within quartz grains are fairly common in Blaylock sandstones.

"Heavy" minerals tend to be concentrated in certain thin laminae which suggest the black sand layers of modern beaches. Some thin sections of Blaylock sandstone contain over a thousand grains of stable minor accessory minerals, mostly leucoxene, zircon, rutile, tourmaline, and staurolite.

Some Blaylock sandstones contain abundant black opaque graphitic material in the interstices between the grains. This material may have been bituminous originally. Where it is present, the metamorphic reaction between sand grains and "clay" matrix is retarded and reconstitution of the "clay" is incomplete.

Missouri Mountain Shale

The Missouri Mountain shale of Silurian age ranges in thickness from 0-300 feet in the Ouachita Mountains of Oklahoma and western Arkansas. It consists largely of red and green clay shale, siliceous shale, chert, and slate, but it also contains some thin layers of sandstone and conglomerate. Miser and Purdue (1929, p. 47) state that

> "The sandstone and quartzite are in layers from 3 to 5 inches thick and occur in places near the base and top of the formation. They are gray and hard and are composed of rounded, translucent quartz grains. On Missouri Mountain, ... a layer of conglomerate 15 inches thick and also thinner ones, all made up of chert pebbles in a siliceous matrix resembling novaculite, occur in the upper part of the slate."

Most of the specimens of Missouri Mountain sandstone and conglomerate collected by the writer came from the Black Knob Ridge area. In this area the sandstones are essentially unmetamorphosed. Most specimens are fairly well-sorted to well-sorted, very fine-grained to fine-grained, quartzose sandstones. Well-developed secondary enlargement of the quartz grains is typical, and most sandstones are cemented by silica. Some rocks are cherty and calcareous, and black opaque (asphaltic?) material in the interstices between the sand grains is fairly common. Quartz is the dominant detrital mineral, but there are also detrital grains of chert, chalcedony, chlorite, argillite, potassic and sodic plagioclase feldspar, and phosphatic and/or chitinous fossil fragments. "Core" analyses made from hand specimens of some of the asphaltic sandstones indicated porosities ranging up to 18% and measurable oil content. A photomicrograph of a thin section of Missouri Mountain sandstone is shown in Pl. 2, Fig. 1.

Thin sections were made of the conglomerate near the top of the Missouri Mountain formation at Grants Gap (center of Sec. 10, T. 1 S., R. 12 E., Atoka County, Oklahoma). This rock is a sandy, calcareous, granule- to pebble-size, chert conglomerate. The matrix of this rock consists of a mixture of anhedral calcite, chalcedonic chert, and abundant rounded to sub-angular quartz sand and silt. The coarser detrital fragments are distinguished by their wide variety. They include the following lithologic types:

1. Medium to coarsely crystalline, slightly silty limestone.
2. Clear cryptocrystalline compact chert.
3. Spiculite chert and radiolarian chert.
4. Siliceous shale.
5. Siliceous limestone with dolomite rhombs.
6. Equigranular finely crystalline dolomite and limy dolomite.
7. Chalcedony.
8. Phosphatic and/or chitinous fossil fragments.

Some of these rock types are similar lithologically to rocks in the Bigfork, Womble, and older formations, but other detrital fragments cannot be traced definitely to any older formation cropping out in the Ouachita Mountains.

Arkansas Novaculite

The Arkansas Novaculite formation of Devonian-Mississippian age varies in thickness from 250 to 900 feet and consists largely of novaculite, chert, siliceous shale, and soft clay shale (see Goldstein and Hendricks, 1953, pp. 428-431). Although sand grains in the novaculite have been reported by most writers, no sandstones have been described from this formation. However, T. A. Hendricks and H. D. Miser, and C. A. Weintz and the writer have collected sandstones from this formation at localities in Black Knob Ridge, in the Potato Hills, and in western Arkansas.

A typical sandstone from the lower member of the novaculite at Black Knob Ridge is poorly-sorted, slightly argillaceous and cherty, and fine-to very fine-grained. The angular to sub-rounded, fine sand-to silt-size grains of quartz, chert, feldspar, quartzite, rock fragments, garnet, zircon, tourmaline, and primary micas are embedded in a matrix of argillaceous material and hydrous iron oxides.

Specimens from the top of the lower member of the novaculite in the Walnut Creek section of the Potato Hills are dolomitic, glauconitic, argillaceous, very fine-grained sandstones and coarse siltstones. The angular to sub-rounded grains of quartz, chert, feldspar, and detrital shale range from medium sand to coarse silt in size. Abundant sublithographic to very finely crystalline, euhedral to subhedral carbonate (dolomite?) rhombs are preserved in some specimens; in other specimens there are numerous rhomb-shaped cavities.

Sandstones 50-60 feet below the top of the novaculite in the Potato Hills are argillaceous, very poorly sorted, very fine-grained quartz sandstones. The subangular to angular grains of quartz, chert, potash and plagioclase feldspar, detrital shale, quartzite, and primary micas are embedded in a matrix of cryptocrystalline silica, silt-size quartz, iron oxides, and argillaceous material that has been reconstituted in part into chlorite and sericite. Traces of glauconite were noted in some of the thin sections.

Hot Springs Sandstone

The Hot Springs sandstone of Mississippian age is exposed only in and near the city of Hot Springs, Arkansas, where it ranges in thickness up to about 200 feet. The Hot Springs sandstone unconformably overlies the Arkansas novaculite and passes by gradual lithologic transition into the overlying Stanley shale.

In the type area, according to Purdue and Miser (1923, p. 5),

> "This formation consists of sandstone and some shale and conglomerate. The shale occurs near the top and the principal bed of conglomerate at the base. The sandstone, which is gray, hard, and quartzitic, is composed of grains of quartz sand that range in size from fine to medium. Its layers are from 3 to 8 inches thick, though here and there they reach a thickness of 6 feet."

Most specimens of Hot Springs sandstones are well-sorted, very fine-to fine-grained, and quartzose. Thin sections show that the individual quartz grains have been "welded" together with secondary silica so that the rock is virtually non-porous. The original outlines of the quartz grains are obscure and poorly defined, possibly indicating some recrystallization. Quartz is by far the most abundant mineral, but there are some grains of potassic feldspar, chert, quartzite, detrital shale, muscovite, ilmenite, leucoxene zircon, and tourmaline. Some of the contacts between interlocked adjacent quartz grains are sutured and serrate, and many grains show strain shadows and undulatory extinction. A photomicrograph of a thin section of Hot Springs sandstone is shown in Pl. 2, Fig. 2.

In the cleaner, clay-free sandstones of the Hot Springs formation there is little indication that the rocks have been subjected to metamorphism. However, those rocks which did contain small to moderate amounts of original interstitial argillaceous material do show definite low-grade metamorphism. In general, the interstitial clay has been reconstituted almost completely into chlorite and sericite, and many of the detrital shale grains have been altered into nests of intercrystallized chlorite, sericite, and cryptocrystalline quartz.

Stanley Shale

The Stanley shale is one of the thickest and most widespread formations in the Ouachita Mountain region. It ranges in thickness from about 6,000 feet in western Arkansas and McCurtain County, Oklahoma, to a maximum of about 12,000 feet in the southeastern part of the area mapped by Hendricks and others (1947) in the Ouachita Mountains of Oklahoma.

The lower part of the Stanley shale has been shown by Hass (1950) to be of Mississippian (Meramec) age; the age of the bulk of the formation is not accurately known. There is a possibility that the Mississippian-Pennsylvanian contact occurs at some undetermined horizon within the Stanley shale (Hass, 1956), but other writers consider the Stanley to be entirely of Mississippian age (Cline, 1956). Details of the stratigraphy of the formation have been given by Harlton (1938), Hendricks and others (1937, 1947), Bokman (1953), Miser and Purdue (1929), Honess (1921, 1923) and many others.

In the Black Knob Ridge area, the typical Stanley sandstone is poorly sorted, argillaceous, chloritic, and weakly metamorphosed to unmetamorphosed. The detrital grains are angular to subrounded, and range in size from medium sand to medium silt. Although quartz is most abundant, there are numerous detrital grains of chert, potassic feldspar, quartzite, shale, and sodic plagioclase feldspar. These detrital grains are embedded in a "mud" matrix which has been reconstituted in part into new chlorite and sericite. Microscopic folds and faults and tiny veinlets of quartz are abundant. On the basis of composition, the average Stanley sandstone is subgraywacke (Goldstein and Reno, 1952, Fig. 9), but texturally the Stanley sandstones may be considered graywacke on the basis of their primary mud matrix (see Pettijohn, 1957, pp. 302-303). A photomicrograph of a thin section of a typical Stanley sandstone from this area is shown in Pl. 2, Fig. 7.

In McCurtain County, Oklahoma and in western Arkansas, the Stanley sandstones show widespread low-grade metamorphism. The Stanley sandstones of this area differ from those near Black Knob Ridge chiefly in that there are more cataclastic structures and that the interstitial "mud" matrix is generally more completely reconstituted into new chlorite and sericite. A typical specimen is sheared and distorted, micaceous, and very fine-grained. The detrital grains are dominantly subangular and, in addition to the dominant quartz, there are considerable amounts of chert, quartzite, and sericitized feldspars. The newly formed micaceous minerals bend around some of the larger detrital grains in a structure intermediate between microscopic "augen" and "mortar" structure. At places the edges of the quartz grains have a corroded appearance due to reaction with the mud matrix during metamorphism. Alignment of the long axes of elongate quartz grains, microscopic folding and faulting, and intrusion of quartz veinlets into the sandstones are common occurrences. A photomicrograph of a thin section of Stanley sandstone from this area is shown in Pl. 2, Fig. 8.

An acidic crystal-vitric tuff named the Hatton tuff lentil of the Stanley formation is widely distributed in McCurtain County, Oklahoma and in western Arkansas (Miser and Purdue, 1929; Honess, 1923). Other tuffs which may or may not be correlative crop out in the Stanley near the Potato Hills, and impure tuffs have been found in the Stanley in cuttings from several wells in southeastern Oklahoma.

Jackfork Sandstone

The Jackfork sandstone of Mississippian age (Cline, 1956, p. 105) or of lower Pennsylvanian age (Hass, 1956, p. 29) ranges in thickness from about 1150 feet up to a maximum of about 7000 feet in the western part of the Oklahoma Ouachita Mountains. Two complete and unfaulted sections of Jackfork sandstone in the Kiamichi Range were measured recently by Cline and Moretti (1956), and were found to be 5600 feet and 6000 feet thick.

In the Black Knob Ridge area the Jackfork sandstone consists largely of sandstone with intercalated shale and siliceous shale. The fresh sandstones range in color from white through gray to blue, and most of the sandstones are fine-to medium-grained. Much of the Jackfork sandstone is massive to thick-bedded in the southeastern part of the area, but at some horizons it is thin-bedded. Mineralogically, these Jackfork sandstones are "dirty" quartzose sandstones and subgraywackes; they are typically low in feldspar and have an introduced mineral cement (usually silica). A typical sandstone is ferruginous to chloritic, argillaceous, fairly well-sorted and fine-grained. Most detrital grains of quartz, chert, quartzite, shale, and feldspar are subangular to subrounded. Laminae consisting largely of minor accessory minerals are present at places, and complete secondary silica cementation was noted in some specimens. The rocks are not obviously metamorphosed, and interstitial "clay" is not completely reconstituted into chlorite and sericite in Jackfork sandstones from the Black Knob Ridge area. In general the Jackfork sandstones are less argillaceous, better-sorted, coarser-grained, and show less cataclastic structures than Stanley sandstones from the same area. Photomicrographs of two thin sections of Jackfork sandstones are shown in Pl. 3, Fig. 1 and Pl. 3, Fig. 2.

In the central anticlinorium of the Ouachita Mountains, the Jackfork sandstones are mostly poorly sorted and are tightly cemented with silica. Most of the rocks show some reconstitution of interstitial argillaceous material into new chlorite and sericite. Quartz is still the dominant mineral, although Bokman (1953) found that such polygranular fragments as chert, quartzite, quartz schist, shale, slate, and phyllite made up about 15-20% by area of his slides of Jackfork sandstones. The writer's slides do not contain as many polygranular grains.

Atoka Shale

The Atoka formation is one of the most variable in the Ouachita Mountain region; there are distinct changes in lithology, composition, and thickness of beds within short distances. Furthermore, dissimilar beds termed "Atoka" are found in the central anticlinorium of the Ouachita Mountains, near Black Knob Ridge and the Potato Hills, in the thrust-faulted frontal zone of the Ouachita Mountains, and in nearby foreland areas such as the McAlester Basin.

In the western part of the Ouachita Mountains in Oklahoma, the Atoka formation is at least 4000 feet thick. It consists mainly of light-gray, silty, micaceous, and flaky shale that contains lenses and beds of sandstone. The sandstone is tan to buff in color, very fine-to medium-grained, and is generally well-bedded in beds a few inches to several feet thick. Thin sections from this area and from western Arkansas indicate that these sandstones are generally poorly sorted, micaceous, and moderately argillaceous. Many specimens contain anhedral, sublithographic to very finely crystalline calcite, which is uncommon in both Stanley and Jackfork sandstones. In addition to the dominant quartz, Atoka sandstones generally contain chert grains, potassic and plagioclase feldspar, detrital flakes of muscovite and biotite, detrital shale, and hydrous iron oxides. In some specimens the interstitial argillaceous material and the detrital shale fragments have been almost completely reconstituted into new chlorite and sericite, whereas the "clay" in other sandstones shows only incipient reconstitution and the rocks would be classified as unmetamorphosed. Atoka sandstones of the central anticlinorium of the Ouachita Mountains in McCurtain County, Oklahoma, and western Arkansas are weakly metamorphosed for the most part.

Thin sections of "Atoka" sandstones in the thrust-faulted blocks of the frontal zone of the Ouachita Mountains lying between the Choctaw and Ti Valley faults also show considerable lithologic variation. Most of these sandstones are poorly sorted, but some are well-sorted. The median grain size of these sandstones ranges from fine sand to coarse silt. Most of the sand grains are subrounded to subangular, with few extremes of either roundness or angularity. Quartz is the dominant mineral, but micas and polygranular fragments such as chert, quartzite, shale, and metamorphic rock grains are abundant locally. A wide variety of feldspars has been noted in these sandstones, although they rarely total more than 3-4% of the rock. The sandstones are weakly to moderately well-cemented with either silica or clay binding material. The interstitial "clay" shows little reconstitution, and quartz veining and cataclastic structures are either minor or absent. Photomicrographs of thin sections of specimens from two of the thrust blocks of the frontal zone are shown in Pl. 3, Fig. 3 and Pl. 3, Fig. 4.

Heavy Mineral Assemblages

Comparatively little work has been published on the minor accessory (heavy) minerals in sandstones of Ouachita facies. Bokman (1953) made a detailed and excellent study of the lithology and petrology of the Stanley and Jackfork formations, including numerous analyses of the heavy minerals. Pitt (1955, p. 22) has discussed briefly the heavy minerals of the Crystal Mountain sandstone. Honess (1923) describes heavy minerals from rocks of Ouachita facies in McCurtain County, but almost all of his descriptions are of heavy minerals in thin sections.

The most comprehensive study of heavy mineral assemblages of sandstones of Ouachita facies known to the writer is an unpublished report of H. D. Winland (1953). A star diagram of some of his results is shown in Fig. 1. It will be noted that sandstones older than Missouri Mountain (Silurian) contain for the most part such extremely stable minerals as zircon, tourmaline, leucoxene, rutile, and iron ore minerals. A flood of garnet appears in the Missouri Mountain sandstones and lasts well into the time of deposition of the Stanley sandstones. Possibly the borderland of crystalline rocks ("Llanoria") adjacent to the Ouachita geosyncline was stripped of its sedimentary cover in some

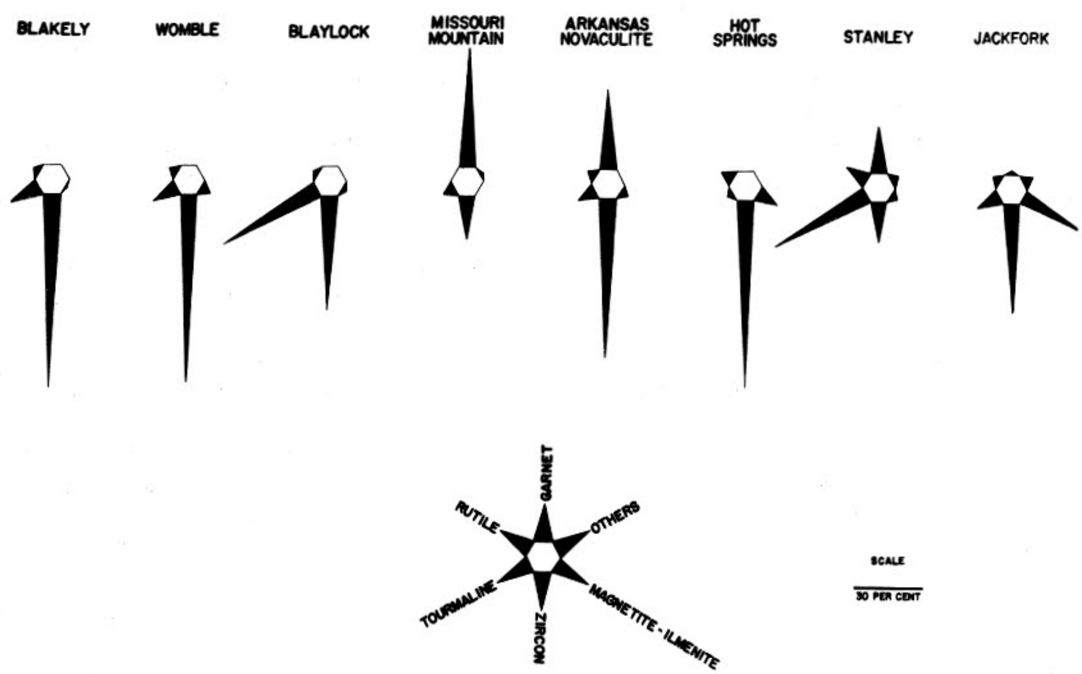


Fig. 1.--Star diagrams of average heavy mineral composition of some Ouachita Mountain sandstones (after H. D. Winland).

late Taconic or Caledonian epeirogeny. and contributed fresh detritus through late Silurian, Devonian, and early Mississippian time. By middle Stanley time and throughout Jackfork time, most of the sedimentary detritus supplied to the Ouachita geosyncline was being derived from erosion and redeposition ("cannibalization") of uplifted previously-deposited sediments of the geosyncline. Erosion of the youngest sediments would contribute much second-cycle garnet; when the pre-Missouri Mountain sediments were eroded, only the resistant multi-cycle heavy minerals of the older sandstones would be available for redeposition. This factor may account for the abundant garnet in early Stanley sandstones and the paucity of garnet in Jackfork sandstones. Bokman (1953) has a different explanation for the differences in heavy mineral assemblages of the Stanley and Jackfork sandstones; he considers that the Stanley is a first-cycle sediment derived from a source composed primarily of crystalline rocks, whereas the Jackfork sandstone was derived principally from meta-sediments.

Theoretical Considerations

The material presented up to this point in this study has been largely factual. Some of the collections of sandstones were not comprehensive, and the sandstones collected may not be truly representative samples of the formations, but petrographic description is generally one of the more exact portions of our inexact science of geology. When these descriptions are fitted into a unified scheme of Paleozoic sedimentation and tectonics in the Ouachita Mountain region, we move from observed facts to theory.

Unfortunately, classification of sandstones has reached a stage of redundant complexity which renders it necessary for a writer either to define each of his terms or follow rigidly some reference work of wide distribution. In the following discussion, the sandstone names are used according to Pettijohn (1957, pp. 283-339). He states (ibid., p. 290) that

> "It seems necessary, . . . , to define sandstone on the basis of those parameters which are the indices of provenance, maturity, and fluidity of the depositing medium. These are the 'source rock index' or ratio of feldspar (plutonic) to rock fragments (supracrustal), the 'maturity index' or ratio of quartz (plus chert) to feldspar plus rock fragments, and the 'fluidity index' or ratio of sand detritus to the interstitial detrital matrix. On the basis of these three parameters sandstones may be grouped into four major classes and several lesser ones (. . .). The major groups are the graywackes, the lithic sandstones, arkosic sandstones, and the orthoquartzites."

The Lukfata, Crystal Mountain, Blakely, Missouri Mountain, and Hot Springs sandstones contain mostly quartz with a little chert. They do not have appreciable amounts of interstitial detrital matrix material, and they contain only small amounts of feldspar and rock fragments. These sandstones would be classified as orthoquartzites.

The Stanley sandstones and the Womble sandstones of McCurtain County have an interstitial primary mud matrix and contain moderate amounts of feldspar and moderate to large amounts of rock grains. These rocks are classified as lithic graywackes. Most of the thin sandstones in the Arkansas novaculite are very much like those in the lower Stanley and would be classified similarly.

The sandstones of the Blaylock, Jackfork, and Atoka formations, although quite variable in lithology and composition, generally are fairly well-sorted and have their intergranular space filled by a combination of clay materials and mineral cement. Most of these sandstones have less than 15% of detrital matrix material and less than 25% of feldspars plus rock fragments, with rock fragments more abundant than feldspars. These sandstones are classified as quartzose subgraywackes or "protoquartzites".

In order to fit these various sandstone types into a unified picture of sedimentation and tectonics in the Ouachita Mountain region, it is necessary to discuss briefly the modern concept of the "geosynclinal cycle".

One of the most fundamental factors influencing sedimentation is tectonism. The relation of sediment types to tectonics is embodied in the theory of the geosynclinal cycle. There is an abundant literature on this subject, which has been summarized ably by Pettijohn (1957, pp. 636-644). He points out that in most geosynclinal basins there is a normal geosynclinal cycle in which the geosyncline is first initiated, then filled, and lastly (but not always) deformed and uplifted. The sediments which accumulate during the various stages of geosynclinal history are more or less petrographically distinct. In the early stages of a geosyncline the first sediments are usually orthoquartzites and carbonates, followed by an euxinic facies of black shales with some siltstone, siliceous limestone, or chert. This facies is succeeded by the "flysch" facies in which large amounts of clastics are poured into a rapidly subsiding geosynclinal trough; the typical deposits of this facies are graywacke and dark shales. As the geosynclinal trough fills up, the "flysch" facies is succeeded by the "molasse" facies in which the sandstones (subgraywacke and protoquartzite) are coarser, cleaner, and cross-bedded. Finally, the geosyncline is deformed, uplifted, and reducing conditions replaced by an oxidizing environment in which red mudstones and red sands become abundant.

On first inspection our classification of Ouachita Mountain sandstones does not seem to fit the classic geosynclinal cycle. We have orthoquartzites in the Missouri Mountain shale and Hot Springs sandstone, followed by a superb example of the "flysch" facies in the Stanley shale of Mississippian age and by the "molasse" facies of the Jackfork sandstone (Mississippian or Pennsylvanian) and Atoka shale (Pennsylvanian). However, there is also an Early Ordovician orthoquartzite facies (Lukfata, Crystal Mountain, Blakely) followed by a "flysch" facies in the Womble shale (Ordovician) and a "molasse" facies in the Blaylock (Silurian). This suggests either our classic geosynclinal cycle is not strictly applicable in the Ouachita Mountain region, or, that we have an earlier, less well-preserved geosynclinal cycle in Ordovician and Silurian time. In my opinion, theoretical considerations based on sandstone types, concealed sediments thicknesses, position of major unconformities, and faunal studies definitely support the concept of the double geosynclinal cycle in the Ouachita Mountains.

Whereas no one doubts the geosynclinal origin of the thick sedimentary accumulations of Mississippian and Pennsylvanian time in the Ouachita Mountains and Marathon Uplift, most recent writers such as Barton (1945), Eardley (1951), and Harlton (1953) question whether the pre-Mississippian rocks in these two areas merit the designation of geosynclinal facies. It has been stated by Barton (1945) that the greatest thicknesses of Ordovician, Silurian, and Devonian sediments are found in rocks of foreland facies lying inward from the Ouachita fold-belt, and that these foreland areas represented the site of the geosyncline of Lower Paleozoic time. It cannot be denied that the Ouachita-Marathon geolyncline had periods of relative quiescence, without active downwarping and with little deposition of sediments. If we merely compare the measured thicknesses of Lower Paleozoic rocks cropping out in the Ouachita Mountains and Marathon Uplift with the thicknesses of Lower Paleozoic sediments encountered in such areas as the Val Verde Basin, Sheffield Channel, Delaware Basin, and McAlester Basin, then we must admit the force of Barton's argument.

Part of this anomalous situation is beginning to clear up as deep wells are drilled within the Ouachita Mountains and Marathon Uplift. Recent drilling in these areas has shown that some of the Ordovician sediments greatly exceed the measured outcrop thicknesses. Furthermore, in the Marathon region, the large amount of intraformational conglomerate in the argillaceous Ordovician limestones suggests that many of the earlier sediments were uplifted, eroded, and redeposited elsewhere. This sort of activity is typical of geosynclinal margins, whereas continuous, uninterrupted deposition of carbonates probably was going on in deeper water in the foreland areas to the north and west. This reworking of geosynclinal sediments thins them considerably; the thickness of the sediments as originally deposited was probably much greater.

Numerous unconformities in the Lower Paleozoic section of the Ouachita Mountains suggest that one or more epeirogenies may have stripped off large amounts of sediments. For example the Polk Creek shale of Ordovician (Richmond) age is always eroded at the top and may be missing entirely. The Blaylock sandstone of Silurian age is found only in a rather restricted portion of the central anticlinorium of the Ouachita Mountains in McCurtain County and western Arkansas, where it may be as thick as 1500 feet; the Blaylock is bounded by unconformable surfaces at both top and bottom and its original thickness at the time of deposition is unknown.

Finally, the faunal evidence published by Wilson (1954) strongly supports the idea of an Ordovician geosyncline in the Ouachita Mountains. Wilson found a pronounced and consistent biofacies change between the geosynclinal Ordovician faunas of the Ouachita-Marathon belt and the foreland-type faunas of the West Texas and Arbuckle Mountain areas.

It is probable that the sediments of Ordovician and Silurian time cropping out in the central anticlinorium of the Ouachita Mountains were distal to the geosynclinal axis of the Early Paleozoic geosynclinal cycle. Our entire picture of Paleozoic history in the Ouachita-Marathon fold-belt suggests progressive, if intermittent, migration of the Paleozoic geosyncline from the old source area ("Llanoria") on the south towards the "Texas craton" of Flawn (1954). The true axial sediments of the Ordovician-Silurian geosynclinal cycle are probably south (gulfward) of our presently-known rocks in the Ouachita-Marathon fold-belt, and these axial sediments are buried beneath younger rocks. Deeper drilling fulfward should eventually encounter these rocks, and sample study of the well cuttings will undoubtedly help to clarify the concept and reveal the details of the Early Paleozoic geosynclinal cycle in the Ouachita-Marathon fold-belt.

References

Barton, J. M., 1945, Pre-Permian axes of maximum deposition in West Texas: Amer. Assoc. Petroleum Geologists Bull., vol. 29, no. 9, pp. 1336-1348.

Bokman, John, 1953, Lithology and petrology of the Stanley and Jackfork formations: Jour. Geology, vol. 61, no. 2, pp. 152-170.

Branson, C. C., 1956, Tentative correlation chart of Ouachita facies rocks with Arbuckle facies and with northeastern Oklahoma: Ardmore Geol. Soc., Guidebook to Ouachita Mountain field conference, p. 15.

Cline, L. M., 1956, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains, Oklahoma: Tulsa Geol. Soc. Digest, vol. 24, pp. 100-106.

Cline, L. M., and Moretti, Frank, 1956, Two measured sections of Jackfork group in southeastern Oklahoma: Oklahoma Geol. Survey, Circular 41, pp. 1-20.

Eardley, A. J., 1951, Structural geology of North America. New York.

Flawn, P. T., 1954, Texas basement rocks; a progress report: Amer. Assoc. Petroleum Geologists Bull., vol. 38, no. 5, pp. 900-912.

Flawn, P. T., 1958, Personal communication, Sept. 8.

Goldstein, August, Jr., and Reno, D. H., 1952, Petrology and metamorphism of sediments of Ouachita facies: Amer. Assoc. Petroleum Geologists Bull., vol. 36, no. 12, pp. 2275-2290.

Goldstein, August, Jr., and Hendricks, T. A., 1953, Siliceous sediments of Ouachita facies in Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 64, pp. 421-442.

Harker, Alfred, 1939, Metamorphism, 2nd ed., London.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists, vol. 22, no. 7, pp. 852-914.

Explanation of Plate 1

Fig. 1. Crystal Mountain sandstone.
This specimen contains almost equal amounts of quartz and calcite. Note how the calcite has eaten into and partially replaced the quartz. Thin section CM-116, plane-polarized light, X 58.

Fig. 2. Womble sandstone.
This rock is a chloritic, medium grained sandstone. The edges of the quartz grains are corroded and frayed due to reaction with the clay of the matrix during metamorphism. Also note the silt-size quartz surrounding the large sand grains. Thin section WO-100, crossed nicols, X 75.

Fig. 3. Womble schistose sandstone.
This specimen is a fine- to medium-grained, poorly sorted, thoroughly metamorphosed, chloritic sandstone. It consists largely of elongated, stretched-out, quartz sand and polygranular (rock) grains in a "clay" matrix that has been completely reconstituted into new chlorite, sericite, and biotite. Thin section WO-110, plane-polarized light, X 101.

Fig. 4. Blaylock siltstone.
This photomicrograph is a micaceous, argillaceous, very coarse-grained siltstone in which reconstitution of the interstitial clay matrix into new chlorite and sericite is not complete. Note the angularity of the silt grains and the indistinct borders shown by many grains. Thin section BL-106, plane-polarized light, X 90.

Fig. 1

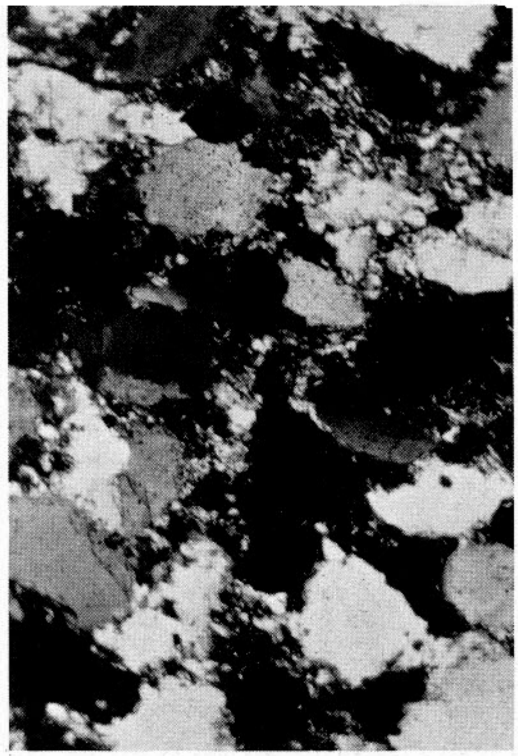


Fig. 2

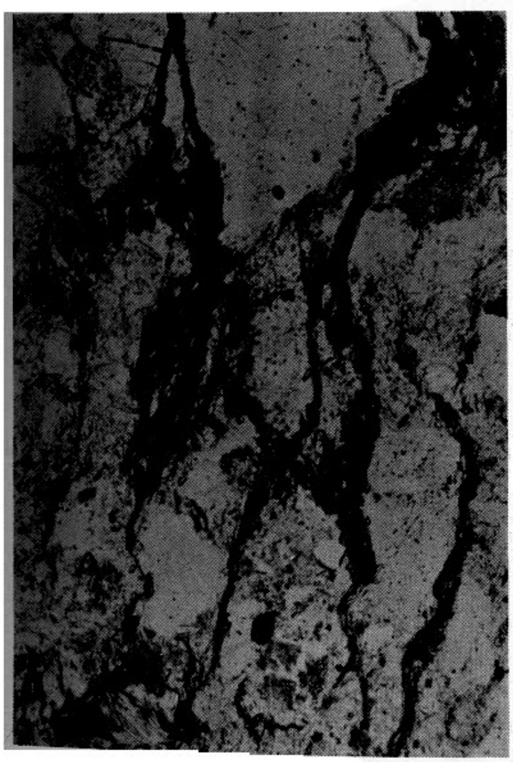


Fig. 3

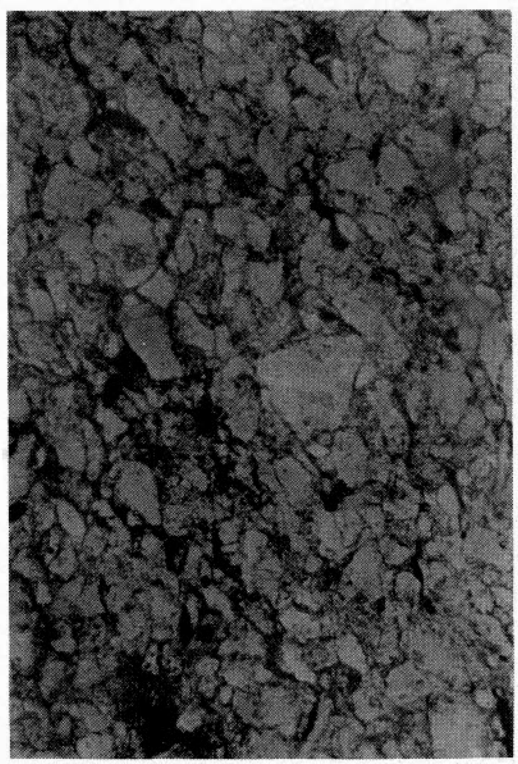


Fig. 4

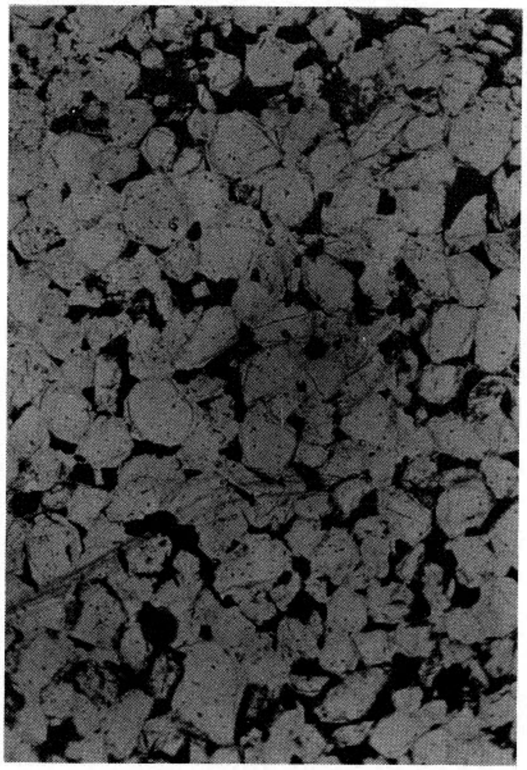


Fig. 1

Fig. 2

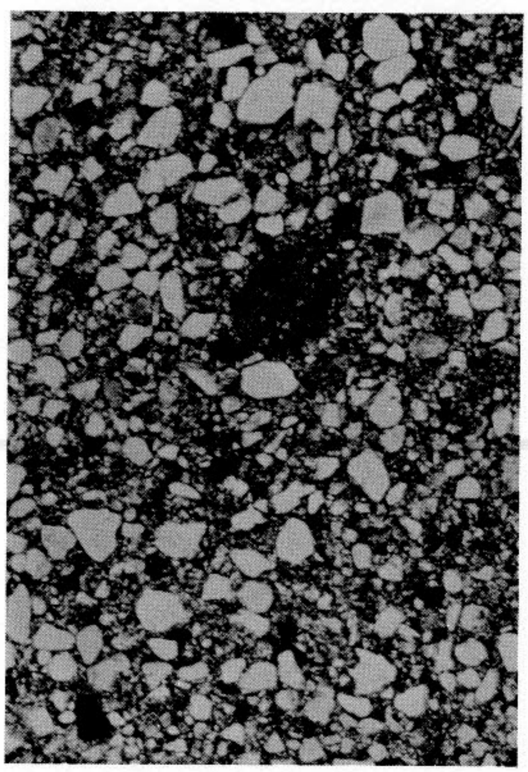


Fig. 3

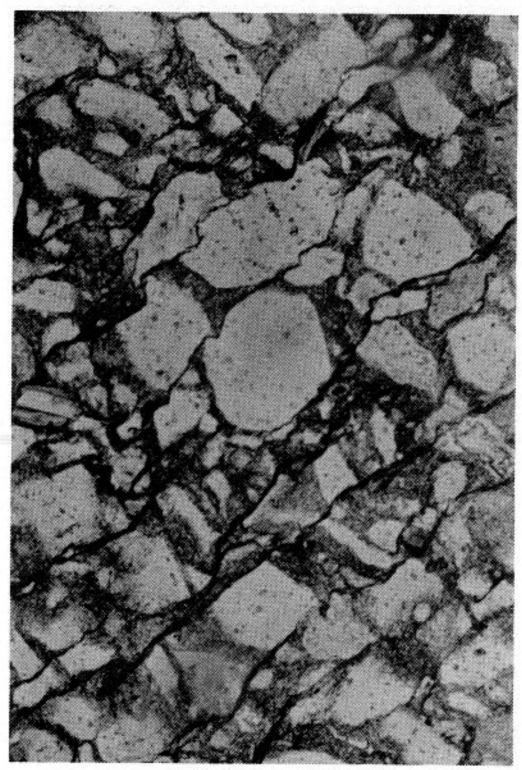


Fig. 4

Explanation of Plate 2

Fig. 1. Missouri Mountain sandstone.
This specimen is a ferruginous, very fine-grained sandstone that contains some black opaque interstitial bituminous (?) material. The rock is well-sorted and almost every quartz grain shows secondary enlargement and development of crystal faces. Dust-size inclusions outline the original borders of the quartz grains. Thin section MM-12, plane-polarized light, X 90.

Fig. 2. Hot Springs sandstone.
This specimen is a well-sorted, very fine-grained, quartzose sandstone. The quartz grains have been secondarily enlarged and completely interlocked with secondary silica cementation. Thin section HS-400, crossed nicols, X 73.

Fig. 3. Stanley sandstone.
This specimen is a relatively unmetamorphosed, very poorly sorted, argillaceous, very fine-grained sandstone from the Black Knob Ridge area. Thin section ST-203, plane-polarized light, X 33.

Fig. 4. Stanley sandstone.
This photomicrograph is of a very poorly sorted, chloritic, argillaceous, very fine-grained sandstone. The rock consists of medium sand- to fine silt-size detrital grains of quartz, chert, quartzite, micas, argillite, and feldspar embedded in a "mud" matrix that has been reconstituted in part into new chlorite and sericite. The specimen has been fractured by a network of sinuous, sub-parallel, discontinuous flaws or micro-faults. Thin section ST-2, plane-polarized light, X 101.

Explanation of Plate 3

Fig. 1. Jackfork sandstone.
This specimen is a poorly sorted, micaceous, argillaceous, fine- to very fine-grained sandstone. Note the mixture of subangular and rounded grains and the relatively small amount of interstitial clay. Thin section JF-211, plane-polarized light, X 27.

Fig. 2. Jackfork sandstone.
This specimen is a poorly sorted, fine-grained, quartz sandstone with silica cement. Note the relatively coarse polygranular (rock) grains and the conspicuous zircons. Thin section JF-206, plane-polarized light, X 27.

Fig. 3. Atoka sandstone.
This specimen is a poorly sorted, fine-grained sandstone that contains only small amounts of interstitial clay. The subangular to subrounded detrital grains are mostly quartz with subordinate amounts of chert, quartzite, and other rock grains. Thin section AT-50B, plane-polarized light, X 33.

Fig. 4. Atoka sandstone.
This rock is a poorly sorted, poorly cemented, very fine-grained, argillaceous sandstone. The clay matrix shows no signs of incipient reconstitution. Thin section AT-55B, plane-polarized light, X 33.

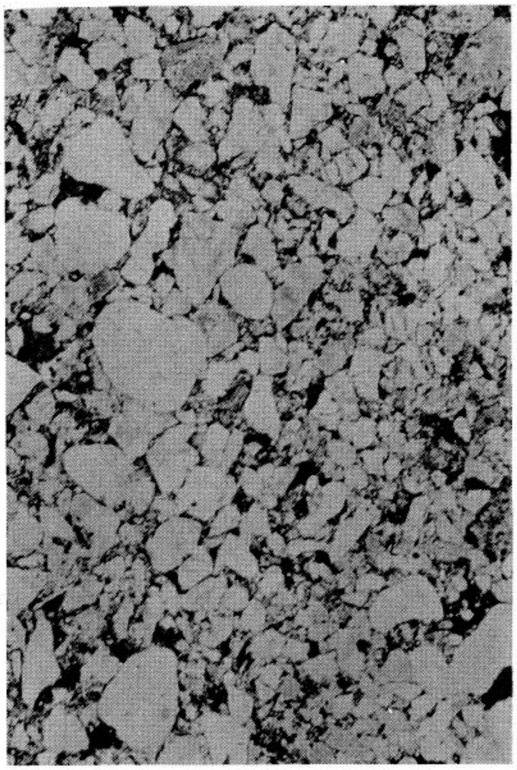


Fig. 1

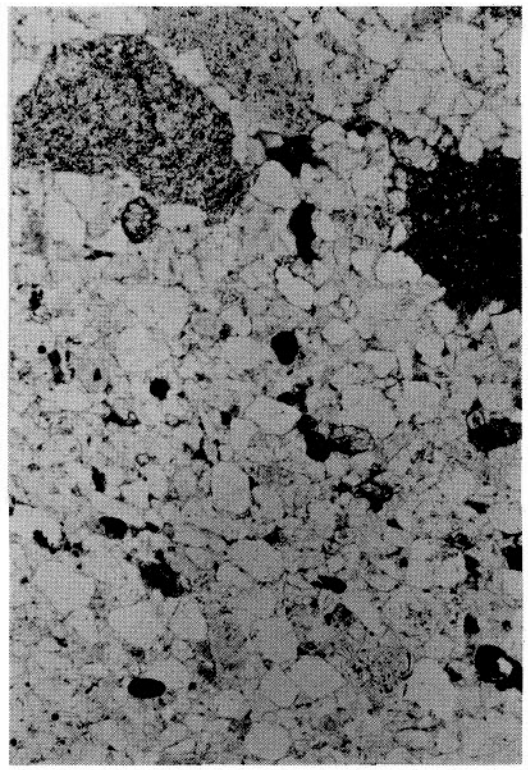


Fig. 2

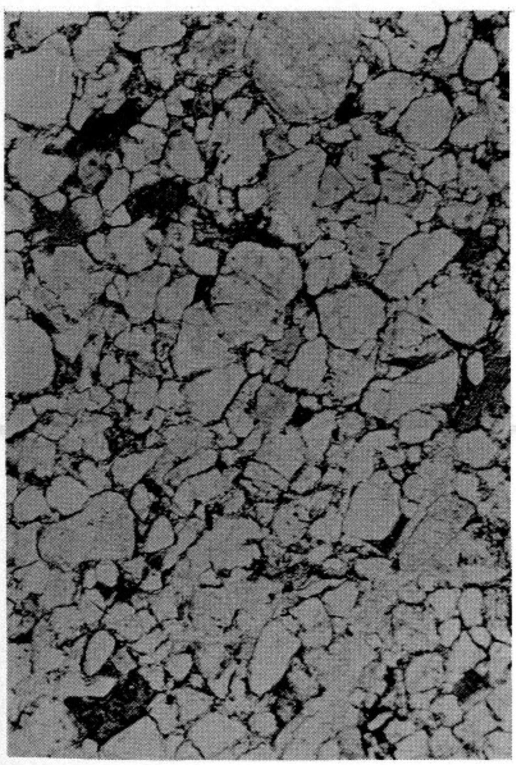


Fig. 3

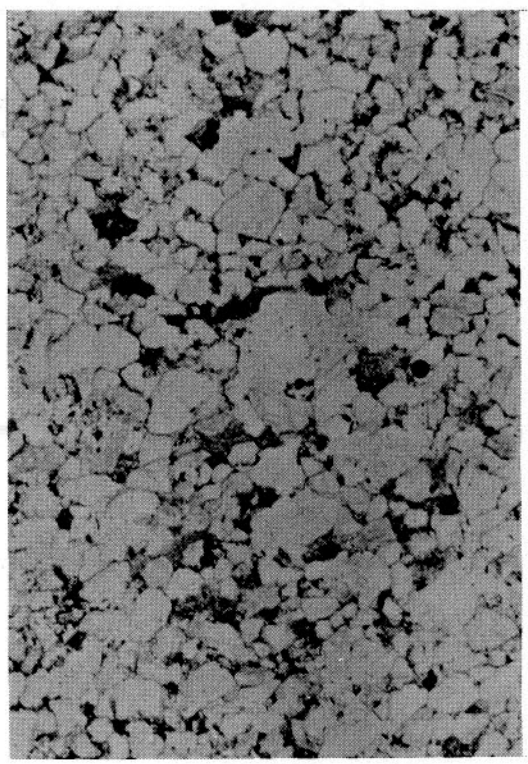


Fig. 4

Harlton, B. H., 1953, Ouachita chert facies, southeastern Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 37, no. 4, pp. 778-796.

Hass, W. H., 1950, Age of lower part of the Stanley shale: Amer. Assoc. Petroleum Geologists Bull., vol. 34, no. 7, pp. 1578-1584.

Hass, W. H., 1956, Conodonts from the Arkansas novaculite, Stanley shale and Jackfork sandstone: Ardmore Geol. Soc., Guidebook to Ouachita Mountain field conference, pp. 27-33.

Hendricks, T. A., Knechtel, M. M., and Bridge, Josiah, 1937, Geology of Black Knob Ridge, Oklahoma: Amer. Assoc. Petroleum Geologists Bull., vol. 21, no. 1, pp. 1-29.

Hendricks, T. A., and others, 1947, Geology of the western part of the Ouachita Mountains of Oklahoma: United States Geol. Survey Prelim. Map 66, Oil and Gas Inv.

Honess, C. W., 1921, The Stanley shale of Oklahoma: Amer. Jour. Sci., 5th ser., vol. 1.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey Bull., vol. 32, pt. 1, 278 pp.

Miser, H. D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles, Arkansas: United States Geol. Survey Bull. 808, pp. 1-195.

Pettijohn, F. J., 1949, Sedimentary rocks, 1st ed., New York.

Pettijohn, F. J., 1957, Sedimentary rocks, 2nd ed., New York.

Pitt, W. D., 1955, Geology of the core of the Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Circular 34, pp. 1-34.

Purdue, A. H., and Miser, H. D., 1923, Description of the Hot Springs district (Arkansas): United States Geol. Survey Geol. Atlas, Folio 215.

Wilson, J. L., 1954, Late Cambrian and early Ordovician trilobites from the Marathon uplift, Texas: Jour. Paleontology, vol. 28, pp. 249-285.

Winland, H. D., 1953, Petrography of Ouachita facies sandstones: Unpublished report, University of Tulsa, pp. 1-55.

SIGMOIDAL FOLDS

(Guidebook mileage 48.45, 1st Day)

Ridge looking southwest from North Boggy River. Right hand ridge is Big Fork chert, left hand ridge is Arkansas novaculite. Valley separating ridges is overlain by Missouri Mountain and Polk Creek shales. Valley to the left of novaculite ridge is Stanley formation.

REGIONAL RELATIONSHIPS OF OUACHITA MISSISSIPPIAN AND PENNSYLVANIAN ROCKS

Carl C. Branson[1]

Introduction

The Ouachita facies has been defined as the clastic, siliceous, metamorphosed representation of Middle Ordovician (Cambrian?) to Early Pennsylvanian rocks in the Ouachita Mountain area. The Ouachita facies here considered is the clastic, abnormally thick Mississippian and Morrowan section exposed in the Ouachita Mountain area.

Late Devonian and earliest Mississippian.- The widespread sedimentary rocks correlative with and similar to the Chattonooga shale of Tennessee have been termed Chattanoogan. Most subsurface geologists and a few other geologists refer to these rocks as Lower Mississippian. As the Devonian system is now defined, Chattanoogan rocks are entirely or almost entirely within that system. The middle division of the Arkansas group is Chattanoogan as shown by conodont studies by Hass. The "Chattanooga" shale is identified throughout the Ozark area in Arkansas, northeastern Oklahoma, and southwestern Missouri (perhaps better referred to as Noel shale), and Hass has noted in the uppermost beds, above typical Upper Devonian conodont faunas, a thin zone bearing what he calls a Lower Mississippian fauna like that of the uppermost 28.5 feet of the middle division of the Arkansas novaculite. The Noel, a black phosphatic fissile shale, is as much as 70 feet thick. A silty facies has been named the Sylamore sandstone, and this lithofacies is widespread in the base of the Noel, its subsurface equivalent is called the misener sand.

The correlative in the Arbuckle Mountain region is the Woodford shale, a black fissile shale containing spores, *Tasmanites*, the fossil wood *Callixylon*, and conodonts. The formation has been called Woodford chert or Woodford novaculite because at a few places it consists of thin-bedded chert, a lithologic type which has not been reported in the subsurface. The cherty phase appears to result from silicification by ground water. The rock is quite similar to the bedded cherts of the Arkansas novaculite of the central Ouachitas. No evidence for an Early Mississippian age of the upper part of the Woodford has been reported. Woodford shale occurs in subsurface in the southeastern portion of the Marietta-Sherman basin, and a single outcrop of similar shale with the Woodford fauna was formerly exposed in the Llano area, but is now under water.

EARLY MISSISSIPPIAN ROCKS OF THE OZARK AREA

NE. OKLA.		SW. MO.	C. MO.	NE. MO.
- - - - - - - - top of *Evactinopora* zone - - - - - - - -				
Reeds Spring		Reeds Spring	Fern Glen	absent
	St. Joe group	Pierson		
- - - - - - - - base of Osagean, Mo. Geol. Survey - - - - - - - -				
St. Joe		Northview		Chouteau
		Sedalia	Chouteau	Hannibal
		Compton		
			Bushberg	

[1] Director, Oklahoma Geological Survey.

TENTATIVE CORRELATIONS OF MISSISSIPPIAN AND EARLY PENNSYLVANIAN OF THE MIDCONTINENT

	SERIES	E. OZARKS and ILL.	SW OZARKS	OUACHITA MOUNTAINS		ARBUCKLE MOUNTAINS		ARDMORE BASIN	FT. WORTH BASIN
PENN.	ATOKAN	Tradewater	Atoka		absent		Atoka	Lake Murray	Smithwick
PENN.	MORROWAN	Caseyville	Bloyd Hale		eroded Morrowan		Wapanucka Union Valley	Golf Course	Marble Falls
MISSISSIPPIAN	CHESTERIAN	Chester	Cane Hill		Johns Valley		Springer	Springer	absent
MISSISSIPPIAN	CHESTERIAN	Chester	Pitkin Fayetteville Hindsville	Jackfork group	Wesley Prairie Mt. Wildhorse Mt.	"Arbuckle Caney"	Sand Branch	Goddard "Arbuckle Caney"	Barnett
MISSISSIPPIAN	MERAMECIAN	Ste. Genevieve St. Louis Salem Warsaw	Moorefield	Stanley group	Chickasaw Cr. Moyers Tenmile Cr.	"Arbuckle Caney"	Delaware Cr. Ahloso Sycamore	"Arbuckle Caney" Sycamore	Barnett
MISSISSIPPIAN	OSAGEAN	Keokuk Burlington	Keokuk	Arkansas group	Upper Division		absent	absent	absent
MISSISSIPPIAN	KINDERHOOKIAN	Chouteau Hannibal	Reeds Springs St. Joe	Arkansas group	Middle Division		Welden	absent	Chappel
DEV.	CHATTANOOGAN	Grassy Creek	Noel	Arkansas group	Middle Division		Woodford	Woodford	unnamed

Early Mississippian. - Rocks of Early Mississippian (Kinderhookian) age are thin and discontinuous in northeastern Oklahoma and in the Arbuckle region. In the Arbuckle Mountains the Welden limestone is but a few feet thick and occurs at few places. The Chappel limestone of the Llano area is also but a few feet thick, but appears to be more uniformly developed. Strata of this age have not been recognized in the Marietta-Sherman Basin. Rocks of Early Mississippian age in the Ozark area are thicker and more widespread.

There is general disagreement as to the position of the Kinderhookian-Osagean contact. Some geologists place the contact below the Pierson and Fern Glen, apparently upon the basis of the range of crinoid species. Another interpretation, based upon the premise that the rocks carrying the Ptychospira sexplicata--Evactinopora sexradiata--Shumardella obsolens fauna are a unit, is that the boundary is above the Reeds Spring and the Fern Glen. The conodonts of the top 28.5 feet of the middle division of the Arkansas novaculite are said by Hass to be of Chouteau-Welden-Chappel age.

Osagean series. - The upper division of the Arkansas novaculite is classified as Osagean, mainly on stratigraphic position. No rocks with Osagean fauna are known in the Arbuckle Mountain area, the Marietta-Sherman Basin, nor the Central Mineral region. Osagean rocks are well developed in the Ozarks, but are truncated southward by pre-Meramecian erosion. The uppermost unit (Keokuk formation) occurs in all but the southwesternmost Ozarks. The Burlington limestone, widespread in the Mississippi Valley, does not extend as far southwest as Oklahoma.

Meramecian series. - Hass in 1950 reported finding Meramecian conodonts in the lower part of the Stanley group. In the Ozark area Meramecian rocks occur everywhere except the northeastern county of Oklahoma. Huffman has distinguished four facies of the Moorefield formation in Oklahoma. Gordon has divided the Meramecian of Arkansas into the Moorefield formation (below) and the Ruddell shale (above). In southwestern Missouri some units of the standard Mississippi Valley section (Warsaw, Salem, St. Louis, Ste. Genevieve) are recognized.

Moorefield strata lie unconformably upon Keokuk and overlap Keokuk knobs. At places the Moorefield rests upon the Reeds Spring and in one area upon the Noel.

The Ahloso and Delaware Creek members of the "Arbuckle Caney" (so-called to distinguish it from type Caney in Johns Valley) are Meramecian. They grade laterally into the sandy Sycamore limestone. These units occur in the Arbuckle Mountain area and in the subsurface in the Marietta-Sherman Basin. The Barnett shale of the Fort Worth basin appears to be of the same age.

No lithologic feature of these Meramecian rocks relates them to the lower part of the Stanley, and they lack the volcanic ash beds.

Chesterian. - Cline (1956) has shown that the Johns Valley shale is Chesterian and that it lies upon the Jackfork group, thus placing in the Chester at least 6,000 feet of rock of the central Ouachitas. In the Ozark area of Oklahoma the Chesterian is represented by the Hindsville limestone (below), the Fayetteville shale, and the Pitkin limestone (above), the latter developed only in the central part of the southwestern Ozarks. At places in the region the Batesville sandstone is developed at the top of the Hindsville and the Wedington sandstone in the Fayetteville. The Chesterian sequence is 0 to 200 feet thick. Exposed near Fayetteville, Arkansas, is a clastic unit called the Cane Hill member of the Hale. This unit increases enormously in thickness southeastward, and it may be a late Chesterian facies which corresponds to the higher Mississippian units of the Ouachitas.

In the Arbuckle Mountain area the Sand Branch member of the "Arbuckle Caney" and the Goddard shale, which Elias equates with the Sand Branch, are Chesterian. All or most of the Springer is also probably Chesterian. These are clastic units of considerable thickness similar to the upper part of the Stanley, the Jackfork, and the Johns Valley. The "Arbuckle Caney" and Goddard occur in the subsurface in the Marietta-Sherman basin, but the Springer has not been positively identified.

The upper part of the Barnett shale in the Fort Worth basin may include Chesterian strata, and the subsurface Comyn formation of Cheney may be late Chesterian.

Morrowan series. - The earliest Pennsylvanian rocks of the region are the Morrowan series. In the Ozarks rocks of this age include the Hale formation, a calcarenite with much quartz sand, and the overlying Bloyd formation, a shale sequence with two persistent limestone beds and with a local coal bed (the Baldwin). On the northern flank of the Arbuckles and northward and eastward, extending into the subsurface, the Morrowan consists of the Union Valley calcareous sandstone (below), a shale unit, and the Wapanucka limestone. In the Ardmore basin the Morrowan Golf Course formation of the Dornick Hills group contains the Primrose sandstone in its basal part and is a shale section with the Joliff and Otterville limestones. Morrowan rocks have been reported locally in the Marietta-Sherman basin.

In the Fort Worth basin, all outcrops of Morrowan age are referred to the Marble Falls limestone group, but in subsurface some of the sediments of the Smithwick and Big Saline facies also appear to be of Morrowan age.

Orogenetic events at the close of Morrowan time are recorded in the Wichita Mountain area, the Ardmore basin, and around the Ozark dome. In the Ozarks the Atoka overlaps the Morrow, and at the water inlet for the city of Wagoner in eastern Oklahoma the basal Atoka is 18 feet of chert pebble conglomerate resting upon the Morrow. Basal Atoka is conglomeratic at places in the McAlester basin and in the Ardmore basin. In the Ouachita Mountains it has been customary to classify the sandstone with Morrowan fossils and the subjacent 200 feet of shale above the Johns Valley as Morrowan, and to map the 3,000 feet of non-fossiliferous clastics above the Morrowan fauna as Atoka. It seems more logical to suppose that these clastics are Morrowan, that there is no Atoka in the Ouachitas, and that the Ouachita orogeny was post-Morrowan and pre-Atokan.

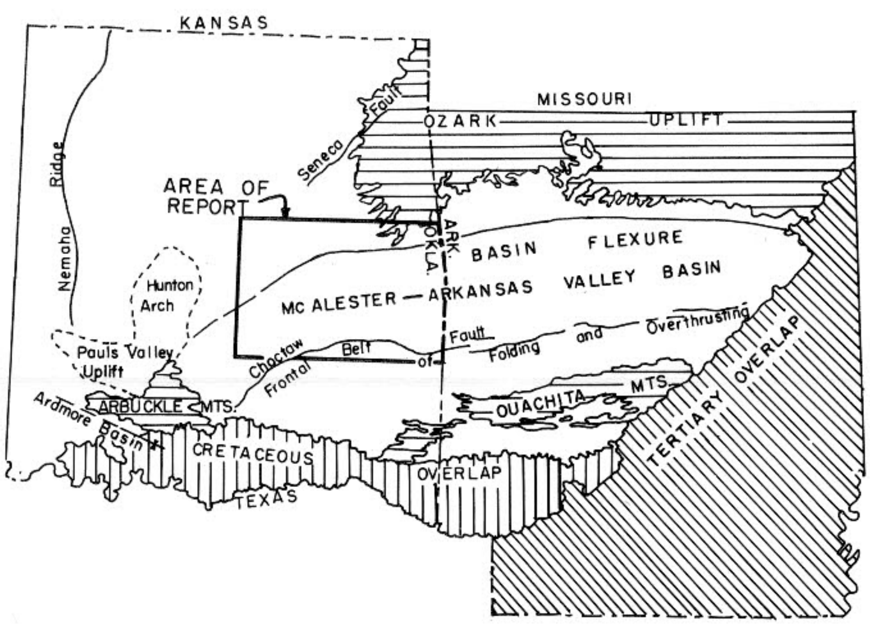
If the correlations suggested here are correct the geosynclinal troughs of Chesterian and Morrowan time occupied the Ardmore basin, the Ouachita Mountain area, the site of the Parkwood group in Alabama and adjacent areas, and the center of the Appalachian trough.

SOME AGE RELATIONSHIPS OF CHESTERIAN AND MORROWAN ROCKS IN EASTERN OKLAHOMA

Richard B. Laudon [1]

The McAlester basin (Fig. 1) of Oklahoma lies immediately north and west of the frontal Ouachitas and in outline is approximately concentric with the traces of the major faults of the Ouachita province. The McAlester basin is separated from the Ouachitas by the Choctaw belt of folding and faulting. The northern boundary of the basin is a hinge-like line of flexing which had a fairly constant position during late Mississippian and early Pennsylvanian time, a point brought out by isopachous studies of each of the formations of the Chesterian, Morrowan, and Atoka. Northwest of the flexure sediments accumulated in a slowly subsiding shelf where deposition was somewhat erratic as indicated by irregular thickness patterns of the formations. The Chesterian and Morrowan sedimentary units thin generally to the north across the platform and finally are truncated by a combination of continuously greater unconformities, decreased deposition, and recent erosion. South of the line of basin flexure, within the McAlester basin proper, the formations become progressively more clastic and thicken rapidly to the south as indicated by the parallelism, uniform spacing, and close spacing of isopach lines. The "basin" region of deposition subsided with a hinge-like motion and with much greater rapidity than the platform area. This subsidence became much more rapid during Atoka time than it had been during Morrow time.

Rocks of Chesterian and Morrow age do not crop out within the McAlester basin. From



MAP SHOWING THE LOCATION OF THE THESIS AREA AND THE MAJOR STRUCTURAL FEATURES ASSOCIATED WITH THE MCALESTER-ARKANSAS VALLEY COAL BASIN.

FIG. I

Fig. 1.--Map showing location of area in relation to major structural features.

[1] Graduate student, University of Wisconsin.

the outcrops on the flanks of the Ozark uplift these rocks thicken in the subsurface and become progressively more clastic toward the south in the direction of the Ouachita belt. The Chesterian and Morrowan rocks are buried beneath a southward-thickening wedge of "Atokan" and Desmoinesian rocks. The near-surface rocks of Desmoinesian age immediately north of the Choctaw fault dip to the north, back into the McAlester basin. However, in the wells immediately north of the fault the deeply buried Chesterian and Morrowan rocks dip and thicken to the south. Thus the Chesterian and Morrowan rocks of the "basin" are in the form of a wedge, which was deposited on the southward dipping limb of a syncline. The deeper part of the McAlester basin displays no corresponding northward dipping limb, as would probably be displayed in a closed sedimentary and structural basin.

In the frontal Ouachitas, south of the Choctaw fault, rocks of Chesterian and Morrowan age are brought to the surface by faulting. These rocks are overwhelmingly clastic and sparsely fossiliferous in contrast to the equivalent rocks of the Ozark and/or Arbuckle facies which contain abundantly fossiliferous marine limestones. The rocks in the subsurface of the McAlester basin display a partial transition in the facies change between the rocks of the Ouachita outcrops and those of the Ozark uplift outcrops.

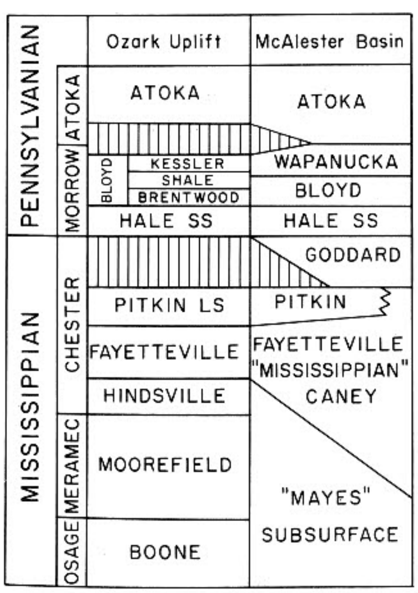


Fig. 2.--Correlation chart.

The thicknesses of the "Caney" (Mississippian Caney shale plus Goddard shale) and the Atoka formations (Fig. 2) are only slightly greater in the outcrops of the frontal Ouachitas than in bore holes immediately north of the Choctaw fault. The continuity of formations, the thickness patterns of the formations, and the lack of dip reversal of the subsurface rocks in the southern part of the McAlester basin suggest that the late Mississippian and early Pennsylvanian sediments of the McAlester topographic basin were deposited in the same geosyncline or basin of sedimentation with those of the frontal Ouachitas. The same evidence suggests that the rocks of the Ouachita region have not been transported a great horizontal distance as a thrust sheet.

In the outcrops around the Ozark uplift and in the platform area of sedimentation to the north and west of the McAlester basin an unconformity separates the Morrowan Hale sandstone from the Mississippian Pitkin limestone. This unconformity dies out into the McAlester basin, and the writer believes that south of the line of basin flexure sedimentation was uninterrupted from Mississippian into Pennsylvanian time (see Figs. 4-6). Even in the area of outcrops it seems that there is only a slight time loss represented by this unconformity, since the formations involved are only slightly thicker at the north edge of the McAlester basin, where there is no unconformity, than in the Ozark uplift outcrops where an unconformity is evident. Similarly the pre-Atoka unconformity of the Ozark uplift and the platform area dies out southward in the McAlester basin.

Goddard-Pitkin relations. - The relationships of the Goddard and Pitkin formations in the McAlester basin is of particular interest. The Pitkin limestone is the youngest Chesterian formation of the Ozark uplift and is overlain unconformably in the Ozark region by the Hale (Cromwell) sandstone, which is basal Morrowan. In the frontal Ouachitas the Goddard shale ("Pennsylvanian Caney") is overlain by Morrowan equivalent Wapanucka limestone. Within the McAlester basin the Pitkin limestone becomes argillaceous southward and grades into the upper part of the "Mississippian Caney". The zone equivalent to Pitkin at the top of the "Mississippian Caney" can be traced from well to well becoming progressively less calcareous until it cannot be distinguished from the rest of the shale formation. The Goddard shale thins rapidly northward and pinches out without reaching the outcrop belt around the Ozark uplift (see Figs. 4-6).

This thinning of the Goddard was caused by slower sedimentation to the north at the time the formation was deposited. The area around the Ozark dome was quite stable and entrapped almost no sediment while the McAlester basin was subsiding with a hinge-like movement and entrapping progressively more sediment to the south. In the outcrop area the Pitkin and Hale are in unconformable contact, and it might be inferred that the Goddard was truncated by this erosional unconformity. Such is not the case to any appreciable degree. Apparently the unconformity dies out without reaching that part of the basin in which the Goddard occurs. In the cross section reproduced as Fig. 4, it can be observed that a few thin beds near the top of the Goddard have slightly higher resistivities than the rest of the formation. These more resistant horizons are found near the top of the formation over the entire area of investigation. This would not be the case if the formation had been uniformly deposited and subsequently tilted and truncated. The predominant cause of convergence of the Goddard must have been a differential rate of sedimentation over the basin at the time of deposition. The Pitkin equivalent zone and Goddard formation both are present and recognizable in only a relatively narrow zone in the subsurface, bounded on the north by the zero thickness edge of the Goddard and on the south by the merging of the Pitkin with the top of the Caney (Mississippian). In this zone the Goddard shale overlies the Pitkin formation. Thus the Goddard is underlain by rocks known to be of Mississippian age and overlain by the Hale formation, which is lower type Morrowan. Elias (1956) assigned the Goddard to the Chester series on the basis of its cephalopod fauna. The present writer believes that the Goddard formation and its equivalents were deposited during the time represented by unconformities in the systemic type areas, and thus the Goddard is not strictly a time correlative of any part of either the Mississippian or Pennsylvanian system as currently defined. The writer would, however, concur in placing these rocks in the Mississippian system although such a placement is necessarily arbitrary. Thus in

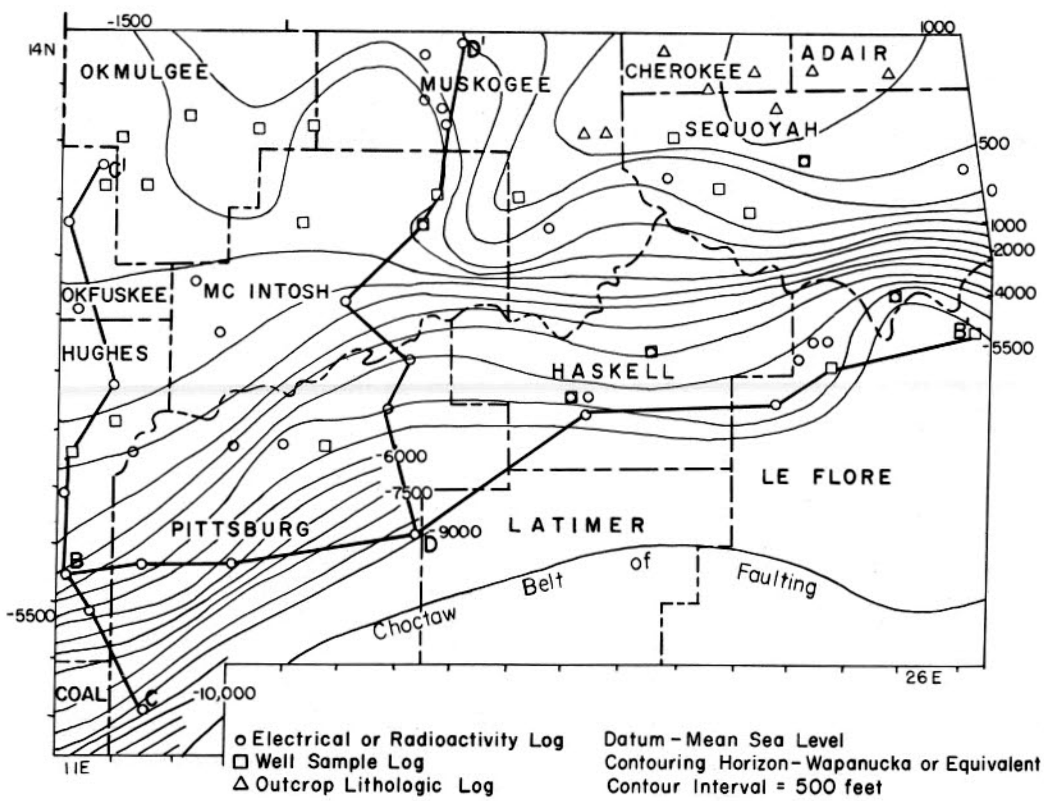


Fig. 3. -- Structure contour map on Wapanucka or its equivalent.

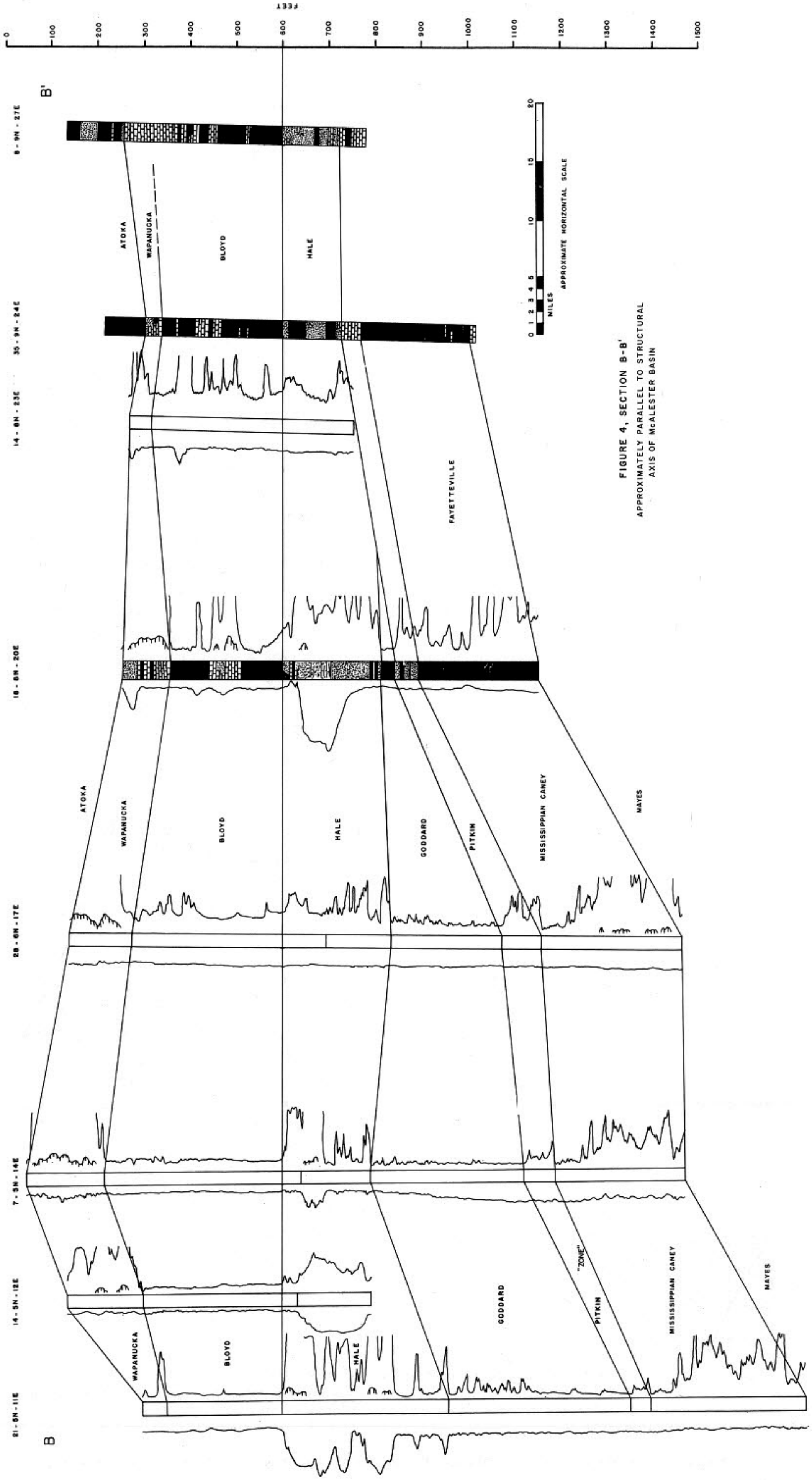


FIGURE 4, SECTION B-B'
APPROXIMATELY PARALLEL TO STRUCTURAL
AXIS OF McALESTER BASIN

a sense these rocks could become a supplement to the type system. This is accepted procedure in formal stratigraphic nomenclature (Ashley, and others, 1933, p. 446) and is exactly the manner in which the Morrow series, after several years of debate, was added to the type Mid-Continent Pennsylvanian system. The writer believes that when such a supplement to the systemic type is proposed, it should be designated in Goddard-equivalent rocks in a more fossiliferous area. The type Atoka[2] series is in sparsely fossiliferous, clastic facies rocks of the Ouachita region and interregional correlation of this series has been confusing and probably inaccurate.

Johns Valley shale. - The age of the Johns Valley shale of the Ouachita Mountains is of critical importance in dating the tremendously thick Stanley and Jackfork groups. The intricacies of this formation with its faunas and erratic boulders are better left to those who have dealt immediately with this problem. It may be worth noting, however, that Ulrich (1927, p. 21) was quite reserved in his dating of the Johns Valley "Morrow fauna":

> "It should be borne in mind, however, that these early Pottsville faunas are more or less modified recurrences of the Spergen fauna. Accordingly it is to be expected that the fauna in these youngest of the Caney boulders also resembles stages of evolution of the Spergen fauna found in the intermediate Chester formations in the Mississippi Valley. But the resemblance in the latter case is clearly inferior in closeness to that observed in comparing the Ouachita Caney (Johns Valley) boulder fossils with Morrow fossils."

If, as the present study indicates, there was uninterrupted sedimentation in the McAlester basin and frontal Ouachitas from Mississippian into Pennsylvanian time, the Johns Valley fauna would probably be closer in time to Morrow than to Spergen as indicated by Ulrich, but still somewhat older than Morrow. As it seems the systemic boundary in the McAlester basin will be placed at the Hale-Goddard contact any fauna even slightly older than Morrow will be placed in the Mississippian.

The convergence patterns of formations in the McAlester basin offer some slight evidence as to the age of the Johns Valley formation. In the subsurface of the McAlester basin the Morrow group ("Wapanucka", Bloyd, and Hale formations) thicken southward to more than 850 feet. In the frontal Ouachitas the Wapanucka formation, contained between the Goddard and the Atoka, is about 300 feet thick and would seem to be equivalent to the entire Morrow group. Since all the other sedimentary units studied thicken toward the Ouachitas, it does not seem likely that the true sedimentary thickness of the Morrow group in the frontal Ouachitas is represented in the Wapanucka limestone. Neither does it seem logical to invoke a pre-Atoka unconformity in the frontal Ouachitas to account for the apparent thinning of the Morrow group, as no evidence of such an unconformity has been reported in the frontal Ouachitas. This problem might be answered if we were to assume that the base of the Atoka formation becomes older to the south and that the lower part of the Atoka formation is partially equivalent to the upper part of the Morrow group of the Ozark uplift. Mather in 1917 recognized a Morrow fauna in the lower part of the Atoka. Harlton (1938) suggested the separation of the Barnett Hill formation containing an "unquestionable Morrow fauna" from the Atoka.

Regardless of the status of the Barnett Hill formation, the existence of Morrow equivalent rocks in the formation above the Wapanucka provides a reasonable explanation for the unexpectedly thin Morrow group in the frontal Ouachitas. Thus it seems that in their type areas the Morrow and Atoka series overlap considerably in time. If the base of the Atoka continues to become older to the south into the central Ouachitas as Cline believes (Cline, 1956), indeed unless this trend is reversed, then the Johns Valley formation underlying the Atoka formation in this area is probably partially equivalent to the Goddard.

[2] Editor's note: A type locality has never been designated for the Atoka formation and, unfortunately, the complete thickness of the formation (rock sense) is not exposed near Atoka.

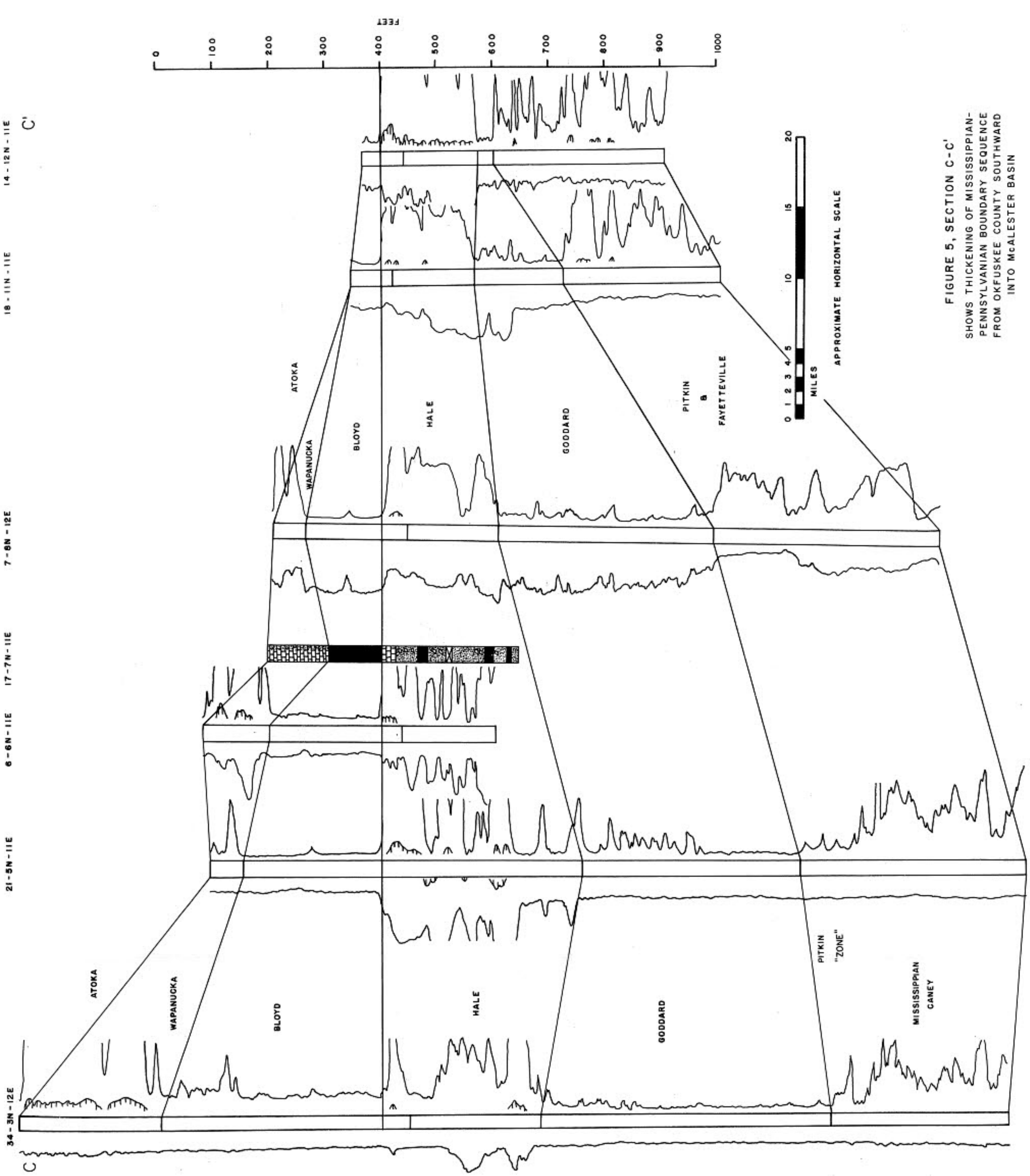


FIGURE 5, SECTION C-C'
SHOWS THICKENING OF MISSISSIPPIAN-PENNSYLVANIAN BOUNDARY SEQUENCE FROM OKFUSKEE COUNTY SOUTHWARD INTO McALESTER BASIN

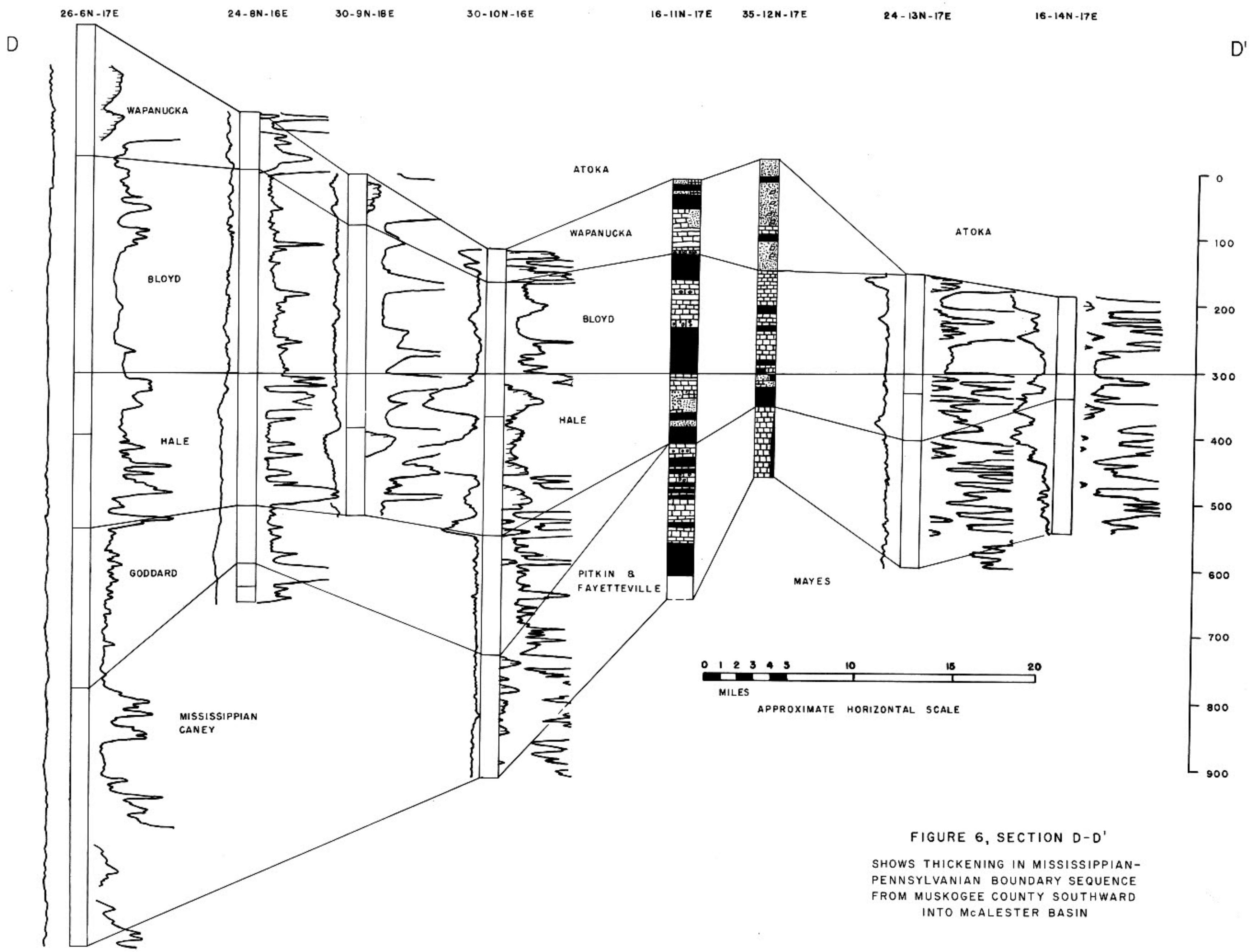


FIGURE 6, SECTION D-D'

SHOWS THICKENING IN MISSISSIPPIAN-
PENNSYLVANIAN BOUNDARY SEQUENCE
FROM MUSKOGEE COUNTY SOUTHWARD
INTO McALESTER BASIN

Acknowledgements. - Dr. L. M. Cline suggested the problem and gave helpful assistance and consultation in all phases of the investigation. The late Mr. Ralph A. Brant of the Atlantic Refining Company, and the staffs of the Research Department of the Carter Oil Company and the Geology Department of the University of Tulsa were consulted and gave assistance in the early stages of the investigation. Many fellow graduate students at the University of Wisconsin gave helpful advice and criticism, and the National Science Foundation gave financial assistance for this investigation.

References

Ashley, G. H., and others, 1933, Classification and nomenclature of rock units: Geol. Soc. Amer. Bull., vol. 44, pp. 423-459.

Cline, L. M., 1956, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains, Oklahoma: Tulsa Geol. Soc. Digest, pp. 100-106.

Elias, M. K., 1956, Upper Mississippian and Lower Pennsylvanian formations of south-central Oklahoma (in Petroleum Geology of Southern Oklahoma): Amer. Assoc. Petroleum Geologists, pp. 56-134.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 22, pp. 852-914.

Laudon, R. B., 1958, A study of Chesterian and Morrowan rocks in the McAlester basin of Oklahoma: Oklahoma Geol. Survey, Circular 46.

Mather, K. F., 1917, Pottsville formations and faunas of Arkansas and Oklahoma: Amer. Jour. Sci., vol. 193, pp. 134-135.

Moore, C. A., 1947, The Morrow series of northeastern Oklahoma: Oklahoma Geol. Survey, Bull. 66.

Ulrich, E. O., 1927, Fossiliferous boulders in the Ouachita "Caney" shale and the age of the shale containing them: Oklahoma Geol. Survey, Bull. 45.

AGE CLASSIFICATION OF THE UPPER PUSHMATAHA SERIES IN THE OUACHITA MOUNTAINS [1]

Bruce H. Harlton [2]

Abstract

In the Ouachita Mountains of Oklahoma and Arkansas, there is a clastic succession of over 13,000 feet of sediments, of which the age classification has been a matter of debate. The rocks have been generally regarded as Mississippian and early Pennsylvanian age. This paper is particularly concerned with the age assignment of the Johns Valley shale. The shales within the Johns Valley with typical "Caney and Goddard" type lithology are considered to be a normal depositional sequence over a wide area. A discussion of the various "bouldery" shales is also presented.

Introduction

The Pushmataha strata of Carboniferous age, are of widespread distribution in the Ouachita Mountains and in the Kiamichi Range. The total thickness exceeds 13,000 feet. The principal concern of this paper is the age assignment of the upper Pushmataha series from the Wesley through the Round Prairie formation, with special emphasis of the Johns Valley shale. There has been a wide divergence of opinion among geologists in regard to the age of this series. In this paper the occurrence of "Caney and Goddard" type rocks in the Johns Valley shale are considered a normal depositional sequence. After a review of previous classifications, the localities are described where critical new evidence has been obtained. The stratigraphic sequence based on this evidence is presented and the paper concludes with a discussion of major diastrophic features which are revealed from the stratigraphic evidence.

Previous nomenclature and earlier interpretations. - The rich fauna from the Caney was described by Girty in 1909. The largest portion of the fauna recorded by him came from the "Ouachita Caney" in Johns Valley. Many of these are transitional upper Mississippian and lower Pennsylvanian forms. The earliest extensive work in the southern Ouachita Mountains was done by Honess (1923) who mapped and described the stratigraphy of the Stanley and Jackfork succession in that area.

The term Johns Valley was introduced by Ulrich (1927) for the "bouldery" shale in Johns Valley in the center of the Tuskahoma syncline, Ts. 1 N. to 1 S., R. 16 E., Pushmataha County, Oklahoma. This is also the type locality of the Caney shale. The name Caney was given by Taff [3] to all shales and was taken from a small settlement 6 miles north of Eubanks in the canoe valley of the Tuskahoma syncline. Fifty years ago this settlement was known as Caney. Ulrich (1927, p. 27) regarded the "Ouachita Caney" as having been derived and transported from the outcrops of the Arbuckle Caney shale by floating masses of shore ice.

Miser (1929) presented the first geologic map of the entire Ouachita Mountain area. In 1933 the writer described the microfauna

[1] Publication authorized by Amerada Petroleum Corporation.

[2] Research geologist, Amerada Petroleum Corporation.

[3] Taff, J. A., as quoted by Gould, Charles N., Index to the Stratigraphy of Oklahoma: Oklahoma Geol. Survey, Bull. 29, 1925.

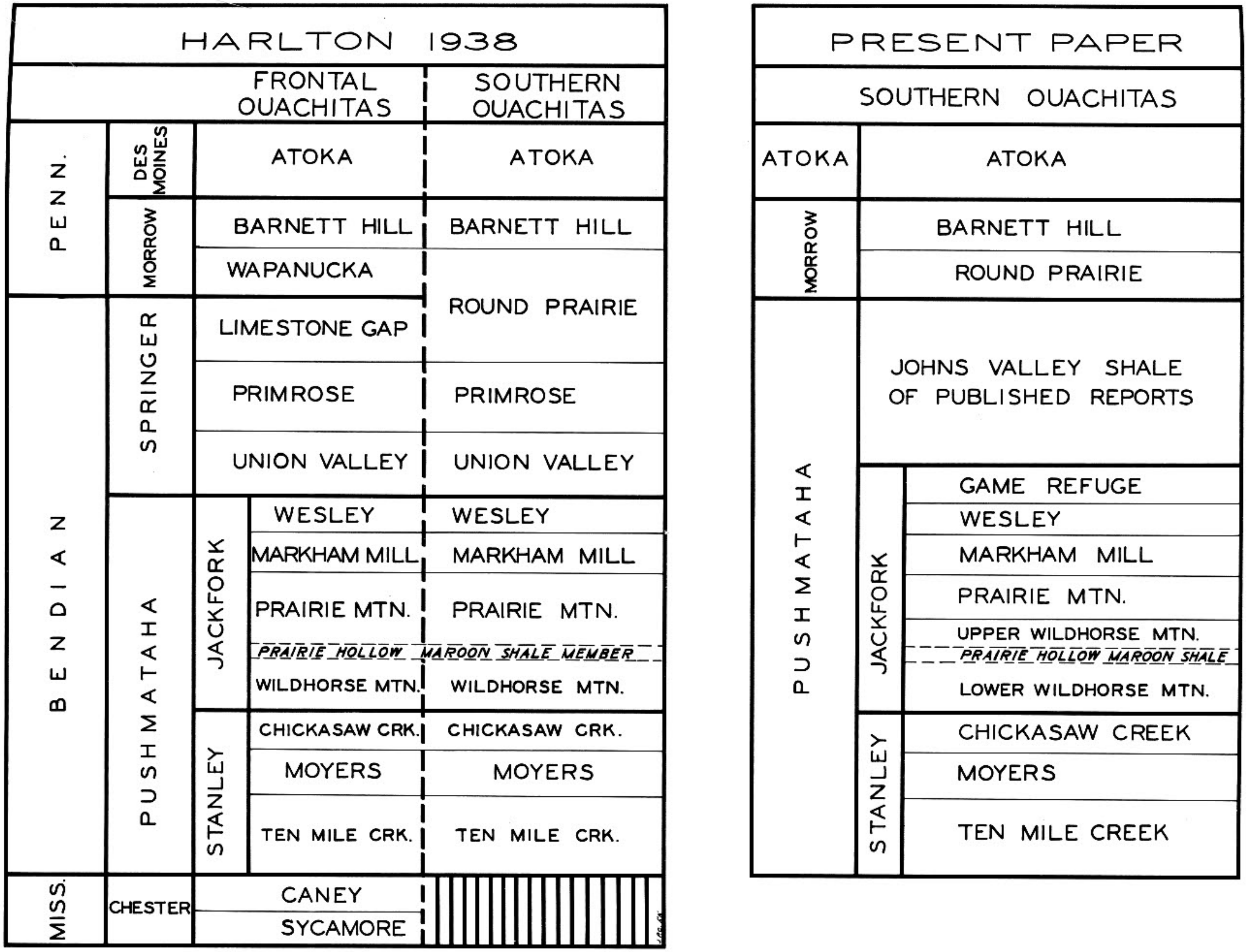


Fig. I - Nomenclenture used in Harlton's 1938 paper compared with that of present paper from the Johns Valley shale and regarded them as transitional forms from the Mississippian and Pennsylvanian. The writer suggested that the term Bendian be intercalated between Mississippian and Pennsylvanian since none of the fossils are persistent nor diagnostic of either period. In the same paper (pp. 5-6) the writer expressed the opinion that the boulders in the Johns Valley shale are remnants of overthrust sheets of limestone masses and that the Caney shale provided a gliding plane for the overriding thrust sheets, and in this manner brought the Caney into Johns Valley. Miser, (1934) discussed in detail the age of the Stanley through the Johns Valley succession. He also believed that the "Ouachita Caney" represents erratic masses, possibly transported by submarine land slips.

In 1938, the writer provided the foundation on which subsequent detailed geologic mapping was conducted. The lithic subdivisions of the Stanley and Jackfork group as shown in chart, (Fig. 1) were mapped and applied by Hendricks in 1947 although some of Harlton's names were not used. In 1956, Cline first recognized and mapped the various formational units in the entire Kiamichi Range, southeast of the Kiamichi River and eastward to the Arkansas state line.

Revised nomenclature. - The usage of the name "Union Valley" sandstone of the Ouachita Mountains which the writer (1938, pp. 889-893) identified entirely on faunal grounds, now is known to be insufficiently diagnositic and should be changed. The best exposures of this formation are in the Round Prairie syncline in sec. 2, T. 2 S., R. 12 E., along Campbell Creek. Other good exposures are in the Game Refuge in the Lynn Mountain syncline, secs. 28 and 29, T. 1 S., R. 18 E., about six to seven miles southwest of Clayton. Since the name Campbell Creek is already preoccupied in the literature, it seems advisable to use Game Refuge as a name for this sandstone unit. The Game Refuge formation now marks the terminal beds of the Jackfork group.

After the selection of the published type locality of the Wildhorse Mountain formation (Harlton, 1938, p. 879) subsequent mapping proved that the Prairie Hollow maroon shale member actually occurred in the covered interval along Wildhorse Creek as indicated on the reproduced map (Fig. 2). Therefore, Cline's (1956a, p. 14) expansion for the inclusion of the Prairie Hollow maroon shale member and the overlying massive sandstones, which are the most prominent ridge and terrace formers everywhere in the Wildhorse Mountain formation, is in accord with the writer. This limits the vertical range of the Prairie Mountain formation.

Cline (1956c) found excellent exposures of a regular undisturbed Jackfork through Atoka stratigraphic sequence in the Lynn Mountain syncline in the western Kiamichi Range, T. 1 S., Rs. 17-18 E. There, the upper Jackfork lies in normal contact with a succession of distinct biostratigraphic types of shales; namely, the basal "bouldery" shale, succeeded upward by "Caney, Goddard, and the Springer" type rocks. The lithologic units of the above succession are persistent in character and remarkably distinct, not only in the Lynn Mountain syncline, but in other areas as well; namely, at Cooper Hollow in the NE 1/4, sec. 32, T. 4 N., R. 20 E., and in Johns Valley, in T. 1 S., R. 16 E. These were re-examined in the light of Cline's findings. This is an emendation of previous views that the "Caney and Goddard" type rocks are exotics.

Distribution of "Caney-Goddard" type rocks in Johns Valley. - The distribution of the "Caney-Goddard" type rocks in Johns Valley is concentrated in an oval-shaped sharply compressed syncline in secs. 4, 5, 9, and 10, T. 1 S., R. 16 E. (See map, Fig. 3). The dips in the inner portion of this oval-shaped area range from about 20° to vertical and in a few instances are overturned. The only exposure having a moderately low dip was observed about 600 feet east of the Baskett boulder locality in sec. 4, where the "Ouachita Caney" dips into the creek bank about 20° to the southeast. In the surrounding area and outward toward the rimrock, the dips become successively shallower, where the average dip ranges from 3-12 degrees. Much of Johns Valley is covered by heavy soil and Quaternary alluvium and the oval outline shown on the map may not represent the complete outline of the more steeply dipping formations. Its ultimate outline, therefore, which cannot yet be clearly portrayed might actually represent the more linear pattern of the syncline.

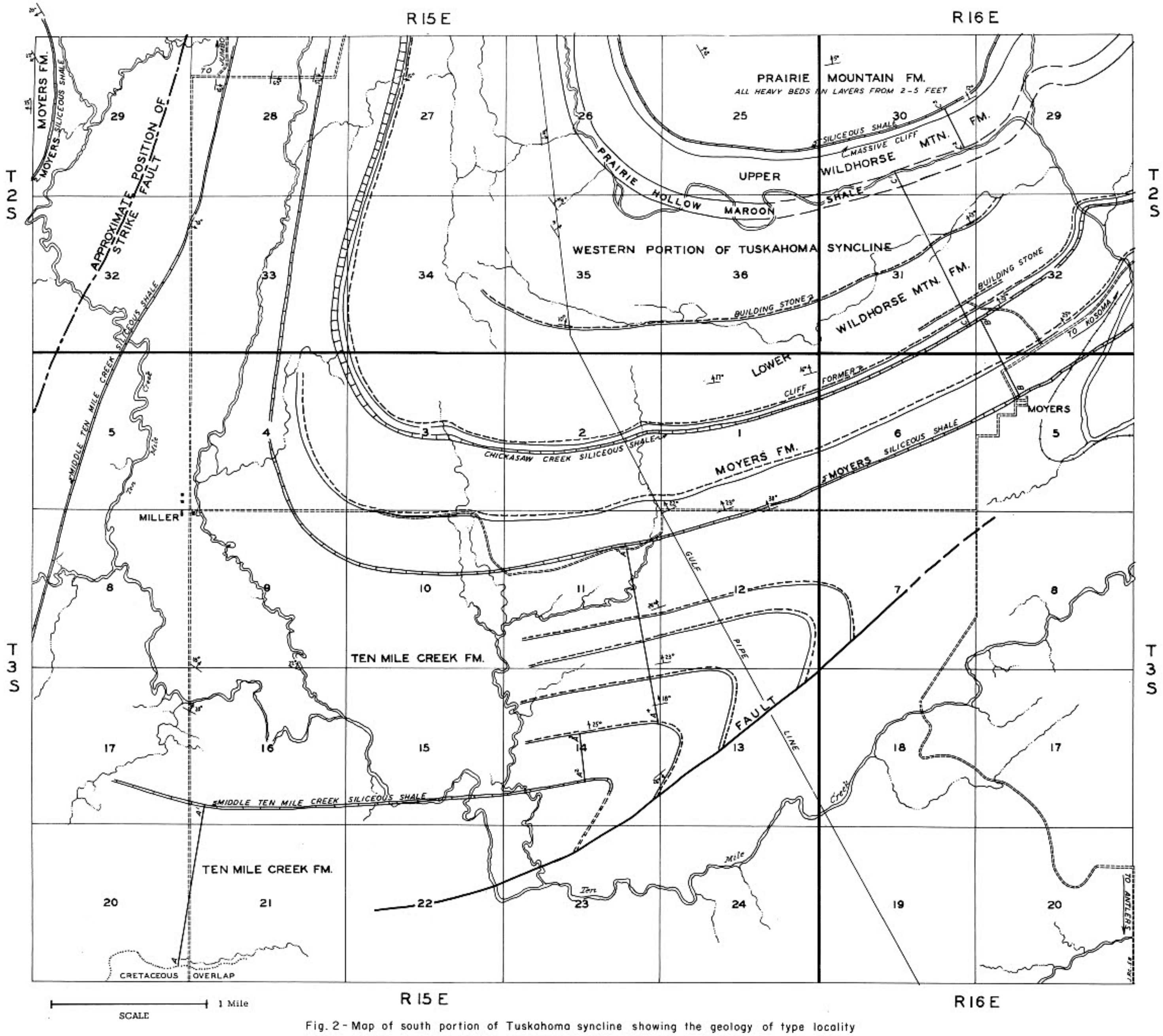


Fig. 2 - Map of south portion of Tuskahoma syncline showing the geology of type locality of Wildhorse Mountain formation

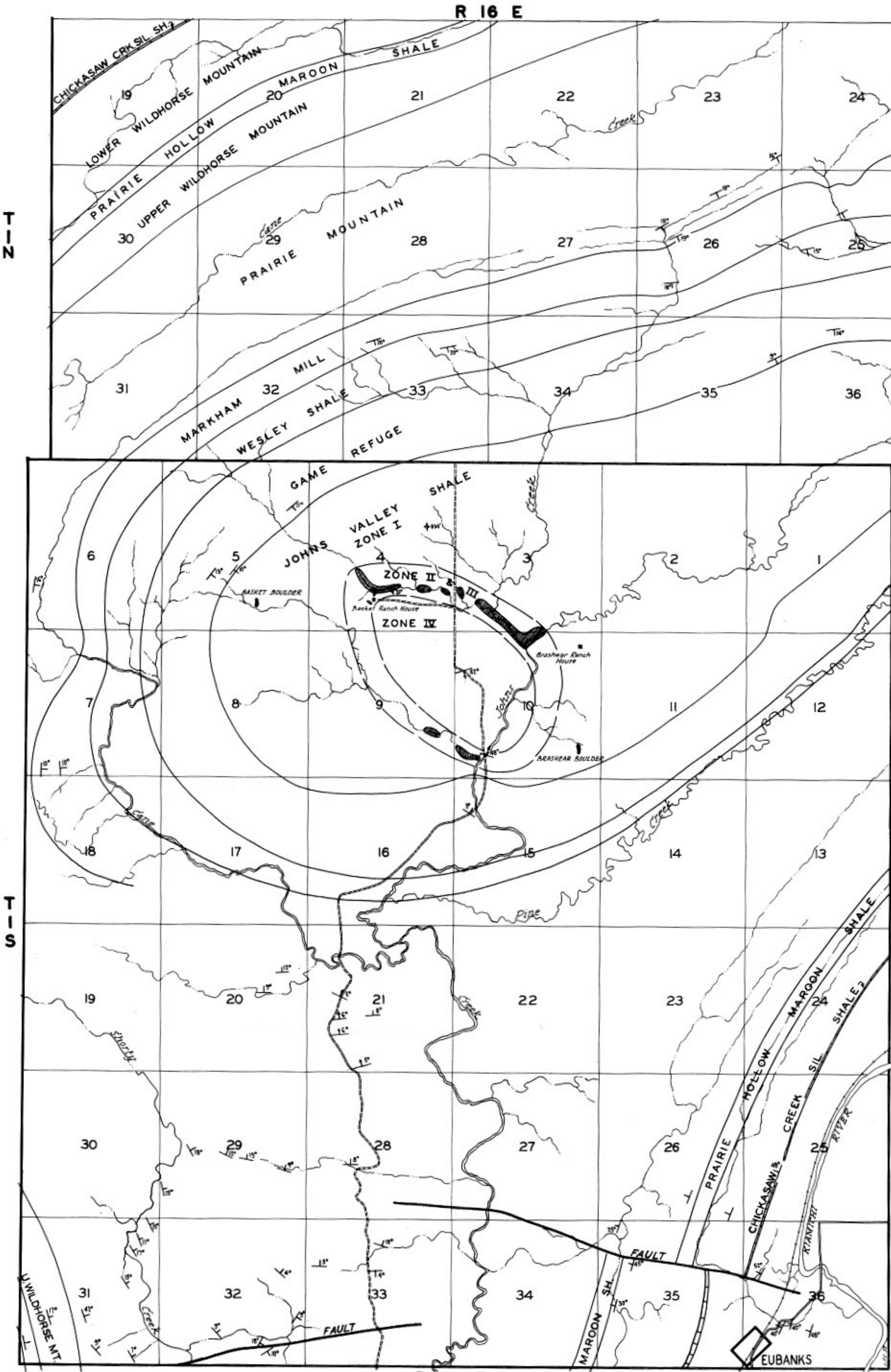


Fig. 3 Geologic map of Johns Valley and vicinity

The sharp vertical depression at the center of the Tuskahoma syncline is one of the most distinctive structural features of the area. This phenomenon probably was brought about by crumpling of the beds at the heart of the syncline. By contrast, the peripheral rim does not reflect the disturbance.

STRATIGRAPHY

Pushmataha Series

The Pushmataha strata comprise a clastic succession of rocks at least 13,000 feet thick. The deposits mark the beginning of a distinctly new regime in physical environment and faunal development. More information is needed to determine the biological features and their faunal relations.

The age equivalence of this great thickness of clastics of the Pushmataha series with other definite Mississippian rocks and with possible early Morrow rocks has not as yet been fully defined. For instance, what is the relationship to the massive Mississippian carbonate facies of the Ozarks? From available subsurface data in northern Arkansas (Maher and Lantz, 1953)there is evidence that the Ouachita and Ozark facies may interdigitate. The facies change of Ouachita sediments can be seen in the immediate Ouachita Mountain front with a gradual yet conspicuous decrease of sandstone interbeds northward, finally becoming predominatly shale. Identical lithologic changes occur in the Ozark carbonate facies going south-southeast basinward (Laudon, 1958). In some cases, the carbonate facies disappear rather sharply, seemingly change to clastics, while in some instances, southward projecting tongues of carbonate facies prevail. In still other examples, the lithologic change can be seen as a gradual one from an increase of argillaceous limestone, to argillaceous calcareous marl and to calcareous shale. Eventually, further subsurface data will determine more accurately the chronological position of the Pushmataha series.

The invertebrate faunas of the Pushmataha series are not large but are far from being completely known. They have been gathered from various levels from only the upper column, namely the Wesley through the Johns Valley formation. Many of the sandstone layers in this interval are characterized by fragmentary remains of brachiopods and gastropods, while others contain an abundance of crinoid columnals and plant remains. The molluscan fauna, namely ammonoids and pelecypods, are largely confined to the Wesley, "Caney and Goddard" type shales and constitute its most distinguishing faunal feature.

Upper Jackfork Group

Only the upper Jackfork, from the Wesley through the Round Prairie formation, will be discussed in the following pages.

Wesley shale. - The Wesley shale is one of the best developed, most widespread, and most easily mappable units throughout the Ouachita Mountains. In Wesley Valley in T. 1 N., R. 13 E., it consists of 500 feet of predominantly siliceous shale. However, the average thickness in the mountains is closer to 350 feet. South and east of Wesley Valley, intercalations of soft, blue-black shales become distinct. In most areas, thin lenses of characteristic chert conglomerate are encountered. Another diagnositic feature of the Wesley shale is the presence of large rounded to sub-rounded chalcedonic masses or concretions which are irregularly intermixed in the shale matrix. Well preserved ammonoids have been found in several of these concretionary masses. Thin lenses of fine-grained sandstone, usually fossiliferous, up to 3 feet in thickness, are encountered at occasional intervals in the lower part of the section. Immediately south of Wesley Valley, the shale body contains rounded to sub-rounded boulders, chiefly of Viola age, ranging in size up to seven feet in length. Southeastward toward the Windingstair fault area, the boulders gradually decrease in size and eventually disappear. The immediate Wesley Valley locality must have been the source area from which these boulders were supplied.

Game Refuge formation. - The Game Refuge formation, previously designated under the name "Union Valley" sandstone, in the future will be known under this new name. It receives its name from the state Game Refuge at Jerusalem Hollow in the western Kiamichi Range in secs. 28 and 29, T. 1 S., R. 18 E., about 6 to 7 miles southwest of Clayton.

The Game Refuge formation consists of approximately 250 to 350 feet of gray to dark gray shale and intercalated massive and thin-bedded fine to medium subangular sandstone. Intercalations of fine to medium-coarse fossiliferous limonitic sandstones with an abundance of crinoid columnals and plant remains make the upper half of the formation remarkably distinct. Some of these fossiliferous horizons are found to consist of brown, highly hematitic, arenaceous, siliceous limestone.

Johns Valley Shale

The stratigraphic relationship of the various shales in Johns Valley has been the subject of considerable controversy. The early history of the controversy has been described by Ulrich, (1927, pp. 21-23) and need not be repeated. Due to the questionable relationship in Johns Valley, the writer in 1938 (pp. 895-900) proposed to substitute the name Round Prairie for the beds of Morrow age for the Johns Valley.

Ulrich originally defined and restricted the name Johns Valley to the "bouldery" shale in Johns Valley. Since only the basal "bouldery" shale is exposed in this valley, its term should eventually be applied only to this formation.

The strata of the Johns Valley can readily be divided into a four-fold zonal arrangement and in ascending order are represented everywhere by an initial bouldery shale, succeeded upward by "Caney, Goddard and Springer" type rocks. The upper bouldery shale of undisputed Morrow age is here referred to the Round Prairie formation. The various zones have been recognized by all investigators in the past two decades and there has been no difficulty in drawing the lithologic boundaries. Furthermore, their distribution is such as to indicate the eventual establishment of independent formations. The zones described below are used only in an informal sense in this paper.

Zone 1. Basal bouldery shale. - In practically every place the basal "bouldery" shale represents the initial clastic cycle of the Johns Valley formation. It is predominantly a dark gray, soft shale. Intercalations of black shale and thin-bedded sandstone layers occur at intervals. Boulders of various dimensions are irregularly intermixed in the shale matrix and do not follow any definite bedding planes. In Johns Valley, all of the boulders of great size dip toward the center of the syncline. The shales weather to a dull gray color of fine granular "mealy" texture.

Zone 2. Caney type rocks. - Zone 2 of the present study comprises the "Caney" type rocks which resemble in all respects the Caney of the Arbuckle Mountains. Cline's (1956) discovery of "Caney" type rocks in undisturbed stratigraphic relationship in the western Kiamichi Range calls for reconsideration of the age of the Johns Valley. His discovery of a conglomerate bed in the "Ouachita Caney" composed of rounded to subrounded cobbles and set in a calcareous bond, has historical importance. The occurrence of the conglomerate in this zone is of undisputed depositional origin and must have come from the same source as the erratic boulders in Zone 1.

The "Caney" type exposures in the Ouachitas are distinct and consist of black, hard fossil, brittle, calcareous shale containing septarian limestone nodules and concretions and abundant phosphatic concretions. The weathered exposures are silvery-gray in color and siliceous and resemble the siliceous Wesley shale.

Zone 3. "Goddard" type lithology. - The contact between zones 2 and 3 is easily recognized by a conspicuous difference in lithology. Zone 3 of "Goddard" type lithology consists of dark gray to black, very soft, thinly laminated to papery shale containing throughout thin lenses of ironstone and limonitic concretions. Many of the concretions contain fossils.

Zone 4. "Springer" type lithology. - Zone 4 which is characterized by "Springer" type lithology marks the youngest unit of the Pushmataha series. It consists of about 350 to 400 feet of predominatly dark gray-greenish to dark gray, soft, flaky to papery shale with occasional interbeds of fine to very fine-grained sandstones up to 6 feet in thickness. Some of the limonitic sandstone layers were originally calcareous as manifested by the poorly preserved faunal casts.

Age of the Johns Valley shale. - In seeking to determine the age of the Johns Valley, one must first consider the evidence of the paleontological data obtained from the "Caney and Goddard" type shales. The richly fossiliferous "Caney" type deposits in the Ouachitas contain species of Caneyella, Orthoceras, Gastrioceras and Goniatites which in the past have been considered of unquestionable Mississippian age. In point of fact, it is known from available records that the range of the above genera is quite variable. Most of them have long ranges, others may be limited to an epoch of time, while still others seem to be restricted only to a certain stage. Such ranges of time have focussed attention on the need for more accurately identifiable forms from a deposit of undisputed age before its exact taxonomic position can be determined. The "Ouachita Caney" may perhaps represent only a minor projecting tongue of the Caney of the Arbuckle Mountains, and again it might represent a recurring simulating black shale deposit of later age.

Morrow Series

Round Prairie formation. - The Round Prairie formation consists of approximately 425 feet of soft gray "bouldery" shale with usually two interbeds of dark gray to black shale and thin sandstone layers, as shown in Fig. 4. The basal boulder bed which ranges in thickness up to 275 feet is succeeded upward by a persistent interval, about 50 feet in thickness, of dark gray to black shale and intercalated thin bedded sandstone, oftentimes very fossiliferous. This zone is succeeded upward by two "bouldery" shale beds, separated near the middle by a zone of dark gray to black, flaky shale with occasional sandstone lentils. The entire thickness of this three-unit zone is slightly over 100 feet in thickness.

The Round Prairie formation thus described is a more restricted term than as originally defined by the writer in 1938.

Diastrophic Features

The succession of major events in the Pushmataha period began with the development of the Ouachita trough. Sedimentation was initiated by the Hot Springs conglomerate and was followed by volcanism as manifested by the Hatton tuff lentils. Tuffaceous beds were deposited as far west as the Jumbo Valley. The deep exploratory test by Signal Oil Company in the cen. SE 1/4 SW 1/4 sec. 9, T. 2 S., R. 15 E., encountered an alternating succession of shale and tuff in the basal Stanley from a depth of 6070 to 6460 feet.

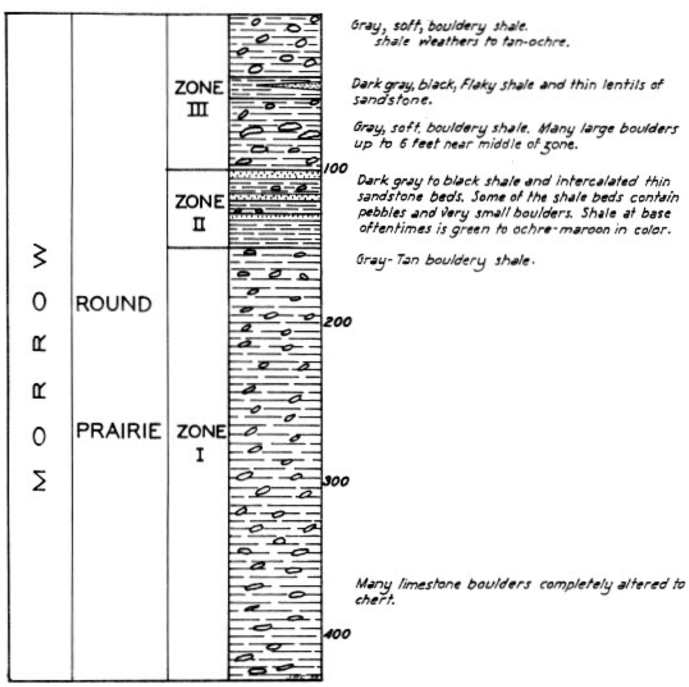


Fig. 4. -- Diagrammatic representation of zones of the Round Prairie formation.

The Stanley epoch which is recorded by the more than 5650 feet of shales and thin and massive bedded sandstones, was brought to a close by uplift within the Stanley area. Its disturbance is recorded in the Chickasaw Creek siliceous shale which contains scattered Viola boulders, now completely replaced by chert.

The Jackfork group consists chiefly of clastic deposits. Its massive sandstone distinguishes it from the preceding Stanley group. The occurrence of boulders in the Wesley shale indicates renewed uplift in the area.

Times of greatest crustal instability occurred from the Johns Valley through Round Prairie time. The deformation became intense during Round Prairie-Morrow time. The tectonic disturbances were profound and widespread in the Mid-Continent and delineated the foundation of horst and graben type structure all over southern Oklahoma.

References

Cline, L. M., 1956a, Regional stratigraphy and history of Ouachita Mountain area (abs.): Amer. Assoc. Petroleum Geologists Bull., vol. 40, no. 12, pp. 427, 428.

Cline, L. M., 1956b, Mississippian-Pennsylvanian stratigraphy of central Ouachita Mountains, Oklahoma: Ardmore Geol. Soc., Guidebook to Ouachita Mountains field conference, southeastern Oklahoma, pp. 47-63.

Cline, L. M., 1956c, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains, Oklahoma: Tulsa Geol. Soc. Digest, vol. 24, pp. 100-106.

Cline, L. M., and Moretti, Frank, 1956, Two measured sections of Jackfork group in southeastern Oklahoma: Oklahoma Geol. Survey, Circular 41, pp. 1-20.

Girty, G. H., 1909, The fauna of the Caney shale of Oklahoma: United States Geol. Survey Bull. 377.

Harlton, Bruce H., 1933, Micropaleontology of the Pennsylvanian Johns Valley shale of the Ouachita Mountains, Oklahoma, and its relationship to the Mississippian Caney shale: Jour. Paleontology, vol. 7, no. 1, pp. 1-29.

Harlton, Bruce H., 1934, Carboniferous stratigraphy of the Ouachitas with special study of the Bendian: Amer. Assoc. Petroleum Geologists Bull., vol. 18, no. 8, pp. 1018-1049.

Harlton, Bruce H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 22, no. 7, pp. 852-914.

Hendricks, T. A., Gardner, L. S., Knechtel, M. M., and Averitt, Paul, 1947, Geology of the western part of the Ouachita Mountains of Oklahoma: United States Geol. Survey, Prelim. Map 66.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Bull. 32.

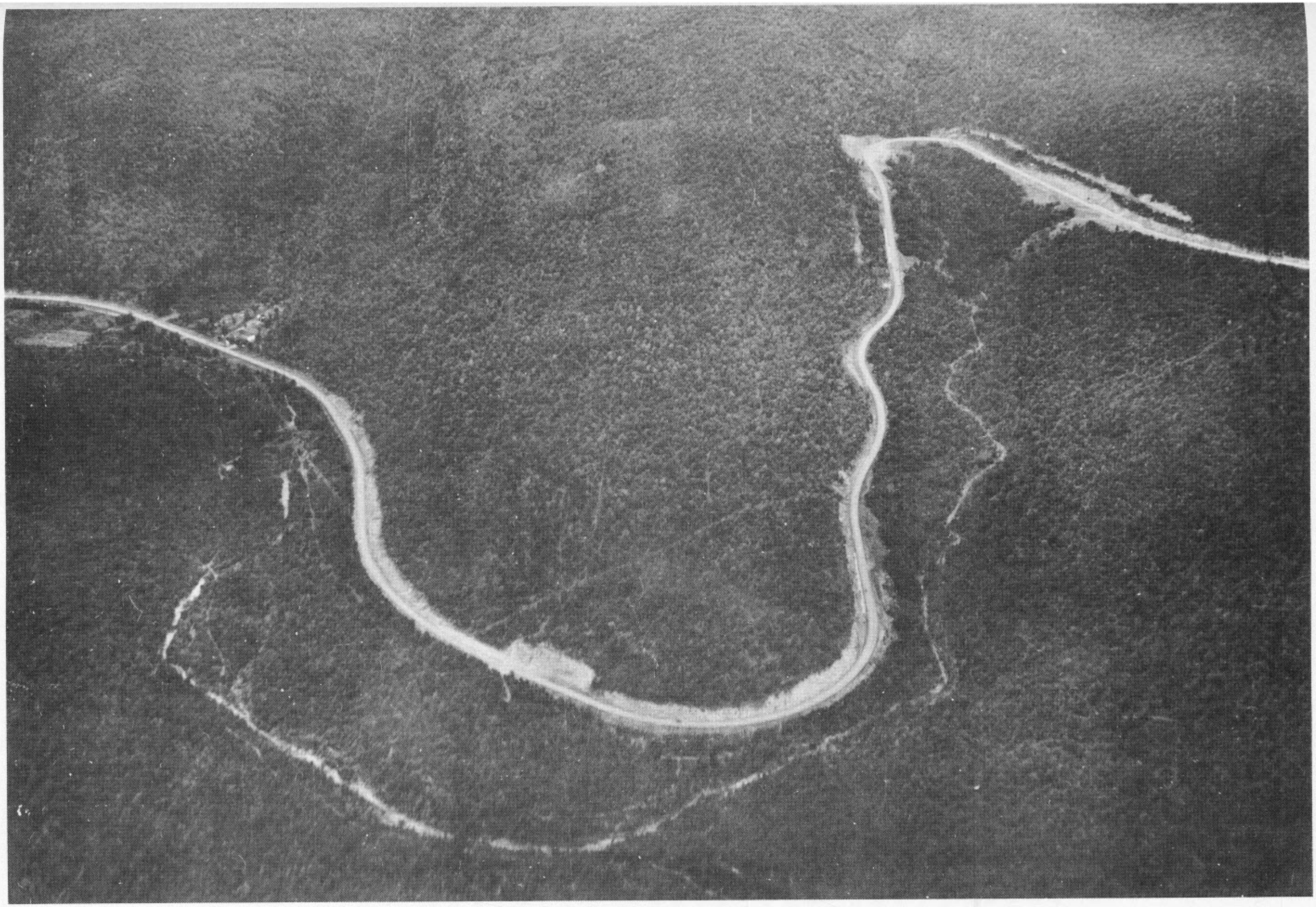
Laudon, Richard B., 1958, Chesterian and Morrowan rocks in the McAlester basin of Oklahoma: Oklahoma Geol. Survey, Circular 46, pp. 1-30.

Maher, John C., and Lantz, Robert J., 1953, Correlation of pre-Atoka rocks in the Arkansas Valley, Arkansas: United States Geol. Survey, Oil and Gas Inv. Chart OC51.

Miser, Hugh D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geol. Survey, Bull. 50.

Miser, Hugh D., 1934, Carboniferous rocks of Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 18, no. 8.

Ulrich, E. O., 1927, Fossiliferous boulders of the Ouachita "Caney" shale and the age of the shale containing them: Oklahoma Geol. Survey, Bull. 45.



HAIRPIN CURVE

(Guidebook mileage 46.55, 2nd Day)

View looking west into Stop 9, the Hairpin Curve on State Highway 2. Details of the geology are discussed in the Guidebook.

SOME MISSISSIPPIAN CONODONTS FROM THE OUACHITA MOUNTAINS

Maxim K. Elias[1]

Introduction

Bruce Harlton was one of the first geologists to find conodonts in the late Paleozoic rocks of the Ouachita Mountains. In 1933 he illustrated a few conodonts from the Johns Valley shale at its type locality and from the Wapanucka formation, which he believed to be contemporaneous. In subsequent stratigraphic papers (1934, 1938, 1947) he mentioned the presence of conodonts in the siliceous shale units, each having, as he claimed, its own distinctive conodont faunas. However, he neither illustrated them nor supplied lists of identifications. Miser, Hendricks and other geologists of the United States Geological Survey collected samples with conodonts from a number of localities in the Ouachita Mountains, particularly from the Stanley shale of Arkansas and Oklahoma, and these together with the additional collections made by Hass furnished the material which Hass identified and for which he established the stratigraphic range of more than a dozen species of conodonts. Because nearly all these conodonts are found only as molds in various rocks Hass published a number of line drawings made from photographs of latex impressions of conodonts from the "middle division of Arkansas novaculite", but only one from the "basal part of the Stanley shale", that of Gnathodus bilineatus (Roundy) (Hass, 1951, pl. 1, fig. 1).

The conodonts from the Arkansas novaculite of Arkansas and its partial equivalent, the Woodford of Oklahoma were previously described by Cooper (1931, 1931a, 1935), who also published an unusually comprehensive and beautifully illustrated monograph on the conodonts of the Bushberg-Hannibal (pre-Welden) horizon in the northern Arbuckle Mountains (1939). Thus, there is a considerable information at hand on the conodonts from the lower Mississippian of southern Oklahoma and southwestern Arkansas--but, as this paper shows, additional important information on these fossils may be expected even from this part of the local Mississippian.

The method of obtaining conodonts by washing them out of shale has an inherent drawback in that the most fragile conodonts invariably break into unidentifiable fragments. For example, in spite of an unusually large number of good conodonts recovered by Cooper from the pre-Welden shales several important conodont groups, such as hindeodells, hindeodelloids, neoprioniods, and others are represented in his conodont plates by comparatively few and greatly fragmented specimens. These same fragile groups are also poorly represented in the small conodont fauna from the Caney shale of the northern Arbuckle Mountains described by Branson and Mehl (1941). The fact that the fragile conodonts are actually present in a good number in the late Paleozoic rocks of the Arbuckle and Ouachita Mountains is now being proved by repeated collecting of samples of shales with conodonts and their molds in shales. This method is admittedly tedious and slow, but it seems the only one that insures, in the long run, a reasonably complete conodont representation from all of the local conodont-bearing stratigraphic horizons. Only when reasonably complete conodont faunas are established it is possible to find out which of the conodonts are truly good horizon markers, and which have long stratigraphic ranges.

Acknowledgments

The conodonts described and illustrated in this paper were discovered and collected in the field by Allan P. Bennison, Lewis M. Cline, Norman L. Johnson and myself, mostly

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in the Kiamichi Mountains and Potato Hills and their vicinities in 1957 and 1958. Our collections are far from complete, but even so they are the first documentary conodont evidence from several conodont bearing horizons in the Stanley shale and other formations, and some of them throw new light on the age of the lower Stanley. Many more conodonts from the Ouachita and Arbuckle Mountains have been collected, prepared, and sketched, but lack of time (and space) does not permit to describe and illustrate them here, and they will be published elsewhere. Most of the work in collecting, preparation, sketching, and identification of the described conodonts was sponsored by Dr. Charles W. Tomlinson, but important new conodonts from the lower Stanley shale of the Potato Hills were collected in the course of field work by Allan P. Bennison and Norman L. Johnson of the Sinclair Oil and Gas Company, in which I participated. I subsequently studied the conodonts for this company. Thanks are due to the Sinclair Company for permission to publish here my conodont identifications.

Method of Study and Sketching

The procedure employed in the study and sketching of the Ouachita conodonts may not be the best possible method, but it is justified by (1) their peculiar state of preservation, (2) the ability of the writer to sketch, and (3) time limitations. If there were at least three or four times as much time available and if complex and delicate illumination for photographing were to be utilized, more exact images of the conodonts could be obtained. Perhaps this same situation confronted Harlton two decades ago, and somewhat later Hass seemingly experienced the same difficulty in spite of having more elaborate photographing equipment at the United States National Museum. Accordingly, only a few conodonts from the Johns Valley and the Wapanucka were illustrated by Harlton, and Hass sketched only one plastotype of a conodont (1951, pl. 1, fig. 1, p. 2530) from the basal Stanley shale (he published superb photographs of the much better preserved Barnett shale conodonts from Texas, 1953).

After obtaining a good number of conodonts and their molds from the shales of different late Paleozoic ages from scores of localities in the Arbuckle and the Ouachita mountains, I was confronted with the need to demonstrate this record graphically in a reasonably short time and in a way that could be useful for stratigraphic purposes. In view of this I decided to utilize my experience in sketching microfossils from their microscopic images and without resorting to camera lucida.

The main reasons for abandoning the use of the camera lucida were again the complexity of technique and time element. The conodonts collected presented a motley group of states of preservation. Fully preserved conodonts were extremely rare. Usually only some portions of a conodont were left embedded in the shale; the others were crushed out, and the surfaces of fractures variedly oriented in space. In a few cases only a certain direction of light produced the clearest possible image of <u>all</u> pertinent details of a conodont in a <u>single</u> light setting; but even in these cases the best light direction was seldom that from the left upper corner, the standard direction used in photographing fossils. Instead of being impressed not upon an ideally flat surface of shale, some good conodont molds were impressed upon an uneven (microscopically) surface due to gritty nature of the shale. In order to make out the complete outline of a conodont (or of its mold) and to observe all its pertinent details (dentition, grooves, etc.), it was necessary in most cases to rotate it completely so that light could illuminate now this and now that detail.

The devised method of sketching is as follows:

(1) The <u>principal</u> <u>dimensions</u> of a conodont (or its mold) were measured by the scale engraved in the ocular of a binocular microscope. In most examples several settings of conodont are needed in order to measure the various distances, because the best possible illumination for each measured distance is essential for exactitude of the various measurements. The light should be sufficiently strong for visibility, expecially when the impression of a conodont is black upon a background of black shale. It is the luster of the impression that distinguishes the conodont impression from the surrounding dull matrix. It was found best <u>not</u> to mount conodonts (or their molds). This allows their free turning in <u>all</u> directions of space. Standard X40 magnification was invariably used.

(2) Details were sketched after the various distances were measured and good understanding of the morphology was reached; and presence of cracks, breaks along which some parts were broken off, and other pertinent details were established. After plotting the distances on sketching paper a free-hand filling of overall outline and of details inside of it followed. Conodont impressions in shale fragments were taken between two fingers of one hand and variously turned for better illumination of one part after another. The other hand was used for sketching.

(3) Shading was done with the light placed in the left upper corner of a properly oriented conodont, --its standard position. Molds were illuminated in the same way.

What other method can be used to illustrate imperfectly preserved conodonts if one has no ability to sketch? Highly magnified camera lucida drawing may be recommended, with light shifted about so as to illuminate parts which are not clearly observable at an initial light position.

Taxonomy of Conodonts

Technical difficulties. - Circumstances conspire to make the taxonomy of conodonts unusually complex and difficult. One difficulty is the existance of a dual nomenclature: one for assemblages and another for different kinds of conodonts found together in assemblages. The genera established for the latter are admittedly artificial "form genera", but they are indispensable for classification of by far the greater majority of conodonts collected. The first conodont genera (introduced by Pander in 1856) are of this kind. The facts that their types are lost, while their original illustrations are generalized and incomplete, add another difficulty to conodont taxonomy. Great susceptibility for becoming electrically charged by even a slight friction, coupled with fragility adds to the difficulties in professional handling of these microscopic fossils. Permanent mounting protects them from breaking or loss, but, on the other hand, prevents complete study and illustration because only one side remains exposed. When more than one individual of a species is collected, combination of differently mounted specimens provides for complete illustration; but in most cases only one specimen of a species is available. Because of the scarcity of individuals of nearly all species it is impossible to establish their variability and ontogeny. Students of conodonts prefer not to mention and usually give no dimensions for their types, and frequently (especially in the past) maintained no standard magnification for illustrations.

Selection of generic characters. - Selection of generic characters for conodont form genera is obviously arbitrary. Although genera gradually became more or less stabilized, many generic concepts are still in a state of flux, and from time to time are being revised. Of course, this is the case in many other groups of fossils, but because of artificiality of the conodont genera, and lack of ontogenetic and variability controls in establishing of conodont species, the generic concepts of conodonts are particularly unstable and open for revision. Just as in the case with other groups of fossils, a critical time for revision of genera arises when intermediate forms between two or more of supposedly sharply different genera are discovered. Such intermediate forms, or connecting links, are particularly apt to occur when fossils, in our case conodonts, are newly collected from stratigraphic intervals where previously only a few, if any were known.

In his important revision of Pennsylvanian conodonts Ellison remarks (1941, p. 109): "Little is known concerning the conodonts either of the upper Mississippian or of the beds of questionable age variously referred to the Mississippian or the Pennsylvanian." It is to this late Mississippian-early Pennsylvanian interval of time where many of the recently collected conodonts of southern Oklahoma belong, and in the course of work on them I became aware of the fact that even the earlier Mississippian conodonts, down to the inception of Carboniferous time, are in need of taxonomic re-appraisal for reception of the numerous conodonts now being collected and studied from the Ouachita and Arbuckle mountains.

Taxonomy and stratigraphy. - Paleontologists are becoming more aware of the necessity of working out the taxonomy of their fossils in intimate contact with advances in detailed stratigraphy. The whole future of paleontology depends on its adjustment to detailed stratigraphy, not only because such adjustment assures its effective use in economic geology, but because it also assures progress in our understanding of fossils as ancient organisms, whose shape and size were dependent on a combination of fluctuating environments and the march of time in the same manner as living organisms; and it is the detailed stratigraphy that provides a paleontologist with a down to earth control that protects him from premature generalizations based on erroneous stratigraphic premises.

Procedure in revision of genera and species. - The primary aim of taxonomy is orderly cataloging. It is an indispensible tool that allows us to handle the millions upon millions of living and fossil species. Revision of taxonomy of even a very small group of these organisms should be expedited with the least disturbance to the previously established orderly cataloging, no matter how imperfect. One of the ways of introducing a new genus or a new species is to first smuggle it as a subgenus or subspecies. This method, frequently practiced in the past, perhaps should not be now recommended for natural genera and species. On the other hand, this practice seems still proper and desirable in handling of artificial genera and species. One of its advantages is that it provides a solid link with previously used and well known generic and specific names, an important help in adjusting our memory to the newly gained knowledge to that previously accumulated. Another advantage is that it provides for a period of time in the course of which validity and practicability of a new taxonomic unit may be tested by various workers on fossils in question.

Basic assumptions in conodont taxonomy. E. O. Ulrich, the founder of the modern taxonomy of Paleozoic bryozoans, was labeled by his contemporaries as a "species maker", meaning that he created an unnecessarily large number of fossil species. In some cases, we are now finding it necessary to recognize two or more species where Ulrich saw only one. When dealing with a great number of conodonts gradually collected anew, we naturally try to identify as many of them as possible with those previously described. Knowing so little about their variability and ontogeny, we are apt to consider some deviations from a type with which we attempt to identify our newly collected form, as being mere individual variations. It is when we collect from a number of successive horizons, and our collection becomes large enough for the recognition of slight but perceptible evolutionary changes in what previously was considered a long ranging and more or less variable genus or species, that a re-evaluation of taxonomic significance of various characters becomes timely; as it may lead to a greater usefulness of a fossil group. Such is the present situation with the Carboniferous conodonts of southern Oklahoma.

The proposed new groupings at sub-generic, specific and subspecific levels are an experiment in testing their seeming validity as horizon markers and as phyletic groups.

In connection with this research some of the pertinent previous taxonomic errors and confusions are analyzed and remedied. The illustration of the conodonts involved in them are republished in form of sketches uniformly executed in a standard x40 scale, and arranged in stratigraphic order.

SYSTEMATIC DESCRIPTION

Genus GNATHODUS Pander (1856)

Subgenus HARLTONODUS Elias, new subgenus

? 1900. Gnatholdus mosquensis Hinde: Nat. Hist. Soc. Glasgow Trans. vol. 5, n. s., pt. 3, p. 342, figs. 2-4 (L. Carbon. of Dalry, etc.).

? 1926. Polygnathus (Gnathodus) mosquensis (Pander) Hinde 1900. Holmes: U. S. Nat. Mus. Proc. vol. 72, p. 11, 18, 37; pl. 7, figs. 2-4 (not pl. 6, fig. 31).

not 1856. Gnathodus mosquensis Pander: K. Akad. d. Wiss. St. Petersburg, p. 33-34, 90; pl. 2A, figs. 10a, b, c.

not 1926. Gnathodus mosquensis Roundy: U. S. Geol. Survey Prof. Pap. 146, p. 12; pl. 2, figs. 6a, b, c, d (Pander's figures re-published)

not 1926. Gnathodus mosquensis Ulrich and Bassler: U. S. Nat. Mus. Proc. vol. 68, p. 54, fig. 5, subfig. 14 (on p. 44) (one of Pander's figures re-published)

not 1928. Gnathod mosquensis Holmes: U.S. Nat. Mus. Proc. vol. 72, p. 10, pl. 6, fig. 31 (Pander's figures re-published)

not 1939. Gnathodus mosquensis Cooper: Jour. Pal. vol. 13, p. 388; pl. 41, figs. 23-25, 30-32.

A new subgenus, Harltonodus, is proposed for conspicuously lopsided bar conodonts previously placed now in Gnathodus, now in Polygnathus. It differs from both of these genera in having outer part of platform two to three times as wide as inner half, the two parts also bearing different oral (upper) sculpture: outer part covered with numerous variously arranged round or elongated (ridge-like) tubercles, while inner part is transversely divided into subequal ridges. The latter is a type of sculpture that characterizes both parts of symmetrical platform in the Devonian Icriodus B. & M.; Devonian-Mississippian Polygnathus Hinde, Mississippian Pseudopolygnathus B. & M., Siphonodella B. & M., and Taphrognathus B. & M.; Mississippian-Permian Cavusgnathus Harris and Hollingsworth; Pennsylvanian Polygnathodella Harlton and Streptognathus Stauffer and Plummer; and Pennsylvanian-Permian Gondolella Stauffer and Plummer.

There seems to be no other known genus or subgenus with a similar asymmetrical sculpture, and thus it may be considered most characteristic of the subgenus Harltonodus.

Because oral (upper) surface of generotype of Gnathodus, G. mosquensis (Pander, 1856, pl. 2A, figs. 10a, b, c) is unknown, we know not what its sculpture is. However, judging by the illustrated aboral (lower) view (Roundy, 1926, pl. 2, fig. 6c) the width of outer and inner sides of its platform is subequal. Because of this it would seem proper to continue to place in Gnathodus sensu stricto, all species which have equal to subequal sides of platform, orally sculptured or unsculptured. Here belong, therefore, Gnathodus texanus Roundy (not his Polygnathus texanus), G. inornatus Hass, and similar Mississippian species placed by various authors either in Gnathodus or Spathognathodus (Spathodus) B. & M. As Huddle judicially remarks (1934, p. 89), "Spathodus (now Spathognathodus, because Spathodus is pre-occupied) is probably a synonym of Gnathodus Pander, but until the true characters of Pander's genus are determined it will be advisable to use the name "Spathognathodus". Hence it would seem proper to accept Spathognathodus as another subgenus of Gnathodus, if we wish to retain this genus at all, as most paleontologists do.

Stratigraphic significance. - The stratigraphic range of Gnathodus (Spathognathodus) is from Silurian to lower Permian, whereas that of Gnathodus (Harltonodus) is strictly Mississippian.

Hass summarized his findings on stratigraphic significance of lopsided gnathods in the Stanley-Jackfork rocks of the Ouachita Mountains thus (1950, p. 1581): "The known range of ....Gnathodus bilineatus or a very closely related species, (is) lower part of the Stanley to the Wesley siliceous shale of Harlton; " that is it ranges throughout Stanley-Jackfork rocks.

My findings are harmonious with this statement, this being also the range of Gnathodus (Harltonodus) bilineatus sensu stricto and of G. (H.) bransoni. These species continued to exist to nearly the end of Mississippian time in the Arbuckle Mountains. To this I may add that a new form, Gnathodus (Harltonodus) minutus n. sp. is now established in the lower part of the Stanley, and is seemingly restricted to it. The presence of Gnathodus (Harltonodus) liratus in the shales south of Clayton is an indication of their probable correlation with the Pella beds of Iowa, the only place where this harltonod has been found previously. A related species, Gnathodus (Harltonodus) multilineatus (Elias, 1956, p. 119, pl. 3, figs. 49-53, reillustrated here in pl. 1, figs. 49, 51, 53) has been found in the upper part of the Sand Branch member of the Caney shale and somewhat similar form (pl. 1, fig. 29) in the lower part of the Sand Branch member.

Incidentally, Gnathodus girtyi Hass (1953; pl. 14, figs. 22-24; republished in Elias, 1955; pl. 3, figs. 30, 31) cannot be referred to Gnathodus (Spathognathodus), but could be placed in a particular group of slender polygnaths, such as P. lacinata Huddle, P. pennata Huddle, and P. rugosa Huddle; it seems also related to some species of Pseudopolygnathus B. & M., and Macropolygnathus Cooper.

GNATHODUS (HARLTONODUS) BILINEATUS (Roundy)

Pl. 1, Figs. 3-12

1926. Polygnathus bilineatus Roundy: U. S.

Geol. Survey Prof. Pap. 146, p. 13; pl. 3, figs. 10a-10c.

part 1941. Gnathodus pustulosus Branson and Mehl: Jour. Sci. Lab. Denison Univ., vol. 35, p. 172; pl. 5, figs. 36 and 37 only.

part 1953. Gnathodus bilineatus Hass: U. S. Geol. Survey Prof. Pap. 243-F, p. 78-79; pl. 14, fig. 25 and 26 only.

part 1956. Gnathodus bilineatus Elias: Petr. Geol. Southern Okla., vol. 1, p. 118; pl. 3, figs. 26 and 29 only (= Hass, 1953; pl. 14, figs. 25 and 26).

part 1956. Gnathodus pustulosus Elias: Petr. Geol. Southern Okla., vol. 1, p. 115; pl. 3, figs. 3 and 5 only (= Branson and Mehl, 1941; pl. 5, figs. 36 and 37).

1957. Gnathodus modocensis Rexroad: Ill. Geol. Survey, Rept. Invest. 199, p. 30-31, pl. 1, figs. 15-17.

1958. Gnathodus modocensis Rexroad: Ill. Geol. Survey, Rept. Invest. 209, p. 17-18; pl. 1, figs. 1 and 2.

Holotype. - Polygnathus bilineatus Roundy, 1926, pl. 3, fig. 10a-c: U. S. N. M. 115103. Barnett sh., Texas; Hass, 1953; pl. 14, fig. 26.

Description. - Gnathods with "shallow V-shaped valley" between principal (median) ridge and ridge-like inner side of platform (cup), a character indicated by specific name "bilineatus". Lesser development of the "valley" and differently sculptured oral side of platform influenced Roundy's decision to separate from bilineatus another species, Polygnathus texanus, which Hass (1953) considers a mature form of bilineatus. I now suggest the restriction of bilineatus to the sense originally meant by Roundy, which makes it possible to add the following characters of this species: irregularly tuberculate surface of outer side of platform, and slightly convex edge of inner side of platform.

Discussion. - Hass's treatment of species P. texanus Roundy as a mere mature form of P. bilineatus Roundy may be questioned. In G. pustulosus identified by Hass (1953, p. 78) with P. bilineatus, the outer side of platform (cup) in mature form (Branson and Mehl, 1941, pl. 5, fig. 38) is not "expanded laterally in its anterior two-thirds" as in P. bilineatus (Hass, 1953, p. 79; pl. 14, figs. 28, 29; republished by Elias, 1955, pl. 3, figs. 23, 25). Such expansion is seen, on the other hand, in a smaller than medium-size example of G. pustulosus (see Branson and Mehl, 1941; pl. 5, fig. 32; republished by Elias, 1955; pl. 3, fig. 4). By keeping Roundy's original understanding of P. bilineatus we may identify with it only those two illustrated specimens of G. pustulosus Branson and Mehl (1941; pl. 5, figs. 36 and 37; republished by Elias, 1955, pl. 3, figs. 1 and 6), which have an irregularly tuberculate sculpture of outer side of platform, and a gently convex edge of its opposite or inner side.

Under thus restricted understanding of Gnathodus (Harltonodus) bilineatus the other two specimens illustrated by Branson and Mehl may be classified as Gnathodus (Harltonodus) bransoni Elias, new name, as illustrated in Branson and Mehl's 1941; pl. 5, figs. 32 and 38.

I agree with Hass (1953, p. 79) that Cooper's (1939, p. 388, pl. 42, figs. 59, 60) "Gnathodus bilineatus" from the pre-Welden shale of Oklahoma (my pl. 1, fig. 2) cannot be put in synonymy with G. bilineatus (even in its broad sense used by Hass) because "Roundy's species possesses many characteristics not recorded by Cooper". I consider it a new species ancestral to both G. bilineatus and G. bransoni, n. sp.

Rexroad (1957, pp. 30-31) admits that his Gnathodus modocensis "closely resembles Gnathodus bilineatus (Roundy), and differs from it "chiefly in that the outer margin of the platform is straight, parapet-like, and is separated from the carina (median ridge), which it very nearly parallels, by a marked trough (Roundy's 'valley')". This statement clearly indicates that Rexroad considers most important the same characters on which I now attempt to separate G. bilineatus from G. bransoni; and his G. modocensis matches well G. bilineatus (Roundy) in a restricted sense here proposed. This same conclusion follows from his additional statement (Rexroad, 1958, pp. 17-18), that G. modocensis "resembles G. bilineatus (Roundy)....especially ....when comparing the adult of the former with the young of the latter". By this he means the original specimen or holotype of Roundy's P. bilineatus, or G. bilineatus in restricted sense.

Occurrence. - Lower and Middle Stanley

shale, Delaware Creek member of Caney shale group, and Barnett shale of Texas.

GNATHODUS (HARLTONODUS) BRANSONI
new species

Pl. 1, Figs. 13-18

1926. Polygnathus texanus Roundy: U. S. Geol. Survey Prof. Pap. 146, p. 14, pl. 3, figs. 13a, b.

part 1941. Gnathodus pustulosus Branson and Mehl: Jour. Sci. Lab. Denison Univ., vol. 35, p. 172; pl. 5, figs. 32 and 38.

1951. Gnathodus bilineatus Hass: Bull. Amer. Assoc. Petr. Geol., vol. 35, pl. 1, fig. 1 (p. 2530-2531).

part 1953. Gnathodus bilineatus Hass: U. S. Geol. Survey Prof. Pap. 243-F, p. 78-79; pl. 14, figs. 28 and 29 only.

part 1956. Gnathodus bilineatus Elias: Petr. Geol. Southern Okla., vol. 1, p. 118; pl. 3, figs. 23-25 only (= Hass, 1953, pl. 14, figs. 28, 29; and = Roundy, 1926; pl. 3, fig. 13a).

? 1956. Gnathodus cf. bilineatus Elias: pl. 3, fig. 40.

part 1956. Gnathodus pustulosus Elias: figs. 2 and 3 only (= Branson and Mehl, 1941, pl. 4, figs. 32, 38).

1958. Gnathodus bilineatus Stanley: Jour. Paleontology, vol. 32, pp. 464-465; pl. 68, fig. 7.

Holotype. - Holotype of Polygnathus texanus Roundy, 1926; pl. 3, figs. 13a, b, whose photographed top view is Hass's Gnathodus bilineatus, 1953, pl. 14, fig. 28. The specimen is U.S.N.M. 115103.

The specific name texanus cannot be used for the species when it is transferred (by Cooper, 1939, p. 388 and later by Hass, 1951, p. 2534, and Elias, 1956, p. 118) to the genus Gnathodus, because there is already Gnathodus texanus, also named so by Roundy (1926, p. 12).

The specific name pustulosus is also unavailable because the holotype of Gnathodus pustulosus Branson and Mehl, 1941; pl. 5, fig. 36, Univ. Mo. C543-1 is here referred to Gnathodus (Harltonodus) bilineatus (Roundy) s. s. Hence G. pustulosus becomes a junior synonym of G. bilineatus.

In view of all this the species is given new name, G. bransoni. Its presence is here recognized in the newly collected material from the lower Stanley.

Description. - Strongly asymmetrical gnathods with posterior end of main ridge (carina) with chevron-shaped nodes, and reinforced by two small short ridges on either side. Outer side of platform (cup) sculptured by parallel rows of nodes, many of which fused into short ridges along these rows, which are straight to curved, and in more or less diagonal orientation to main ridge. Inner side of platform about half as wide as outer side, with single ridge divided transversely into small parallel ridges; its edge is angularly convex in middle, and concave at a flaring anterior.

Discussion. - G. (H.) bransoni differs from G. (H.) bilineatus (Roundy) by the regular arrangement of nodes and ridges in outer side of platform and concavo-convex outline of the outer edge of inner side of platform.

In view of younger age of the Barnett shale of Texas than of the Delaware Creek member of the Caney shale of Oklahoma (Elias, 1956, p. 69-70), the larger size of Texas over Oklahoma examples of G. (H.) bransoni may be of stratigraphic significance. With it is combined somewhat greater anterior expansion of the platform in Texas examples. Furthermore, transverse ornamental ridges in inner side of platform in Oklahoma examples are more nearly subequal than the corresponding ridges in Texas examples. An apparent increase of size of G. (H.) bransoni with an advance of geologic age is supported also by the fact that four examples of a closely related species, now found in the lower Stanley shale in association with typical lower Mississippian conodonts, are smaller than the two Delaware Creek shale specimens of N. bransoni: illustrated by Branson and Mehl (1941, pl. 5, figs. 32, 38; Elias, 1956, pl. 3, figs. 2, 3). Of same small size is Cooper's (1939, pl. 42, figs. 59, 60) "Gnathodus bilineatus" from the pre-Welden shale.

Occurrence. - Delaware Creek member of Caney shale group of Oklahoma; basal Stanley shale, 121 feet above base, at Caddo Gap, Arkansas; and Barnett shale of Texas.

A poorly preserved specimen (pl. 1, fig. 18) from the basal part of Goddard shale at its type-locality in Oklahoma seemingly matches the species satisfactorily.

GNATHODUS (HARLTONODUS?) LIRATUS Youngquist and Miller

Pl. 1, Figs. 19-21

1949. Gnathodus liratus Youngquist and Miller: Jour. Pal., vol. 23, pp. 619-620; pl. 101, figs. 15-17.

Description. - Asymmetrical gnathods with "considerably expanded" outer side of platform (cup), "relatively flat and ornamented by irregularly spaced nodes" (Youngquist and Miller, 1949, p. 620). In top view of holotype pl. 101, fig. 17, these same nodes are shown arranged in three rows, the nodes diminishing in size toward interior of platform. Inner side of platform is shown ornamented by a single marginal nodose ridge instead of being transversely divided into small ridges, as characteristic for subgenus Harltonodus; hence the species is questionably referred to it.

Discussion. - Youngquist and Miller admit that their species "somewhat resembles" G. pustulosus Branson and Mehl, but differs "in having the inner oral margin of the platform sharply curved orad forming a narrow furrow between the edge of the platform and the carina; also the latter structure is stouter than in G. pustulosus".

The species is like G. (H.) bilineatus (Roundy) in having furrow ("valley" of Roundy) between main ridge (carina) and marginal (inner) ridge, whose nodes, prior to retouching in the photograph (mentioned by the authors) could have been more like the short transverse ridges in the type of G. bilineatus (Hass, 1953, pl. 14, fig. 26): the whole outline of this marginal ridge is also much like that in Roundy's species. On the other hand the arrangement of nodes in rows in outer side of platform seems transitional between the sculpture in G. (H.) bilineatus and G. (H.) bransoni.

In view of all this G. (H. ?) liratus may be considered, at least at the present, a distinct species. This view is strengthened by the presence of two molds of a gnathod quite similar to it in the shale exposed in Jerusalem Hollow, south of Clayton, Oklahoma. In the mold (pl. 1, fig. 20) the outer side of platform bears three parallel, slightly curved rows of nodes fully comparable in size and arrangement with those in the species from the Pella beds; and the mold of aboral (lower) side of platform (cup) (pl. 1, fig. 21) is almost exact replica of the corresponding view of the Pella bed species (Youngquist and Miller, 1949, pl. 101, fig. 16).

Occurrence. - Pella beds (just above St. Louis limestone) four miles south of Pella, Iowa; and now in the shale south of Clayton, correlated (on the evidence of goniatites) with the Delaware Creek shale of the Arbuckle Mountains, Oklahoma.

GNATHODUS (HARLTONODUS) MINUTUS new species

Pl. 1, Figs. 22-25

Holotype. - Specimen of Pl. 1, fig. 24.

Description. - Very small, strongly arched gnathods, with outer side of platform two to two and a half times as wide as inner side; outer side of platform with three to four antero-laterally curved ornamental ridges, paralleling its curving outer edge; main (median) ridge prominent, nodose, posteriorly acuminate and fused with subequally acuminate ends of outer and inner sides of platform. Inner side of platform about as narrow as main ridge, approximating it in height and nodosity, separated from it by deep furrow, shallowing posteriorly. Remains of a flattened and tangentially cleavaged specimen (pl. 1, fig. 23) reveal inner structure of outer side of platform, showing roots of all four curving ornamental ridges. Posterior part of main ridge narrow, time and a half as long as platform.

Discussion. - Resembles G. (H.) bilineatus in outline of platform and deep furrow (valley) separating main ridge from nodose crest of inner side of platform; but has quite different ornamentation of outer side of platform, which approaches in its curving lineation that in G. (H.) bransoni n. sp. However, this ornamentation differs from the latter in solid (and deep-seated) nature of curving ornamental ridges, while corresponding ornamental ridges in G. (H.) bransoni are broken into short ridges and nodes. Other difference from the latter is absence of anterior concavity in the edge of inner side of platform. Besides, Gnathodus (Harltonodus) minutus differs from both G. (H.) bilineatus and G. (H.) bransoni in smaller size and strong arching of platform. Increase in size from G. (H.) minutus to G. (H.) bransoni to G. (H.) bilineatus may have stratigraphic significance.

Occurrence. - Known only from the lower Stanley shale, 475 feet above its base. Southwestern part of Potato Hills, Oklahoma.

GNATHODUS (HARLTONODUS) MULTILINEATUS Elias

Pl. 1, Figs. 26-28

1956. Gnathodus multilineatus Elias: Petr. Geol. Southern Okla., vol. 1, p. 119, pl. 3, figs. 49, 51-53 (only).

Holotype. - Specimen illustrated in 1956, pl. 3, fig. 49.

Description. - Asymmetrical gnatods whose outer side of platform is sharply truncated and sculptured orally by subequal small nodes densely spaced in each of four (in holotype) rows, which are strictly parallel to main ridge, and sharply separated from each other by narrow linear depressions. Inner side of platform about half the width of outer side, and extends only slightly longer anteriorly, ornamented by one (?) long nodose ridge (outer edge incomplete). Main ridge serrated into fused subequal dents, with slight tendency toward chevron shape at posterior end.

Two younger individuals have fewer rows of subequal nodes ornamenting outer side of platform, and one nodose ridge ornamenting inner side of platform (Elias, 1956, pl. 3, figs. 51, 53).

Discussion. - G. (H.) multilineatus is nearest to G. (H.) bransoni but differs from it by sharp lateral truncation of its outer side and greater regularity of its ornamentation, consisting of sharply differentiated and strictly parallel to main ridge rows of subequal nodes instead of combination of short ridges and nodes in G. (H.) bransoni.

Genus NEOPRIONIODUS Rhodes and Muller, 1956

Generotype: Prioniodus conjunctus Gunnell, 1931.

The revision of Pander's genus Prioniodus by Öpik (1936), and particularly by Lindström (1954), cleared the way for introduction of a new genus Neoprioniodus, which embraces the species ranging from the Ordovician to the Lower Triassic formerly assigned to Prioniodus in a broad sense. Lindström's (1954, p. 589) definition of Prioniodus: "Compound conodonts with a subcentral cusp, from the base of which diverge three denticulate edges or processes, one posteriorly, one anteriorly, and one laterally" clearly excludes from the old genus the familiar "prioniods" with an anteriorly placed instead of subcentral cusp or fang, from which extends only one bar or "process", (posterior to cusp) instead of three in three different directions.

In an attempt to preserve the well established use of Prioniodus, Branson and Mehl suggested to continue its broad application, while admitting that their study of the new material from Pander's type locality indicates that the generotype Prioniodus elegans is an "atypical species" (1944, in Shimer and Shrock's Index Fossils of North America, p. 241). Rhodes and Muller (1956) disagreed with this view and introduced the new generic name Neoprioniodus for the forms excluded from Prioniodus Pander under the emended generic definition by Lindström (1954), and selected for the generotype Prioniodus conjunctus Gunnell.

They define Neoprioniodus as follows: "Compound conodonts consisting of a denticulate posterior bar, at the anterior end of which a large fang (main cusp) is developed. The base of this fang may or may not extend downward below the level of the bar to form an 'anticusp', the anterior edge of which may or may not be denticulated. There is normally a basal cavity below the fang, which may be extended as a shallow groove on the aboral surface of the posterior bar" (Rhodes and Muller, 1956, p. 698).

The definition gives sufficient latitude for an inclusion in Neoprioniodus of nearly all species previously described as Prioniodus except very few strictly Ordovician forms with "three denticulate edges or processes, one posteriorly, one anteriorly, and one laterally" (Lindström, 1954, p. 589).

However, the selection of N. conjunctus for generotype makes the greatly expanded base of the main cusp in this species most typical for Neoprioniodus. Ellison describes it thus: "anticusp plow-shaped, strongly extended below general aboral outline, widely expanded on inner side into a flaring apron, anterior edge of many specimens show germ denticles in transmitted light" (1941, p. 114). Close to N. conjunctus in this and other respects is Hass's species Prioniodus inclinatus from the Barnett of Texas, where anterior denticles are well developed (pl. 4, fig. 25), or an anterior shelf-like projection for their development is evident (pl. 4, fig. 26).

The species with flaring apron and incipient to fully developed anterior denticles comprise only a small group of species customarily placed in Prioniodus (now Neoprioniodus), while most of them have a laterally compressed main cusp with no flaring basal expansion and no anterior denticles, nor shelf-like anterior projection. Pending further division of numerous prioniods, other than Prioniodus and Neoprioniodus in restricted sense, into smaller genera, tentative grouping of some species here described and revised appears desirable. Three groups are suggested: the Neoprioniodus ligo (Hass), group, the N. cassilaris (Branson and Mehl) group, and the N. alatoideus (Cooper) group.

Group of NEOPRIONIODUS LIGO (Hass) of "lazy T" prioniods

The group is characterized by perpendicular orientation of bar to a straight tusk (cusp-anticusp combination), the whole resembling "lazy T" of a cattle brand or T in reclining position.

Stratigraphic range of group is from the Delaware Creek member of the Caney shale to basal part of the Springer group in the Ardmore basin. It occurs in the Barnett shale of Texas, and in Illinois it ranges from the Keokuk to the Renault of the Chester series.

Additional characters of the group are as follows:

2. Bar sharply differentiated from cusp-anticups.
3. Anticusp shorter than cusp, usually half as long as cusp, and not less than one-third of cusp.
4. Denticles subequal, discreet to widely spaced.

It seems that the group originated from Neoprioniodus cassilaris (Branson and Mehl, 1941-B, p. 186; pl. 6, figs. 11, 12, 16, 17) or related species through sharper differentiation of bar from "tusk" or cusp-anticusp, as can be seen in what I believe to be a variety of N. cassilaris, which may be labeled as N. cassilaris var. keokukensis Elias, n. var.

Type of group. - Neoprioniodus ligo (Hass), Barnett shale, Texas.

Other species of group. - Neoprioniodus erectus Rexroad, N. scitulus Branson and Mehl, s.s. and N. rynikeri Elias, n. sp.

NEOPRIONIODUS LIGO (Hass)

Pl. 2, Figs 12-14

part 1926. Prioniodus peracutus Roundy: U. S. Geol. Survey Prof. 146, p. 10, pl. 4, figs. 7, 8 only.

1953. Prioniodus ligo Hass: U. S. Geol. Survey Prof. Paper 247-F, p. 87-88, pl. 16, figs. 1-3.

1956. Prioniodus ligo Elias: Petroleum Geol. Southern Oklahoma, vol. 1, p. 109, pl. 2, figs. 16-18 (redrawn from Hass's photograph figs. 1-3).

1956. Prioniodus cf. ligo Elias: Petroleum Geol. Southern Oklahoma, vol. 1, pl. 3, fig. 51.

Holotype. - U. S. N. M. 115172; illustrated in Hass's pl. 16, fig. 1.

Description. - Species is well characterized by Hass, and judging by three examples illustrated is only slightly variable. For its difference from closely related N. erectus and N. scitulus s.s see discussion under description of Neoprioniodus erectus Rexroad.

NEOPRIONIODUS ERECTUS Rexroad

Pl. 2, Figs. 8-11

1957. Neoprioniodus erectus Rexroad: Illinois Geol. Survey, Rept. Invest. 199, p. 34, pl. 2, figs. 23, 25 (Renault fm.).

Holotype. - Rexroad illustrated both of his "cotypes, two specimens", Illinois Geol. Survey 2P58. For reasons stated below I designate specimen illustrated in his pl. 2, fig. 25 (my pl. 2, fig. 11) as holotype.

Description. - Because all sides of liberated (from rock) specimens from Illinois are open for examination, Rexroad's description of the species is complete in all details: "Posterior bar short, thin, arched, bowed inward; denticles probably seven or eight in number, slightly compressed laterally, free. Terminal fang (cusp) long, narrow, strongly compressed laterally with sharp edges fore and aft; outer side more convex in cross section; from lateral view anterior margin (of cusp) slightly convex, posterior margin straight; viewed anteriorly fang is concave inward; tip slightly twisted. Aboral projection (anticusp) long, pointed, postero-aboral margin convex, meeting aboral margin of bar at low obtuse angle. Lateral tips of pit not flared, extending from tip of aboral projection onto aboral margin of posterior bar, making pit exceptionally long and narrow..." (italics mine, to indicate principal characters of species which can be recognized in the ordinarily encountered conodont molds in the siliceous Stanley shales). Specimen illustrated in Rexroad's pl. 2, fig. 25 is selected holotype because posterior edge of its anticusp ("postero-aboral margin" of "aboral projection", in Rexroad's description) is not merely convex, as described, but shows a slight rounded angularity at a point nearer to posterior bar than to tip of anticusp. Similar, but less conspicuous angularity is observable in the corresponding convex margin of Rexroad's figure 23, only here it is closer to tip of anticusp than to bar. Specimen of figure 25 is also preferable for holotype because its denticles are better preserved.

Two neoprionioids in my collections are remarkably similar to N. erectus, particularly the specimen (pl. 2, fig. 9) from the shale south of Clayton correlated with the Delaware Creek. The other of my specimens is from the Delaware Creek shale in the Henry House Creek of the Ardmore basin. Most characteristic features common with N. erectus are as follows: straight, knife-like main cusp, with gently convex anterior edge, with maximum convexity opposite posterior bar; posterior bar narrow, perpendicular to main cusp, and slightly arched; denticles few, subequal, discreet, perpendicular to bar and parallel to main cusp, anticusp wider than cusp, its posterior edge angularly convex. These two specimens from Oklahoma differ from N. erectus only in wider spacing of denticles.

Discussion. - Rexroad points out that "N. erectus has an outline almost identical with that of Prioniodus ligo Hass", but finds difference in their anticusps. Perhaps most important difference between the two species is practically straight anterior edge of cusp-anticusp in Neoprioniodus ligo instead of a slight but quite distinct convexity of same in N. erectus, and nearly twice as large size of N. ligo over N. erectus.

Similar to both of these species is also Neoprioniodus scitulus (Branson and Mehl) from the Delaware Creek shale. As Hass (1953, p. 88) expresses it, "in gross features Prioniodus ligo closely resembles P. scitulus Branson and Mehl, but differs in that it is larger and has the aboral side of its main cusp grooved instead of excavated". Besides, anterior edge of cusp-anticusp in Neoprioniodus scitulus is distinctly convex, being in this respect similar to that in N. erectus. Another difference in N. schitulus is palmate arrangement of denticles instead of their parallel arrangement in both N. ligo and N. erectus; and anticusp in N. scitulus is as wide as cusp, instead of being wider than cusp in N. ligo and N. erectus.

The two neoprionioids from the Vienna-Menard and Golconda of Illinois illustrated by Rexroad (1957) (in pl. 2, figs. 22 and 26) do not match either N. ligo or N. scitulus, and I consider them to represent a new species whose principal characteristics are pointed out below.

Occurrence. - Renault limestone of Illinois, and Delaware Creek shale member of Caney shale in the Arbuckle and Ouachita Mountains of Oklahoma.

NEOPRIONIODUS SCITULUS
Branson and Mehl

Pl. 2, Figs. 6, 7

1941-A. Prioniodus scitulus Branson and Mehl: Denison Univ. Sci. Lab. Bull., vol. 35, p. 173, pl. 5, figs. 5, 6.
1956. Prioniodus scitulus Elias: Petroleum Geol. Southern Oklahoma, vol. 1, p. 109, pl. 2, figs. 9, 10.
not 1957. Neoprioniodus scitulus Rexroad: Ill. Geol. Survey, Rept. Invest. 199, pl. 2, figs. 22, 26.
not 1958. Neoprioniodus scitulus Rexroad: Ill. Geol. Survey, Rept. Invest. 209, pl. 5, figs. 10-14.

Holotype. - Specimen illustrated by Branson and Mehl (1941-A) in pl. 5, fig. 6; my designation (re-published in Elias, 1956, pl. 2, fig. 9). Both original cotypes numbered Univ. Mo. C545-4.

Description. - Branson and Mehl describe N. scitulus as follows: "Posterior bar straight or very slightly arched, laterally straight or slightly concave inward, short, thin; aboral (lower) edge truncated and medially grooved. Bar denticles slightly inclined backward, laterally compressed, small, slender, subequal, with gradually tapering sharp free apices. Terminal fang (cusp-anticusp) erect, straight or somewhat curved, concave inward, laterally compressed with sharp edges, exceptionally long and slender, gradually tapering to a sharp point; produced aborally (below) in a sharply pointed process (anticusp) that extends considerably below the aboral edge of the bar; anterior edge, viewed laterally, presenting a regular, gently convex outline. Excavation beneath the fang shallow, narrow, with sharp lateral edges neither of which is greatly flared. Syntypes: - Univ. Missouri, C545-4." Italics are mine, to indicate the characters observable in lateral view of the species.

Discussion. - The best way to understand N. scitulus in its original narrow sense is to remember its close similarity to N. ligo, the type of "lazy T" group of prioniods. The present selection of holotype of N. scitulus helps to emphasize the two principal differences between these two species: N. scitulus has a distinctly convex edge of anterior instead of straight to basally slightly concave edge in N. ligo; anticusp in N. scitulus is proportionally shorter; and its denticles are slightly palmate, particularly so in holotype.

For other comparisons see the discussions under N. erectus and N. cassilaris.

NEOPRIONIODUS RYNIKERI new species

Pl. 2, Fig. 15

Holotype. - Figured specimen, pl. 2, fig. 15.

Description. - Typical "lazy T" prioniod of Neoprioniodus ligo group, with cusp-anticusp straight, laterally compressed, anterior edge straight to curved backward in upper part of cusp. Anticusp nearly as long as cusp. Bar straight, perpendicular to cusp-anticusp; denticles spaced their own width apart, slender, in palmate arrangement.

Discussion. - N. rynikeri differs from other species of N. ligo group by nearly equal length of its cusp and anticusp, and by palmate arrangement of denticles. It is nearest to N. ligo, but the latter is twice the size of N. rynikeri, and its denticles are subparallel and densely spaced.

Occurrence. - Subsurface Delaware Creek shale, Gulf Oil Corp. Riner # 1 well, 6215 feet, near Overbrook, Ardmore basin, Oklahoma.

Group of NEOPRIONIODUS CASSILARIS (Branson and Mehl)

Prioniods with imperceptible transition from fang to posterior bar, that is with no clear cut differentiation between the two structures.

Stratigraphic range of group is from Middle (?) Devonian to basal Springer group.

2. Fang dominates (except in very late species).
3. Base of fang (cusp) with antero-posterior flare.
4. Bar arching to posteriorly deflected.
5. Denticles more or less decreasing in size posteriorly, confluent or not confluent at base.

Oldest representative of the group is Neoprioniodus alatus Hinde. The group seemingly differentiated from Neoprioniodus s.s. through loss of flaring apron in anterior.

Type of group. - Neoprioniodus cassilaris Branson and Mehl.

Other species of group. - Neoprioniodus solidiformis Elias, N. miseri, new species, and two undescribed species in the middle Goddard and basal Springer of the Ardmore basin.

NEOPRIONIODUS CASSILARIS (Branson and Mehl) em.

Pl. 2, Figs. 17-21

part 1941-B. Prioniodus cassilaris Branson and Mehl: Denison Univ. Sci. Lab. Bull. vol. 35, p. 186, pl. 6, figs. 11, 12, 16, 17 only.

1950. Prioniodus cassilaris Youngquist, Miller, and Downs: Jour. Paleontology, vol. 24, p. 528, pl. 67, figs. 23, 24.

1957. Neoprioniodus scitulus Rexroad: Illinois Geol. Survey, Rept. Invest. 199, pl. 2, figs. 22, 26.

not 1941-A. Prioniodus scitulus Branson and Mehl: Denison Univ. Sci. Lab. Bull., vol. 35, p. 173, pl. 5, figs. 5, 6.

Holotype. - Univ. Missouri C575-3. Specimen illustrated by Branson and Mehl, 1941-B, pl. 6, figs. 11, 12: aboral (bottom) and side views.

Description. - Branson and Mehl described their holotype and cotypes, liberated from rock, as follows: "Fangs laterally compressed with sharp anterior and posterior edges, gently tapering from an antero-posteriorly wide base; in lateral view straight or slightly recurved with slightly concave anterior outline, in some cases convex in its proximal half; moderately extended aborally. Posterior bar arched, slightly curved laterally, of moderate length, comparatively thick and narrow; denticles not confluent at their bases, but offset with the sides of the bar, moderately compressed, closely crowded but distinct, with short free pointed apices. Aboral excavation beneath the fang without laterally flaring lips, very shallow, bilaterally almost symmetrical, greatly extended antero-posteriorly and extending as a groove on the flat aboral surface of the posterior bar." Italics are mine, to indicate principal characters of species which can be recognized in side view.

Branson and Mehl demonstrated variation in anterior outline of P. cassilaris in lateral views of four specimens, and considered that the specimen of figures 16 "may represent another species" (explanation to plate 6). The present attempt to segregate "lazy T" group of prioniods influences my judgment that still another of four original specimens of P. cassilaris may be considered its variety keokukensis, n. var., and ancestral to this group. The remaining three specimens, including holotype, may be characterized by:

1. Expansion of cusp (fang) into its base with slight anterior flare that results in slight concavity of anterior edge near its base.
2. Straight anterior edge, except mentioned slight concavity near base.
3. Expansion of base posteriorly with greater than anterior flare, which tends to minimize differentiation from it of posterior bar.
4. Palmate arrangement of denticles.

When thus characterized the species is recognizable in the unpublished collections from the middle of Goddard shale in the Ardmore basin, apparently its stratigraphically highest occurrence.

The more complete (fig. 23) of two Burlington conodonts identified with N. cassilaris by Youngquist, Miller, and Downs (1950) seems nearest to N. cassilaris var. keokukensis than to typical forms of the species; the second (fig. 24) is like typical forms.

Two specimens of Neoprioniodus scitulus from the Vienna-Menard and Golconda illustrated by Rexroad (1957, pl. 2, figs. 22, 26) seem to be referrable to N. cassilaris s.s., as here emended.

Discussion. - Comparison of the mentioned two illustrations by Rexroad with three illustrations from the Kinkaid of what Cooper identified as Prioniodus scitulus (1946, pl. 20, figs. 1-3) shows that understanding of Neoprioniodus scitulus Branson and Mehl became overly broad. When it is restricted to narrow, original sense, neither Rexroad nor

Cooper's identifications will be acceptable. Five specimens from the Glen Dean of the Chester series identified by Rexroad (1958, pl. 5, figs. 10-14) as Neoprioniodus scitulus represent a species, seemingly new, that is more nearly like N. cassilaris. Nearest to the latter are two younger examples (figs. 10 and 11), which may indicate recapitulation of ancestral characters. N. cassilaris s.s. differs from N. scitulus s.s. by basal antero-posterior flaring of cusp (fang), so that the base of cusp can be hardly considered an anticusp. N. scitulus could have developed from N. cassilaris through greater differentiation of posterior bar such as seen in N. casillaris var. keokukensis, n. var. (Branson and Mehl, 1941-B, pl. 6, fig. 15).

NEOPRIONIODUS CASSILARIS var. KEOKUKENSIS, n. var.

Pl. 2, Fig. 22

part 1941. Prioniodus cassilaris Branson and Mehl: Denison Univ. Sci. Lab. Bull. vol. 35, p. 186, pl. 6, fig. 15 only.

Holotype. - Specimen illustrated by Branson and Mehl, 1941, pl. 6, fig. 15.

New varietal name keokukensis is given to prioniod which differs from typical N. cassilaris in differentiation of antero-posterior flaring of basal part of cusp into a distinct anticusp whose width equals that of cusp. Another difference of new variety is greater length of posterior bar.

The mentioned differences are considered transitional from N. cassilaris group of prioniods to N. ligo group.

Occurrence. - Known only in the Keokuk formation of Illinois.

NEOPRIONIODUS MISERI, new species

Pl. 2, Figs. 23, 24

Holytype. - Specimen illustrated in pl. 2, fig. 23, from about 160 feet above base in Delaware Creek member of Caney shale. Henry House Creek, Ardmore basin, Oklahoma.

Description. - Laterally compressed prioniod, with conspicuously wide and thin cusp-anticusp, with two or three longitudinal to diagonal gentle corrugations. Cusp passing imperceptibly into laterally compressed, anteriorly flaring bar. Denticles nearest to cusp appear to cleave from its posterior edge, first four or five subequal, descending along posterior edge of cusp in offset position to each other. Subsequent denticles very gradually diminishing in size on gradually tapering bar.

Discussion. - Two nearly identical specimens were recovered: one in the Delaware Creek shale of the Ardmore basin, and the other in the lower conodont-bearing bed of an equivalent to the Delaware Creek shale in Hershel Craig farm, south of Clayton, Oklahoma. The type of species from the Ardmore basin has a more acuminate cusp than the example from the Ouachita Mountains, but the latter has a more acuminate bar than in the former.

N. miseri differs from all other species of N. cassilaris group in having anteriorly flaring and steeply posteriorly deflected bar, and in cleave-like development of denticles nearest to cusp.

Occurrence. - Neoprioniodus miseri has been found only in the Delaware Creek shale and its equivalents of the Arbuckle and Ouachita Mountains. A somewhat similar but distinct undescribed species occurs in the middle Goddard shale, and another similar, also undescribed species, occurs in the basal part of the Springer group, both in the Ardmore basin.

NEOPRIONIODUS SOLIDIFORMIS Elias

Pl. 2, Figs. 25-27

1956. Prioniodus solidiformis (in explanation to pl. 2 errornously called solidifundus) Elias: Petroleum Geol. Southern Oklahoma, vol. 1, p. 109-110, pl. 2, figs. 28 and 29.

Holotype. - Specimen illustrated in pl. 2, fig. 28 of Elias, 1956 (this paper pl. 2, fig. 25).

Description. - Original description is as follows: "The new species differs from other known species of the genus by having the basal edge of the bar running into the large anterior tooth without appreciable change in direction, so that this straight or nearly straight common edge makes about a 45 degree angle with the anterior edge of the large anterior tooth".

It may be added that anterior edge is convex over greater length, but becomes concave at anticusp. Cusp seemingly oval in cross-section. Denticles five, large, subequal, subparallel to cusp. Bar very short, laterally compressed, practically undifferentiated from cusp-anticusp.

Occurrence. - About 150 feet from base of the Goddard shale at its type-locality (Oil Creek) in the Ardmore basin, Oklahoma.

Group of NEOPRIONIODUS ALATOIDEUS Cooper

No attempt is made here to set limitations of this large, quite tentative group, except it may be broadly defined as laterally compressed prioniods, with long and slender, straight or curved cusp-anticusp; and well differentiated, straight or arched bar, inclined to the cusp-anticusp at various less than 90 degrees angle.

The group seemingly originated in late Devonian time.

Typical species of the group in Neoprioniodus alatoideus Cooper sensu lato of early Mississippian age.

NEOPRIONIODUS aff. ALATOIDEUS (Cooper)

Pl. 2, Figs. 1-3

1931a. Prioniodus alatoideus Cooper: Jour. Paleontology, vol. 5, 232 pp., pl. 28, fig. 1.
1934. Prioniodus alatoideus Huddle: Bull. Amer. Paleontology, vol. 21, pp. 37-38, pl. 1, figs. 4, 5.
1935. Prioniodus alatoideus Cooper: Jour. Paleontology, vol. 9, p. 310, pl. 27, fig. 3 (same as 1931a, fig. 1; in smaller, x30 magnification).
part 1947. Prioniodus alatoideus Bond: Ohio Jour. Sci., vol. 47, pp. 34-35, pl. 1, fig. 6 only.

Holotype. - Single specimen illustrated by Cooper, 1931a, pl. 28, fig. 1 (my pl. 2, fig. 1) from the Woodford formation of Oklahoma.

Description. - Cooper states that "single available specimen" of his species is incomplete, lacking posterior part of bar. His characteristic is as follows: "The bar is straight and somewhat thin...the denticles are broad, fused, sharp pointed, and relatively long. The main cusp is more slender than in P. alatoides Holmes and more inclined forward. The posterior side of the downward projection (anticusp) of the cusp is at right angles to the bar; the whole projection (anticusp) is short and thick as compared with the width of the base of the cusp."

Huddle illustrates two specimens of similar species from the upper New Albany shale of Indiana, admittedly of considerably larger size than Cooper's type (1934, pl. 1, figs. 4 and 5, fig. 4 sketched here as pl. 2, fig. 2), and it seems that his specimens may be accepted as somewhat variable adults of Cooper's juvenile type. If so, Huddle's description constitutes a more complete one for the species.

"Tooth small with short, thin, laterally compressed bar; cusp long, narrow, gradually tapering, sharp edged, rounded on one side, and flattened on the other side; denticles numerous, 20 to 24, long, slender, apparently deeply inserted, and closely appressed. Anticusp subtriangular, with anterior edge forming a straight line with the cusp and the posterior edge perpendicular to the bar. Length .8-1.6 mm."

"The species differs from P. alatus and and P. alatoides in having a narrower cusp and numerous slender denticles" (1934, pp. 37-38).

Only one of two specimens illustrated by Bond (1947, pl. 1, fig. 6) may be accepted as another example of N. alatoideus.

Single mold obtained from the lower Stanley shale is certainly closer to P. alatoideus than to P. alatus and P. alatoides, but if identified with P. alatoideus would constitute another variety of the species, Its size is intermediate between that of Cooper and Huddle's specimens, the angle between lower (aboral) edges of anticusp and bar is much larger than 90 degrees, and the denticles are less numerous (15) and diminish in size suddenly after three first ones behind the cusp.

Discussion. - If more than one species are represented by the illustrated examples placed here in the synonymy of Neoprioniodus alatoideus, they may be at least considered members of a group of closely related conodonts of lower Mississippian age, their delicate cusp and denticles contrasting conspicuously in comparison with all other contemporaneous prioniods. Somewhat similar prioniods collected from younger beds in the Ouachita and Arbuckle Mountains have differences beyond those within just registered variability of N. alatoideus s.l.

Occurrence. - Woodford formation of Oklahoma; middle and upper Arkansas novaculite (identified but not illustrated by Cooper, 1935, p. 310); upper and middle New Albany shale (identified but not illustrated by Huddle from middle New Albany shale, 1934, p. 38); Ohio shale of Ohio; and lower Stanley of Oklahoma.

Genus HINDEODELLOIDES Huddle, em.

1934. Hindeodelloides Huddle, Bull. Amer. Paleontology, no. 72, p. 48.
1947. Hindeodelloides Hass, Jour. Pal., vol. 21, pp. 132, 134.

Huddle defines the genus as follows (p. 48):

"Bar laterally compressed, thin, straight or gently curved, with an anticusp which gives the tooth a T-shaped appearance similar to the pick shape of Ligonodina. The plane including the denticles on the anticusp is inclined to the plane of the bar and cusp. Denticles laterally compressed, closely appressed, and usually alternating in size.

"The genus differs from Hindeodella in having an anticusp; from Ligonodina in the close appression of the denticles; from Hamulosodina in that the anticusp is entirely denticulate, and from Falcodus in that the denticles on the anticusp are at an angle to the plane and cusp.

"Genoholotype: Hindeodelloides bicristatus, new species."

Although Ellison considers this genus a junior synonym of Hindeodella, other authors, and particularly Hass (1947), recognize its validity, and so do I. The recognition of Hindeodelloides is particularly important in view of its narrow stratigraphic range: it has been found only in the upper Devonian and lowermost Mississippian.

It seems that Huddle has not quite clearly indicated in what respect Hindeodelloides differs from Hindeodella, thus seemingly giving a reason for Ellison's decision to lump these two genera together. The so-called anticusp in the genus Hindeodelloides is not a direct prolongation of the cusp (main tooth) but is in more or less off-set position to it. One of the original seven species included by Ulrich and Bassler in their genus Hindeodella, H. decurrence (1957, pl. 8, fig. 13) has nearly the same kind of an off-set anticusp-like frontal prolongation of the cusp, and thus could easily be referred to the genus Hindeodelloides. On the other hand, all three species originally included by Huddle in Hindeodelloides have the following character which is absent in Hindeodella decurrence and all other species of Hindeodella: the posterior of their bar sags down slightly but abruptly, while its height or only the height of dents arising from it, increase. In this respect Hindeodelloides is much like two other of Huddle's genera, Angulodus and Metaprioniodus, whose stratigraphic range is as narrow as that of Hindeodelloides. All three form genera could have come from a jaw of the same biological genus, their narrow Upper Devonian - lowermost Mississippian range being the same. The highest recorded occurrence of Angulodus is a single fragmentary specimen in an extraordinarily large conodont-fauna from the pre-Welden "Bushberg-Hannibal horizon" in the northern Arbuckle Mountains. It was identified by Cooper (1939, p. 385, pl. 47, fig. 78) as Angulodus cf. spissus Huddle, a species originally established in the Middle Albany shale (Upper Devonian) where it is rare.

The present finding of Hindeodelloides is an important indication of a much older

Mississippian age for the lower part of the Stanley than previously believed (Hass, 1951, pp. 2526-2527).

HINDEODELLOIDES BICRISTATUS Huddle

Pl. 2, Figs. 28-30

1934. Hindeodelloides bicristatus Huddle, Bull. Amer. Pal., no. 72, p. 48, pl. 7, figs. 2, 3; pl. 12, fig. 6.

Huddle described three species of Hindeodelloides, and has left without names three additional potential species, known only from fragments. The species desig nated as genotype, H. bicristatus has been found "rare throughout the New Albany shale"; he illustrated three specimens of this species, one from the lower, and two from the upper part of the New Albany shale. Greater angularity of the anticusp to the bar can be observed in the specimens from the upper than from the middle New Albany shale, especially in the specimen illustrated in pl. 7, fig. 3; and it is this same specimen that in this and other respects is much like the specimen from the Stanley. If the magnification indicated by Huddle for his plate 7, figure 3 were exactly correct it would be 1.25 mm. long, but in the text (p. 48) he gives the variability of the length for H. bicristatus from .6 to 1.2 mm., a limit into which the length of the Stanley specimen fits.

The Stanley specimen may seem to have sharper differentiated dents of the anticusps than what can be seen in Huddle's photographs of plate 7, but this may be due to imperfection of the photography, as he shows the dents clearly differentiated in the sketch of his plate 12.

The Stanley specimen has one of the posterior dents larger than the others, and somewhat similar tendency in similar dent can be detected also in Huddle's plate 7, figure 2 (this paper pl. 2, fig. 28).

Taking all the similarities and differences into account, the specimen from the Stanley is well within the variability of Hindeodelloides bicristatus, as per the description and illustrations by Huddle.

Occurrence. - 475 feet above base of Stanley shale. Womble anticline. NE 1/4 NE 1/2 sec. 31, T. 3 N., R. 20 E.

Age Consideration. - The upper part of the New Albany shale, from which H. bicristatus has been collected, is now stratigraphically differentiated from the underlying parts of the New Albany. The latter is named the Blackiston formation and referred to the upper Devonian, and the former is divided, in ascending order, into the Sanderson, Underwood, and Henryville formations. Campbell refers Huddle's upper New Albany conodonts to these three formations (1946, pp. 840, 854-855).

Huddle concluded (1934, p. 23) that "the upper 5 or 10 feet of the formation, including the upper conodont fauna, afford the only likelihood of a zone of Mississippian age", the underlying greater part of the Albany shale remaining in the upper Devonian.

Branson and Mehl (1941-A, pp. 201-204) who examined Huddle's types and collected from the New Albany, have discounted part of the conodonts included by Huddle in his upper New Albany conodont fauna as coming from the weathered underlying Devonian shale. They identify the restricted upper New Albany conodont fauna with the Bushberg ("basal Mississippian") conodont fauna of Missouri. They consider the conodont fauna described by Cooper (1939) from the one foot thick shale under the Welden limestone of Oklahoma as the Middle Mississippian, or possibly "a mixed fauna, Bushberg and Welden, or that the Bushberg of South Central Oklahoma is equivalent to not only the Bushberg of Missouri but also to the Chouteau in its broadest sense and probably the Burlington and the Keokuk as well" (1941a, pp. 204-205). It may be mentioned, in connection with this, that my identification of the numerous bryozoans from the lower part of the Sycamore in the Ardmore basin indicates Burlington-Keokuk age for this part of the limestone, which is an approximate equivalent of the Welden in the northern Arbuckle Mountains.

Summarizing this discussion it may be suggested that the lower part of the Stanley with Hindeodelloides bicristatus cannot be as young as the Sycamore or Welden, that is its age is substantially older than that of the Delaware Creek shale and even more so than the Barnett of Texas.

Genus LIGONODINA Ulrich and Bassler

Thanks to the research by C. L. Cooper and J. W. Huddle, the understanding of the genus has been clarified. For the purpose of the following brief record of the genus in the lower Stanley shale, Huddle's (1934, pp. 58-59) characteristics of the genus are quoted:

"Conodonts with rounded cusp, denticulated bar and anticusp. The denticles on the anticusp are at right angles to the plane of bar and cusp, and are inclined upward. The anticusp extends downward from one side of the cusp, and a deep groove extends from the tip of the anticusp on the undenticulate side to the end of the bar along the aboral (lower) side.

"The essential characters of the genus are the lateral attachment of the anticusp and the fact that the denticles on the anticusp are at right angles to the plane of the bar and cusp." (Huddle, 1934, pp. 58-59).

The position of the denticulated anticusp at a right angle to the plane of denticulated bar complicates identification of the genus when (as frequently is the case), only its bar, or only its anticusp are observed. When such broken from each other parts are well preserved, their assignment to a species is only seldom possible. Huddle illustrated many of such well preserved parts without specific identification.

Because only such broken apart parts are preserved as molds in the lower Stanley shale, they are here classified as _Ligonodina_ sp. They are satisfactorily preserved, not only for this generic assignment, but also for comparison with the previously described species of the genus, and this comparison indicates their close relation to some upper Devonian, but particularly to lower Mississippian species.

LIGONODINA sp., cf. LIGONODINA sp. Huddle, 1934, pl. 12, fig. 8

Pl. 2, Figs. 32-35

A surprisingly large percentage of the conodont molds recovered from the lower Stanley shale can be placed in the genus _Ligonodina_, the frequency of occurrence in itself suggesting an early Mississippian age. The detached denticulated anticusps, such as illustrated here on plate 2, figures 33 and 34 are quite like some of those from the Upper Albany shale illustrated by Huddle (1934, pl. 12, figs. 17, 19, 21); and they are quite unlike the rarely described, much smaller anticusps from later Mississippian ages, such as from the Barnett of Texas, -- see _Ligonodina fragilis_, Hass, 1953, pl. 15, fig. 1.

Molds of bars typical of genus _Ligonodina_, the best illustrated here in plate 2, figure 35, are also fairly comparable to the examples of the genus from the New Albany shale, one of which is shown here side-by-side.

Occurrence. - Common in lower Stanley shale, 475 feet above its base, southwestern part of Potato Hills, Oklahoma.

Explanation of Plates and Various Comments

Until about the middle of 1958 I felt strongly that the conodonts collected by me and others in the Ouachita Mountains tended to support a late Mississippian to early Pennsylvanian age for the Stanley-Jackfork sequence, an opinion reached by some paleontologists in the past on the evidence of a few other marine invertebrates and terrestrial plants. This opinion conflicts with the field geologic evidence accumulated within the past few years by Cline and accepted by Tomlinson and Bennison (with all of whom I have been associated in the study of the geology and paleontology of the region).

A small but stratigraphically very important recent collection of conodonts from the lower part of Stanley shale provided strong evidence in favor of an early instead of medial or late Mississippian age for the lower Stanley, and this necessitates revision and re-evaluation of other paleontologic evidence, particularly that of the earlier collected conodonts. Results of an attempted revision and re-evaluation are expressed in the offered taxonomic revision of some conodont groups, the material on which appears to be more complete than on others. Stratigraphic evidence provided by these groups is assembled on two appended plates. In the lower part of each are shown all conodonts of the described groups known from the middle and upper Arkansas novaculite and the lower and middle Stanley of the Ouachita Mountains, and from their equivalents from elsewhere. In the upper parts of each plate are shown similar conodonts from the Delaware Creek shale and higher Mississippian of the Arbuckle Mountains, and some conodonts from the Mississippian of Illinois. To these are added the following conodonts from elsewhere: a few collected by Cline and me from a stretch of shale exposures in a meridional valley of the Kiamichi Mountains south of Clayton; and from the Barnett shale of Texas. In both of the latter cases the stratigraphic position of the added conodonts was determined on the evidence of the associated goniatites. Cline and I readily agreed that the few goniatites from the shale south of Clayton are of undoubted Caney age (I think they are Delaware Creek---Middle Caney). The goniatites of the Barnett shale associated with the conodonts described from it are a natural mixture of some characteristic latest Visean (P_2) and some equally characteristic earliest Namurian (E_1) of British Isles and western Europe. The Barnett fits into a postulated unconformity (Elias, 1956) between the Delaware Creek shale and the Goddard shale in the Ardmore basin.

The information on the middle and upper Mississippian conodonts of Illinois, Iowa, and Missouri, particularly that on the Chester conodonts by Rexroad; and my recent (unpublished) identification of bryozoans and other invertebrates from the lower part of the Sycamore limestone of the Ardmore basin collected by Jeff Prestwich, are used here in an attempt to correlate the middle and upper Mississippian of the Arbuckle Mountains with the classical Mississippian.

Although many conodonts from the upper Stanley and lower part of Jackfork were also collected and partly prepared and identified, additional work on these is needed. Conodonts were also collected by me, but not studied, from two shale intervals in the Sycamore formation at the spillway of Goddard ranch dam. These and additional conodonts from the ever growing number of horizons and localities in the Ouachita Mountains will undoubtedly provide us with a much more complete information on these microfossils, and will assure their use for stratigraphic purposes in the surface and subsurface of the Mississippian and early Pennsylvanian of southern Oklahoma.

References

Bond, R. H., 1947, Ohio shale conodonts: Ohio Jour. Sci. vol. 47, pp. 21-37, 2 pls.

Branson, C. C., and Mehl, M. G., 1933, A study of Hinde's types of conodonts preserved in the British Museum: Missouri Univ. Studies, vol. 8, pp. 133-156, pls. 11-12.

Branson, C. C., and Mehl, M. G., 1941, Caney conodonts of Upper Mississippian age: Denison Univ. Sci. Lab. Jour., vol. 35, pp. 167-178, pl. 5.

Branson, C. C., and Mehl, M. G., 1941a, The recognition and interpretation of mixed conodont faunas: Denison Univ. Sci. Lab. Jour. vol. 35, pp. 195-209.

Branson, C. C., and Mehl, M. G., 1941b, Conodonts from the Keokuk formation: Denison Univ. Sci. Lab. Bull. vol. 35, pp. 179-188, pl. 6.

Branson, C. C., and Mehl, M. G., 1944, Conodonts, in index fossils of North America, by H.W. Shimer and R. R. Shrock, N. Y., Wiley & Sons; pp. 235-246, pls. 93-94.

Campbell, Guy, 1946, New Albany shale: Geol. Soc. America Bull., vol. 57, pp. 829-908, 7 figs. 3 pls.

Cline, L. M., and Moretti, Frank, 1956, Two measured sections of Jackfork group in southeastern Oklahoma: Oklahoma Geol. Survey, Circ. 41, 20 pp., 1 fig.

Cooper, C. L., 1931, Conodonts from the Arkansas novaculite, Woodford formation, Ohio shale, and Sunbury shale: Jour. Paleontology, vol. 5, pp. 143-151, pl. 20.

Cooper, C. L., 1931a, New conodonts from the Woodford formation of Oklahoma: Jour. Paleontology, vol. 5, pp. 230-243, pl. 28.

Cooper, C. L., 1935, Conodonts from the upper and middle Arkansas novaculite Mississippian at Caddo Gap, Arkansas: Jour. Paleontology, vol. 9, pp. 307-315, pl. 27.

Cooper, C. L., 1939, Conodonts from a Bushberg-Hannibal horizon in Oklahoma: Jour. Paleontology, vol. 13, pp. 379-422, pls. 39-47, 2 figs.

Cooper, C. L., 1947, Upper Kinkaid (Mississippian) microfauna from Johnson County, Illinois; Jour. Paleontology, vol. 21, pp. 81-94, pls. 19-23 (also Illinois Geol. Survey, Rept. Invest. 122).

Cooper, C. L., and Sloss, L. L., 1943, Conodont fauna and distribution of a lower Mississippian black shale in Montana and Alberta: Jour. Paleontology, vol. 17, pp. 168-176, pls. 28-29, 1 fig.

Elias, M. K., 1956, Upper Mississippian and Lower Pennsylvanian formations of south-central Oklahoma: Petroleum Geol. Southern Oklahoma, vol. 1, pp. 56-134, 6 pls., 2 figs.

Ellison, Samuel, 1941, Revision of the Pennsylvanian conodonts: Jour. Paleontology, vol. 15, pp. 107-143, pls. 20-23, 4 figs.

Ellison, Samuel, 1946, Conodonts as Paleozoic guide fossils: Amer. Assoc. Petroleum Geologists Bull., vol. 30, pp. 93-110, 3 figs.

Harlton, B. H., 1933, Micropaleontology of the Pennsylvanian Johns Valley shale of the Ouachita Mountains, Oklahoma, and its relationship to the Mississippian Caney shale: Jour. Paleontology, vol. 7, pp. 3-29, 7 pls.

Harlton, B. H., 1934, Carboniferous stratigraphy of the Ouachitas with special study of the Bendian: Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 1018-1049, 8 figs.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 22, pp. 852-914, 2 pls., 23 figs.

Harris, R. W., and Hollingsworth, R. V., 1933, New Pennsylvanian conodonts from Oklahoma: Amer. Jour. Sci., ser. 5, vol. 25, pp. 193-204, 1 pl.

Hass, W. H., 1950, Age of lower part of Stanley shale: Amer. Assoc. Petroleum Geologists Bull., vol. 34, pp. 1578-1584.

Hass, W. H., 1951, Age of Arkansas novaculite: Amer. Assoc. Petroleum Geologists Bull., vol. 35, pp. 2526-2541, 1 pl.

Hass, W. H., 1953, Conodonts of the Barnett formation of Texas: United States Geol. Survey Prof. Paper 243-F, pp. 69-94, pls. 14-16.

Hinde, G. J., 1900, Notes and descriptions of new species of Scotch Carboniferous conodonts: Nat. Hist. Soc. Glasgow Trans., vol. 5, n. s., pp. 338-346, pls. 9-10.

Holmes, Grace B., 1928, A bibliography of the conodonts with descriptions of early Mississippian species: United States Natl. Mus. Proc., vol. 72, art. 5, 38 pp., 11 pls.

Huddle, J. W., 1934, Conodonts from the New Albany shale of Indiana: Bull. Amer. Paleontology, vol. 21, 136 pp., 13 pls., 2 figs.

Lindström, Maurits, 1954, Conodonts from the lowermost Ordovician strata of south-central Sweden: Geol. Fören. Förhandl, bd. 76, heft 4.

Öpik, Armin, 1936, Konodontidest: On the conodonts: Äratrükk Eesti Loodesest, No. 3, pp. 105-107, figs. 57-59, Tartu.

Pander, C. H., 1956, Monographie der fossilen Fische des silurischen Systems der russisch-baltischen gouvernements: K. Akad. d. Wiss. St. Petersburg, 91 pp., 9 pls.

Rexroad, C. B., 1957, Conodonts from the Chester series in the type area of southwestern Illinois: Illinois Geol. Survey, Rept. Invest. 199, 43 pp., 4 pls., 21 figs.

Rexroad, C. B., 1958, Conodonts from the Glen Dean formation (Chester) of the Illinois basin: Illinois Geol. Survey, Rept. Invest. 209, 27 pp., 6 pls., 14 figs.

Rhodes, F. H. T., and Muller, K. J., 1956, The conodont genus Prioniodus and related forms: Jour. Paleontology, vol. 30, pp. 695-699.

Roundy, P. V., 1926, "Introduction", and "The micro-fauna" in P. V. Roundy, G. H. Girty, and M. I. Goldman, Mississippian formations of San Saba County, Texas: United States Geol. Survey Prof. Paper 146, pp. 1-2 and 5-23, 4 pls.

Stanley, E. A., 1958, Some Mississippian conodonts from the high resistivity shale of the Nancy Watson No. 1 well in northeastern Mississippi: Jour. Paleontology, vol. 32, pl. 68, fig. 7.

Stauffer, C. R., and Plummer, Helen J., 1932. Texas Pennsylvanian conodonts and their stratigraphic relations: Texas Univ. Bull. 3201, pt. 1, pp. 13-15, 4 pls.

Ulrich, E. O., and Bassler, R. S., 1926, A classification of the tooth-like fossils, conodonts, with descriptions of American Devonian and Mississippian species: United States Natl. Mus., Proc., vol. 68, art. 12, No. 2613, 63 pp., 11pls., 5 figs.

Youngquist, W. L., and Miller, A. K., 1949, Conodonts from the late Mississippian Pella beds of south-central Iowa: Jour. Paleontology, vol. 23, pp. 617-622, pls. 101-102.

Youngquist, W. L., Miller, A. K., and Downs, H. R., Burlington conodonts from Iowa: Jour. Paleontology, vol. 24, pp. 525-530, pl. 67.

Explanation of Plate 1

Figures

1. *Gnathodus delicatus* Branson and Mehl. After Hass, 1951, Pl. 1, Fig. 4, Middle Arkansas novaculite, Arkansas.
2. *Gnathodus* new species? (After Cooper, 1939, *Gnathodus bilineatus*, Pl. 42, Fig. 59. Pre-Welden shale, northern Arbuckle Mountains, Oklahoma.

3-12. *Gnathodus* (*Harltonodus*) *bilineatus* (Roundy), *sensu stricto*.
   3. From lower Stanley shale, 475 feet above base: south-western Potato Hills, Oklahoma.
   4. From middle Stanley shale: 8 miles east of Clayton, Oklahoma.
   5, 6. From Delaware Creek shale: 5 miles south of Ada, Oklahoma. After Branson and Mehl, 1941, Gnathodus pustulosus, Pl. 5, Figs. 36, 37.
   7. From Delaware Creek shale: subsurface, Gulf Oil Corp. # 1 Riner well, 6513.5 feet, Ardmore basin, Oklahoma.
   8. From Paint Creek formation, Illinois; after Rexroad, 1957, *Gnathodus modocensis*, Pl. 1, Fig. 17a.
   9. From Glen Dean formation, Illinois; after Rexroad, 1958, *Gnathodus modocensis*, Pl. 1, Fig. 2.
   10, 11. From Barnett shale, Texas; after Hass, 1953, Pl. 14, Figs. 25, 26.
   12. From middle Goddard shale: Grindstone Creek, Ardmore basin, Oklahoma.

13-18. *Gnathodus* (*Harltonodus*) *bransoni*, new species.
   13. From basal Stanley shale, 120 feet above base: Caddo Gap, Arkansas. After Hass, 1951, *Gnathodus bilineatus*, Pl. 1, Fig. 1.
   14, 15. From Delaware Creek shale: 5 miles south of Ada, Oklahoma. After Branson and Mehl, 1941, *Gnathodus pustulosus*, Pl. 5, Figs. 32, 38.
   16, 17. From Barnett shale, Texas. After Hass, 1953, Pl. 14, Figs. 28, 29.
   18. From basal Goddard shale, 150 feet above base. Goddard Ranch, Oil Creek, Ardmore basin, Oklahoma. After Elias, 1956, *Gnathodus bilineatus*, Pl. 3, Fig. 25.

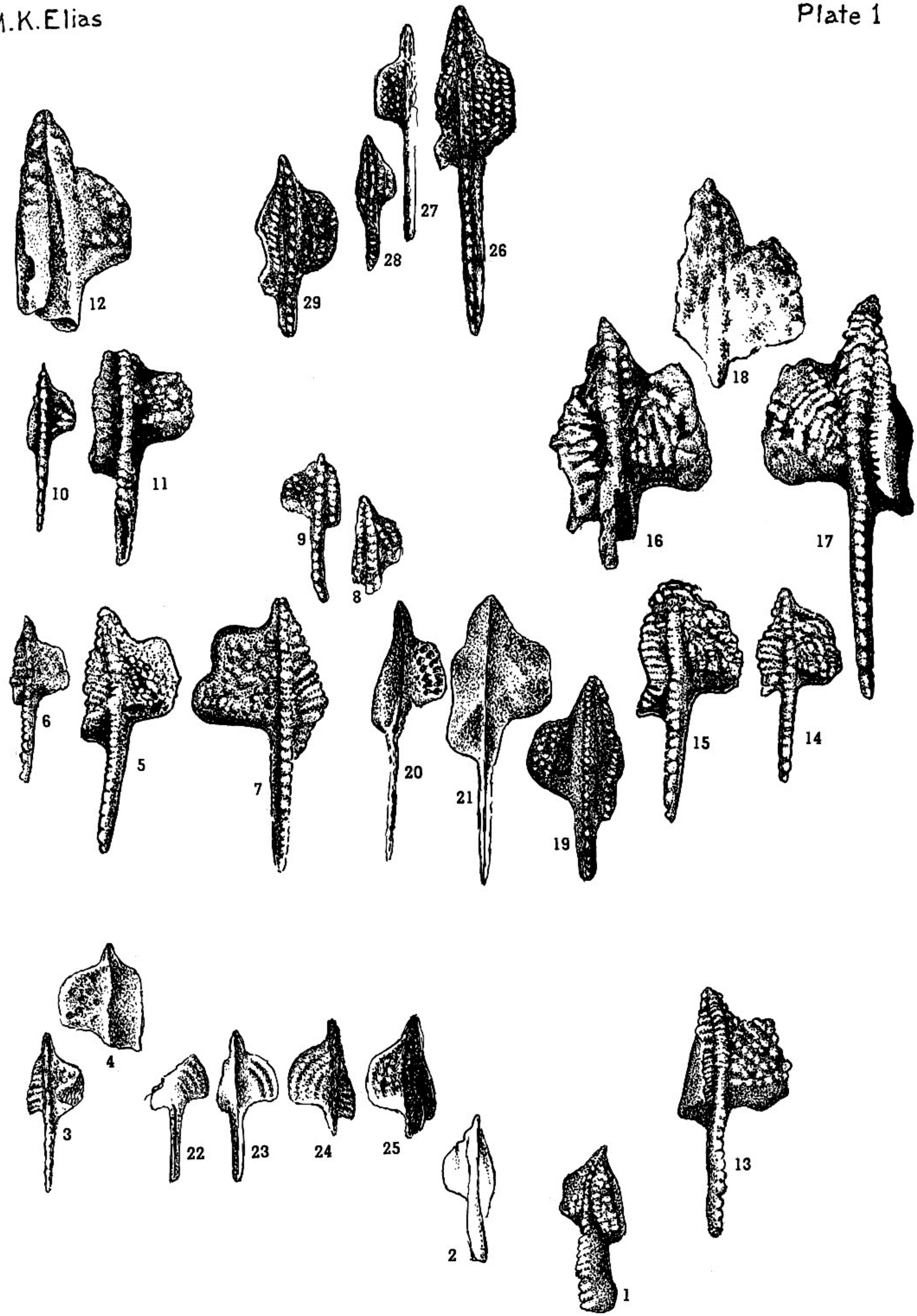
19-21. *Gnathodus* (*Harltonodus?*) *liratus*, Youngquist and Miller.
   19. Pella beds, Iowa. After Youngquist and Miller, 1949, Pl. 101, Fig. 15.
   20, 21. From shale south of Clayton, Oklahoma; correlated with Delaware Creek shale of Arbuckle Mountains.

22-25. *Gnathodus* (*Harltonodus*) *minutus*, new species. Lower Stanley shale, 475 feet above base: southwestern Potato Hills, Oklahoma.

26-28. *Gnathodus* (*Harltonodus*) *multilineatus* Elias. Upper Sand Branch shale, northern Arbuckle Mountains. After Elias, 1956, Pl. 3, Figs. 49, 51, 53.

29. *Gnathodus* (*Harltonodus*) cf. *multilineatus*. Lower Sand Branch shale, northern Arbuckle Mountains.

All magnifications 40X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Explanation of Plate 2

Figures

1-3. Neoprioniodus alatoideus Cooper, sensu lato.
- 1. From Woodford formation, northern Arbuckle Mountains, Oklahoma. After Cooper, 1931, Pl. 28, Fig. 1.
- 2. From upper New Albany shale, Indiana. After Huddle, 1934, Pl. 1, Fig. 5.
- 3. N. aff. alatoideus. From lower Stanley shale, 200 feet above base; southwestern Potato Hills, Oklahoma.

4. Neoprioniodus sp., of N. alatoideus group. Middle Stanley shale; 8 miles east of Clayton, Oklahoma.

5. Neoprioniodus sp., of N. ligo group. Middle Stanley shale; 8 miles east of Clayton, Oklahoma.

6, 7. Neoprioniodus scitulus Branson and Mehl. Delaware Creek shale; 5 miles south of Ada, Oklahoma. After Branson and Mehl, 1941, Pl. 5, Figs. 5, 6.

8-11. Neoprioniodus erectus Rexroad.
- 8. From Delaware Creek shale; Henry House Creek, Ardmore basin, Oklahoma.
- 9. From shale south of Clayton, Oklahoma; correlated with Delaware Creek shale of Arbuckle Mountains.
- 10, 11. From Renault formation, Chester series of Illinois. After Rexroad, 1957, Pl. 2, Figs. 23, 25.

12-14. Neoprioniodus ligo Hass. Barnett shale of Texas. After Hass, 1953, Pl. 16, Figs. 1-3.

15. Neoprioniodus rynikeri, new species of N. ligo group. Subsurface Delaware Creek shale, Gulf Oil Corp. # 1 Riner well, 2615 feet; near Overbrook, Ardmore basin, Oklahoma.

16. Neoprioniodus sp. of N. ligo group. Basal Springer group, Oil Creek, Ardmore basin, Oklahoma.

17-21. Neoprioniodus cassilaris (Branson and Mehl).
- 17-19. From Keokuk formation of Illinois. After Branson and Mehl, 1941-B, Pl. 6, Figs. 12, 16, 17
- 20, 21. From Vienna-Menard and Golconda formations of Illinois. After Rexroad, 1957, Neoprioniodus scitulus, Pl. 2, Figs. 22, 26.

22. Neoprioniodus cassilaris var. keokukensis, new variety. Keokuk formation of Illinois. After Branson and Mehl, 1941-B, Pl. 6, Fig. 15.

23, 24. Neoprioniodus miseri, new species.
- 23. From Delaware Creek shale, 160 feet above base. Henry House Creek, Ardmore basin, Oklahoma.
- 24. From shale south of Clayton, Oklahoma; correlated with Delaware Creek shale of Arbuckle Mountains.

25, 26. Neoprioniodus solidiformis Elias. Goddard shale, 150 feet above base, Oil Creek, Ardmore basin, Oklahoma. After Elias, 1956, Pl. 2, Figs. 28, 29.

27. Neoprioniodus cf. solidiformis Elias. Shale south of Clayton, Oklahoma; correlated with Delaware Creek shale of Arbuckle Mountains.

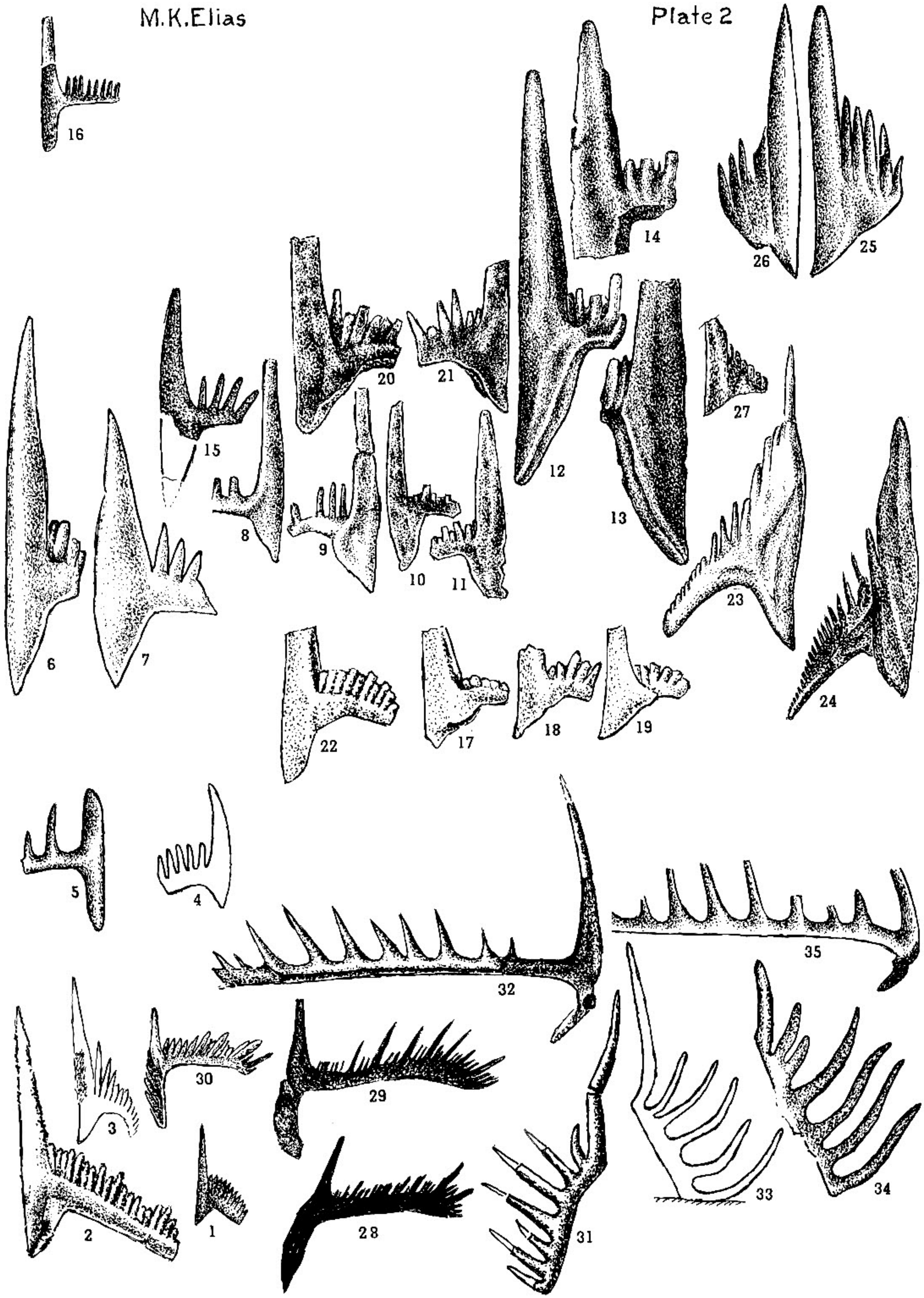
28-30. Hindeodelloides bicristatus Huddle.
- 28. From lower New Albany shale (Upper Devonian), Indiana.
- 29. From upper New Albany shale (Lower Mississippian), Indiana.
- 28 and 29 after Huddle, 1934, Pl. 7, Figs 2 and 3.
- 30. From lower Stanley shale, 475 feet above base; southwestern Potato Hills, Oklahoma.

31. Ligonodina cryptodens Huddle. End view of anticusp; lower New Albany shale, Indiana.

32. Ligonodina sp., Huddle. Lateral view of bar; upper New Albany shale, Indiana. 31 and 32 after Huddle, 1934, Pl. 12, Figs. 21 and 8.

33-35. Ligonodina sp. Lower Stanley shale, 475 feet above base; southwestern Potato Hills, Oklahoma.

All magnifications 40X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THE ATOKA OF THE McALESTER BASIN-ARKANSAS VALLEY REGION

B. J. Scull [1], G. D. Glover [1], and Roger Planalp [2]

Introduction

The Atoka strata in the McAlester Basin of eastern Oklahoma and the Arkansas Valley in northwestern Arkansas are the linkage between the Ouachita facies, the Ozark facies and the Arbuckle facies. They form an integral part of each of these facies, hence the history of the Atoka deposition of this region must be clearly understood before the structural and age relationships of these provinces can be established.

The McAlester Basin-Arkansas Valley synclinorium extends over 250 miles east-west and over 50 miles north-south. Within specific areas, lithologic zones of the Atoka can be correlated for many miles but any correlation attempt throughout the synclinorium would be meaningless. On our cross-sections we have arbitrarily designated lower, middle and upper Atoka zones. These designations are simply guides based on general configurations and should not be used as definitive correlations.

Although the fact that the bulk of the Atoka of eastern Oklahoma and western Arkansas was derived from an eastern or southeastern source area has been established for many years (Croneis, 1930), no graphic illustrations have been published. The primary purpose of this presentation is to graphically illustrate the sand distribution patterns in the Atoka of the synclinorium. The written discussion is limited to material which helps clarify concepts gained from the cross-sections or adds to the understanding of the regional setting of the Atoka.

The Atoka formation in eastern Oklahoma and western Arkansas represents a larger number of environments of deposition (as indicated by lithotopes) than any other formation in this region. This is because the Atoka lithologies are readily recognized in the Ouachita Mountains, in the synclinorium north of the Ouachitas and on the cratonal shelf areas from the Ozark Platform to the Hunton Arch. Several environments and types of sedimentation were present in each of these tectonic provinces during the Atoka deposition. Paleozoic sediments above and below the Atoka have a more limited areal distribution because of non-deposition (Hartshorne), unrecognized correlations (Morrow), or facies differences that warrant separate formational names (Viola-Bigfork). The Atoka sediments exhibit only minor lithologic variations even though they thin by depositional patterns from over 19,000 feet to a few feet in the outcrop area and were laid down in different tectonic environments - in an incipient orogenic belt (Ouachita), in a subsiding trough (McAlester Basin-Arkansas Valley), and on unstable cratonal elements (Ozark, Hunton).

This paper is specifically concerned with the Atoka unit in the McAlester Basin-Arkansas Valley synclinorium (the Arkansas Coal Basin of some writers). The Atoka is variously described as a formation or a series (Blythe, 1957). In this report the Atoka is regarded as a lithologic entity which unconformably overlies the Morrow and unconformably underlies the Hartshorne of the synclinorium. However, for purposes of description and continuity, the Atoka units in the Ouachita and cratonal zones will be considered.

The Border Rocks

The Atoka formation north of the synclinorium in eastern Oklahoma has been studied in detail by Blythe (1957). This work is

[1] Sun Oil Company, Richardson, Texas.

[2] Athletic Mining and Smelting Company, Ft. Smith, Arkansas.

essentially a northern extension of the studies of Wilson and Newell (1937) along the southern shelf area, and an analysis of the regional position of the Atoka. Henbest (1953) contributed to the understanding of the Atoka with his studies in the Ozark Highlands of Arkansas.

The material in the forementioned reports which is relevant to this presentation is summarized in the following paragraphs.

The Atoka sediments of the unstable cratonal shelf areas are mostly shallow marine sandstones and shales. However, marine limestone and conglomerate are present. In Oklahoma the conglomerates are characterized by chert pebbles and in Arkansas by quartz pebbles. Some of the Atoka of these areas is terrestrial and some was laid down below the base of wave action in the Atokan seas. The sands of the Atoka are fine-grained and sub-angular. Patches of medium to coarse-grained sand are irregularly distributed through the formation.

The younger Atoka units overlap the older units northward. The volume of sand has a relatively even distribution, which decreases northward at a considerably lesser rate than the volume of shale so that the sand-shale ratio approaches unity. These sediments were deposited on an eroded surface and were subjected to differential erosion before the overlying units were deposited.

The Atoka sands of the shelf area in Oklahoma form the Dutcher-Gilcrease zone of the subsurface. Apparently, these sands were for the most part derived from northern and western sources and did not spill over the shelf into the synclinorium. The shelf facies, except in broad zonation, is not correlatable with the basin facies. The lithologic patterns of the basin suggest that only middle and upper Atokan units were deposited on the shelf areas. This is depicted by the arbitrary subdivision lines on the accompanying cross-sections.

The Atoka formation in the Ouachita Mountains is little known and much disputed. The classical works are those of Croneis (1930), Harlton (1934), Hendricks and others (1947), Honess (1923, 1924), Miser (1934)and Miser and Purdue (1929). The distribution, even the definition, of the Atoka formation has been clouded by the contentions revolving around the stratigraphic and tectonic position of the Johns Valley shale. The various factors and interpretations will not be reviewed here because other papers in this symposium present the problem in detail. The solution of the Johns Valley problem is important to the study of the Atoka in that the base of the Atoka in the Ouachita Mountains cannot be established until the relationships of the Johns Valley are determined.

We would like to inject some factors generally ignored in the arguments. In the De-Queen and Caddo Gap quadrangles, south of the core of the Ouachitas, Miser and Purdue (1929) mapped units which they called Atoka. These units rest conformably on Jackfork sandstone and, lithologically, are identical with units mapped as Atoka in the southern part of the Arkansas Valley. Areally, the Atoka of the complexly faulted frontal Ouachitas in Oklahoma is minor compared to the essentially unmapped exposures of the Atoka in the frontal Ouachitas of Arkansas. In Arkansas, Reinemund and Danilchik (1957) measured over 19,000 feet of Atoka in the Waldron quadrangle; Scull and White (1952) measured over 14,000 feet of Atoka south of Rover where the upper part of the formation has been removed by erosion; and Croneis (1930) measured over 9,000 feet of Atoka in the section near Perryville where neither the top nor the bottom of the formation is exposed.

These relationships pose questions that must be answered before the regional position of the Atoka can be established. Are the Atoka units in the southern Ouachitas relatable to the Atoka of the frontal Ouachitas? Did the axis of maximum sedimentation progressively migrate northward from the time of deposition of the Arkansas novaculite (Middle and Upper Devonian and Lower Mississippian) until the upper Desmoinesian time (Weirich, 1953; Scull 1954) ? If so, where is the zone of maximum accumulation of the Johns Valley and its equivalents? Is the possibility, suggested by Croneis (1930, that the lower Atoka of the frontal Ouachitas was deposited contemporaneously with the upper Morrow more fact than fiction? How much, if any, of the Atoka was derived from older Ouachita sediments?

Several quadrangles in the Ouachitas must be mapped in detail before these questions can be adequately answered. However, the distribution of Atoka lithologies within the McAlester Basin-Arkansas Valley synclinorium aids somewhat in determining the regional setting of the Atoka.

General Lithology

The surface exposures of the Atoka within the synclinorium have been discussed in numerous reports which are listed under "References" and so will not be repeated here. For regional subsurface studies the reports of Caplan (1954, 1957) and Sheldon (1954) for Arkansas, and Weirich (1953) for Oklahoma are the most useful.

Lithologic descriptions of the Atoka tend to be monotonous. One finds that in Oklahoma the formation consists of dark shales, some silty, some micaceous, some carbonaceous, with intercalated gray, tan, brown and black siltstones, very fine-grained sandstones and fine-grained sandstones. Limonite, calcite and silica cements are common to the sandstone. With respect to volume, only traces of limestone and conglomerate occur. In Arkansas the section is different in that more medium-grained sandstones and traces of coal are present. These lithologies comprise over 1250 cubic miles of strata and have such lenticular or irregular distributions that only in rare cases can an individual bed be traced from well to well. However, as mentioned above, gross lithologic patterns can be carried both locally and regionally.

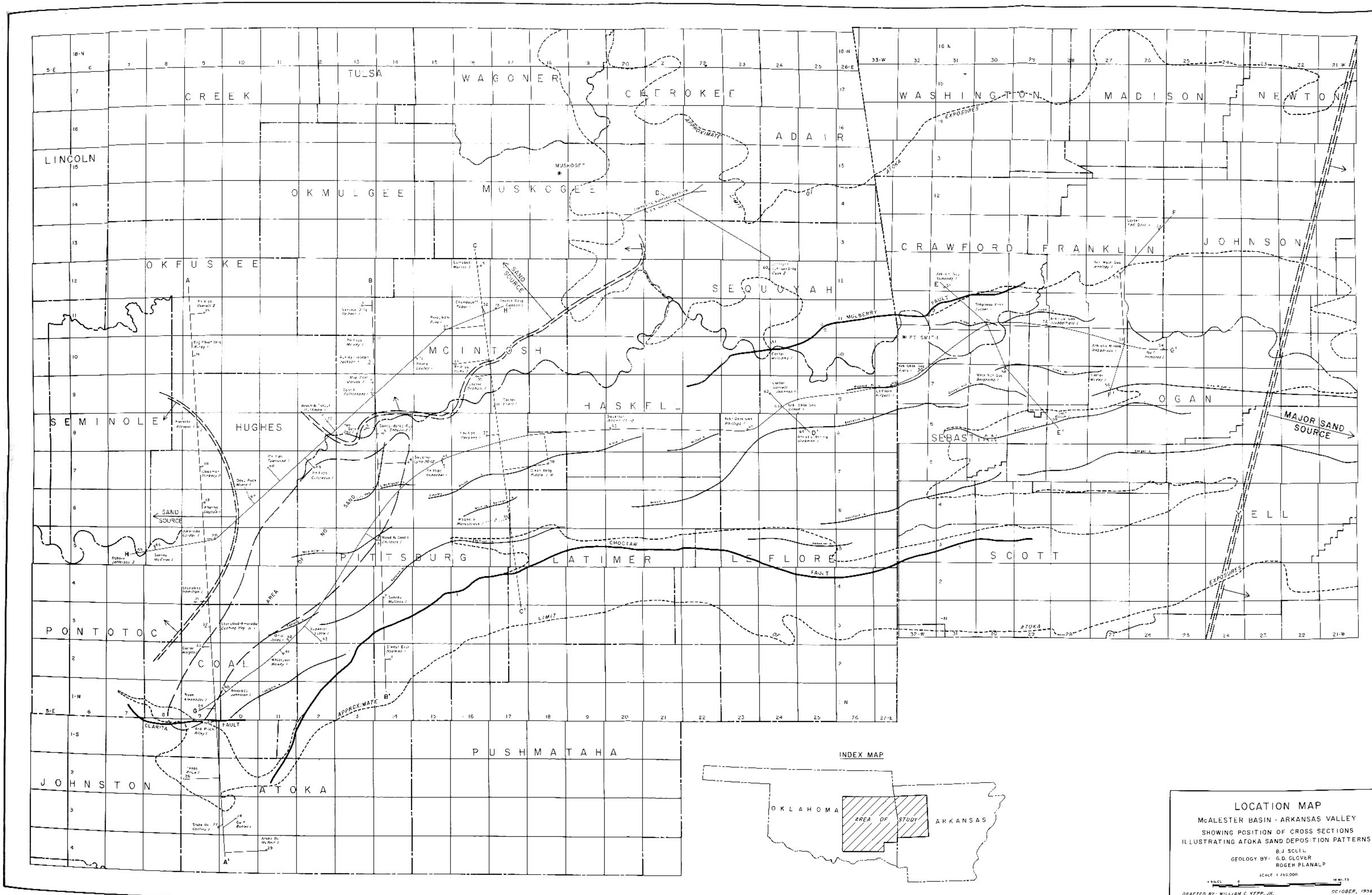
The gross lithologic characteristic is uniformity and the local lithologic characteristic is diversity. These characteristics, of course, are keys to petrologic classification of Atoka deposition. However, we will defer classification to a later section so that the regional distribution patterns are meaningful to the classification.

Tectonic Setting

The axis of the geosyncline in which the Atoka accumulated is marked by the extremely thick sections mentioned above. This axis of deposition is within the frontal Ouachita belt so that basin sediments discussed in this report were deposited on the north limb of the geosyncline and on the bordering shelf areas. The paleo-current indicators mapped by Reinemund and Danilchik (1957) and noted by Scull and White (1952), as well as the grain size increases recorded by Croneis (1930), show that the general current movement during the deposition of the Atoka was from east to west along the axis of the downwarp. Post-depositional deformation has distorted the directional properties of paleo-current indicators with the result that northern or southern directional components can not be reliably ascertained from the comparatively small areas which have been studied. If one reads the description of the Atoka in the Waldron Quadrangle (Reinemund and Danilchik, 1957) and accepts Pettijohn's (1957) concept of the graywacke suite, the Atoka of the frontal Ouachitas is automatically classified as bathyl or abyssal marine geosynclinal strata.

For those geologists who cannot accept coal and carbonaceous shales as logical components of an abyssal assemblage, an alternate interpretation is readily available. Prorating of geologic time yields one to three million years that might be alloted to the deposition of the Atoka. Using the smaller time interval for sake of argument, the Atoka of this thick section could have accumulated at an average rate of two feet per century or .2 of an inch per year in a topographically shallow, subsiding trough. Whether or not this is rapid deposition is a matter of definition. The rhythmic nature of the sediment pattern might be attributed to major climatic cycles.

The Atoka along the shelf areas described by Blythe (1957), Croneis (1930), Henbest (1953), and Wilson and Newell (1937), in Pettijohn's classification, are of the cratonic, neritic and intertidal suite. These rocks are situated tectonically much as they were at the time of deposition. The Atoka which crops out in the frontal Ouachitas was deposited in the deeper portions of the geosyncline and was later subjected to strong structural displacements. The frontal Ouachitas and the cratonal elements bound the McAlester Basin-Arkansas Valley synclinorium. The synclinorium is the downwarped north limb of the geosyncline. The major axis of the synclinorium (the hinge line of Weirich, 1953) is followed rather faithfully by the course of the Canadian River in Oklahoma and the Arkansas River in Arkansas. North of this axis the folds are open and have gentle dips; south of the axis folding was intense and dips are steep to overturned. The main axis of subsidence



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APPROXIMATE LIMIT OF ATOKA EXPOSURES
SAND AREA
INDEX MAP
OKLAHOMA
AREA OF STUDY
ARKANSAS
LOCATION MAP
McALESTER BASIN - ARKANSAS VALLEY
SHOWING POSITION OF CROSS SECTIONS
ILLUSTRATING ATOKA SAND DEPOSITION PATTERNS
GEOLOGY BY: B.J. SCULL
G.D. GLOVER
ROGER PLANALP
SCALE 1:250,000
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DRAFTED BY: WILLIAM C. KEPP, JR.
OCTOBER, 1958

has been displaced northward about 45 miles since Atoka deposition was initiated.

Tectonic activity within the synclinorium began before the end of Morrow time, and by the time the middle Atoka units were being laid down, structural features were sufficiently developed to affect sedimentation. Evidence for this is provided by Hendricks and Parks (1950) who noted large shale slivers in the middle Atoka sandstones. There was little tectonic activity during upper Atoka time. However, toward the close of the Atokan both the cratonal shelves and the Ouachita complex were subjected to uplift.

The Hartshorne, which unconformably overlies the Atoka, was not deposited on the shelf areas (Weirich, 1953) and perhaps is not represented in the southern part of the synclinorium (Reinemund and Danilchik, 1957). The limited distribution of the Hartshorne, which is mostly depositional rather than erosional, shows that the boundaries of the synclinorium had been defined and delimited by the end of Atoka time. There is insufficient information presently available to construct palinspastic maps to show the displacement of the southern boundary by the orogenic movements. The major orogenic compression was post-Atokan and crustal shortening was several miles at least.

Cross Sections Showing Lithologic Distribution

In a study of this type it is sometimes difficult not to dwell on the better known areas and postulate feebly about those areas of limited control. We have attempted to scatter our control points so that a maximum of information could be shown. The tectonic history of the region is such that the only feasible method of showing the distribution pattern in proper perspective is to use the top of the Atoka as the datum line. This has serious drawbacks in that the top of the Atoka as deposited can not everywhere be accurately determined. Also, with this approach, unconformities within the formation are ignored. We have deliberately refrained from indicating known or probable correlations because they are local in nature and distract from an attempt to develop regional patterns.

We have prepared eight cross-sections using data from over 200 wells, 70 of which appear on the cross-sections. Because of its length, the section G-G', along the main axis of the synclinorium, is divided into an east half (Fig. 9) and a west half (Fig. 8). This section depicts the gross lithologic pattern of the Atoka in the synclinorium while the other sections yield supporting evidence. Since isopach maps of the Atoka appear in recent reports (Weirich, 1953; Caplan, 1957) none were prepared for this report. Sand-shale ratio maps are not presented because in the few areas where optimum control is available the local lithotopes so dominate the configuration that regional continuity is not apparent.

On cross-section G-G' (Figs. 8 and 9) the following features may be noted. In the eastern part of the section the sands and shales of the Atoka were deposited rhythmically but not harmonically. Each major zone consists of a predominant sand section and a predominant shale section although each section is marked by local variations. A few of the more prominent zones are connected with lines on the cross-sections so that the rhythmic aspect is emphasized. In individual wells, including several not shown on the cross-section, the Atoka consists of 40 to 60 percent sand and coarse silt. These percentages are rather persistent westward in the Arkansas Valley although 40 percent is the more normal sand content.

Just west of the LeFlore County Gas Company # 1 Kilgore well, near the eastern border of Oklahoma, the lithologic make-up of the Atoka changes considerably. There the lower 1500 feet of the formation contains less than five percent of sand and coarse silt, in contrast to the much more sandy and apparently equivalent eastern section. The upper 4500 feet of the formation contains sand percentages comparable to the more eastern zones. The lower zone of very sparse sand persists westward and thickens upsection at an increasing rate so that in western Pittsburg, southeastern Hughes and northeastern Coal Counties, the Atoka as a whole is represented by over 5000 feet of shale, with essentially no sand content. The sand content of the upper Atoka in the vicinity of the Kilgore well changes noticeably west of the Arkansas-Oklahoma Gas Company # 1 Phillips in northwestern LeFlore County. When traced westward, the sand content is less than 25 percent in south central Haskell County, less than 10 percent in northern Pittsburg County, and as mentioned above, no sand is present in western Pittsburg County.

There are two features of note illustrated by cross-sections E-E' (Fig. 6) and F-F' (Fig. 7). One feature is the southward thickening of the Atoka and the other is the thickening of the shale zones between sand zones of comparatively uniform thickness. Hendricks and Parks (1950) have presented a reasonable analysis of the rates of thickening of the Atoka in the Arkansas Valley. Their values for the middle and upper parts of the formation are no longer completely valid because much more data have become available since their report was written. The rate of thickening of the lower Atoka as given by Hendricks and Parks must be almost doubled to fit the information now available. Our figures differ from those of Hendricks and Parks partly because the subdivisions are arbitrarily defined.

The Atoka thickens from 4500 feet near the Mulberry fault (see Fig. 1) to over 19,000 feet in the Waldron quadrangle almost fifty miles to the south. The total rate of thickening of the formation over this distance is roughly 290 feet per mile. However, since the rate of thickening is inconsistent, each zone must be treated separately.

Each set of wells used to calculate the rates of thickening in the Atoka give different results. Near average ratios for the northern part of the province can be obtained by using two wells on cross-section F-F' (Fig. 7), namely, the Carter Oil Company # 1 Federal Government and the Athletic Mining and Smelting Company # 1 Robberson. The # 1 Federal Government well is located near the northern margin of the synclinorium and the Robberson well is located near the main axis of the synclinorium. Thickness increments are given in hundreds of feet because precise measurements do not materially change the ratios, and dip control for thickness corrections ordinarily cannot be determined.

The Atoka increases in thickness from 2600 feet in the # 1 Federal Government well to 5900 feet in the # 1 Robberson (see cross-section F-F', Fig. 7). The rate of thickening for the formation is 183 feet per mile. Zonal increases are: upper part, 55.5 feet per mile, middle part, 88.8 feet per mile, and lower part, 38.8 feet per mile.

Southward from the main axis the gross rate of thickening is 405 feet per mile to the area of the Waldron quadrangle. Tentative regional correlations suggest that the rate of thickening is rather uniform south of the axis of the synclinorium. The generalized sketch below shows the thickening pattern of the Atoka in the Arkansas Valley. The change of slope probably occurs south of the vicinity of the Robberson well. However, there is the suggestion that the present northern limb of the synclinorium was part of the cratonal shelf during Atoka deposition and that the southern limb represents the "continental slope" of the Atoka geosyncline. The resistance of the cratonal margin to deformation localized the zone of maximum downwarp of the synclinorium.

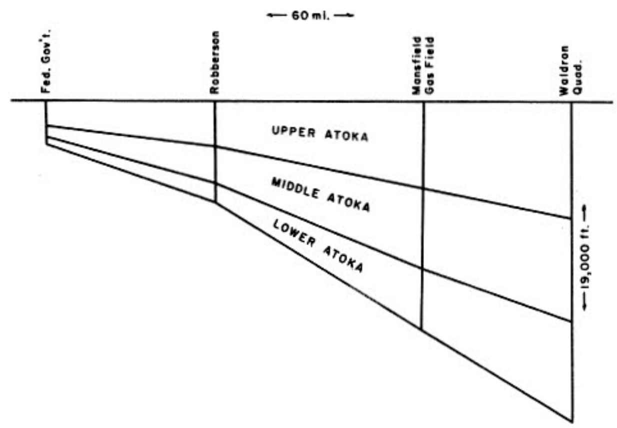

Fig. 11. - Generalized sketch of Atoka thickness in Arkansas Valley.

On cross-section D-D'(Fig. 5) it can be noted that the shale facies accruement shown on cross-section G-G'(Figs. 8 and 9) is restricted to the general area of the main axis of the synclinorium. The Athletic Mining and Smelting Company # 1 Hickman well is located on the Backbone structure which, as stated by Hendricks and Parks (1950), marks a zone of structural and sedimentation change in the Paleozoic sediments, particularly in the Atoka. The zone north of the # 1 Hickman well probably is the zone of cratonal resistance to the Ouachita deformation.

Whether the absence of the shale facies in the Carter Oil Company # 1 Williams well (cross-section D-D', Fig. 5) is the result of nondeposition or faulting is not known. Possibly the Mulberry fault zone was active and the Atoka was not deposited on the upthrown side. However, an almost identical Atoka section, with the shale facies missing, is present in the Athletic Mining and Smelting Company # 1 Bertha Dunn well a little over a mile southeast of the Arkansas-Oklahoma Gas Company # 1 Cloud well.

The Lohman and Johnson # 2 Cook (cross-section D-D', Fig. 5) is located on one of the major faulted prongs of the Ozark uplift. In the township east of this well the measured sections of Blythe (1957) show that the Atoka of this area does not contain a sufficient amount of chert or limestone pebbles to indicate an Ozark source. However, northeastward in Cherokee County and on into Kansas the Atoka contains material indicating that some if not most of the sediment washed off the Ozark uplift. The Atoka of these northern areas comprises the inner cratonal shelf facies and is lithologically distinct from the Atoka of the outer cratonal shelf and of the "continental slope" to the south. The outer cratonal shelf and "continental slope", which existed during Atoka time were later incorporated into the McAlester-Arkansas Valley synclinorium.

The lower shale facies as seen on section C-C' (Fig. 4) terminates northward in the vicinity of the Carter Oil Company # 1 Graham well. The # 1 Graham and the Phillips Petroleum Company # 1 Ruby wells are located on the extension of the Warner uplift as shown by Arbenz (1956) on the tectonic map of Oklahoma. Wilson and Newell (1937) have supplied the information that the Warner uplift and associated structures were moderately active during Atoka deposition. This activity was associated with Ozark movements but was coupled with encroaching Ouachita displacements.

The sand limit lines on cross-section C-C' (Fig. 4) reflect the sediment contribution of the cratonal elements to the Atoka seas. It is likely that the sand components received from the north do not extend south of the # 1 Graham well. This has not been proven, hence the sand limit lines were not terminated within the cross-section.

On cross-section B-B' (Fig. 3) the separation of the sand components derived from the cratonal areas is well illustrated. For pictorial purposes the sand limit lines crossing the Superior Oil Company # 36-12 Lytle well are restricted and those crossing the Sunray Oil Corp. # 1 Mullens well are expanded. The zones shown in the # 1 Mullens do contain appreciable amounts of siltstone and sandy shale as well as some sandstone.

The Southwest Exploration Company # 1 Hochman well (cross-section B-B', Fig. 3) is located south of the Choctaw fault. There was a minimum of ten miles of crustal shortening between the # 1 Hochman and the # 1 Mullens wells. Rather than ignore this shortening we have interpolated the sand facies northward to indicate our concept of the sand distribution pattern of this area. The complete dominance by the shale facies in the McAlester Basin makes an interesting contrast with the well developed sandstones of the shelf and Ouachita elements.

The shelf and slope configurations are well developed on cross-section B-B' (Fig. 3). Although there is no obvious structural feature which represents the zone of cratonal resistance, an area of change is indicated by formational patterns on the Geologic Map of Oklahoma (Miser, 1954). In the south-central part of McIntosh County the Senora, Stuart and Thurman outcrop patterns are sharply deflected north of the Canadian River. Ries (1954) has shown that there is little thickness variation of the post-Atoka formations just west of the deflection zone. Therefore, the deflection probably represents a flattening of the regional dip controlled by a subsurface flexure of minor vertical dimensions. This flexure is undoubtedly responsible for the change in course of the Canadian River. The flexure is the only indication of cratonal stability, as its southern border approximates the main axis of the synclinorium.

Ries (1954) in a cross-section, illustrates quite well the pre- and post-Atokan unconformities on the Atoka shelf area of Okfuskee County.

The three wells at the southern end of section A-A' (Fig. 2) are located in the complex where the Ouachita structural elements abut or override the Arbuckle structural elements. The Atoka was completely eroded from the area at the southern end of the section before the Cretaceous strata were deposited. The Gulf Oil Corporation # 1 Butler well presumably was drilled north of the Choctaw fault but the high sand content is suggestive of Ouachita facies. Comparison of the Atoka in the # 1 Bulter and in the State-Cameron # 1 Collins wells shows relationships similar to those observed by comparison of the Atoka in the # 1 Hochman and # 1 Mullens wells of cross-section B-B' (Fig. 3). The sites of deposition may have been several miles farther apart than the present locations.

North of the # 1 Collins well (cross-section A-A', Fig. 2), in the southern part of the McAlester Basin, the shale facies is dominant. The Clarita fault was active prior to Atoka deposition but apparently was inactive during much if not all of Atoka time. Later movements along this fault are related to the Arbuckle tectonics.

The Atoka shelf element is strongly developed north of the Carter Oil Company # 1 Morgan well (cross-section A-A', Fig. 2). The sand components were mostly derived from the Hunton and Lawrence uplifts and other exposed cratonal elements in Seminole, eastern Okfuskee, and Lincoln Counties.

Cross-section H-H' (Fig. 10) was compiled to show the irregular distribution of the Dutcher-Gilcrease sand series along the cratonal shelf of the Atoka.

Conclusions

The Atoka formation of eastern Oklahoma and western Arkansas was deposited in a geosyncline which had a wide northern limb extending to the cratonal arc, and reaching from the Ozark to the Hunton uplift. Little is known of the southern limb of the geosyncline. The axis of major deposition was displaced northward in post-Atoka time. The migration localized against comparatively stable cratonal elements. South of the zone of stabilization the Atoka strata were strongly deformed but north of the zone the folding was much less intense. The thickest known Atoka section is exposed in the frontal Ouachitas. How much lateral displacement of this section took place during the Ouachita orogeny is not known. The Atoka thins rather regularly from the frontal Ouachitas to the stabilized zone bordering the craton. Farther northward the thinning is quite irregular.

The paleo-current indicators in the Atoka of the frontal Ouachitas and the southern part of the Arkansas Valley show that most of the sediments were derived from the east. The sand distribution patterns also show that the bulk of the Atoka sand throughout the Valley was derived from the east. The progressive southward thickening of the shale zones between relatively uniform sand zones indicates that much of the shale was derived from south and southeast of the Arkansas Valley. In Oklahoma, the general absence of sand in the thick Atoka of the McAlester Basin is another indication of an eastern source for the sands. The Dutcher-Gilcrease sand series is obviously derived from the craton to the north, and forms only a minor part of the total volume of the Atoka.

By description, the lithology of the thickest Atoka section fits into the graywacke suite of Pettijohn (1957), and the Atoka of the "slope" fits into the subgraywacke suite. We hesitate to use the terms "flysch" and "molasse", respectively, for these rock units. The presence of coal and carbonaceous shales in the thicker sections and the general lack of coarse conglomerates and red shales place the Atoka too much out of phase with the type sections of these facies to warrant using them. Furthermore, the deformational history of the Ouachitas is too poorly known to make orogenic or tectonic sedimentary classifications.

References

Arbenz, J. Kaspar, 1956, Tectonic map of Oklahoma.

Blythe, Jack G., 1957, The Atoka formation on the north side of the McAlester basin: Ph. D. thesis, University of Oklahoma.

Branner, George C., and others, 1929, Geologic map of Arkansas: Arkansas Geol. Survey.

Caplan, William M., 1954, Subsurface geology and related oil and gas possibilities of northeastern Oklahoma: Arkansas Geol. and Conserv. Comm., Bull. 20.

Caplan, William M., 1957, Subsurface geology of northwestern Arkansas: Arkansas Geol. and Conserv. Comm., Inform. Circular 19.

Cline, L. M., 1955, Regional stratigraphy and history of the Ouachita Mountain area: Shale Shaker, vol. 6, pp. 26, 27.

Croneis, Carey, 1930, Geology of the Arkansas Paleozoic area with especial reference to oil and gas possibilities: Arkansas Geol. Survey, Bull. 3.

Dane, C. H., Rothrock, H. E., and Williams, James Steele, 1938, Geology and fuel resources of the southern part of the Oklahoma coal field, Pt. 3, the Quinton-Scipio district, Pittsburg, Haskell, and Latimer counties: United States Geol. Survey Bull. 874-C.

Harlton, Bruce H., 1934, Carboniferous stratigraphy of the Ouachitas with special study of the Bendian: Amer. Assoc. Petroleum Geologists Bull., vol. 18, pp. 1018-1049.

Henbest, Lloyd G., 1953, Morrow group and lower Atoka formation of Arkansas: Amer. Assoc. Petroleum Geologists Bull., vol. 37, pp. 1935-1953.

Hendricks, T. A., 1937, Geology and fuel resources of the southern Oklahoma coal field, Pt. 1, The McAlester district, Pittsburg, Atoka, and Latimer counties: United States Geol Survey Bull. 874-A.

Hendricks, T. A., 1939, Geology and fuel resources of the southern Oklahoma coal field, Pt. 4, The Howe-Wilburton district, Latimer and LeFlore counties: United States Geol. Survey Bull. 874-D.

Hendricks, T. A., Gardner, L. S., Knechtel, M. M., and Averitt, Paul, 1947, Geology of the western part of the Ouachita Mountains of Oklahoma: United States Geol. Survey, Oil and Gas Inv. Prel. Map 66.

Hendricks, T. A., and Parks, Bryan, 1950, Geology of the Ft. Smith district, Arkansas: United States Geol. Survey Prof. Paper 221-E.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Bull. 32.

Honess, C. W., 1924, Geology of southern LeFlore and northwestern McCurtain counties, Oklahoma: Oklahoma Geol. Survey, Circular 3.

Huffman, George G., 1958, Geology of the flanks of the Ozark uplift: Oklahoma Geol. Survey, Bull. 77.

Knechtel, M. M., 1937, Geology and fuel resources of the southern Oklahoma coal field, Pt. 2, The Lehigh district, Coal, Atoka, and Pittsburg counties: United States Geol. Survey Bull. 874-B.

Knechtel, M. M., 1949, Geology and coal and natural gas resources of northern LeFlore County, Oklahoma: Oklahoma Geol. Survey, Bull. 68.

Miser, Hugh D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles, Arkansas: United States Geol. Survey Bull. 808.

Miser, Hugh D., 1934, Carboniferous rocks of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists Bull., vol. 18, no. 8, pp. 971-1009.

Miser, Hugh D., and others, 1954, Geologic map of Oklahoma: Oklahoma Geol. Survey and United States Geol. Survey.

Moore, Carl A., 1947, The Morrow series of northeastern Oklahoma: Oklahoma Geol. Survey, Bull. 66.

Oakes, Malcom C., and Knechtel, M. M., 1948, Geology and mineral resources of Haskell County, Oklahoma: Oklahoma Geol. Survey, Bull. 67.

Pettijohn, F. J., 1957, Sedimentary rocks: Harper and Brothers, New York.

Reinemund, John A., and Danilchik, Walter, 1957, Preliminary geologic map of the Waldron quadrangle and adjacent areas, Scott County, Arkansas: United States Geol. Survey, Oil and Gas Inv. Map OM 192.

Ries, Edward R., 1954, Geology and mineral resources of Okfuskee County, Oklahoma: Oklahoma Geol. Survey, Bull. 71.

Scull, B. J., and White, K. D., 1952, Private report to the Continental Oil Company.

Scull, B. J., 1954, Northern Arkansas: oil and gas possibilities: World Oil, Sept. and Oct.

Sheldon, Mary G., 1954, Sample descriptions and correlations for selected wells in northern Arkansas: Arkansas Geol. and Conserv. Comm., Inform. Circular 17.

Weaver, O. D., Jr., 1954, Geology and mineral resources of Hughes County, Oklahoma: Oklahoma Geol. Survey, Bull. 70.

Weirich, Thomas E., 1953, Shelf principle of oil origin, migration and accumulation: Amer. Assoc. Petroleum Geologists Bull., vol. 37, no. 8, pp. 2027-2045.

Wilson, C. W., Jr., and Newell, N. D., 1937, Geology of the Muskogee-Porum district, Muskogee and McIntosh counties, Oklahoma: Oklahoma Geol. Survey, Bull. 57.

LATE MISSISSIPPIAN - EARLY PENNSYLVANIAN STRATIGRAPHY OF THE OUACHITA MOUNTAINS, OKLAHOMA

L. M. Cline [1] and O. B. Shelburne [2]

Introduction

Objectives Of This Paper

This paper describes some of the stratigraphic features of the Ouachita geosyncline of Oklahoma as they are believed to have been during late Mississippian and early Pennsylvanian time. There is considerable evidence that pre-Atoka sediments were deposited in rather deep water. The lithologic characteristics of the late Mississippian and early Pennsylvanian Stanley-Jackfork-Johns Valley-Atoka stratigraphic sequence are comparable to the typical black shale flysch facies of the Eocene of the Alps and the Eocene and Cretaceous of the Carpathians. The conclusion is reached that a predominately deep water black shale and radiolarian chert environment was periodically interrupted by turbidity currents flowing down the steep sides of the depositional trough and that these currents brought in sands foreign to the black shale environment. The presence of convolute bedding, graded contacts of sandstones and overlying shales, of abundant flow casts and groove casts on the under surfaces of the sandstones, the general lack of cross bedding and ripple marks, and the scarcity of fossils except for planktonic and nektonic forms, supports this thesis.

The Ouachita Facies

The arcuate pattern of folds which comprises the Ouachita Mountains of southeastern Oklahoma and southwestern Arkansas is but one portion of a sinuous foldbelt which extends from western Texas eastward to within approximately 60 miles of the buried extension of the southern Appalachian Mountains. The foldbelt is now largely buried beneath relatively undisturbed Mesozoic and Cenozoic rocks but it is exposed in two arcuate uplifts, the Marathon uplift of Tran-Pecos Texas and the Ouachita Mountains. The Ouachita facies has been encountered in the subsurface by numerous deep wells which start in Mesozoic or Cenozoic rocks (Morgan, 1952, pp. 2266-2274). Notwithstanding the fact that the Ouachitas and Marathons are several hundred airline miles apart, and much farther when measured along the winding course of the folds, the two geographic provinces have so many things in common that their rocks are collectively referred to a Ouachita facies, in contrast to shelf or platform types of sediments which comprises the Arbuckle facies to the north or west, as the case may be. The stratigraphic column within the foldbelt also shows some vertical variations in facies, the pre-Stanley section being relatively thin (in fact, it may be considerably thinner than the Arbuckle facies) and being characterized by cherts and dark graptolitic shales, the Stanley-Jackfork-Johns Valley-Atoka sequence being much thicker and containing much sandstone. In the minds of some geologists the pre-Stanley section with the cherts and novaculites constitutes the Ouachita facies, whereas, in the opinion of others, the abnormally thick shale and sandstone sequence of the Stanley and later rocks characterizes the Ouachita facies. We regard all portions of the Ouachita stratigraphic column as belonging to the Ouachita facies because at all levels the lithology differs appreciably from the Arbuckle facies, although there are transitional zones coinciding approximately with the faulted belt of the frontal Ouachitas. The earlier Ouachita

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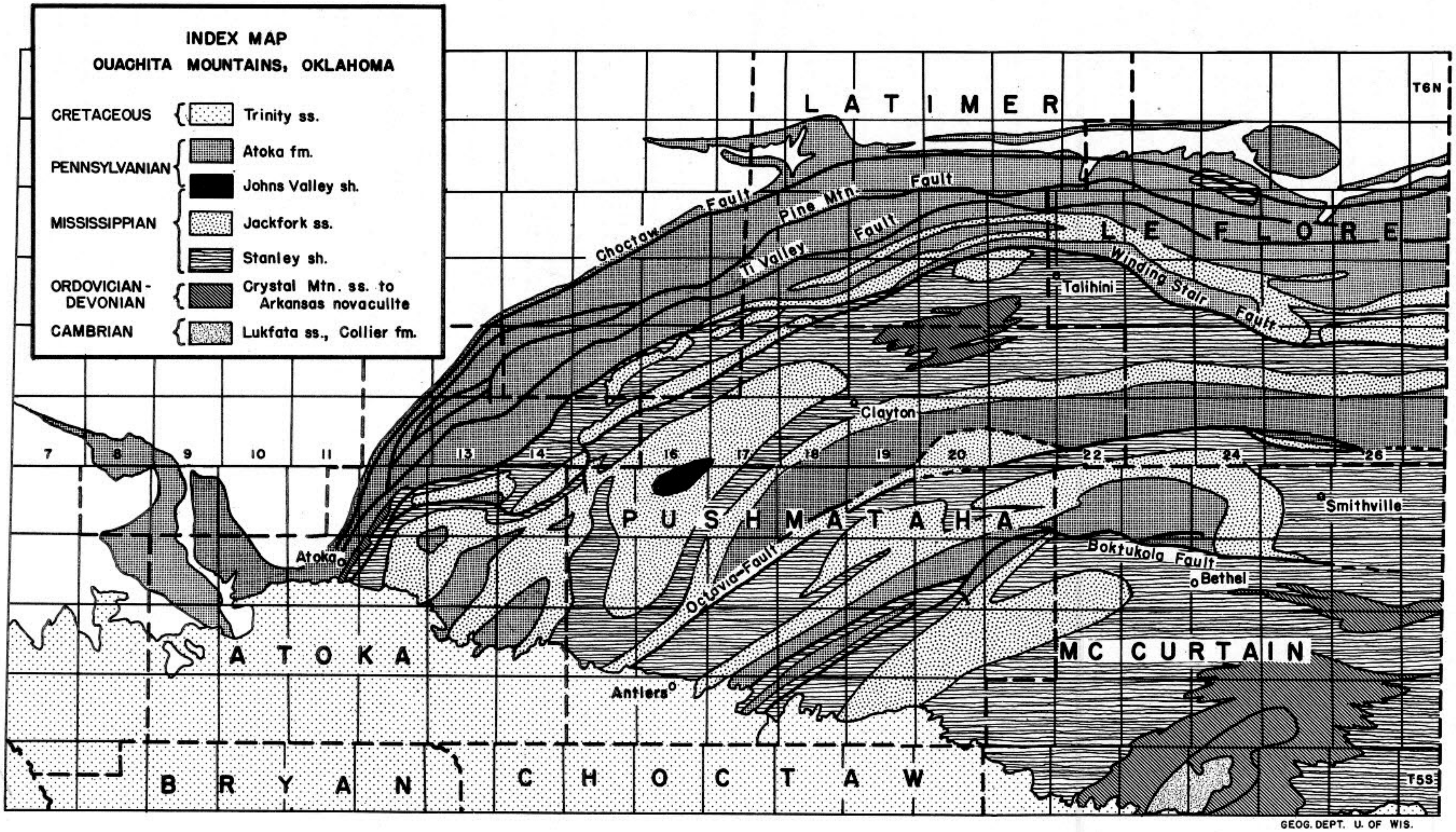

Fig. 1. -- Index Map, Ouachita Mountains, Oklahoma.

deposits seem to represent a period of very slow sedimentation in essentially a starved trough and with the post-Arkansas novaculite sequence representing a much more rapid period of sedimentation during a period of more active tectonism.

Arcuate pattern of the Ouachitas and Marathons. - Sedimentation appears to have been initiated in both the Ouachita and Marathon areas in deep arcuate troughs. One is tempted to suggest that the arcuate troughs were in reality deep sea trenches lying at the south margin of a stable continental area which supported the shallow seas in which the miogeosynclinal Arbuckle carbonate-bearing facies accumulated. If such was the case, we should then note that the Marathon and Ouachita arcs were convex toward the continent. However, we should not overlook another possibility; the possibility that a stable platform area existed to the south of the Ouachita-Marathon geosynclinal belt and the possibility that deep drilling south of the present foldbelt will encounter relatively undisturbed rocks of shallow water origin, but not necessarily belonging to the Arbuckle facies.

Relatively flat-lying fossiliferous rocks of early Paleozoic age are known to occur beneath Mesozoic and Cenozoic rocks in nearly 50 wells in the coastal plain of southeastern Georgia and northwestern Florida (Applin, 1951, pp. 1-28). It is noteworthy that this area of relatively undisturbed Paleozoic rocks lies southeast of the crystalline rocks that constitute the Piedmont province. Several recent workers (i.e., King, 1950) regard the Piedmont as including metamorphosed Paleozoic sediments, if indeed the bulk of the metamorphics are not Paleozoic. It is thus entirely possible that Paleozoic rocks exist south of the Ouachita foldbelt and that ultimately they will prove to be connected with the Paleozoics southeast of the Piedmont in Georgia and Florida. To the writers' knowledge, the relationship of the southern Appalachians to the eastern Ouachitas has not been definitely established, neither by geological nor geophysical methods. One cannot help but speculate as to whether they meet approximately at right angles in the manner that the arcuate Palau and West Caroline trenches of the western Pacific or whether the two join by virgation, with the eastern Ouachitas curving southward to parallel the buried southern Appalachians.

Siliceous sediments and graptolitic shales of the older Paleozoics. - The Ouachita facies contains some characteristic rock types that contrast strongly with the lithology of the Arbuckle facies, with the gradation taking place in a relatively narrow belt marking the junction of the northern and western margin of the Ouachita geosyncline and the south and east margin of the platform or shelf. Thin but laterally persistent cherts, such as the Big Fork and Arkansas novaculite, characterize the early Paleozoic rocks of the Ouachita geosyncline, whereas their equivalents on the platform, the Viola and Hunton are limestones.

Dark graptolitic shales, which occur separately and also interbedded with siliceous sediments, are perfectly at home in the Ouachita environment, although they also occur in shales and in fine-grained carbonates (some horizons of the Viola and Bromide) in the Arbuckle facies.

At higher stratigraphic levels black siliceous shales such as the Chickasaw Creek, and Wesley, some including thin beds of radiolarian chert, occur at numerous horizons as high as the lower Atoka.

Black shale flysch facies of the late Mississippian and early Pennsylvanian Stanley-Jackfork-Johns Valley-lower Atoka sequence.- In the heart of the Ouachita Mountains in southeastern Oklahoma, late Mississippian and early Pennsylvanian strata embraced in the Stanley-Jackfork-Johns Valley-Atoka sequence have an aggregate thickness of approximately 22,000 feet. The rocks are almost entirely of clastic origin with shale predominating in the Stanley, sandstone being most important in the Jackfork, and with the two rock types being in subequal proportions in the Atoka. Carbonate rocks are noticeably absent in this upper part of the Ouachita section, whereas they are fairly common in its stratigraphic equivalents in the platform to the north and west. The Ouachita equivalents of this part of the stratigraphic column are substantially thinner on the platform area of northeastern Oklahoma.

The most characteristic lithologic feature of the Stanley-Jackfork stratigraphic sequence is the repeated alternation of dark shales and gray sandstones. Details of lithology and sedimentary structures closely resemble the black shale flysch facies of the Alps and Carpathians of Europe.

Progress Of This Investigation

In the summer of 1953 Cline began a study of the late Paleozoic strata of the central Ouachita Mountains under the sponsorship of Dr. C. W. Tomlinson and the work has continued during the summer months and vacation periods since that time. Cline's initial interest in the area stemmed from his interest in Mississippian and Pennsylvanian stratigraphy and in the belief that the Ouachita geosyncline should contain strata transitional to the two systems. Mapping by Cline has shown that most of Harlton's map units (Harlton, 1938) of the Tuskahoma syncline are also present in the Lynn Mountain syncline northeast of Antlers and that they extend almost to the Arkansas line, some 70 airline miles to the east (Cline, 1956a, p. 427). Persistent beds were traced principally by surface mapping aided by aerial photographs. In 1956 Cline and Moretti described two complete and unfaulted stratigraphic sections of the Jackfork sandstone near the east end of the Kiamichi Range and showed several of Harlton's map units to be present. Shelburne mapped the Boktukola syncline in the summers of 1956 and 1957 and his mapsheet is reproduced as Plate 2 of this paper.

At this writing that portion of the Lynn Mountain syncline lying west of Oklahoma Highway 2 (which crosses the syncline between Clayton and Nashoba Junction) and extending west-southwest to Antlers has been mapped in some detail by Cline (Plate 1 of this paper) and reconnaissance mapping has extended eastward almost to the Arkansas line. South of the Lynn Mountain syncline the next major syncline is the Boktukola, which syncline contains a Stanley-Jackfork-Atoka sequence that correlates with a similar sequence in the Kiamichi Range (Cline, 1956b, pp. 50-53). Shelburne's mapping has proved the persistence of several stratigraphic markers in the Boktukola syncline east of Little River and he has successfully traced them into Rod Johnson's (Johnson, 1954) map area in the syncline west of Little River.

Acknowledgements. - We extend our sincere thanks to Dr. C. W. Tomlinson for sponsoring the work and for his continuing interest in the geology of this area. He has reviewed many of the more controversial problems in the field with us, has offered many helpful suggestions, but he has at no time sought to impose his own ideas on us, rather allowing us complete freedom to do the type of work that seemed most expedient at each stage of the project. Dr. M. K. Elias has contributed substantially to our work, principally through the identification of fossils. Mr. Bruce Harlton was kind enough to aid us in a restudy of some of his type sections in the Prairie Mountain and Lynn Mountain synclines and in checking some of our correlations along the Indian Service road. Dr. Carl Branson, Director of the Oklahoma Geological Survey, has encouraged us in our work, has reviewed some of the field work on two occasions, and the Oklahoma Survey defrayed some of Shelburne's field expenses. Dr. Thomas A. Hendricks, who has done so much map work in the frontal Ouachitas, spent several days in the field with Cline in the summer and fall of 1958 and has made some very helpful suggestions; we appreciate the encouragement he has given us. Several men interested in Ouachita geology have accompanied us in the field at various times and made helpful observations; among them are: Dr. Norman Williams, Director of the Arkansas Geological Survey, Mr. Hugh Miser and Dr. Chalmer Cooper of the United States Geological Survey, Mr. Horace Griley, Mr. B. W. Miller, Dr. Allan Bennison, Dr. Norman Johnson, Mr. E. C. Parker, Mr. William Hilseweck, Dr. Daniel Feray, Dr. William Jenkins, Mr. H. A. Sellin, and others.

Stratigraphy Of The Stanley-Jackfork-Johns Valley-Atoka Stratigraphic Sequence

The rocks which crop out to form the main portion of the central Ouachita Mountains of southeastern Oklahoma belong principally to the Stanley shale, Jackfork sandstone, Johns Valley shale, and Atoka sandstone formations. This stratigraphic succession of late Mississippian and early Pennsylvanian age has an aggregate thickness of nearly 22,000 feet in contrast to the thin pre-Stanley section (about 5,000 feet) exposed in the Choctaw anticlinorium in the "core" of the Ouachitas. The Stanley and Jackfork are referred to the Pennsylvanian system by most workers but Cline (1956a, p. 428; 1956c, p. 103) has published the opinion that both formations should be assigned to the Mississippian. The evidence for the Mississippian age of the Jackfork is that the overlying Johns Valley shale contains abundant Mississippian Caney fossils in its lower portion. Some workers have contended that fault slices of Caney were brought up from an underlying Arbuckle facies. Cline has pointed to the widespread occurrence of Mississippian fossils, seemingly always at about the same stratigraphic position, and believes that a portion of the Mississippian Caney intertongues with the lower Johns Valley.

The stratigraphic section with which we are dealing is given in the correlation chart, Table 1.

Stanley Group

The Stanley group consists of perhaps 10,000 feet of strata, prevailingly shaly but with sandstone becoming prominent in the upper 1,500 feet. The exact thickness is not known because there is no complete well-exposed and undisturbed stratigraphic section. This thick body of shale is weak, both from the standpoint of topographic expression and structural competence. It is the great valley maker of the Ouachitas, characteristically forming the flat floors of long, linear anticlinal valleys such as the Stanley and McGee valleys.

There are so few key beds in the entire 10,000 feet of Stanley that stratigraphic

Table 1. -- Mississippian-Pennsylvanian Stratigraphic Section

Central Ouachita Mountains, Oklahoma

System	Series	Group or Formation	
Pennsylvanian	Atokan	Rocks of Atoka lithology	
	Morrowan	Johns Valley shale	
Mississippian	Chesterian	Jackfork ss.	Game Refuge sandstone
			Wesley siliceous sh.
			Markham Mill fm.
			Prairie Mountain fm.
			Wildhorse Mountain fm.
	Meramecian	Stanley sh.	Chickasaw Creek siliceous sh.
			Moyers fm.
			Ten Mile Creek sh.

partition is difficult. This is particularly true of the lower portion of the group which, according to Honess (1923), has only one persistent recognizable stratigraphic marker (a volcanic tuff) in the lower 3,000 feet in the southern Ouachitas. In the central Ouachitas the Stanley contains several relatively thin but persistent dark blue-gray to black siliceous shales which are good key beds in mapping. In the western and northwestern portions of the frontal Ouachitas the Stanley thins rapidly and seems to be represented by a few feet (or at the most, a few hundred feet) of strata in the lower part of the Mississippian Caney shale. This spectacular thinning may result from overlap at the base of the Stanley; T. A. Hendricks (1947) is of the opinion that the Stanley rests unconformably on the Woodford chert (upper Arkansas novaculite equivalent) in the western Ouachitas. However, Cline is of the opinion that most, if not all, of the thinning is depositional and that there is rapid westward convergence at all stratigraphic levels within the Stanley.

Making use of the distinctive and persistent dark siliceous shales as stratigraphic boundaries, Harlton (1938) divided the Stanley group into three formations; named in upward order they are the Ten Mile Creek, Moyers, and Chickasaw Creek formations.

Ten Mile Creek Formation

Embracing the greater part of the Stanley group, the Ten Mile Creek formation measures in excess of 5,650 feet in its type section at the south end of the Tuskahoma syncline (Harlton, 1938, p. 864) where its base is hidden by overlapping Cretaceous strata.

Fig. 2.--Aerial mosaic showing south end of Tuskahoma syncline and edge of Gulf Coastal Plain. Wildhorse Mountain comprises the south part of the syncline; the type section of the Ten Mile Creek formation lies at base of mountain and immediately south.

Lithologically the formation consists of laminated shales and platy siltstones interbedded with thin sandstones. Sandstone becomes increasingly important upward in the section but is subordinate to shale in any considerable part of the formation. The shales are dark gray when fresh but a moderate amount of weathering produces a fissile olive-green product. The dirty, poorly sorted sandstones are dark gray when fresh but they become successively green-gray, yellowish, and in some cases buff, as weathering becomes more advanced. The gray shales are devoid of megascopic fossils but probably contain spores and other microscopic plant materials.

The best stratigraphic markers in the Ten Mile Creek formation, and for that matter throughout the Stanley and Jackfork groups, are some hard blue-gray siliceous shales. These siliceous shales carry a marine microfauna consisting of conodonts, radiolaria, and sponge spicules. Honess (1923, p. 152) called attention to a 25-foot zone of black cherty shales and thin-bedded cherts in the middle of the Stanley, stating that this is a well defined formation which extends from Little River eastward almost to Arkansas. Later Miser and Honess (1927, p. 11) proposed the name Smithville chert lentil for these cherts. Shelburne believes the Smithville chert to be equivalent to

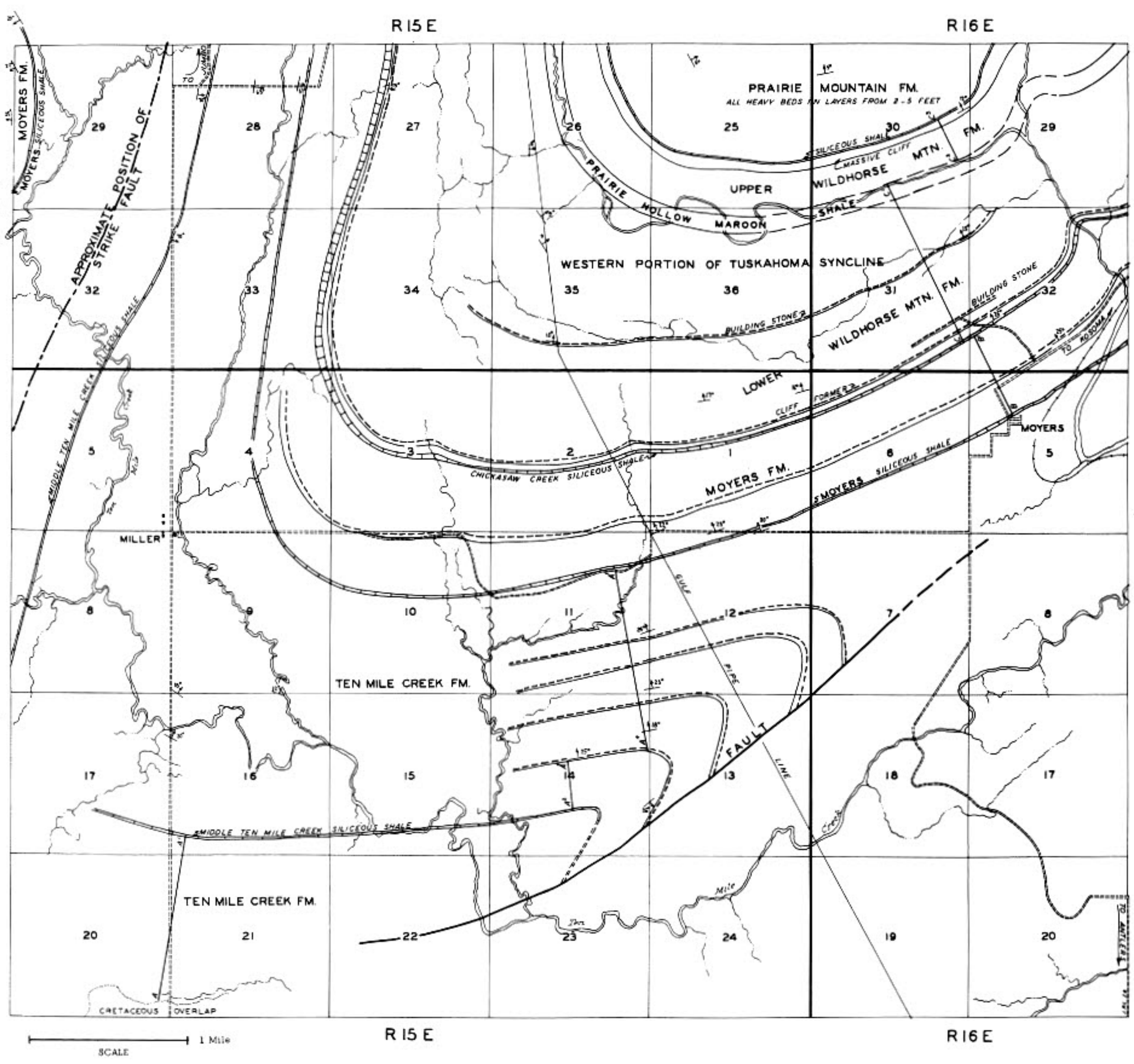

Fig. 3. -- Geologic map of type localities of Ten Mile Creek and Wildhorse Mountain formations (After Harlton, 1959, this symposium.)

Harlton's (1938, p. 868) Middle Ten Mile Creek siliceous shale of the central and western Ouachitas. The persistence of this chert has been proved by the mapping of Harlton in the western Ouachitas, Willis (1956) and McCollough (1954) in the central Ouachitas, and Honess (1924) and Shelburne in the southern Ouachitas.

In the southern Ouachitas the Hatton volcanic tuff, the base of which lies 270 feet above the base of the Stanley, attains a thickness of 90 to 228 feet and Honess' (1923) map depicts it as cropping out in narrow bands along sharp folds throughout an area of considerable size about the Choctaw and Cross Mountains anticlinoria. In addition to the Hatton tuff, Cline has seen two thinner beds on the north side of the Choctaw anticlinorium that he thinks are tuffs, and Shelburne has seen two tuffs, and possibly four depending on structural interpretation, in the lower Stanley on the north flank of the Cross Mountains anticlinorium.

Moyers Formation

At the type locality of the Moyers formation, in the lower slopes of Wildhorse Mountain, just north of the village of Moyers in T. 2 S., R. 16 E., the formation consists of 1,100 feet of alternating sandstones and shales. The sandstones form massive beds as thick as 20 feet (occasionally more), and are essentially devoid of bedding, are poorly sorted, and weather spheroidally. The shales are soft, thinly bedded, easily weathered, and vary in color from dark gray to green-gray. The alternation of shales and sandstones produces conspicuous rock benches and terraces which rise prominently above the valley-forming shales of the Ten Mile Creek formation to form the lower slopes of the synclinal mountains.

The boundaries of the formation are distinct and ordinarily well exposed, the top being defined by the overlying siliceous and cherty shales of the Chickasaw Creek formation, the base being sharply set off from the underlying Ten Mile Creek by the Moyers siliceous shale, a 20-foot bed of very dark hard siliceous shale (Harlton, 1938, p. 870). The Moyers siliceous shale is dark blue-gray to green-gray, depending on degree of weathering, thinly bedded but weathering as a unit to produce a low ridge that is ordinarily easy to recognize and map, although the outcrop is usually at the margins of valley flats. At first glance the siliceous shale appears to be unfossiliferous but a hand lens examination shows numerous conodonts and high magnification reveals the presence of radiolaria. Harlton has previously noted that this excellent key bed appears to have great lateral persistence, and we have extended the known outcrop eastward almost to the Arkansas line. East of the settlement of Ludlow in the central Ouachitas the Moyers siliceous shale consists of 6 feet of dark gray cherty shale which is intruded by sandstone dikes.

Chickasaw Creek Shale

Lying stratigraphically above the terrace-forming sandstones and shales of the Moyers formation is the Chickasaw Creek formation consisting of 130 to 300 feet of dark blue-gray shales with some thin-bedded black siliceous shales and cherts. Notwithstanding the fact that the Chickasaw Creek is thin, it is widespread throughout the Ouachitas and is the best stratigraphic marker in the entire Stanley-Jackfork succession.

Thin-bedded dark blue-gray shale is the predominating rock type in the formation but there are some evenly bedded quartzitic siltstones and sandstones. The most distinctive rock type is black siliceous shale, with intercalated black chert. The chert occurs in even beds ranging in thickness from one-half inch to six inches, is dark gray to black and is mottled with white globular to almond-shaped areas of white silica ranging from a fraction of a millimeter to several millimeters in diameter.

A microscope examination of the black siliceous shales shows them to contain abundant radiolaria and a profusion of sponge spicules. Individual laminae in some of the chert beds are a microcoquina of tiny monaxon and tetraxon sponge spicules. The distinctive white globules referred to above are clearly visible to the naked eye and obviously are much larger than radiolaria. Harlton (1938) has suggested that radiolaria served as nuclei about which silica was deposited. It seems that there are all gradations of these concentrations of white silica from spherical globules to almond-shaped masses flattened parallel to thin tabular masses interlaminated with black shale and chert. The structure, texture, and fabric argue strongly for a syngenetic or penecontemporaneous origin for the silica. Pure, clear opaline silica appears to have been deposited concurrently with black organic-rich muds which also were rich in radiolarian tests and mechanically concentrated sponge spicules.

Jackfork Group

In its typical development the Jackfork group consists of 5,600 to 5,700 feet of alternating sandstones and dark gray shales with minor amounts of black siliceous shales which include some thin chert beds. Sandstone is the prevailing rock type, hence the Jackfork is more resistant to erosion than the underlying Stanley shale. Taff (1902) named the Jackfork for Jackfork Mountain in the frontal Ouachitas. Harlton (1938) classified the Jackfork as a group and divided it into four formations, which named in ascending order are: the Wildhorse Mountain, Prairie Mountain, Markham Mill, and Wesley. He excluded the upper 400 to 500 feet of Taff's Jackfork from the group in the mistaken belief that it is a correlative of the Union Valley sandstone of the area near Ada, where that sandstone has a Morrow fauna. We are restoring this sandstone, now called the Game Refuge sandstone by Harlton, to the Jackfork group.

Harlton made the significant observation that several black siliceous shales in the Jackfork are widespread in their occurrence and he demonstrated that they are valuable stratigraphic markers by using them as datum planes in mapping. Hendricks and his coworkers Gardner and Knechtel (1947) also demonstrated the value of the siliceous shales as stratigraphic datum planes by mapping two of them but they did not use Harlton's terminology.

A restudy of the type sections of Harlton's Jackfork formations by Cline (1956a, p. 427) has revealed that the lower portion of the Prairie Mountain formation at the type locality is the stratigraphic equivalent of the upper Wildhorse Mountain formation at its type locality. Whereas this stratigraphic duplication resulted in an excessive estimate of the thickness of the Jackfork in the Tuskahoma syncline by as much as 3,600 feet, we believe that the interests of stratigraphy are best served by retaining both formation names, although the thickness of the Prairie Mountain formation is thereby considerably reduced and the Prairie Hollow maroon shale is now included in the Wildhorse Mountain formation rather than in the Prairie Mountain formation.

The "Union Valley" sandstone of Harlton (Game Refuge sandstone) has been restored to the Jackfork group because it is Mississippian in age (Cline, 1956c, p. 103) and is not equivalent to the type Union Valley which is Pennsylvanian in age. We have suggested to Harlton that a new name should be given to this sandstone which lies between the Wesley and Johns Valley shales and he has done so in his paper in this symposium.

Jackfork exposures in the Kiamichi Range. - The type localities of Harlton's Jackfork formations are widely scattered, one being in the Tuskahoma syncline in Pushmataha County, another being in the Round Prairie syncline northeast of Atoka in Atoka County, a third lying between these localities, and the fourth being in the frontal Ouachitas northeast of Stringtown. The most complete exposures of the Jackfork group seem to be in the north face of the Kiamichi Range, in a belt extending from Clayton eastward to the Arkansas line. Two complete, unbroken, and essentially unfaulted stratigraphic sequences have recently been described in some detail by Cline and Moretti (1956). The sections are 22 miles apart, are structurally aligned, and probably are about on sedimentary strike. One is exposed in a series of deep new road cuts between Big Cedar and Octavia, the other in an older series of cutbanks on the Indian Service road southeast of Albion, thus affording one an opportunity to compare relatively fresh strata with their somewhat weathered equivalents. Zone by zone descriptions have been published in Circular 41 of the Oklahoma Geological Survey.

Jackfork of southern Ouachitas. - Honess (1924) reported the Jackfork sandstone to be 10,000 to 13,618 feet thick in the southern Ouachitas and divided it into "Lower Jackfork" and "Upper Jackfork" separated by a fossiliferous sandstone which he reported as having a Morrow fauna. He noted that the 7,000 feet of Upper Jackfork is apparently

Fig. 4.-- Looking east by southeast at north escarpment of Kiamichi Range near Kiamichi Post Office. Photography by L. M. Cline.

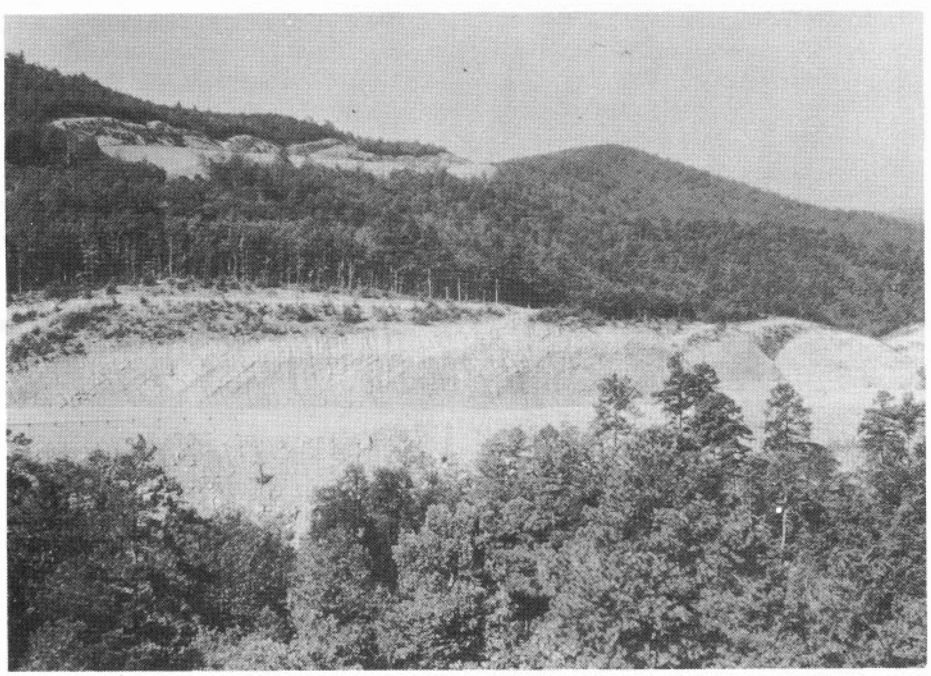

Fig. 5.-- Jackfork sandstones and shales exposed in newly constructed road up north face of Kiamichi Range between Big Cedar and Octavia. Lower Wildhorse Mountain in lower cutbanks, upper Wildhorse Mountain in upper level. Photography by L. M. Cline.

a sandy shoreward phase of the Atoka and the map accompanying his 1924 report shows the trace of the fossiliferous bed separating the Lower and Upper Jackfork on the north side of the Boktukola syncline. Miser (1925, p. 37) pointed out that inasmuch as Honess' Upper Jackfork is Atoka in age, it should be called Atoka, and it was so labeled on Miser's 1926 edition of the geologic map of Oklahoma.

Wildhorse Mountain Formation

At the type locality of the Wildhorse Mountain formation at the south end of the Tuskahoma syncline, in secs. 30, 31, and 32, T. 2 S., R. 16 E., the formation consists of 3,550 feet of massive sandstones interbedded with alternating shales and thin sandstones (Harlton, 1938, p. 878). A 250 to 300 foot interval of variegated red and green shales and soft green-gray massive sandstones occurs in the middle of the formation in Wildhorse Mountain and the interval has proved to be a persistent stratigraphic unit. Harlton (1938, p. 880) named the unit the Prairie Hollow shale from outcrops in Prairie Hollow on Prairie Mountain northeast of Atoka and he considered the shale as a member of the Prairie Mountain formation. Because the shale occurs in the midst of the type Wildhorse Mountain formation, Cline (1956b, p. 56) has proposed that the Prairie Hollow be regarded as a member of the Wildhorse Mountain formation for the reasons that: (1) the type locality has better exposures than the type Prairie Mountain, (2) the original description of the Wildhorse Mountain formation, and (3) because Harlton's original description of the Wildhorse Mountain is given in greater detail than that of his Prairie Mountain.

The upper part of the Wildhorse Mountain formation contains much sandstone. Massive sandstones near the top of the formation form prominent ridges and rock terraces that are easily traced on aerial photographs. It is not implied that individual beds are persistent over any considerable distance, but rather that the sandstones collectively exert a strong influence on the topography. These sandstones commonly form the crest of the Kiamichi Range from near Antlers and Moyers eastward into Arkansas. From field examinations it would appear that the sand-

Fig. 6.--Aerial mosiac of south portion of Tuskahoma syncline showing Wildhorse Mountain and Harlton's line of traverse in measuring the type section of the Wildhorse Mountain formation.

Fig. 7.-- Thinly bedded sandstones and dark shales in lower Wildhorse Mountain formation, Big Cedar-Octavia road, Kiamichi Range. Photograph by L. M. Cline.

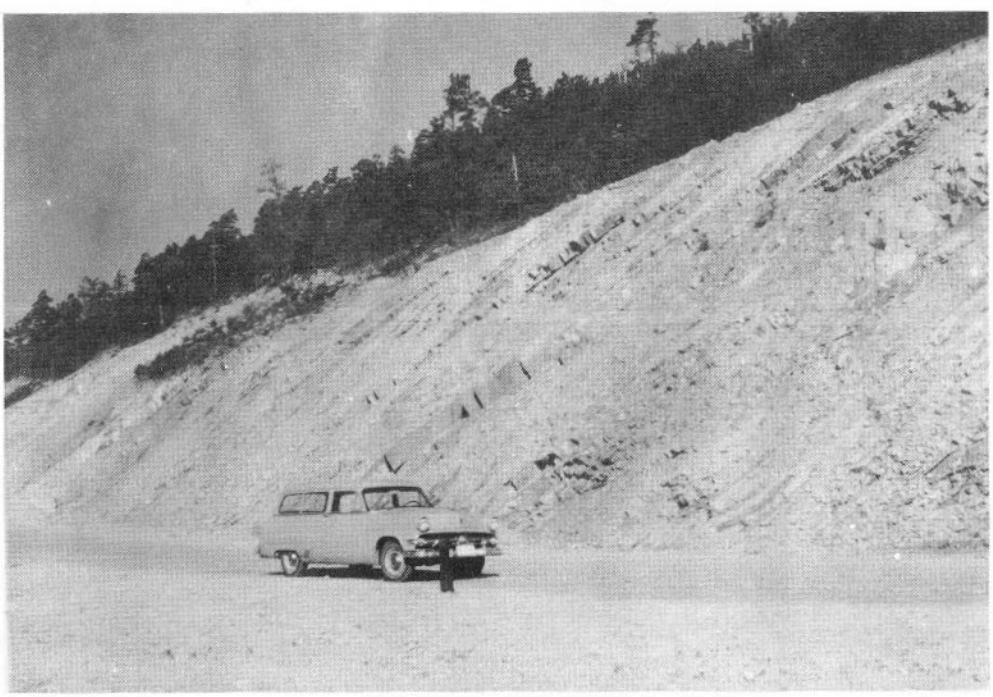

Fig. 8.--Resistant sandstones in upper Wildhorse Mountain formation. Big Cedar-Octavia road, Kiamichi Range. Photograph by L. M. Cline.

stones in the upper Wildhorse Mountain formation are whiter, cleaner, and better sorted than those in the lower part of the formation. Where sandstone rests on shale the lower surfaces of many sandstone beds have well developed flute casts, flow casts, load casts, and groove casts. From weathered outcrops one gets an impression that the formation is predominately composed of sandstone but fresh exposures reveal a high proportion of shale. Cline and Moretti (1956) have published a detailed description of fresh roadcut exposures of the formation in the eastern part of the Kiamichi Range where it is revealed that of the 3,542 feet of the formation, 880 feet is sandstone, 737 feet is sandstone with subordinate shale, 1,579 feet is shale with subordinate sandstone, and 328 feet is shale. Most of the section consists of intricately interbedded dark blue-gray to black laminated shale and thin, even beds of sandstone. The interlaminations appear to be so perfect that they could perhaps be described as rhythymic, each cycle being composed of a couplet of shale and sandstone.

It is noteworthy that the Prairie Hollow variegated shale and sandstone persists throughout much of the Ouachitas, having been identified and mapped in the frontal Ouachitas by Hendricks (Hendricks, and others, 1947), in the central Ouachitas by Cline (1956a, b, c), and in the southern Ouachitas by Shelburne.

Prairie Mountain Formation

The Prairie Mountain formation lies conformably on the Wildhorse Mountain formation and, except for the presence of the thin Prairie Mountain siliceous shale at the base, could not be separated from it. In Cline's map of the western part of the Lynn Mountain syncline, which is reproduced as Plate 1 in this paper, the Wildhorse Mountain and Prairie Mountain formations are mapped together because it was not expedient to map them separately in the time available. Harlton (1938) lists a thickness of 1,350 feet for the Prairie Mountain formation at the type locality in Prairie Mountain northeast of Atoka, but he erroneously included the Prairie Hollow shale and other Wildhorse Mountain equivalents in the type Prairie Mountain. When due allowance is made for this duplication of section, the restricted Prairie Mountain formation can be scarcely more than 400 feet thick in the type area. Harlton also gives a thickness of 4,300 feet for the formation in the Tuskahoma syncline north of Moyers but we believe this thickness to be greatly overestimated. The only published stratigraphic sections where the formation has been measured and described in detail are those by Cline and Moretti (1956) for two localities in the eastern part of the Kiamichi Range; the maximum thickness that can be assigned the formation in the thicker of the two exposures is less than 2,000 feet.

Lithology. - We have been unable to make significant progress in zoning the main body of the Prairie Mountain formation. Sandstone and shale occur in approximately equal proportions and individual beds do not seem to be very persistent. The most nearly complete exposures of the Prairie Mountain formation known to us are in cutbanks of a new road over Kiamichi Mountain, southeast and south of Big Cedar in southeastern LeFlore County, Oklahoma. Here Cline and Moretti (pp. 15, 16) measured 1,892 feet of strata which they assigned to the formation; this includes approximately 340 feet of sandstone, 655 feet of sandstone with some shale, 845 feet of shale with subordinate sandstone, and 52 feet of shale. A comparison with the lithology of the Prairie Mountain where the formation is exposed along the Indian Service road, about 20 miles westward in the Kiamichi Range, indicates that even the thicker sandstone beds do not achieve great areal extent. Throughout the thickness of the formation there are bodies of dark blue-gray sandy shales with intercalated thin beds of hard, quartzitic, light gray sandstone which average less than a foot in thickness. The bedding surfaces of the sandstones are parallel and sets of closely spaced joints result in the production of rhomboidal or tabular blocks upon weathering.

Prairie Mountain siliceous shale member. - A one to two-foot bed of dark gray siliceous shale marks the base of the Prairie Mountain formation in the type area in the western Ouachitas. Harlton (1938, pp. 880, 881) named this member the Prairie Mountain siliceous shale and stated that it is widespread in most of the Ouachitas. It seems to be persistent in the Round Prairie and Tuskahoma synclines where it has been used as a datum in mapping by Harlton. We doubt that it is as widespread

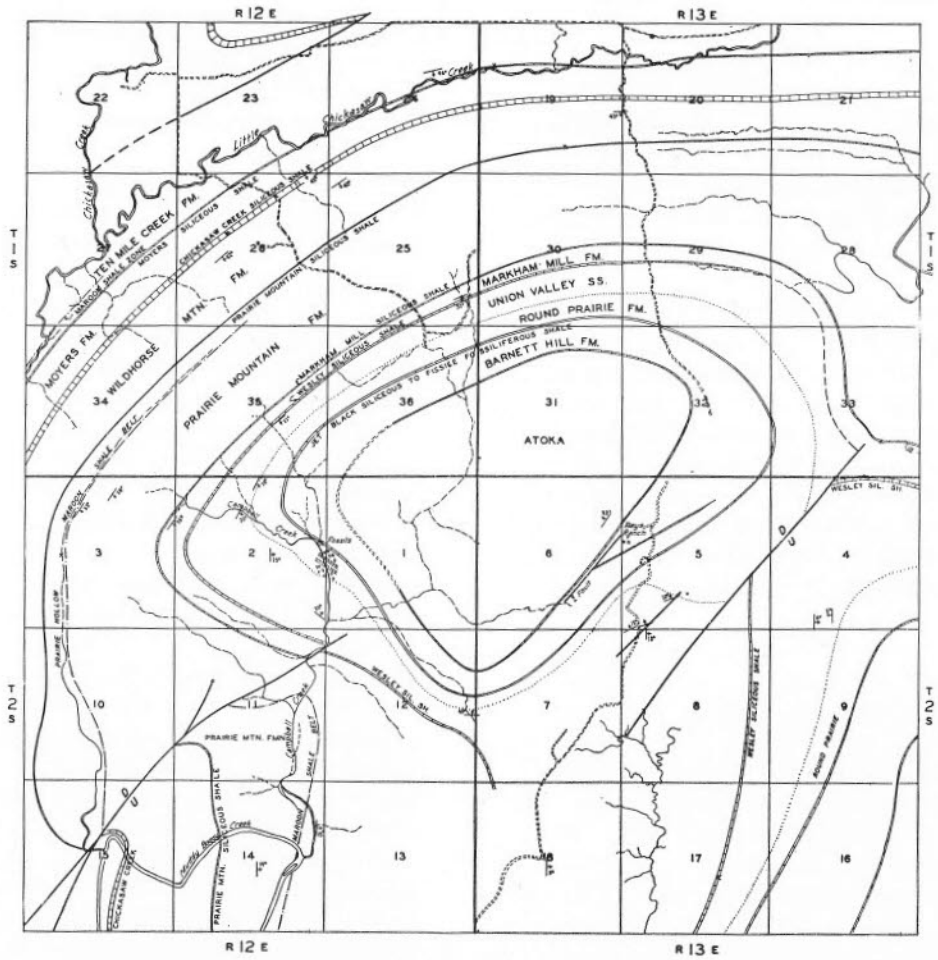

Fig. 9. -- Geologic map of Round Prairie. After B.H. Harlton, 1938.

Fig. 10. -- Sandstones and shales in lower part of Prairie Mountain formation. Big Cedar-Octavia road. Photograph by L. M. Cline.

as Harlton thought because Cline's mapping in the Lynn Mountain syncline indicates that it is not present (or is too thin to be a practical key bed), and Shelburne has been unable to find it in his mapping in the Boktukola syncline farther south.

This siliceous shale is prominently banded, the exceedingly fine laminae resulting from changes in the relative amounts of clay and iron oxides (Goldstein and Hendricks, 1953, pp. 432, 433) which alternate with carbonaceous and/or sapropelic material. Harlton (p. 882) reports that a microscope examination reveals the presence of numerous capsules and spines of radiolaria, sponge spicules, and Goldstein and Hendricks report spore exines.

Markham Mill Formation

Conformably above the Prairie Mountain formation is the Markham Mill formation. The name was given by Harlton (1938, p. 884) for outcrops near the old Markham sawmill (no longer in existence) in the NE cor. SW 1/4 sec. 21, T. 2 S., R. 14 E. on the west side of the Farris syncline. A second type locality was chosen in sec. 2, T. 2 S., R. 12 E., on Campbell Creek in the Round Prairie syncline. In its type area the formation consists of 400 feet of strata chiefly composed of sandstone in heavy beds below and becoming thinly bedded above. The base is marked by 15 to 20 feet of dark gray siliceous shale which is the most characteristic member of the formation.

The thickness seems to be fairly constant in the central Ouachitas. Cline and Moretti (1956, pp. 5, 15) recorded thicknesses of 318 and 328 feet respectively for two well exposed sections in the western part of the Kiamichi Range. The formation thins rapidly northward from the type section into the frontal Ouachitas. Harlton reported only 10 feet of sandstone above the thin basal siliceous shale in the fault block between the Ti Valley and Windingstair faults.

Markham Mill siliceous shale. - In the western Ouachitas 15 to 20 feet of hard, platy to blocky, dark gray siliceous shale marks the base of the formation. Some of the harder beds closely approach chert in general appearance, and on weathered outcrops float of this characteristic material are an aid in identifying and mapping the formation. The shale has inclusions of carbonized plant fragments which give it a mottled appearance. Weathered fragments of the more siliceous beds have a pinkish-gray to reddish-gray color.

Harlton (1938, p. 885) reported lenses of a cherty sandstone conglomerate in the lower part of the siliceous shale at Prairie Mountain and he noted that the conglomerate coarsens northward (presumably he means that the individual components of the conglomerate become coarser) to include boulders in the frontal Ouachitas. He stated that the boulders were "dropped" into the sea during deposition of the siliceous shale.

Areal distribution of the formation. - The basal siliceous shale is a valuable stratigraphic datum in the western Ouachitas. Harlton used it as a key bed in his mapping in the Round Prairie, Farris, and Tuskahoma synclines. Cline has noticed its presence south of the Kiamichi River in the north limb of the Lynn Mountain syncline, although he has not positively identified it in this range east of the Indian Service road. R. H. Johnson (1954) mapped a siliceous shale in the Medicine Springs area east of Antlers believing it to be the Markham Mill. The writers believe that this siliceous shale is in the upper Wesley formation and that it correlates with a siliceous shale farther east in the Boktukola syncline which Shelburne has mapped as the Wesley (see Shelburne's map, Plate 2, this report). Cline has mapped the Markham Mill with the Prairie Mountain formation in the western part of the Lynn Mountain syncline (see Cline's map, Plate 1, this report) for the reason that in the time available to map the area it did not seem practical to search for an obscure siliceous shale just to be able to differentiate 300 to 400 feet of prevailingly sandstone strata from the much thicker underlying Prairie Mountain sandstones.

Wesley Shale Formation

In the central Ouachitas from 150 to 500 feet of soft blue-gray and dark gray to black siliceous shale intervenes between the sandy strata of the Markham Mill formation below and the Game Refuge sandstone above. This shale is the Wesley shale of Harlton (1938,

Table 2. Correlation of Wesley-Johns Valley stratigraphic sequence between the central and frontal Ouachita provinces, southeastern Oklahoma

<table>
<tr><th>Belt southeast of Choctaw fault</th><th>Belt southeast of Katy Club fault</th><th>Belt southeast of Pine Mountain fault</th><th>Belt southeast of Ti Valley fault</th><th>Central Ouachitas</th></tr>
<tr><td>Atoka formation</td><td>Atoka formation</td><td>Atoka formation
(spiculite zone 200 feet above base)</td><td>Atoka formation
(spiculite zone)</td><td>Atoka formation
(spiculite zone)</td></tr>
<tr><td>Wapanucka limestone</td><td>Chickachoc chert</td><td rowspan="2">Springer formation</td><td rowspan="3">Johns Valley shale</td><td rowspan="2">Johns Valley shale</td></tr>
<tr><td>Springer formation</td><td>Springer formation</td></tr>
<tr><td rowspan="4">Caney shale</td><td rowspan="4">Caney shale</td><td rowspan="4">Caney shale</td><td>Game Refuge sand-stone</td></tr>
<tr><td>Wesley shale</td><td>Wesley shale</td></tr>
<tr><td>Jackfork sandstone (undifferentiated)</td><td>Jackfork sandstone</td></tr>
<tr><td>Stanley shale</td><td>Stanley shale</td></tr>
<tr><td rowspan="2"></td><td rowspan="2"></td><td>Woodford shale and chert</td><td rowspan="2">Woodford chert</td><td rowspan="2">Arkansas novaculite</td></tr>
<tr><td>Pinetop chert</td></tr>
</table>

p. 886) and the upper Jackfork siliceous shale of Hendricks et al (1947). The Wesley shale is one of the most widespread and useful stratigraphic markers in the thick Jackfork-Johns Valley-Atoka stratigraphic sequence.

Areal distribution and stratigraphic relations. - South of the Windingstair fault, in the central Ouachita Mountains, the stratigraphic relations of the Wesley are straightforward, but north of the fault there are some stratigraphic problems in connection with the Wesley and its correlatives. Where Harlton (1938) and Hendricks et al (1947) have mapped the Wesley in the Round Prairie, Farris, and Tuskahoma synclines and adjoining areas, the Game Refuge sandstone normally intervenes between the Wesley below and the Johns Valley shale above. The outcrop belt of the Game Refuge is ordinarily somewhat thicker than that of the Wesley but the Game Refuge thins northward (and westward) and in the western part of the frontal Ouachitas the Wesley and Johns Valley have been mapped as being in contact.

The Wesley is present throughout the Lynn Mountain syncline and is readily recognized in outcrops by its characteristic lithology and its strike valleys can be traced on aerial photographs (Cline, 1956b). Shelburne (Plate 2, this report) has mapped the Wesley throughout the Boltukola syncline to the south.

In the fault block north and northwest of the Windingstair fault the Game Refuge sandstone thins rapidly and the width of its outcrop belt becomes less than that of the Wesley. The westward thinning is illustrated by Hendrick's map which shows the Game Refuge to be well developed in sec. 23, T. 1 N., R. 13 E., but thinning rapidly southwestward along the outcrop in secs. 22, 27 and 28 and disappearing altogether in sec. 33 to bring the Wesley and Johns Valley shales in contact. Farther northwest, adjacent to the Ti Valley fault, Harlton and Hendricks and his coworkers have mapped the Wesley in contact with the Johns Valley. An exception to this general rule is found in the small fault block in which is located the type locality of the Wesley; both Harlton and Hendricks et al have mapped the Atoka as resting unconformably on the Wesley in this area.

North of the Ti Valley fault Hendricks et al have employed a stratigraphic terminology that corresponds more to Arbuckle facies terminology than to that of the Ouachita facies. There are significant differences in the geology on opposite sides of the fault, the chief difference being that the Jackfork formation is not recognized north of the fault and that in place of Wesley and Johns Valley (beneath the Atoka) the writers believe these differences are more apparent than real. In the first place, the northward disappearance of the Jackfork is forecast by the very rapid thinning in all of the members of this group as they are traced northward from the Round Prairie, Farris, and Tuskahoma synclines. A close study of Hendrick's map shows convincing evidence bearing on this point. For example, the complete thickness of the Jackfork group in the type area of the Wesley shale can not be more than 1,700 feet, whereas a few miles south, in the Round Prairie syncline, the interval approximates a mile in thickness. If this rapid rate of northward convergence continues to the Ti Valley fault, it then becomes unnecessary to postulate great horizontal movement along the fault to explain the absence of the Jackfork sandstone to the northwest. As for the use of the terms Springer and Caney in the area north of the fault, the writers are convinced that the Wesley represents some part of the Caney and that the Springer is the partial equivalent of the Johns Valley. Cline now has some field studies under way which he hopes will clear up this point. It should not be overlooked that in Harlton's original discussion of the Wesley (Harlton, 1938, p. 886) he stated that everywhere along the Choctaw fault the Wesley is in depositional contact with "Chester Caney"; in the vicinity of Pine Top School, strata mapped as Springer by Hendricks were called Wesley by Harlton. Our difference of opinion with Harlton comes from our belief that Wesley represents some part of the Caney.

Lithology and thickness. - At the type locality the Wesley consists of 500 feet of gray shale with some subrounded masses of dark gray to black chalcedony and many intercalated lenses of a peculiar chert conglomerate (Harlton, 1938, p. 886). Hendricks (1947, p. 17) has noted a facies change from soft blue-black and hard black siliceous shale in the northwest part of the Ouachitas to a brown clay shale containing scattered thin beds of hard black siliceous shale in the southeast. In the Lynn Mountain syncline Cline (Plate 1, this report) has mapped as Wesley about 400 feet of soft gray shale containing some thin dark siliceous (cherty) layers. Eastward in this syncline, in cutbanks along the Indian Service road where it crosses the Kiamichi Range, the formation consists of about 150 feet of soft blue-gray shale with intercalated thin beds of siltstone and fine-grained sandstone and at least one conspicuous chert bed near the top (Cline and Moretti, 1956, p. 5). The chert is dark gray to black and weathering produces 6-inch rhombohedrons with a characteristic green-gray weathered surface. This chert is present in the Kiamichi Range eastward to the Arkansas line. Shelburne has also found chert beds to be present in the top of the Wesley in the Boktukola syncline where that formation attains a thickness of at least 223 feet.

Erratics in the Wesley. - Harlton has emphasized the occurrence of the thin beds of chert conglomerate in the Wesley and called attention to the presence of rock fragments derived from an Arbuckle facies. Hendricks (1947, p. 17) noted the presence of erratic boulders of chert and flint with diameters up to seven feet. According to Hendricks, Ordovician trilobites and graptolites have been collected from some of the erratics. From the fact that there is a progressive decrease in the size of the boulders from northwest to southeast, Hendricks concluded that they had been derived from the northwest; we would add that this is in harmony with the size-grade distribution of the erratics in the Johns Valley formation.

Fauna. - Ammonoid cephalopods have been reported to occur in many of the "siliceous masses" (Harlton, 1938, p. 889) and in some of the limonitic (sideritic) concretions (Hendricks, 1947, p. 34). According to Harlton the non-siliceous portions of the shale contain foraminifera of which *Hyperammina*, *Cornuspira*, *Bigenerina*, *Haplophragmoides*, *Agathammina*, and *Sherbonites* are the most plentiful. Hendricks noted that conodonts and radiolaria are abundant (most likely in the siliceous members), in places the radiolaria appearing as spheres of clear silica approximating one millimeter in diameter.

Correlation. - Harlton (1938, p. 889) correlated the Wesley with the "Pennsylvanian Caney" (Goddard-Springer) and Hendricks (1947, correlation chart, p. 8) evidently intended to correlate it with the upper part of the Springer. Inasmuch as the basal portion of the Johns Valley shale has Mississippian Caney goniatites and typical Caney lithology (Cline, 1956c, p. 103), the Wesley shale of the central Ouachitas must be late Mississippian in age. It is entirely possible that the upper part of the Wesley shale at its type locality includes some Goddard or Springer equivalents because neither Harlton nor Hendricks *et al* recognized Johns Valley in this small fault block; rather, they mapped Atoka on Wesley. Notwithstanding the need for additional studies of the Wesley-Johns Valley sequence in the frontal Ouachitas, we have ventured some tentative correlations between the various fault blocks in the chart below.

Game Refuge Sandstone

Overlying the Wesley shale and underlying the Johns Valley shale is a persistent fossiliferous sandstone containing abundant fragments of crinoid columnals, brachiopods, bryozoans, and plant fragments. Harlton (1938) correlated this sandstone with the Union Valley sandstone (the "Cromwell" of the subsurface), which at its type locality east of Ada contains a goniatite fauna of basal Morrow age (Miller and Owen, 1944). Inasmuch as the lower portion of the overlying Johns Valley shale contains a Caney goniatite fauna of late Mississippian age (Cline, 1956c, p. 103), this fossiliferous sandstone must be older than the type Union Valley. Recognizing the inappropriateness of the term Union Valley, Harlton has proposed the name Game Refuge sandstone in his paper which is included in this symposium.

Type locality. - Harlton states that the best exposures of the formation known to him are along Campbell Creek in sec. 2, T. 2 S., R. 12 E., in the Round Prairie syncline northeast of Atoka but that the name Campbell Creek is preoccupied. He has chosen the name Game Refuge sandstone because it is available and because there are good outcrops in the State Game Refuge about 6 to 8 miles southwest of Clayton in Pushmataha County. There are good outcrops of the formation along the north side of Jerusalem Hollow in the SW 1/4 sec. 28 and the N 1/2 sec. 32, T. 1 N., R. 18 E. where its relations to the Wesley and Johns Valley are quite clear.

Lithology and thickness. - The Game Refuge is predominately sandstone but it includes some intercalated gray shale. It contains some thin siliceous shale in the Kiamichi

Range, the Boktukola syncline, and the Medicine Springs area. The sandstone has good topographic expression and produces good outcrops, lying as it does between the relatively weak Wesley and Johns Valley shales.

The Game Refuge is characterized in having zones of limonitic sandstone containing the plant Calamites and a fauna composed of external and internal molds of marine invertebrates. The invertebrate fossils were originally included in a quartzose sandstone as fragments of calcium carbonate but it seems that by the time diagensis had been completed the calcium carbonate had largely been dissolved; some of the molds have since been filled with argillaceous material.

Ripple marks, including some interference ripples, and occasional small scale cross bedding are about as characteristic of the Game Refuge as the mold fauna. It seems significant that these sedimentary structures are not common in the Jackfork sandstones, whereas the flow casts, flute casts, and other bottom markings so typical of the sandstones of the Jackfork group have rarely been noted in the Game Refuge sandstone.

The formation approximates a thickness of 350 to 400 feet in the western Ouachitas, except that it thins northward in the frontal Ouachitas to disappear altogether north of the Ti Valley fault. It is about 500 feet thick in the Kiamichi Range and in the Boktukola syncline. It should be noted that in their description of the Jackfork-Atoka strata of the Kiamichi Range, Cline and Moretti (1956, pp. 4, 5, 15) referred only 160 feet of sandstone to the Game Refuge ("Union Valley" of that report, following Harlton's earlier terminology). Cline and Shelburne are now of the opinion that the following zones described for the Indian Service road stratigraphic section should also be included in the Game Refuge formation: zone 106, which at that time was identified as Johns Valley?, and zones 107 and 108 which were included in the basal Atoka. In making the original correlation Cline and Moretti were strongly influenced by the opinion that the sandstones of zones 107 and 108 were the direct lateral equivalents of the fossiliferous sandstone in the Boktukola syncline from which Honess had obtained his "Morrow fauna". However, Shelburne has found that Honess' type "Morrow fauna" lies 800 feet above the Wesley shale and 300 feet above the Game Refuge sandstone in that area.

Fauna. - The abundance of Calamites fragments in the formation has been noted above. Because the invertebrate fossils are mostly molds it is difficult to pursue identification as far as species rank. Nonetheless, Harlton (1938, p. 893) has been able to identify the molluscs Aviculopecten, Bellerophon, Gastrioceras, and Orthoceras (more likely Pseudorthoceras), the brachiopod Productus (probably Dictyoclostus), and the trilobite Griffithides.

Johns Valley Shale

Definition and history of the nomenclature. - Lying stratigraphically above the Jackfork sandstone and below the Atoka formation in the central Ouachita Mountains is the Johns Valley shale. The formation name Johns Valley was introduced by Ulrich (1927, pp. 21, 22) for outcrops in bowl-shaped Johns Valley in the trough of the Tuskahoma syncline, particularly in the north one-half of T. 1 S., R. 16 E. In recent years there has been a tendency to overlook the fact that Taff's (1901, p. 3) type locality of the Caney shale is also in Johns Valley, which is precisely the same locality which Ulrich proposed for his type Johns Valley formation. The name was derived from the Cane Creek (now Caney Creek) which drains a portion of the valley. Also, generally ignored, is the fact that many of the fossils that were illustrated by Girty (1909) in United States Geological Survey Bulletin 277, The fauna of the Caney shale formation of Oklahoma, were collected from his localities 2075, 3983, 3984, and 3986 in secs. 2, 3, and 4, T. 1 S., R. 16 E., in Johns Valley.

It should be noted that Girty's collecting localities referred to above are now known to be in the lower part of Taff's Caney shale, and that Taff intended that the name Caney should be applied to all of the shale between the Jackfork and Atoka in the central Ouachitas; Ulrich was well aware of this. Furthermore, all geologists acquainted with Girty's work on the Caney fauna have known that both Girty and Taff regarded the Caney fauna in Johns Valley as coming from rocks identical in age with the now better known Mississippian Caney shale of the Arbuckle facies to the west. Miser and Honess (1927, pp. 11, 12, fig. 2) were firmly convinced that the Arbuckle Caney is the same age as the shale containing the Caney fossils in Johns Valley. Ulrich acknowledged the contemporaneity of the fossils but he believed that the fossils in Johns Valley came from exotic or erratic boulders which had been ice-rafted into the enclosing shale. Inasmuch as Ulrich found what he believed were Wapanucka (lower Pennsylvanian) fossils in a boulder-bearing portion of the Johns Valley shale, he concluded that all of the Johns Valley is Pennsylvanian in age and therefore younger than the Caney of the Arbuckle region. We should note that Ulrich did not find Pennsylvanian erratics in the Johns Valley at the type locality.

The goniatite-pelecypod fauna that Girty described from the lower Caney is now generally regarded by paleontologists as late Mississippian and as a fauna typical of a black shale facies. The upper, softer, lighter gray shale occurring above the black Mississippian Caney in the Arbuckle facies has long been referred to as "Pennsylvanian Caney" by subsurface workers but more recently it has been named Goddard and the term seems to be coming into general use.

We should emphasize that the presence of the limestone erratics in the lower Caney in Johns Valley did not mislead Miser and Honess into correlating all of the "Ouachita Caney" with Pennsylvanian equivalents as it had misled Ulrich. Miser and Honess published diagrams (1927, fig. 2) showing two possible interpretations of the relationship of the Arbuckle to the Ouachita facies. Regardless of which of their interpretations is the correct one, both diagrams show the Caney of the Arbuckles as having been physically continuous with the Caney of the Ouachitas at the time of deposition; both diagrams show the Caney of the Ouachitas as lying stratigraphically above the Jackfork. We are firmly convinced that Miser and Honess were correct in assigning the Caney of the Ouachitas a stratigraphic position above the Jackfork and we would like to see them given proper credit for being so discerning. Ulrich (1927), in his characteristic verbose manner, went to great lengths to convince the reader that the "Ouachita Caney" (Johns Valley shale) must be younger than Arbuckle Caney.

Because of uncertainties concerning the use of the terms Caney, Mississippian Caney, Pennsylvanian Caney, Ouachita Caney and Johns Valley, Harlton (1938, p. 896) proposed that the name Johns Valley be abandoned and that the name Round Prairie be adopted for the shale between the Jackfork and Atoka. Outcrops in the Round Prairie syncline northeast of Atoka, in the NE 1/4 NE 1/4 SE 1/4 sec. 2, T. 2 S., R. 12 E., were designated as the type locality. Reasons for choosing Round Prairie over Johns Valley were given as, (1) the poor exposures in Johns Valley, and (2) the chaotic mixture of the exotic boulders and the crumpled shale which precluded determination of the true character of the "normally deposited shales".

Subsequent workers, including Hendricks (1947), Hendricks, Gardner, and Knechtel (1947), Rea (1947), Howell (1947), Cline (1956a, 1956b, 1956c), and Cline and Moretti (1956), have not followed Harlton in the use of the term Round Prairie, preferring instead the older term Johns Valley. If the law of priority were to be invoked, the name Johns Valley would be suppressed as a synonym of Caney (original usage), but in this case the name Johns Valley is so thoroughly established in the literature and so firmly entrenched in the minds of geologists working in the Ouachitas that the interests of geology are best served by the continued use of the name Johns Valley. Also, the name Caney is now generally understood to refer only to the lower, Mississippian portion, of Taff's original Caney.

<u>Areal distribution, lithology, and thickness</u>. - In its characteristic development the Johns Valley formation includes from 425 to 900 feet of prevailingly shaly strata. The lower portion contains dark to black shales with typical Mississippian Caney lithology and fossils, whereas the upper portion contains considerable amounts of lighter gray clay shale with some thin interbedded sandstones whose lithology resembles that of the sandstones of the overlying Atoka. At the few localities where both contacts of the formation have been seen, there are zones in both lower and upper portions of the shale in which limestone and chert erratics occur. The erratics include numerous rounded pieces ranging in size from pebbles to boulders, and less numerous angular blocks of limestone, some of which attain diameters of 40 feet or more. The boulder-bearing shales have been referred to by some geologists as the "lower boulder bed" and the "upper boulder bed". In reality, there are intervals in both the upper and lower portions of the Johns Valley in which exotics occur in several beds, some of which are shale beds separated by well

defined sandstone beds; so it is not inappropriate to refer to an upper and a lower boulder bed.

Johns Valley shale containing erratics is known to occur in a belt about 25 to 30 miles wide and about 125 miles long, extending from near Atoka, Oklahoma, in the western part of the frontal Ouachitas eastward to Boles, Arkansas. Hendricks, Gardner, and Knechtel did not map the Johns Valley north of the Ti Valley fault in the frontal Ouachitas but we believe that it almost certainly is represented north of the fault by the Springer formation and by the upper part of their mapped Caney. We have not found boulder-bearing Johns Valley south of Hardy Creek in the Lynn Mountain syncline but on the basis of stratigraphic position Shelburne has assigned shales to the Johns Valley which lie above the Game Refuge sandstone in the Boktukola syncline.

Limestone erratics occur in the Johns Valley in the Windingstair Range almost as far east as the state line but in the next range south, the Kiamichi Range, we have been unable to find erratics east of Oklahoma State Highway 2 which crosses the range between Clayton and Nashoba Junction. Nonetheless, a well-developed strike valley, carved from soft shales which lie stratigraphically above the Game Refuge sandstone, can be traced eastward on aerial photographs. A ground check at several localities reveals that these shales have a stratigraphic position immediately above the Game Refuge sandstone, which in turn rests on easily recognized Wesley shale. Shale also occupies this same stratigraphic interval above the Game Refuge sandstone in the Boktukola syncline and we see no reason to call it anything other than Johns Valley, notwithstanding the lack of exotic boulders.

Noteworthy Johns Valley Outcrops

Johns Valley. - Various aspects of the stratigraphy of the Johns Valley in its type area have been discussed by previous writers and we therefore will limit our observations to a few points which seem pertinent to our paper.

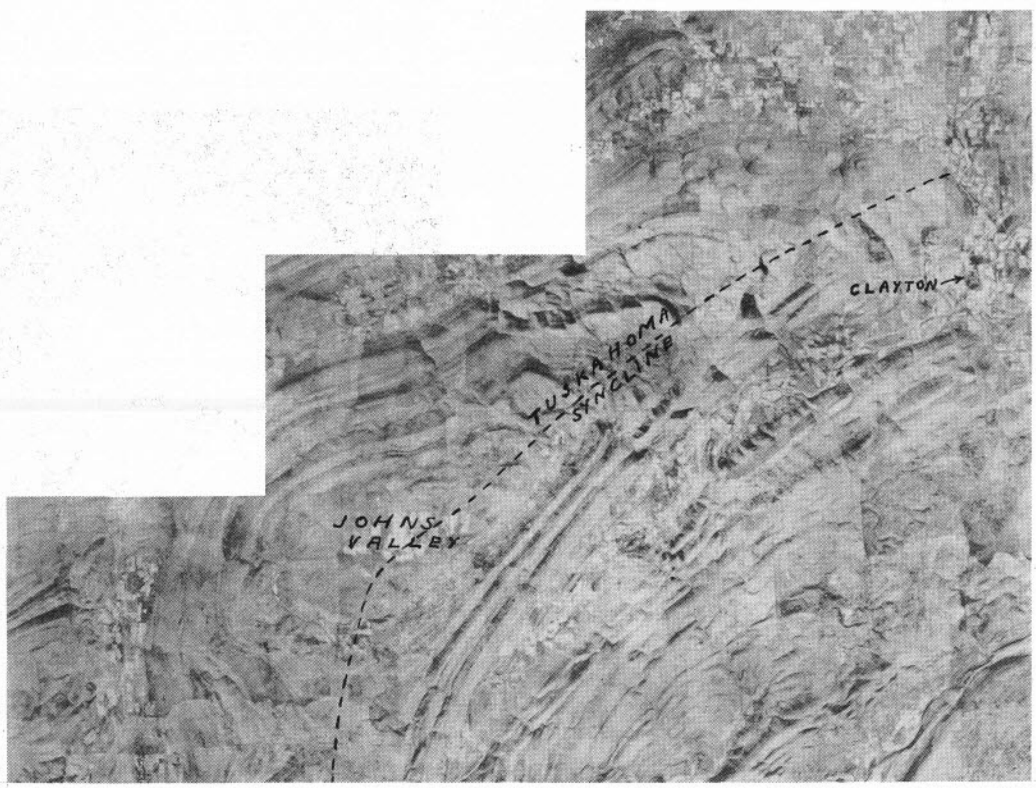

Fig. 11. -- Aerial mosiac showing the northeastern portion of the Tuskahoma syncline and bowl-shaped Johns Valley. (see Tobin Photo Map for regional setting of Johns Valley)

At the time Ulrich (1927) named the Johns Valley shale he stated that in Johns Valley it is underlain by the Jackfork sandstone and overlain by sandy shales of the Atoka formation. Fortunately he mentioned (p. 22) that the presence of Atoka in Johns Valley had been questioned by "recent investigators". Hendricks et al (1947) did not map Atoka in Johns Valley, and the sandy shales occupying the trough of the syncline are labeled Springer and included in the upper Johns Valley by Harlton on the map which accompanies his paper which is included in this symposium. On Harlton's map he had differentiated a wide belt of "bouldery shale" lying stratigraphically above the Game Refuge sandstone (uppermost Jackfork), a much thinner band of "Caney-Goddard" lying inside the bouldery shale, and an oval-shaped area of "Springer" inside this and occupying the deepest part of the Tuskahoma syncline. Although the observed dips in the inner oval area range from 20 degrees to vertical, according to Harlton, and locally are overturned, we believe that it is safe to say that almost the complete thickness of the Johns Valley shale is represented in its type locality. It seems like a stroke of fate that erosion should have stripped all of the Atoka and possibly a small amount of the upper Johns Valley, but that seems to be the case. Recently discovered outcrops across the Kiamichi River, only a few miles to the east, reveal the Johns Valley shale between the Atoka and Game Refuge in such simple structural relations as to leave no doubt as to its true stratigraphic position beneath the Atoka.

Incidentally, the last time the writers visited the Johns Valley it could be reached only walking, on horseback, or by jeep or logging truck (although one of the few remaining residents made occasional trips to town on a tractor). The road has recently been improved and is accessible by automobile.

Farris syncline. - The Johns Valley crops out in two continuous belts on opposite sides of the Farris syncline several miles southwest of the type locality of the Johns Valley. An interesting boulder-bearing shale outcrop, which is easily reached by automobile, lies east of the road and not far from the center of sec. 22, T. 2 S., R. 14 E., the locality being about one mile due east of the type locality of the Markham Mill formation. The stratigraphic relations of the Johns Valley to the Game Refuge and the Atoka forma-

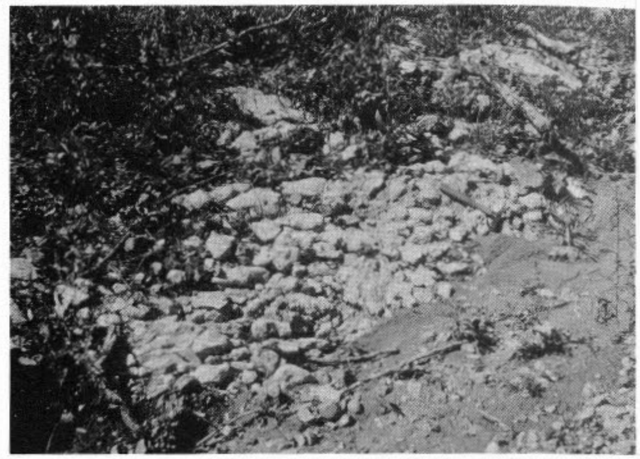

Fig. 12. -- Erratic block of Silurian limestone embedded in Johns Valley shale. Type locality in Johns Valley. Photograph by L. M. Cline.

Fig. 13. -- Typical black limestone concretion interbedded with black shale of Caney lithology which is a part of the Johns Valley formation. The concretion contains Mississippian goniatites. Type section of Johns Valley shale in Johns Valley. Photograph by L. M. Cline.

tions are easily determined. There is a gentle dip-slope developed on the upper surface of the fossiliferous Game Refuge sandstone, and sandstones in the lower Atoka form a low cuesta on the opposite side of the strike valley developed in the Johns Valley shale. Limestone erratics ranging in size from pebbles, cobbles, and boulders, and a few large blocks, are abundant in the dark shale matrix. Two large blocks of Silurian limestone are shown in the photograph reproduced as figure 14.

Fig. 14. -- Limestone erratics in Johns Valley shale near center of Sec. 22, T. 2 S., R. 14 E., in the Farris syncline. Upper photograph shows large blocks of Silurian limestone. Lower photograph shows boulders and cobbles embedded in soft, dark gray shale. Photographs by L. M. Cline.

Round Prairie syncline. - The Johns Valley forms an almost continuous outcrop belt around Round Prairie in the inner portion of the Round Prairie syncline northeast of Atoka, being interrupted only on the south side by faulting. In 1938, when Harlton proposed to substitute the name Round Prairie for Johns Valley, he designated outcrops in the NE 1/4 NE 1/4 SE 1/4 sec. 2, T. 2 S., R. 12 E. as the type locality for his new formation. He recognized a lower boulder-bearing shale separated from an upper bouldery shale by two or three beds of fossiliferous sandstone. A black siliceous shale was said to mark the top of the formation. The aggregate thickness of the various members of the Johns Valley is about 800 feet in Round Prairie. Both Harlton and Hendricks et al mapped Atoka in the center of the Round Prairie syncline. Round Prairie is accessible by automobile by a road from Atoka which enters from the south and another road entering from the north which is reached from Stringtown.

Hairpin Curve locality. - The most publicized of Johns Valley localities is the Hairpin Curve exposure along Oklahoma State Highway 2, about midway between Clayton and Wilburton, in the south part of sec. 3, T. 3 N., R. 19 E. The locality has been visited by many geologists including those in attendance during the 1947 Tulsa Geological Society Field Conference in the Ouachita Mountains. Hendricks and Averitt published an excellent description of the Johns Valley in this exposure and reproduced two photographs of the boulder beds in the guidebook (pp. 32-34). In referring to the Hairpin Curve section, Cline (1956c, p. 104) stated that "the exposures of the lower and upper Johns Valley appear to be on opposite sides of an anticline". In commenting on the structure at this same locality Misch and Oles (1957, p. 1903) state that the Johns Valley outcrops are in an "unbroken anticline which is succeeded on the north and on the south by unbroken synclines". More recently Cline has restudied the Hairpin Curve locality in greater detail than before and has found that the beds in the lower Johns Valley, which are exposed south and southeast of the sharp curve, are overturned. Near the top of the artificial cut, south of the curve, there are some sandstone beds that are nearly horizontal and a superficial examination would lead one to suppose that they are in their normal depositional position.

Closer study reveals that load casts, flute casts, groove casts, and other sedimentary structures characteristic of the under surface of sandstones are present on the upper surfaces of the sandstone as they are now exposed; thus, the stratigraphic sequence is upside down. Confirmation that the beds south of the curve have been overturned by folding (probably accompanied by faulting) is found in a close study of the contacts of some of the sandstone and shale beds; in a normal depositional sequence the contact of a sandstone bed with the underlying shale is sharper than the contact with the overlying shale, which contact has a tendency to be somewhat gradational. Inasmuch as the rocks south of the curve are overturned, the structure is not that of a simple anticline as Cline formerly thought, and as Misch and Oles recently stated so emphatically, but it is probably faulted in much the way that Hendricks mapped it. How much of the lower part of the Johns Valley is omitted by faulting we do not know, but we suspect that the 200 feet of section shown as covered in the Hendricks-Averitt measured section (their zone 26) approximated the amount of missing strata. Cline and Laudon have redescribed the section in some detail and this description is reproduced in the guidebook which has been prepared for the Ouachita field trip held in connection with the Dallas meeting of the American Association of Petroleum Geologists in March, 1959. The strata are somewhat better exposed now than at the time Hendricks and Averitt described the section and, of course, we interpret the south part of the exposures as being overturned. A traverse eastward along the highway substantiates our contention that the beds south of the curve are overturned. It also establishes the fact that these rocks represent the lowermost strata in the Johns Valley; the Game Refuge sandstone, also overturned nearest the curve but gradually returning to its normal structural position as it is traced eastward, underlies the Johns Valley and, in turn, it is underlain by the Wesley shale at the place where the road curves southward.

The overturned section south of the Hairpin Curve contains some interbedded strata with typical black shale Caney lithology and with phosphatic concretions containing goniatites of late Mississippian age, among which are *Gastrioceras choctawensis* and *Lyrogoniatites* (see Fig. 15). This tongue of Caney shale is in its proper stratigraphic position above the Game Refuge sandstone and does not represent a fault slice brought up by the fault which is present in the stream valley immediately north.

These beds with Caney lithology and Caney fossils also contain numerous exotic boulders and pebbles of limestone which were dropped into the shale at the time it was mud. They are not tectonic boulders or part of a friction carpet. Precisely this same lithology is characteristic of the lower Johns Valley in outcrops in the State Game Preserve southwest of Clayton where the formation is found in continuous outcrops between the Game Refuge and Atoka sandstones and under such simple structural conditions as to leave no doubt that the lower Johns Valley includes beds of Mississippian Caney age and lithology.

North of the Hairpin Curve the upper Johns Valley shale occurs in its normal position and dips northward beneath the Atoka sandstone. The base of the exposed section is composed of alternating shales, siltstone, and sandstones and erratics have not been observed. Upward in the section scattered pebbles of limestone erratics occur in shales sandwiched in between well-bedded sandstones. In the upper part of the section there is a striking concentration of limestone boulders, blocks, and cobbles and they rest in a channel which clearly bevels about 11 1/2 feet of the underlying strata (see Fig. 15). There is a noticeable decrease in the size of the erratics upward in this deposit, the overall effect being not unlike that of graded bedding, but it is of course on a somewhat larger scale than the usual examples. The erratics in the lower part of the channel fill include well-rounded boulders with diameters in excess of a foot, slightly rounded blocks of similar dimensions, the whole being embedded in a clay-shale matrix. Upward the boulders give way to cobbles and they in turn give way to pebbles which are widely separated in the gray clay-shale matrix as to give the effect of plums in a pudding. Throughout this deposit there are rounded masses of a hard, brownish quartzitic sandstone. We interpret this particular "boulder bed" as the product of a single turbidity flow or submarine slide. The flow initially attained a high velocity during which phase it was able to transport boulders and scour previously deposited muds and sands. As the peak of the flow was reached and the

Fig. 15.--Johns Valley shale in cutbanks along Oklahoma Highway 2, south part Sec. 3 T. 3 N., R. 19 E., about midway between Clayton and Wilburton. Upper left: boulder beds in upper part of Johns Valley; Atoka sandstones dipping right (north) on right side of photograph. Upper right: typical black, concretionary Caney shale lithology in lower part of Johns Valley; south of curve; beds are overturned and dip 35 degrees to left (south). Lower left: boulder-bearing shale resting in channel cut through shales and sandstones in upper Johns Valley; possibly cut by turbidity current which deposited the boulder beds. Lower right: boulder-bearing channel fill. Photographs by L. M. Cline.

velocity trailed off, pebbles began to drop out and were deposited with the muds of the flow and those obtained from the reworked bottom. The sandstone masses represent lenses of sand torn from the bottom and rolled along the flow. Some geologists have interpreted these rounded sandstone masses here and at other localities as "tectonic boulders" in the friction carpet of an advancing thrust sheet. We doubt it.

Lynn Mountain syncline. - Persistent search has revealed two belts of Johns Valley shale cropping out in the Lynn Mountain syncline between Clayton and Antlers. One of the belts is 20 miles in length and it contains excellent outcrops of the Johns Valley under such simple structural conditions as to leave no doubt that it is part of a depositional sequence that includes the Wesley shale and Game Refuge sandstone below and the Atoka sandstone above. Equally important is the fact that black shale with typical Mississippian Caney lithology and typical Caney pelecypods and goniatites constitutes the lower part of the Johns Valley shale, a point that can be checked at numerous localities. Because these outcrops have a key role in the interpretation of Ouachita stratigraphy, Cline has given them considerable attention in mapping the western part of the Lynn Mountain syncline (Plate 1, this report).

The more accessible of the two outcrop belts is crossed by Oklahoma Highway 2 at a place a few miles south of Clayton, just north of Hardy Creek and about 1 1/2 miles north of Nashoba Junction. The Game Refuge sandstone forms the crest of an asymmetrical anticline immediately north of the creek and north of this sandstone the Johns Valley shale is exposed intermittently in ditches and cutbanks along the highway for about a quarter of a mile. Erratics are not numerous in the Johns Valley at this locality but a few limestone cobbles and pebbles and at least one boulder (Fig. 16) of undisputed Arbuckle facies have been discovered by very careful search. Rolled sandstone masses are very common, some of them attaining diameters of several feet (Fig. 17). This locality is noteworthy in being the southeasternmost known outcrop of boulder-bearing Johns Valley, although it is not the easternmost nor the southernmost occurrence. Measured from the northwestern margin of the Stanley-Jackfork-Johns Valley depositional trough, this locality is farther out in

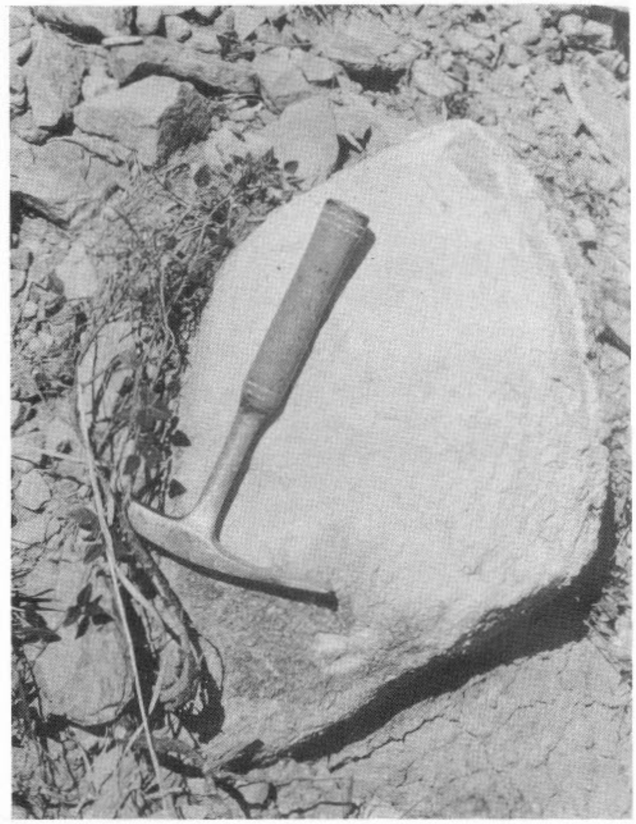

Fig. 16.--Boulder of Bromide limestone in upper portion of Johns Valley shale. Cutbank north of Hardy Creek, Oklahoma Highway 2 between Clayton and Nashoba Junction. Photograph by L. M. Cline.

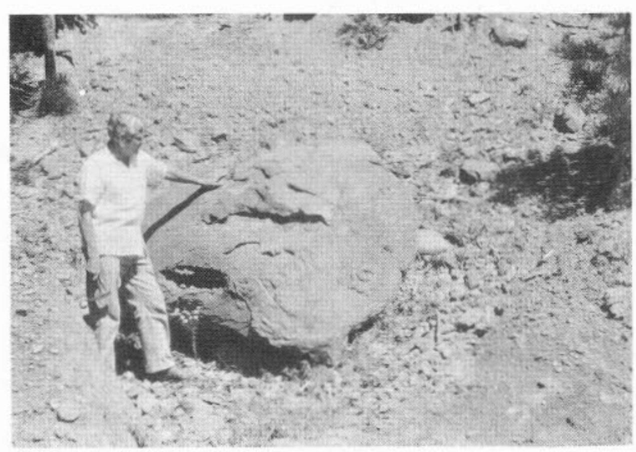

Fig. 17.--Rolled sandstone mass in lower Johns Valley shale. Cutbank north of Hardy Creek, Oklahoma Highway 2 between Clayton and Nashoba Junction. Dr. M. K. Elias standing alongside. Photograph by L. M. Cline.

the trough and farther from the presumed source of erratics than any other known occurrence of the boulder-bearing beds.

The best and most continuous exposures of the Johns Valley shale known at present lie in the other belt in the Lynn Mountain syncline, in the hills south and southeast of the Kiamichi River, between Clayton and Kosoma. It is most unfortunate that the three most instructive outcrop areas in this belt are comparatively inaccessible.

One interested in studying the Johns Valley shale in the Lynn Mountain syncline should first see the outcrops in the State Game Refuge southwest of Clayton and southeast of Stanley. Permission to enter should first be obtained from Conservation Headquarters in the preserve and once this is obtained, it is possible to drive fairly close to the outcrops in a Jeep or logging truck.

The steep valley wall along the south side of Jerusalem Hollow is capped by Atoka sandstones and shales dipping southeastward into the Lynn Mountain syncline at about 40 degrees. Numerous gullies and ravines trending northwestward down the slope to join Jerusalem Creek expose various portions of the Johns Valley shale and we are certain that a composite section showing the complete thickness of the formation could be pieced together by studying outcrops beginning in the SW 1/4 sec. 28, T. 1 N., R. 18 E. and continuing southwestward across the NW cor. sec. 33, diagonally across sec. 32, and through the south part of sec. 36. The Game Refuge sandstone is well exposed in the opposite wall of the valley, the dips being about 40 degrees and the strike the same as that of the Atoka. The most significant feature of this series of outcrops is that black laminated shale containing marble-size phosphatic concretions and large black limestone concretions consistently forms the lower 200 feet (estimated) of the Johns Valley shale. The dark, dense, sideritic limestone septaria enclose numerous well-preserved <u>Goniatites choctawensis</u> and <u>Caneyella</u> and less numerous <u>Rayonoceras</u>. The black shale in which they are embedded also contains numerous specimens of <u>Caneyella</u> flattened parallel to bedding and some of the harder more siliceous layers contain conodonts. The fauna is a typical Caney fauna and offers irrefutable evidence of the late Mississippian age of the lower portion of the Johns Valley shale and evidence for a Mississippian age for the very thick underlying Stanley-Jackfork succession; inasmuch as the upper part of the Arkansas novaculite, which underlies the Stanley, has been judged to be Mississippian on the basis of its contained conodonts (Hass, 1956). It is also noteworthy that there are limestone erratics enclosed in this Caney portion of the lower Johns Valley shale in these outcrops. By no stretch of the imagination can these exotics be construed as part of a friction carpet of an advancing thrust sheet; they are as much a part of the lower Johns Valley as the black shale that contains them.

The second instructive series of outcrops in this belt lies southwest of the Jerusalem Hollow outcrops and is offset from them by a dip fault at a point almost due south of Stanley (see Fig. 18). Where Jerusalem Creek intersects this fault it makes an abrupt turn and flows north following the strike of the fault. The fault is a tear fault bounding the northeast side of an overturned block which begins south of Stanley and continues southwestward to a position opposite Kosoma, an outcrop distance of about 14 miles. Throughout this distance the Jackfork sandstone and Johns Valley shale are overturned to the southeast, a direction opposed to the principal direction of movement along the major thrust faults in the central and frontal Ouachitas.

The first good outcrops of Johns Valley shale southwest of the tear fault lie in a strike valley excavated by the headwaters of Beuhler Creek. The outcrops are reached by fording the Kiamichi River at a point about midway between Stanley and Dunbar and following Beuhler Creek upstream to the place where it branches into two tributaries whose directions of flow are at right angles to that of the main stream. The traverse up the main stream reveals excellent exposures of the formations of the Jackfork group, each overturned and dipping northwest, the dips averaging approximately 80 degrees. The Johns Valley shale has the same dip and strike but it is in fault contact with the Atoka sandstone which dips southeast at an angle of about 25 degrees. This strike fault merges with the tear fault at the northeast end of the block. The total horizontal movement southeastward along the fault can hardly be more than a few

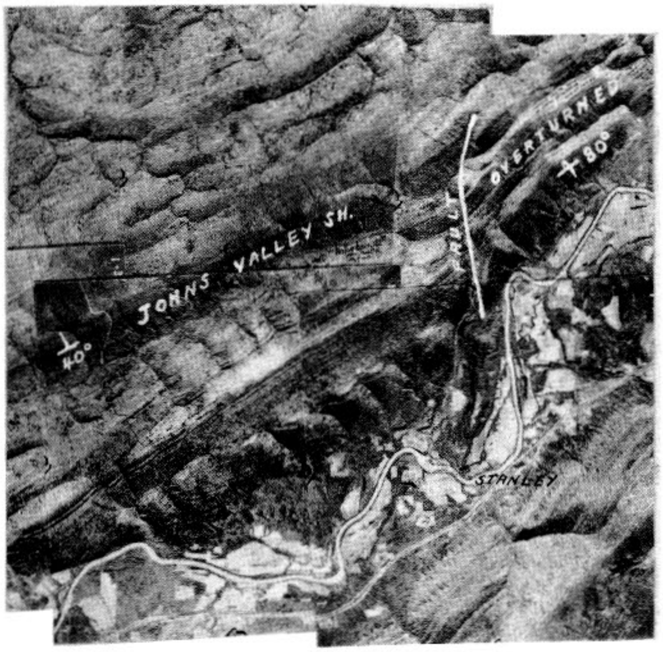

Fig. 18. -- Aerial mosiac of area south and southeast of Stanley, Oklahoma. Note that top of photograph is to the south. The area southeast of Stanley and southeast of the Kiamichi River contains a normal depositional sequence of Jackfork, Johns Valley, and Atoka, all dipping southeast into the trough of the Lynn Mountain syncline. In the block west of the tear fault the strata also occur in their proper stratigraphic order but are overturned to the southeast, opposed to the northwest movement along most of the major faults in this part of the Ouachitas.

hundred feet; the abrupt change from the overturned beds in the Johns Valley to the gentle southeast dips of the Atoka into the Lynn Mountain syncline was accomplished principally by adjustments in the incompetent shales in the upper Johns Valley and the lower Atoka. The belt of Johns Valley strikes southwestward across the SW 1/4 sec. 1 and the W 1/2 sec. 12, T. 1 S., R. 17 E. The lower part of the formation exhibits Caney lithology, including the laminated black shale containing the small marble-size phosphatic concretions and the larger limestone septaria with the Caney goniatites. The black shale contains numerous limestone erratics, some of which are very large; one exotic block of Ordovician limestone has a length of about 40 feet. These outcrops are very instructive in that the bedding planes of the laminated black shale pass over and under the erratics, demonstrating conclusively that the erratics were incorporated in the black shale at the time it was deposited. Large pieces of a limestone conglomerate may have been emplaced by a submarine slide. Some of the angular pieces of limestone have a lithology not unlike that of the Caney concretions and some resemble the Sycamore limestone; a feature suggesting that these shelf limestones were debouched into a black mud environment, possibly by sliding down a slope greased by the accumulating muds.

Another interesting series of Johns Valley exposures lies in a strike valley in the east-central part of sec. 32 and the adjacent NW 1/4 sec. 33, T. 1 S., R. 17 E. The area, lying about 2 1/2 miles due east of Eubanks and 3 miles southeast of Dunbar, may be reached by fording the Kiamichi River south of Dunbar and walking about 1 1/2 miles up the canyon of a northwest-flowing stream. This traverse crosses all of the Jackfork formations, all of which are overturned to the east-southeast and dip west-northwest about 75 to 80 degrees. Just east of the watergap which cuts through the Game Refuge sandstone, the stream branches, each branch following a strike valley carved from the Johns Valley shale. The south fork has numerous exposures of laminated black Caney shale containing dark limestone concretions bearing _Goniatites choctawensis_ and _Caneyella_. There are numerous limestone erratics completely enclosed by the bedding planes of the black shale. A most interesting feature is the inclusion of some large blocks of a limestone conglomerate which we believe may have been emplaced as a submarine slide. The bedding planes of the shale parallel the strike of the underlying Game Refuge sandstone but locally the shales are badly crumpled, as one might expect in incompetent shales lying between competent sandstones in a slightly overturned section.

Boktukola syncline. - In the Boktukola syncline from 500 to 700 feet of strata have been mapped as Johns Valley by Shelburne. The formation is conformable with the underlying Game Refuge sandstone and gradational with the overlying friable sandstones and gray silty shales of the Atoka. The Johns Valley lithology is similar to that in the Lynn Mountain syncline to the north, but limestone erratics have not been found notwithstanding a careful search for them. Rounded sandstone masses rnaging in diameter from three inches to six feet, depending on the thickness of the parent bed, are very characteristic of the formation.

It is noteworthy that the fossiliferous beds from which Honess (1924, p. 23) obtained his "Morrow fauna" lie within the Johns Valley formation and about 330 feet above the top of the Game Refuge sandstone. This particular fossiliferous zone can not be traced outside of Section 6, T. 1 S., R. 23 E. Other occurrences cited by Honess, and presumed by him to be the same bed, range as low stratigraphically as the Markham Mill formation and as high as the lower Atoka.

Origin Of Erratic Boulders And Blocks In The Johns Valley Shale

The method by which the limestone erratics were emplaced in the Johns Valley shale has intrigued all geologists who have mapped in the Ouachitas. There is an extensive literature dealing with the subject but not all of the problems have been solved. The only point on which there is general agreement is that the erratics are Arbuckle and/or Ozark types and that they must have been derived from the south, southwest, west, north, or northeast, or possibly from more than one of these directions. However, the view has also been expressed that the Ouachita facies is allochthonous and that it overlies an autochthonous Arbuckle facies from which the boulders were derived by faulting.

Theory that the erratics were brought up by faults from an underlying Arbuckle facies. - Pointing to the large size of the Caney erratics in the Johns Valley shale in Johns Valley, at the Hairpin Curve, and in Cooper Hollow, it has been held by some geologists that they are fault slices brought up from an underlying Arbuckle facies. Enthusiasts for long-distance low-angle overthrusting have been intrigued with this idea. The erratics have been visualized by some workers as having constituted part of a friction carpet at the base of an advancing thrust sheet.

As we began our studies in the Ouachitas we visited Johns Valley and most of the other outcrops of the Johns Valley cited in the literature. Our first and lasting impression was, and is, that the Johns Valley shale occupies a constant stratigraphic position between the Jackfork and Atoka. It seems odd to us that tectonic boulders brought up by thrusting should invariably choose the same stratigraphic horizon to come to rest. Howell (1947) has pointed out that the weak Johns Valley shale is a natural place for faulting to develop and that the (greasy) shales offered a lubricant for such thrusts. There are other thick and persistent shales stratigraphically lower in the Stanley and Jackfork groups. Why have the limestone exotics bypassed these lower shales on their way to a higher plane? Along the Boktukola and Octavia faults, and the other major thrusts in the central Ouachitas, the relatively competent Jackfork sandstones are in fault contact with the thick, soft, incompetent Stanley shales. If the Johns Valley erratics are part of a friction carpet, why have not the Stanley shales stopped some of the erratics in these shear zones? Such faults surely must have served as feeders to supply the tectonic boulders of the Johns Valley, if indeed they are tectonic.

We should add that we are firmly convinced that the supposed Caney fault slices at the Hairpin Curve and Cooper Hollow localities are as much a part of the normal stratigraphic succession as any of the rocks with which they are in contact. It is true that faulting has disturbed the beds at both localities but one familiar with the details of the stratigraphic sequence can work out the structural details.

The boulders were deposited concurrently with the Johns Valley shale. - In 1947 Rea (pp. 47-49) stated the belief that the Johns Valley lies conformably between the Jackfork and Atoka sandstones in all of the localities in the western Ouachitas where it had been observed. From the beginning of our Ouachita studies we have worked on the assumption that the Johns Valley shale has constant stratigraphic position. Using aerial photographs we have traced belts of Johns Valley from known occurrences and subsequent ground checks have revealed many previously unknown outcrops of the shale, some of them containing limestone erratics. In recognizing an upper and a lower boulder-bearing zone within the Johns Valley, Harlton (1938) expressed his faith in their having constant stratigraphic position and in their horizontal continuity. We have observed the Johns Valley at many localities, i.e., Hairpin Curve and Beuhler Creek, where the bedding planes of the shale pass over and under erratic boulders without evidence of structural disturbance. There is not the slightest doubt in our minds that the boulders were emplaced in the Johns Valley at the time the containing shales were being deposited. How they were transported to their depositional sites is not so apparent!

Theory of ice-rafting. - Ulrich (1927, pp. 34-36, 44) put forth the theory that the erratics in the Johns Valley were transported to their depositional sites by ice floes derived from shore ice. Other writers sharing this view were Powers (1928, p. 1046), and Rea (1947, p. 49). Rea's summary of the arguments favoring this theory is especially good.

Theory of submarine sliding. - Miser (1934) regarded the "Ouachita Caney" as large masses possibly transported by submarine slides or slips. Whereas we like this method of transportation to explain movement of some of the large limestone blocks, we should point out that the "Ouachita Caney" masses to which Miser referred, are actually near their depositional sites and are a part of the Johns Valley.

The writers' viewpoint. - Of the various theories advanced to explain transportation of the erratics to their depositional site, ice-rafting seems to best explain most of the observed features. It explains the great range in size grade in the material, and in particular it offers a method of transportation for the smaller erratics, many of which are granule and pebble size; it also explains the decrease southeastward in size away from the shoreline. It explains the apparent concentration at particular stratigraphic levels. It explains the lack of deformation in the shales which enclose many of the erratics.

The largest blocks may have been transported in a different manner. There is a possibility that they slid down the steep western and northern slopes of the Ouachita trough on a bottom greased with accumulating black muds. The rapid northwestern convergence of the Stanley and Jackfork groups and their disappearance near the Ti Valley fault argues for rapid subsidence within the Ouachita trough and probably an appreciable southeastward slope into the trough. Thinning from approximately 16,200 feet in Johns Valley to almost nothing at the Ti Valley fault about 13 1/2 miles northwest, the bedding planes at the base of the Stanley had a minimum southeastward dip of about 13 degrees at the end of Jackfork time. It could be argued that deposition kept pace with subsidence and that the depositional surface at any given time never had an appreciable slope, but the turbidity flow characteristics of the Jackfork sandstones suggests strong currents and, therefore, appreciable bottom slope during deposition. A slope of as little as one degree would have been sufficient to cause the large limestone blocks to slide over the slippery black muds.

Age of the Johns Valley shale. - Our views concerning the age of the Johns Valley shale have been set forth in some detail in the preceeding pages. In summary, the lower Johns Valley includes a depositional tongue of the Caney shale and is therefore of late Mississippian age. The middle and upper portions of the formation almost certainly contain Springer equivalents and lithologically are very similar to it. At the Hairpin Curve, on the Indian Road, and in Honess' type locality of his "Morrow fauna" in the Boktukola syncline, there are fossiliferous sandstones which probably are Morrow in age (Dr. M. K. Elias, personal communication; also letters to Dr. C. W. Tomlinson). Our opinion is that the uppermost Johns Valley is equivalent to some part of the Wapanucka formation, possibly the lower and middle portions. Thus, it seems that the Johns Valley shale bridges the Mississippian-Pennsylvanian time line.

At the Hairpin Curve locality the boulder-bearing upper Johns Valley occupies a channel which seems to have been scoured simultaneously with the deposition of the boulder-bearing clay-shale. There is a strong suggestion that the transporting agent was a turbidity flow or submarine slide.

Atoka Formation

We have not examined the main body of the Atoka formation in any detail, but in connection with mapping the Johns Valley-Atoka contact we have studied the lower several hundred feet of the Atoka rather carefully. There is a dark siliceous shale containing abundant sponge spicules in some of the harder more siliceous beds which occurs from 100 to 200 feet above the base of the formation from the Ti Valley fault southeastward into the central Ouachitas. Hendricks and his co-workers (1947) regard this as a good stratigraphic datum and state that it is widespread. We believe that the spiculite may correlate with one of the spiculitic beds in the upper part of the Wapanucka limestone in the frontal Ouachitas and the McAlester basin. There is considerable field evidence that the upper Wapanucka grades into sandstone southeastward from the frontal Ouachitas. Thus, the lowermost Atoka of the central Ouachitas may include beds somewhat older than the lowermost Atoka of the frontal Ouachitas. Cline is of the opinion that the base of the Winslow sandstone (Atoka equivalent) of Arkansas also crosses time lines from its northernmost outcrops in Arkansas southward into the Ouachita geosyncline, its base becoming progressively lower to the south.

We have seen no evidence of a depositional break between the Johns Valley shale and the Atoka sandstone in the central Ouachitas. Sandstones in the upper Johns Valley are lithologically like those in the lower Atoka. The strike is the same in both formations. It therefore seems that the unconformity which has been reported at the base of the Atoka on the northeastern Oklahoma shelf dies out southward into the Ouachita geosyncline.

The complete thickness of the Atoka formation is not represented in the Ouachita Mountains, the upper portion having been removed by erosion. Considerable thicknesses of the formation are preserved in the Lynn Mountain and Boktukola synclines where downfolding has protected the lower part of the formation from being eroded. Shelburne estimates that 6,800 feet of Atoka is present in the trough of the Boktukola syncline and Cline estimates that the Atoka is equally thick in the Lynn Mountain syncline. Our present state of knowledge does not permit a precise zone by zone correlation with the Atoka of the McAlester basin, so there is no way of knowing how much of the upper part of the formation has been eroded. We are not absolutely certain that the complete Atoka section was ever represented in the Ouachitas because the possibility exists that the central Ouachitas were somewhat elevated in pre-Desmoinesian time and thus became source areas in late Atoka time.

CONCLUSIONS

Tectonic Setting Of Ouachita Geosyncline

The arcuate pattern of folds which comprises the Ouachita Mountains of Oklahoma and Arkansas is but one portion of a sinuous foldbelt which extends from western Texas eastward to within approximately 60 miles of the buried extension of the Appalachian Mountain system. Paleozoic sedimentation seems to have been initiated in the Ouachitas in a deep arcuate trough lying at the south margin of the stable shelf area in which the carbonate-bearing Arbuckle facies accumulated. The pre-Stanley sedimentary sequence is relatively thin, possibly being thinner than the Arbuckle facies, and is characterized by cherts (some of which contain radiolaria) and dark graptolitic shales; the rocks represent a period of very slow sedimentation in essentially as starved trough. In post-Arkansas novaculite time, active tectonism resulted in rapid subsidence and rapid sedimentation. The late Mississippian-early Pennsylvanian Stanley, Jackfork, Johns Valley, and Atoka sequence was deposited in a rapidly subsiding trough. Subsidence and sedimentation were so rapid that long before the end of Atoka time some 24,000 feet of late Paleozoic shales and sandstone were trapped in the Ouachita trough. Some time after the deposition of lower Atoka sediments the geosyncline was severely compressed, its sediments were folded into long, linear folds and many of the folds were ruptured to produce thrust faults.

Environment of Sedimentation

Lithologically the late Mississippian and early Pennsylvanian rocks are comparable to the typical black shale flysch facies of the Eocene of the Alps and the Eocene and Cretaceous of the Carpathians. Interpretations may vary as to the significance of a flysch facies but if the literature gives us an accurate description of black shale flysch, this facies exists in the Ouachitas. The Johns

Valley boulder-bearing shale is reminiscent of what Alpine geologists call wildflysch, although again interpretations of its origin have varied.

The conclusion is reached that a predominately deep water black mud environment was periodically interrupted by invasions of sand. Turbidity currents flowing down the steep sides of the Ouachita trough may have debouched sands derived from a nearby shelf environment. The presence of abundant flow casts, flute casts, load casts, and groove casts on the under surfaces of sandstone beds, the presence of well-developed convolute bedding in some strata, the general lack of cross bedding and ripple marks, the presence of planktonic fossils in the black shale and the absence of benthonic forms in these same shales, support this thesis. Some of the sandstones have crinoid columnals, bryozoan and brachiopod fragments, fragments of invertebrates that are normally at home in a shelf environment, but their occurrence in the sandstones is spotty and may be explained on the assumption that their environment of life was not that of deposition. There is a correlation between grain size in the sandstones and the size of the contained shell fragments; this indicates a relationship between competency of the transporting currents and the size of shell fragments picked up and transported.

Directional studies in progress permit the tentative conclusion that the turbidity flows were resolved into powerful currents which flowed westward and southwestward down the axis of the depositional trough. A few dozen compass readings on flute casts, flow casts, and groove casts indicate that the prevailing current direction was west 12 degrees south.

Erratic boulders in the Johns Valley. - We favor ice-rafting as the most plausible mechanism for transporting the limestone erratics to their depositional sites in the accumulating Johns Valley muds. The bedding planes of the containing shales pass under and over most of the pebbles, cobbles, and boulders without appreciable disturbance. The very large limestone blocks may have had a different origin. The enclosing shales are disturbed adjacent to the blocks and this suggests the possibility that the blocks slid down a depositional slope greased by the accumulating black muds. It must be admitted, however, that a 40-foot long block of limestone should have disturbed muds even if it were dropped by ice.

The evidence is conclusive that most of the erratics are depositional, not tectonic, in origin. Many of the exotics are completely enclosed in undisturbed shales, which in turn are enclosed (interbedded) in well-bedded sandstones which show no evidence of structural movement other than convolute bedding contemporaneous with deposition. The hypothesis that the erratics represent part of a friction carpet at the base of an advancing thrust sheet is untenable. Firstly, there is absolutely no evidence for bedding plane faults in many of the "boulder beds". Secondly, it seems odd that boulders brought up from an underlying Arbuckle facies by faulting should invariably choose the same stratigraphic datum to come to rest. It is true that the thick Johns Valley shale is a natural place for structural adjustment to take place, and the greasy shale would offer a lubricant for such thrusts. However, there are other thick, laterally persistent shales in the underlying Jackfork and Stanley groups and they have not been so favored by visits by their erratic neighbors from the Arbuckle facies. Is it reasonable to expect "tectonic boulders" completely to bypass these lower shales on their way to higher shales? Along the Boktukola and Octavia faults, and other major thrusts in the central Ouachitas, Jackfork sandstones are commonly in fault contact with Stanley shales. If the Johns Valley boulders are part of a friction carpet, why have not some of them lodged in these shear zones? Surely such thrusts must have served as feeders to supply the "tectonic boulders" of the Johns Valley, if indeed they are tectonic!

Correlation Of The Stanley, Jackfork, Johns Valley, and Lower Atoka

The geologic age of the Stanley, Jackfork, Johns Valley, and lower Atoka rocks of the central Ouachita Mountains has been the subject of much speculation. Although there are some fossil plants in all of these units, paleobotanical evidence has not been especially helpful in deciding whether the sequence is Mississippian or Pennsylvanian. Uncertainties regarding the age stem from the scarcity of marine invertebrates in this thick flysch facies. There are abundant macerated plant fragments and an occasionally well-preserved Calamites or Lepidodendron, but it is an understatement to say that the evidence from paleobotany has not been especially helpful. Apart from the plant fossils, the next most common fossils are some radiolaria, sponge spicules, and conodonts found in some layers of the hard siliceous shales. Literature containing descriptions and stratigraphic occurrences of late Paleozoic radiolaria and sponge spicules is too meagre to be very helpful in dating these strata. There is more information on the conodonts, but here again much cataloguing of species, descriptive work, and stratigraphic work remains to be done.

Hass (1956) regards the Stanley conodonts as being early and middle Mississippian. Thusfar the Jackfork has yielded very little paleontologic evidence regarding its age. On the basis of stratigraphic position it has to be Mississippian because it underlies the Johns Valley shale, the lower part of which contains late Mississippian goniatites. The lower Johns Valley contains a depositional tongue of Mississippian Caney shale which contains typical late Mississippian goniatites.

We have mapped the Johns Valley shale in a belt some 20 miles long in the Lynn Mountain syncline between Clayton and Kosoma where it occurs under such simple stratigraphic conditions as to leave no doubt that it stratigraphically overlies the Jackfork and underlies the Atoka. Throughout this distance a tongue of Caney shale comprises the lower part of the formation. Thus, the lower Johns Valley is Mississippian and the underlying Jackfork must be Mississippian.

The middle and upper portions of the Johns Valley almost certainly contain Springer equivalents and lithologically they are very similar to it. The uppermost Johns Valley contains Honess' "Morrow fauna" and probably is equivalent to some part of the Wapanucka formation, possibly the lower and middle portions. The Mississippian-Pennsylvanian boundary lies within the Johns Valley shale.

Northwestward thinning of the Stanley-Jackfork groups. - Attaining an aggregate thickness of 16,200 feet in Johns Valley in the Tuskahoma syncline, the Stanley and Jackfork groups thin to practically nothing at the Ti Valley fault some 13 1/2 miles to the northwest. We believe that this rapid thinning is primarily depositional. Convergence at all stratigraphic levels in both groups is rapid and argues for a rapidly subsiding trough with a steep northwestern margin. The latter point is supported by the presence of the many turbidity flow features of the Stanley and Jackfork sandstones. Some of the northwestward thinning may have resulted from horizontal movement along the thrust sheets but to us it seems that the rates of thinning of individual units is not significantly interrupted by the faulting. This is one of the few points on which we have a professional difference of opinion with our good friend T. A. Hendricks. He has studied the amount of overthrusting along the major faults of the frontal Ouachitas and he has some strong arguments that horizontal movement has been considerable.

Lower Atoka strata in central Ouachitas. - A siliceous shale containing abundant sponge spicules lies from 100 to 200 feet above the base of the Atoka formation in the central Ouachitas. We believe that this spiculite correlates with some part of the Wapanucka spiculite, logically the upper part, and, if we are correct, the base of the Atoka formation in the central Ouachitas is Morrow in age. The unconformity that marks the base of the Atoka in the northeastern Oklahoma platform or shelf has not been observed by us in the central Ouachitas and we believe that deposition was continuous from Johns Valley into Atoka time.

References

Applin, Paul L., 1951, Preliminary report on buried pre-Mesozoic rocks in Florida and adjacent states: United States Geol. Survey Circ. 91.

Cline, L. M., 1956a, Regional stratigraphy and history of Ouachita Mountain area (abs): American Association Petroleum Geologists Bull., vol. 40, no. 12, pp. 427-428.

Cline, L. M., 1956b, Mississippian-Pennsylvanian stratigraphy of central Ouachita Mountains, Oklahoma: Ardmore Geol. Soc., Guidebook to Ouachita Mountains field conference, southeastern Oklahoma, pp. 47-63.

Cline, L. M., 1956c, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains, Oklahoma: Tulsa Geol. Soc. Digest, vol. 24, pp. 100-106.

Cline, L. M., and Moretti, Frank, 1956, Two measured sections of Jackfork group in southeastern Oklahoma: Oklahoma Geol. Survey Circ. 41, pp. 1-20.

Girty, G. H., 1909, Fauna of the Caney shale of Oklahoma: United States Geol. Survey Bull. 377.

Goldstein, August, Jr., and Hendricks, T. A., 1953, Siliceous sediments of Ouachita facies in Oklahoma: Geol. Soc. America Bull., vol. 64, pp. 421-442.

Harlton, B. H., 1933, Micropaleontology of the Pennsylvanian Johns Valley shale of the Ouachita Mountains, Oklahoma, and its relationship to the Mississippian Caney shale: Jour. Paleont., vol. 7, pp. 3-29.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: American Association Petroleum Geologists Bull., vol. 22, pp. 852-914.

Hass, W. H., 1956, Conodonts from the Arkansas novaculite, Stanley shale, and Jackfork sandstone: Ardmore Geol. Soc., Guidebook to field conference in Ouachita Mountains, pp. 25-33.

Hendricks, T. A., Gardner, L. S., Knechtel, M. M., and Averitt, P., 1947, United States Geol. Survey, Oil and Gas Inv. Prelim. Map no. 66.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Bull. 32.

Honess, C. W., 1924, Geology of southern Leflore and northwestern McCurtain counties, Oklahoma: Bur. Geology, Norman, Oklahoma, Circ. 3.

Howell, J. V., 1947, The origin of imbricate structure: Tulsa Geol. Soc., Guidebook to Field Conference in western part of Ouachita Mountains in Oklahoma, pp. 50-53.

Johnson, R. H., 1954, Geology of the Medicine Springs area, Pushmataha County, Oklahoma: unpublished M. S. thesis, University of Oklahoma.

King, P. B., 1950, Tectonic framework of southeastern United States: American Association Petroleum Geologists Bull., vol. 34, no. 4, pp. 635-671.

Kramer, Wm. B., 1933, Boulders from Bengalia: Jour. Geol., vol. 41, pp. 590-621.

McCollough, E. L., 1954, The geology of the Finley area, Pushmataha County, Oklahoma: unpublished M. S. thesis, University of Oklahoma.

Miller, A. K., and Owen, J. B., 1944, The cephalopod fauna of the Pennsylvanian Union Valley formation of Oklahoma: Jour. Paleont., vol. 18, no. 5, pp. 417-428.

Miser, Hugh D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geol. Survey, Bull. 50.

Miser, Hugh D., 1934, Carboniferous rocks of Ouachita Mountains: American Association Petroleum Geologists Bull., vol. 18, no. 8, pp. 971-1009.

Miser, Hugh D., and Honess, C. W., 1927, Age relations of the Carboniferous rocks of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geol. Survey, Bull. 44.

Morgan, H. J., 1952, Paleozoic beds south and east of Ouachita folded belt: American Association Petroleum Geologists Bull., vol. 36, no. 12, pp. 2266-2274.

Powers, Sidney, 1928, Age of folding of the Ouachita Mountains: Geol. Soc. America Bull., vol. 30, pp. 1031-1072.

Taff, Joseph A., 1901, Description of the Coalgate Quadrangle, Indian Territory: United States Geol. Survey Geol. Atlas, Folio 74.

Taff, Joseph A., 1902, Description of the Atoka Quadrangle, Indian Territory: United States Geol. Survey Geol. Atlas, Folio 79.

Ulrich, E. O., 1927, Fossiliferous boulders in the "Ouachita Caney" shale and the age of the shale containing them: Oklahoma Geol. Survey, Bull. 45.

Willis, P. D., 1954, Geology of the Snow area, Pushmataha County, Oklahoma: unpublished M.S. thesis, University of Oklahoma.